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Latest Pleistocene and Holocene glaciation of Baffin Island, Arctic Canada: key patterns and chronologies

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ABSTRACT

Melting glaciers and ice caps on Baffin Island contribute roughly half of the sea-level rise from all ice in Arctic Canada, although they comprise only one-fourth of the total ice in the region. The uncertain future response of arctic glaciers and ice caps to climate change motivates the use of paleodata to evaluate the sensitivity of glaciers to past warm intervals and to constrain mechanisms that drive glacier change. We review the key patterns and chronologies of latest Pleistocene and Holocene glaciation on Baffin Island. The deglaciation by the Laurentide Ice Sheet occurred generally slowly and steadily throughout the Holocene to its present margin (Barnes Ice Cap) except for two periods of rapid retreat: An early interval ~12 to 10 ka when outlet glaciers retreated rapidly through deep fiords and sounds, and a later interval ~7 ka when ice over Foxe Basin collapsed. In coastal settings, alpine glaciers were smaller during the Younger Dryas period than during the Little Ice Age. At least some alpine glaciers apparently survived the early Holocene thermal maximum, which was several degrees warmer than today, although data on glacier extent during the early Holocene is extremely sparse. Following the early Holocene thermal maximum, glaciers advanced during Neoglaciation, beginning in some places as early as ~6 ka, although most sites do not record near-Little Ice Age positions until ~3.5 to 2.5 ka. Alpine glaciers reached their largest Holocene extents during the Little Ice Age, when temperatures were ~1–1.5 °C cooler than during the late 20th century. Synchronous advances across Baffin Island throughout Neoglaciation indicate sub-Milankovitch controls on glaciation that could involve major volcanic eruptions and solar variability. Future work should further elucidate the state of glaciers and ice caps during the early Holocene thermal maximum and glacier response to climate forcing mechanisms.

1. Introduction

The Earth has been warming for over a century. This warming has been most pronounced in the Arctic, especially over the last few decades (IPCC, 2007). Planetary warming is having dramatic effects, with increasing rates of melt and iceberg discharge from the Greenland Ice Sheet (Rignot and Kanagaratnam, 2006; Howat et al., 2007), and pronounced melting of mountain glaciers (Meier et al., 2007). In addition, the summer arctic sea ice pack is declining at rapid rates, with the smallest extent yet recorded in 2007 (Stroeve et al., 2007, 2008). Arctic warming has important ramifications for the rest of globe via the Earth's energy budget and sea-level rise, in particular because of strong positive feedbacks that amplify warmth and its effects at high latitudes (Serreze and Francis, 2006; IPCC, 2007; Meier et al., 2007).

With ~151,000 km^2 of the Canadian Arctic covered by glaciers and ice caps, and 36,800 km^2 of ice on Baffin Island and 4800 km^2 on Bylot Island, there is more than twice as much ice in the eastern Canadian Arctic as in Alaska (Andrews, 2002). Although Alaska's glaciers are thought to be a larger contribution to sea-level rise at present (Meier et al., 2007), the significant sources of ice in the eastern Canadian Arctic will undoubtedly be an equally if not more important contributor to future sea-level rise. Of all the ice in the Canadian Arctic, glaciers and ice caps on Baffin Island are providing roughly half the total contribution to sea-level rise, despite significantly less ice there (Abdalati et al., 2004).

Our understanding of ongoing and future changes in the cryosphere can benefit by taking a longer-term view of glacier fluctuations. For example, future glacier response to continued arctic warming (e.g., 5–6 °C by 2100 AD; ACIA, 2005) may be similar to how glaciers responded to past periods that were warmer than today. Records of temperature change in the eastern Canadian Arctic reveal a Holocene thermal maximum period that was warmer than present. Several records of Holocene climate change

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(e.g., pollen, marine mollusks, ice core oxygen isotopes; Kaufman et al., 2004) reveal a broad period of mildly warmer-than-present summer temperatures that span the middle Holocene. Other records, however, like the ice core melt-layer record from the Agassiz Ice Cap on Ellesmere Island and chironomid-inferred summer temperature records from lakes on northeastern Baffin Island reveal a thermal maximum spanning from ~10 to ~7 ka that was up to 5 °C warmer than present (Fisher et al., 1995; Briner et al., 2006a), temperatures similar to the Arctic Climate Impact Assessment (ACIA, 2005) projections for 2100 AD. In addition, a new chironomid-inferred summer temperature record that is higher resolution than prior lacustrine-based reconstructions spanning the early Holocene shows two pronounced periods of cooling that interrupted the thermal maximum at ~9.3 and ~8.5 ka (Axford et al., in press). Both episodes of abrupt cooling are apparent in ice core records from the Greenland Ice Sheet (Vinther et al., 2006). Although different proxies may be recording different environmental factors, all existing proxy data generally indicate that the eastern Canadian Arctic experienced warmer summers than today in the early Holocene, punctuated by discrete cold intervals and followed by colder-than-present summer temperatures in the late Holocene. Glacier changes during these temperature swings provide a basis for evaluating the sensitivity of the cryosphere to future climate change.

Here, we summarize key sites that constrain the extent and timing of latest Pleistocene and Holocene glaciation on Baffin Island. Although several recent papers summarized the Holocene climate history of the North American Arctic (CAPE, 2001; Gajewski and Atkinson, 2003; Kaufman et al., 2004; Kerwin et al., 2004; Miller et al., 2005), there has yet to be a synthesis of Holocene glaciation from the eastern Canadian Arctic. Miller et al. (2005) discussed Holocene glaciation of Baffin Island, but mainly focused on results from four lacustrine sites, two of which are in proglacial environments. Thus, we summarize what is known, and unknown, about both Laurentide and alpine glaciation on Baffin Island from the latest Pleistocene through the 20th century.

Ages in this paper are reported in ka (thousands of years ago). All radiocarbon ages were calibrated using CALIB 5.0.1 (Stuiver et al., 2005) and are reported to the nearest 100 years with 95% confidence limits. Radiocarbon ages younger than 2 ka are reported as years AD. The cosmogenic exposure age-based chronologies summarized in this paper were based on the following: AMS measurements at Lawrence Livermore National Laboratory, ^{10}Be production rate of ~5.1 atoms g^{-1} yr^{-1}, standardized with KNSTD3370, and ^{10}Be half-life of 1.5×10^6 yr. Balco et al. (in press) determined a locally calibrated ^{10}Be production rate for Baffin Island that is 12% lower (corresponding to ages 12% older) than used in previous publications.

2. Continental deglaciation from the Laurentide Ice Sheet to the Barnes Ice Cap

The Laurentide Ice Sheet (LIS) deglaciation history of Baffin Island and surrounding areas is broadly known. Prior to the application of cosmogenic exposure dating, the deglaciation chronology was based almost solely on radiocarbon dating of glacio-marine deposits, which are younger than 10 ka almost everywhere across the eastern Canadian Arctic. Early work by Blake (1966, 1970, 1975, 1992) suggested that the LIS on southern Baffin Island and an Innuitian Ice Sheet in the Canadian High Arctic extended onto the continental shelves during the late Wisconsinan. But, the lack of radiocarbon-dated glacio-marine deposits nevertheless led to a paradigm shift in the 1960s from depicting a massive ice sheet (Fling, 1943) to a much more limited extent of late Wisconsinan ice in the Canadian Arctic to sub-arctic Labrador, with ice margins commonly believed to be at fiord heads (Ives, 1964, 1977, 1978; Andrews, 1966, 1987, 1989; Løken, 1966; England, 1976a,b, 1983, 1992; Dyke, 1979; Dyke et al., 1982; Dyke and Prest, 1987).

Offshore marine studies (Jennings, 1993; Jennings et al., 1996) in the Cumberland Sound area (Fig. 1), followed by cosmogenic exposure dating studies that took place over the last decade (e.g., Marsella et al., 2000; Kaplan et al., 2001; Miller et al., 2002; Briner et al., 2005, 2006b; Davis et al., 2006) led to a second paradigm shift for a more extensive late Wisconsinan ice extent on Baffin Island, although not as large as had been earlier postulated. Further work from the Canadian High Arctic (England, 1998, 1999; Zreda et al., 1999; Dyke and Savelle, 2000; Dyke and Hooper, 2001; Dyke et al., 2002, 2005; England et al., 2005) and in Labrador (Clark, 1988; Clark et al., 1989, 2003; Marquette et al., 2004) led to similar conclusions, and improved previous knowledge on the Last Glacial Maximum (LGM) configuration and subsequent latest Pleistocene retreat chronology. And, the compilation of radiocarbon ages by Dyke et al. (2003) has provided a relatively well-constrained retreat history of the LIS over the Holocene as it diminished in size to what is now called the Barnes Ice Cap. Details are still lacking, however, and the chronology of ice margin history during periods of documented abrupt climate changes in the early Holocene remains poor (Vinther et al., 2006; Axford et al., in press).

From an LIS margin that mostly terminated on the continental shelves around Baffin Island during the LGM, deglaciation exposed landscapes first on Cumberland Peninsula and the distal lowlands along the northeast coast (Dyke et al., 2002; Miller et al., 2002; Briner et al., 2005). At the onset of the Holocene, there remained a sizeable Foxe Dome of the LIS on Baffin Island. Many of the large and deep sounds and fiords rimming Baffin Island rapidly deglaciated between 12 and 10 ka (Andrews et al., 1970; Andrews, 1987, 1989; Kaplan and Miller, 2003; Briner et al., 2007). As one example, an ice stream in Lancaster Sound (Fig. 1) retreated over 1000 km in probably 500 years or less, from ~10.8 to 10.2 ka (Dyke et al., 2003).

In Cumberland Sound (Fig. 1), which was once believed to be ice free since the early Wisconsinan (Dyke, 1979; Dyke et al., 1982), recent work suggests that this area, including the adjacent fiords, was full of Laurentide ice during the late Wisconsinan, including the Younger Dryas (Jennings, 1993; Jennings et al., 1996; Kaplan et al., 1999, 2001; Marsella et al., 2000; Kaplan and Miller, 2003). Work over the past two decades also suggests extensive late Wisconsinan ice in the Frobisher Bay area, including Hudson Strait and Meta Incognita Peninsula, with a complex sequence of advances and retreats (Miller, 1980; Miller et al., 1988; Miller and Kauffman, 1990s).

Despite overall retreat, LIS outlet glaciers experienced several advances during the early Holocene. The best-dated moraines were deposited during the Cockburn Substage, defined by Andrews and Ives (1978) to be between 9000 and 8000 ^{14}C yr BP. Little progress has been made in further dating or mapping of early Holocene moraines, and Andrews and Ives (1978) remains the most comprehensive review to date. Moraines of Cockburn age are widely traceable across southern, eastern, and northern Baffin Island (Ives and Andrews, 1963; Falconer et al., 1965; Miller and Dyke, 1974). The chronology for moraines deposited during the Cockburn Substage is based on radiocarbon ages on marine mollusks where ice margins are associated with fossiliferous marine deposits. Calibrating the radiocarbon ages reported in Andrews and Ives (1978) reveals that the LIS advanced between ~9.5 and 8.5 ka (Table 1), with possible modes centered around 9.1 and 8.7 ka (Fig. 2).

At the time of the Andrews and Ives (1978) compilation, moraines of Cockburn age were regarded as the maximum limit of the LIS during the late Wisconsinan, and thus were thought to record a major ice advance. As currently understood, moraines of Cockburn age mark several well-expressed (geomorphically) and well-dated advances during overall LIS retreat from an LGM limit at

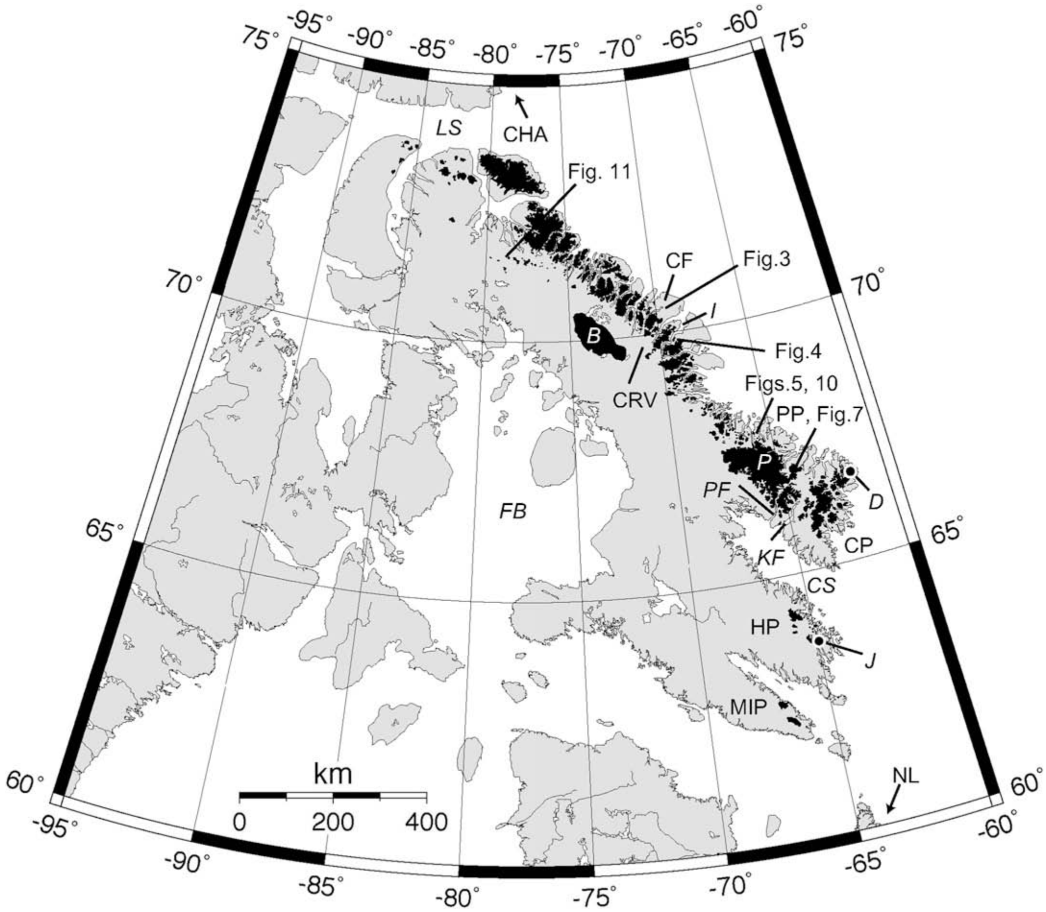

Fig. 1. Baffin Island and its current (ca 1960s) cover of alpine glaciers and ice caps. B, Barnes Ice Cap; P, Penny Ice Cap; LS, Lancaster Sound; CF, Clyde Foreland; I, Inugsuin Fiord; D, Donard Lake; CP, Cumberland Peninsula; CS, Cumberland Sound; PF, Pangnirtung Fiord; KF, Kingnait Fiord; PP, Pangnirtung Pass; MIP, Meta Incognita Peninsula; HP, Hall Peninsula; FB, Foxe Basin; J, Lake Jake; CRV, Clyde River valley; CHA, Canadian High Arctic; NL, northern Labrador.

the outer coast. Although there is a rich morphostratigraphic record of LIS retreat between the heads of fiords and sounds on eastern and southern Baffin Island and the Barnes Ice Cap (the Baffinland Drift), it is generally only where ice limits can be tied to fossiliferous marine deposits that the ice margins are well dated (Andrews and Ives, 1978; Dredge, 2004).

Following the deposition of moraines during the Cockburn Substage, continued LIS deglaciation was repeatedly interrupted by stillstands and advances. Many fiord heads along eastern Baffin Island still contained outlet glaciers, and ice margin deposits have been dated in a few locations that suggest standstills or advances of ice margins around 8 ka. Briner et al. (2007) report an ice-contact

Table 1
Radiocarbon ages on bivalves from Baffin Island raised marine deposits.

Lab. No.	Relationship to ice margin[a]	Radiocarbon age (^{14}C yr BP ± 2S.D.)	Correction[b] (yr)	Calibrated and corrected age (ka ± 2S.D.)[c]	Reference
CURL-7038	Ice-contact/minimum	7620 ± 80	0	7.89 ± 0.08	Briner et al. (2007)
Y-1834	Ice-contact	7820 ± 280	0	8.08 ± 0.28	Andrews and Ives (1978)
GSC-1064	Unknown	7890 ± 320	410	8.56 ± 0.22	Andrews and Ives (1978)
I-1932	Minimum	7940 ± 260	410	8.66 ± 0.36	Andrews and Ives (1978)
I-1673	Minimum	7970 ± 680	410	8.73 ± 0.80	Andrews and Ives (1978)
CURL-7046	Minimum	8050 ± 70	0	8.33 ± 0.08	Briner et al. (2007)
GSC-1060	Unknown	8090 ± 320	410	8.86 ± 0.24	Andrews and Ives (1978)
I-193	Maximum	8210 ± 260	410	9.09 ± 0.48	Andrews and Ives (1978)
GSC-462	Ice-contact/minimum	8230 ± 580	410	9.06 ± 0.42	Andrews and Ives (1978)
GaK-3092	Ice-contact	8290 ± 340	0	8.60 ± 0.46	Andrews and Ives (1978)
I-724	Ice-contact	8350 ± 600	410	9.41 ± 0.76	Andrews and Ives (1978)
GSC-1638	Ice-contact	8410 ± 680	410	9.26 ± 0.38	Andrews and Ives (1978)
Y-1830	Ice-contact/maximum	8430 ± 280	0	8.77 ± 0.36	Andrews and Ives (1978)
GX-0930	Maximum	8435 ± 210	0	8.77 ± 0.30	Andrews and Ives (1978)
GSC-813	Ice-contact/maximum	8630 ± 380	410	9.51 ± 0.20	Andrews and Ives (1978)
GSC-2183	Ice-contact/minimum	8660 ± 220	410	9.52 ± 0.12	Andrews and Ives (1978)
St-3816	Ice-contact	8760 ± 700	0	9.11 ± 0.88	Andrews and Ives (1978)
GaK-5479	Unknown	8930 ± 360	0	9.34 ± 0.44	Andrews and Ives (1978)

[a] This column indicates the age relationship of the datable material to the ice margin.
[b] 410 yr was added to GSC and I laboratory ages because they were originally reported with assumed δ^{13}C values of −25 per mil instead of 0 per mil.
[c] Radiocarbon ages were calibrated using CALIB 5.0.1 (Stuiver et al., 2005); bivalves were corrected 550 years for the reservoir offset.

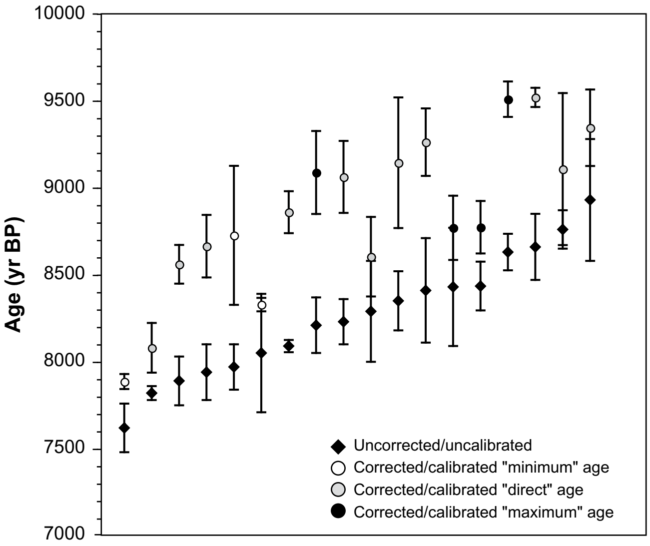

Fig. 2. Radiocarbon ages reported in Andrews and Ives (1978) that constrain the age of LIS outlet glacier advances during the Cockburn Substage. Two ages reported in Briner et al. (2007) have been added. Radiocarbon ages that date the moraines deposited during the Cockburn Substage have never before been calibrated, thus the marine-reservoir-corrected and calibrated ages (CALIB 5.0.2) presented here demonstrate that moraines assigned to the Cockburn Substage were formed at various times spanning between ∼9.5 and ∼8.2 ka, with clusters of ages around 9.1 and 8.7 ka. Minimum, direct and maximum age constraints for ice limits of the Cockburn Substage are labeled as they are described in Andrews and Ives (1978).

delta that was deposited between 8.5 and 7.9 ka. Ice-contact deltas in adjacent Inugsuin and Sam Ford fiords have similar ages (e.g., Løken, 1965). Thus, there appear to be ice margins that potentially represent a response to cooling associated with the 8.2 ka event. On the other hand, it remains difficult at present to evaluate the significance of any single ice margin because few have been dated.

On northwestern Baffin Island, Dyke and Hooper (2001) mapped LIS margin retreat at unprecedented detail. Over 50 radiocarbon ages and detailed ice margin mapping from 1:60,000 air photos delimit deglaciation of the entire northwestern sector of Baffin Island between 11 and 6 ka. A significant advance in Navy Board Inlet culminated ∼10.5 ka. The pattern of retreat on the millennial timescale is generally steady, although on shorter timescales retreat is interrupted by stillstands and advances. Sizeable moraines were built during the Cockburn Substage, and at least one advance is dated to ∼8.2 ka. However, moraines formed during these two intervals are part of a landform assemblage consisting of other morainal deposits formed between 11 and 6 ka (Dyke and Hooper, 2001).

Fiords that penetrate deeply into Baffin Island contain post-Cockburn-age moraines that are associated with marine deposits. For example, in Sam Ford and Kangok fiords, prominent ice margins are as young as ∼7 and ∼6 ka, respectively, after which time the shrinking eastern margins of the LIS terminated almost entirely on land (Andrews et al., 1970; Andrews and Ives, 1978). Following the retreat of the LIS out of most fiord heads on east-central Baffin Island at 7–6 ka, and a sizeable collapse of the ice sheet in Foxe Basin, the residual Foxe Dome became isolated as the terrestrial-based Early Barnes Ice Cap. Through the middle Holocene, the Early Barnes Ice Cap slowly and fairly steadily retreated to its present configuration by ∼2 to ∼1 ka. Andrews and Barnett (1979) summarize the middle–late Holocene retreat chronology of the Early Barnes Ice Cap, and Dyke et al. (2003) provide the most up-to-date maps of LIS retreat and disintegration. The Barnes Ice Cap lies west of the drainage divide on Baffin Island, damming several significant drainage systems, creating the large Conn and Bierler ice-dammed lakes; additional large proglacial lakes existed in the past when the Barnes Ice Cap extended farther north and south.

A Holocene chronology for the Barnes Ice Cap (Fig. 1) stems largely from lichenometrically dated moraines and associated proglacial lake shorelines (Andrews and Barnett, 1979). To the south of the Barnes Ice Cap, Andrews and Barnett (1979) propose a retreat chronology with distinct ice limits and associated lakes between 5 and 3 ka, and an ice margin within 10 km of the present Barnes Ice Cap by 2–3 ka. Briner et al. (2008) recently reported new ^{10}Be ages of deglaciation from striated bedrock in the upper Clyde River valley (Fig. 1), where Andrews and Barnett (1979) propose an ice margin ∼5 ka. The ^{10}Be chronology depicts a thinning lobe of ice that retreated from ∼540 m asl at ∼5.0 ka, from ∼400 m asl at ∼4.7 ka, and from the valley floor (320 m asl) by 3.5 ka. These ^{10}Be ages of deglaciation are remarkably close in age to radiocarbon- and lichen-determined ages from Andrews and Barnett (1979). Recession of the southern Barnes Ice Cap in the late Holocene arises from dated macrofossils in lacustrine deposits of ice-dammed Generator Lake (Barnett and Holdsworth, 1974; Andrews and Barnett, 1979; Dredge, 2004). Several phases of the lake are dated between ∼4 and ∼1 ka, and reveal that the Barnes Ice Cap retreated to near its present ice margin within the last millennium. One portion of the Barnes Ice Cap surged about 100 years ago; factors responsible for the advance remain uncertain (Barnett and Holdsworth, 1974; Andrews and Barnett, 1979)

On the north and west sides of the Barnes Ice Cap, several studies delimit the retreating Early Barnes Ice Cap in the middle and late Holocene (Ives and Andrews, 1963; Ives, 1964; Andrews, 1966; Dyke, 2008). Dyke (2008) mapped the detailed retreat pattern and chronology of Steensby Inlet, western Baffin Island, between 6 and 4 ka, and shows that the Early Barnes Ice Cap retreated out of Steensby Inlet and onto Baffin Island ∼5 ka. Several levels of ice-dammed lakes, around which a detailed geomorphic record is preserved, are linked to fluctuations of the northern Barnes Ice Cap. Lake shorelines are dated by lichenometry, and reveal smaller ice-dammed lake extents resulting from a retreating ice margin between ∼4 ka and the present (e.g., Ives and Andrews, 1963; Andrews and Barnett, 1979). Finally, lichenometry suggests that an advance of the Lewis outlet glacier occurred between 1500 AD and 1900 AD during the Little Ice Age (LIA; Andrews and Barnett, 1979).

An important feature of the deglaciation of the LIS/Barnes Ice Cap is the dynamic component of ice flow. The rapid deglaciation of sounds and fiords was in part driven by high calving rates (e.g., Andrews et al., 1985; Kaplan et al., 2001). Similarly, the development of ice streams on Baffin Island likely controlled the retreat pattern in at least some sectors. In particular, several transient ice streams operated on northern Hall and Cumberland peninsulas (Fig. 1), and in northernmost Foxe Basin during the early–middle Holocene (De Angelis and Kleman, 2007). Although data are sparse on the age and style of these ice streams, a prominent ice stream (Steenbsy Inlet Ice Stream) apparently operated for only ∼500 years, responding to the collapse of the LIS ∼6.5 ka in northern Foxe Basin (Dyke, 2008). Finally, major surges of ice sourced in the Labrador sector crossed Ungava Bay and the mouth of Hudson Strait (Fig. 1), reaching southern Hall Peninsula during the Gold Cove advance (11.3–11.0 ka) and southern Meta Incognita Peninsula during the Noble Inlet advance (10.0–9.5 ka; Kaufman et al., 1993; Veillette et al., 1999; Andrews and MacLean, 2003).

3. Early Holocene alpine glacier extent

Evidence from locations on Baffin Island most distal to Foxe Basin suggests that most alpine glaciers were behind their LIA margins during LIS deglaciation. For example, a sediment core from Donard Lake, Cumberland Peninsula (66°39′50″ N, 61°47′10″ W;

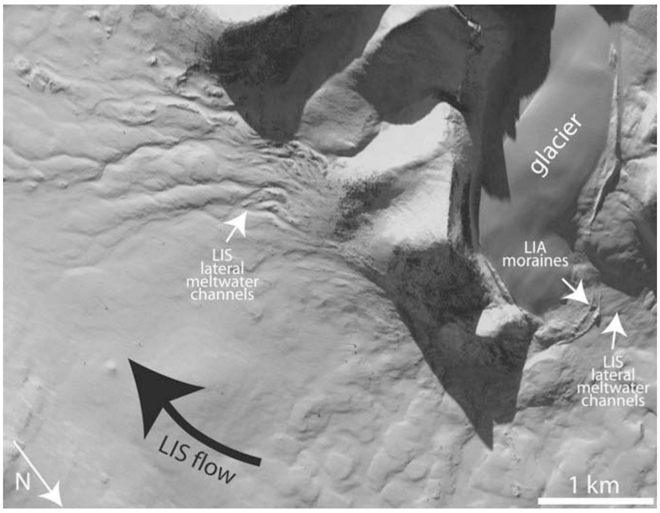

Fig. 3. Image from Google Earth showing the relationship between an extant glacier on the Clyde Foreland (Fig. 1) and its LIA moraine with lateral meltwater channels created by the LIS during an ice marginal position dated at 14.0 ± 0.7 ka (Briner et al., 2005). The relationship indicates that the glacier has not been more extensive than its LIA position since ∼14 ka, a time span that includes the Younger Dryas, Cockburn Substage, and 8.2 ka event.

Fig. 1), which lies a few hundred meters beyond a voluminous LIA moraine (fronting the Caribou Glacier), has basal radiocarbon ages of ∼14 ka (Moore et al., 2001; Miller et al., 2005). The absence of moraines between the LIA moraine and Donard Lake indicates that the Caribou Glacier reached its most extensive position since ∼14 ka during the LIA. Similarly, on the Clyde Foreland (Fig. 1), northeastern Baffin Island, LIA moraines from a small valley glacier (70°29′ N, 69°21′ W) bury lateral meltwater channels formed by the LIS as it retreated from the Clyde Foreland during deglaciation (Fig. 3). Cosmogenic exposure ages of the meltwater channel set, including one cobble resting on bedrock just beyond the LIA moraines, indicate that LIS deglaciation of the area took place 14.0 ± 0.7 ka ($n = 5$; Briner et al., 2005). While the geomorphic relationship between this LIA moraine and LIS deglaciation has been long recognized (Miller, 1976), this new deglaciation chronology updates the previous work. Thus, alpine glacier records from distal locations on Baffin Island that deglaciated prior to the Younger Dryas indicate that the most extensive glacier advance since ∼14 ka, including during the Younger Dryas, occurred during the LIA (Mangerud and Landvik, 2007).

Additional evidence from northeastern Baffin Island that alpine glaciers reached their largest Holocene extents during the LIA comes from radiocarbon ages of basal sediments in lakes within 2 km of two different glacier snouts near Inugsuin Fiord (Fig. 1). In the first setting, two separate lakes that lie within 2 km of Igloo Door Glacier (informal name; 69°50′ N, 68°43′ W) have basal radiocarbon ages on macrofossils of 10.4 ± 0.2 ka and 10.2 ± 0.4 ka (Briner, unpublished data); in both sediment cores, the macrofossils rest immediately above glacier-derived sand. There are no moraines between the LIA moraine and the lakes, suggesting that the largest extent of the glacier since the earliest Holocene was during the LIA. The ages likely represent the timing of LIS deglaciation of the valley because the ages are similar to ages of deglaciation in adjacent fiords (Briner et al., 2007). At the second site, an ice-cored LIA moraine is nested within a well-vegetated moraine with subdued morphology that dams Lake PL3 (informal name; 69°48′ N, 69°09′ W; Fig. 4) in a trunk valley tributary to Inugsuin Fiord. The radiocarbon age of macrofossils in a core from Lake PL3 immediately above diamicton is 10.3 ± 0.1 ka, and there is no significant minerogenic signature in the lake sediments between the dated level and the sediment surface (Briner, unpublished data). Thus, the pre-late Holocene moraine is older than 10.3 ka, and unlike at the other glaciers discussed above, this glacier appears to have built a prominent moraine beyond its LIA extent during the earliest Holocene. Neither glacier was more extensive than they were during the LIA since ∼10.2 to 10.4 ka, demonstrating maximum middle–late Holocene glacier extent during the LIA. The precise age of the pre-10.3 ka moraine and its climatic significance remain unknown.

There is some evidence that alpine glaciers may have advanced after 10 ka but prior to Neoglaciation. Miller (1973a,b) reported two settings near Okoa Bay, northern Cumberland Peninsula, where alpine glaciers built moraines during the early Holocene. At one site, an alpine glacier moraine is coeval with the local marine limit,

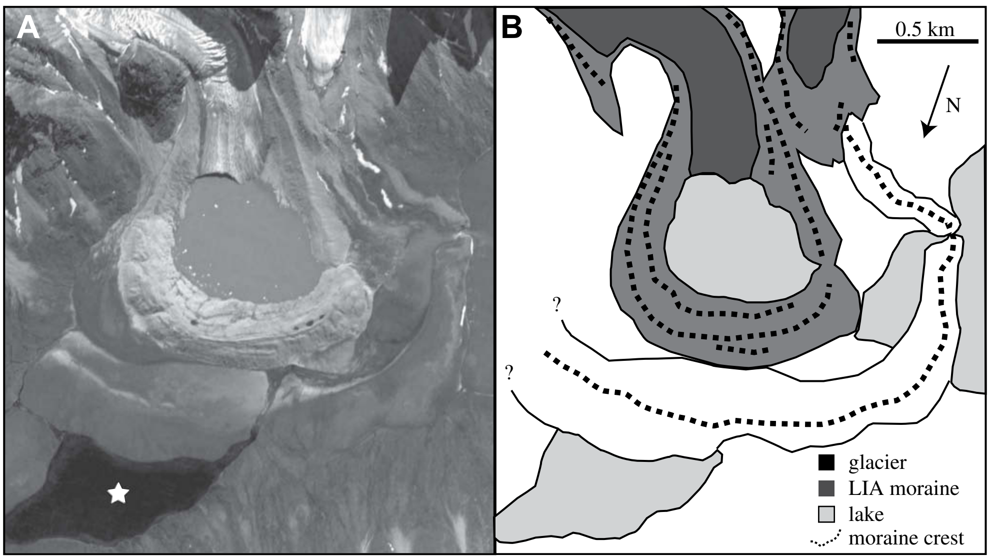

Fig. 4. Image from Google Earth (A) and schematic trace (B) showing two mountain ice cap outlet glacier snouts and their LIA moraines. Location near Inugsuin Fiord (Fig. 1). The glaciers merged in the past and formed an older moraine just outboard of the LIA moraines. This moraine is more subdued and vegetated than the LIA moraines. Lakes are dammed by the LIA moraine and older moraine. Basal lake sediments from a core (star) from the lake dammed by the older moraine are 10.3 ± 0.6 ka (Briner, unpublished data), indicating that the older moraine formed prior to this time.

which dates to 9.2 ± 0.1 ka (average of two radiocarbon ages; Miller, 1973a,b), suggesting local glacier activity during the Cockburn Substage (Andrews and Ives, 1978). At a second site, a series of prominent moraines were deposited ~5 km downvalley from a compound cirque with nine extant cirque glaciers (Fig. 5) that have been considered to be coeval with early Holocene LIS fluctuations (Miller, 1973a,b).

Evidence for early Holocene glaciation is also contained in lake sediment records. Although there is no moraine record of early Holocene activity of the Caribou Glacier, Donard Lake cores contain an interval of laminated minerogenic sediments (bounded by organic-rich sediments) between ~10 and 8 ka (Moore et al., 2001; Miller et al., 2005). Lake Jake lies 2.5 km downvalley from an outlet glacier of the Cornelius Grinell Ice Cap, Hall Peninsula (Fig. 1), and 1.5 km downvalley from its LIA moraine. Radiocarbon-dated basal sediments from the distal basin in the two-basin lake suggest that deglaciation occurred ~9.5 ka. The proximal basin in Lake Jake was glaciated during an advance of the outlet glacier. Three radiocarbon ages from terrestrial macrofossils reworked into diamicton in a sediment core from the proximal basin average 8.2 ± 0.2 ka (Miller et al., 2005). The advance is expressed as an increase in minerogenic sediments in the distal basin. These data suggest that local glaciers advanced in response to the 8.2 ka event for least one location on Baffin Island (Alley and Ágústdóttir, 2005).

4. Late Holocene alpine glaciation

After experiencing generally restricted conditions in the early Holocene, mountain glaciers on Baffin Island expanded in the late Holocene, beginning in some localities as early as the middle Holocene. The Barnes Ice Cap, on the other hand, was generally retreating throughout the entire Holocene, although it experienced stillstands in the late Holocene that correspond to alpine glacier advances elsewhere on Baffin Island (Andrews and Barnett, 1979). Thus, the expansion of glaciers on Baffin Island during the late Holocene mainly concerns alpine glaciers and mountain ice caps.

The re-advance of alpine glaciers during Neoglaciation is clearly illustrated by voluminous fresh moraines, usually ice cored, that front extant glaciers across Baffin Island. In addition, prominent lichen-kill trimlines are widespread, surrounding small cold-based mountain ice caps and sizeable portions of high-elevation plateaus on north-central Baffin Island. Smaller scale patches of lichen-kill terrain are also widespread, and likely document persistent summertime exposure of previously permanently snow-covered ground. Because ice extent during the LIA exceeded the extent of glaciers during other pulses of expansion in the late Holocene, evidence for the timing of pre-LIA maxima is sparse.

The most detailed records of late Holocene alpine glaciation are from Cumberland Peninsula, where Miller (1973a,b) and Davis (1985) constrained the timing of late Holocene glacier maxima with lichenometry. Fig. 6 provides a growth curve for *Rhizocarpon geographicum s.l.* (sensu lato is used because taxonomy of the group is too complex to distinguish to the species level in the field) on Baffin Island based on five ages from markers for gravestones of European whalers and six calibrated radiocarbon ages (Table 2).

Despite the LIA being the maximum late Holocene advance, pre-LIA moraines are preserved in some settings. Pre-LIA moraines may be preserved either where they occur on the upvalley lateral side of a glacier that flows into a trunk valley, or where large ice-cored moraines were simply too bulky to be overwhelmed by younger advances. For example, Spire Glacier (Fig. 7) near the head of Kingnait Fiord (Fig. 1) records an extensive sequence of large, ice-cored Neoglacial moraines and Tuktu Glacier in Pangnirtung Pass (Fig. 1) records an outer lateral moraine of early Neoglacial age, besides the common LIA moraines (Fig. 7).

The 85 moraines fronting 22 glaciers that Davis (1980, 1985) studied southeast of the Penny Ice Cap (Fig. 1) yielded several modes of lichen diameters that correspond to glacial advances culminating at ~3.5 ka, ~2.3 ka and ~900 AD. Maximum diameters of *R. geographicum s.l.* for these Neoglacial advances are summarized and compared with three other areas in the eastern Canadian Arctic in Fig. 8. LIA advances culminated at ~1350 AD, ~1600 AD and in the early 20th century (central Cumberland Peninsula; Fig. 9). The 110 moraines fronting 49 glaciers that Miller

Fig. 5. A prominent set of alpine glacier moraines on Cumberland Peninsula (Fig. 1) may have been deposited in the early Holocene. Moraines mapped as the equivalent to those shown here are tied to a marine limit dating to 9.16 ± 0.7 ka (Miller, 1973b).

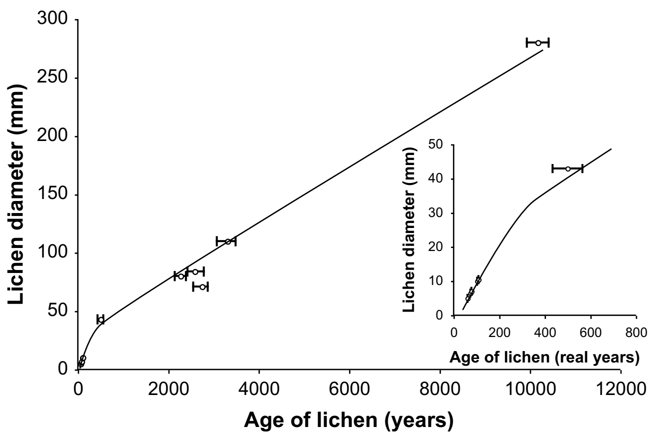

Fig. 6. Growth curve for *Rhizocarpon geographicum s.l.* on Baffin Island. Six calibrated radiocarbon ages used as control points for growth curve shown by 1 sigma error bars derived from CALIB 5.0.1, as summarized in Table 2. Five control points from gravestone markers for European whalers for younger part of growth curve discussed in Miller and Andrews (1972) and Miller (1973a,b).

Table 2
Radiocarbons and gravestone ages used as controls for growth curve of Rhizocarpon geographicum sensu lato on Baffin Island.

Sample	Radiocarbon age (^{14}C yr BP ± 2S.D.)[a]	Correction (yr)[b]	Mean calibrated age (2S.D.)	Minimum calibrated age (2S.D.)	Maximum calibrated age (2S.D.)	Lichen diameter (mm)	Material, location, and interpretation	Reference
Radiocarbon ages (Lab No.)								
GaK-3722	680 ± 160	250	1450 AD	1580 AD	1320 AD	43	Seal bone from floor of Thule house ruins	Miller (1973a)
GaK-1992	2400 ± 180	250	2.27 ka	2.03 ka	2.51 ka	80	Seal bone from floor of Dorset house ruins	Miller and Andrews (1972)
GSC-1304	2520 ± 300	0	2.59 ka	2.19 ka	2.99 ka	84	Detrital organics in deltaic sediments, Generator Lake	Andrews and Barnett (1979)
GSC-1315	2620 ± 300	0	2.75 ka	2.50 ka	2.99 ka	71	Detrital organics in deltaic sediments, Generator Lake	Andrews and Barnett (1979)
GSC-1276	3090 ± 340	0	3.31 ka	2.94 ka	3.67 ka	110	Detrital organics in deltaic sediments, Generator Lake	Andrews and Barnett (1979)
GaK-5479	8530 ± 360	400	10.17 ka	9.69 ka	10.65 ka	280	Shells in marine sediments related to moraine	Andrews and Barnett (1979)
Whaler gravestones (age AD)[c]								
1908 AD	NA	NA	61 yr	NA	NA	5	Date on wooden cross	Miller and Andrews (1972)
1896 AD	NA	NA	73 yr	NA	NA	6.5	Date on wooden cross	Miller and Andrews (1972)
1891 AD	NA	NA	78 yr	NA	NA	7	Date on wooden cross	Miller and Andrews (1972)
1866 AD	NA	NA	103 yr	NA	NA	10	Date on wooden cross	Miller and Andrews (1972)
1860 AD	NA	NA	109 yr	NA	NA	10.5	Date on wooden cross	Miller and Andrews (1972)

[a] Marine material ages were calibrated with a 450-yr reservoir correction.
[b] Correction added to the radiocarbon age before calibrating because GaK ages did not have their δ^{13}C ratios normalized to wood (−25 per mil).
[c] Calibrated ages in yr were determined for lichen diameter measurements made in 1969.

(1973a,b, 1975) and others (Carrara and Andrews, 1972; Mears, 1972) studied northeast of the Penny Ice Cap (Figs. 1 and 10) yielded generally similar modes of lichen diameters, although not entirely, to those measured by Davis (1985). The oldest Neoglacial advance culminated ~3.5 ka, with subsequent advances culminating ~500 AD, ~1300 AD, ~1600 AD and 1900 AD (northern Cumberland Peninsula; Fig. 9). The 22 moraines fronting six lobes of the Barnes Ice Cap studied by Andrews and Barnett (1979) and 11 moraine fronting four glaciers in the Torngat Mountains in northern Labrador (Fig. 1) studied by McCoy (1983) also indicate several pre-LIA advances (Fig. 8). Taken together, the lichenometrically-dated moraine record from Cumberland Peninsula shows glacier lengths only slightly smaller than LIA maxima as early as ~3.5 ka and subsequent similar maxima at ~2.3 ka, ~500 AD and ~900 AD. Alpine glaciers advanced in several discrete periods during the last millennium, reaching maxima at ~1350 AD and 1600 AD, and the most extensive late Holocene advance ~1900 AD.

The sediment record from Donard Lake (Fig. 1) provides a more continuous view of Neoglaciation on Cumberland Peninsula. Strongly laminated minerogenic sediments and corresponding high magnetic susceptibility values of Donard Lake sediments appear abruptly ~6 ka (Moore et al., 2001; Miller et al., 2005). Because the Caribou Glacier does not drain into Donard Lake until it thickens enough to cross a topographic threshold, the changes in lake sediments at ~6 ka document the thickening of the glacier at that time. The magnetic susceptibility of Donard Lake sediments increases dramatically, and the organic matter content decreases dramatically, between 2.5 and 2 ka. Thus, it appears that glaciers grew on Cumberland Peninsula as early as 6 ka, and Neoglaciation intensified 2.5–2 ka. If Caribou Glacier is representative of other alpine glaciers, then alpine glaciers on Cumberland Peninsula have been fluctuating more or less near their present margins since ~6 ka, with their most extensive position reached late during the LIA.

On northern Baffin Island, extensive areas of high-elevation plateaus lack lichen cover, suggesting extensive recent permanent snow and ice cover (Fig. 11). This lichen-kill has long been recognized, and assumed to represent vast landscapes that have been only recently deglaciated by expansive LIA ice caps and permanent snowfields (Ives, 1962; Falconer, 1966; Andrews et al., 1976; Locke and Locke, 1977). Davis and Wright (1975) suggested that LIA lichen-kill on Baffin Island represented an aborted initiation of full glaciation, now championed as the "overdue-glaciation" hypothesis (Ruddiman et al., 2005). Although an alternative hypothesis regards the lichen-free terrain as a steady-state feature of seasonally snow-covered landscapes (Koerner, 1980), there is little doubt that the extent of plateau ice caps and permanent snowfields is decreasing (see below). Building on previous research (Falconer, 1966; Andrews et al., 1976) that dated fossilized tundra vegetation exposed from beneath the retreating ice caps to constrain the timing of the lichen-kill event, Anderson et al. (2008) combined 50 new radiocarbon ages from such vegetation with radiocarbon-dated lake sediment cores and in-situ ^{14}C measurements in newly exposed quartz bedrock. The oldest tundra vegetation radiocarbon ages occur in clusters at 400 AD and 900 AD, suggesting that at least some ice caps have persisted on the plateau for well over one millennium and through the so-called Medieval Warm Period (Lamb, 1965). The largest modes in the radiocarbon ages, however, occur at ~1250 AD and ~1450 AD, suggesting significant expansion of plateau ice caps at these times (Fig. 9). Although the timing of initial recession of the northern plateau ice caps is not known, the lack of radiocarbon ages since ~1650 AD suggests that the plateau was extensively snow covered from the 17th century until the start of the 20th century. This timing is consistent with a lichenometry-constrained age of ~1700 AD for terrain just inside the trimline reported by Andrews et al. (1976). The lake sediment cores and in-situ ^{14}C measurements from quartz reported by Anderson et al. (2008) reveal that LIS deglaciation of the study area occurred ~6 ka, that the plateau remained ice free until 2.8 ka, and that the plateau has been glaciated for a total of 2000 years since then.

Although no detailed record of Neoglaciation exists from northeastern Baffin Island, two radiocarbon age constraints corroborate what has been found elsewhere. A birch leaf from lake sediments reworked into the LIA moraine of an alpine glacier on the Clyde Foreland (Fig. 3) dates to 3.6 ± 0.1 ka (Miller et al., 2005; the age reported here differs from that reported in Miller et al. (2005) because the former was incorrectly assigned a humic acid radiocarbon age and thus was corrected for an old carbon effect). Although a maximum age for Neoglaciation, the timing is consistent with the oldest moraine crests on Cumberland Peninsula. A second site, a sediment core from Igloo Door Lake, which lies ~1 km beyond Igloo Door Glacier, records the onset of the glacier's maximum Neoglacial advance. Approximately one decimeter of finely laminated silts and clays overlies several decimeters of organic-rich sediment, indicating that the glacier advanced across a topographic threshold and delivered sediment-laden meltwater into the lake, much like at Donard Lake. The radiocarbon age on humic acids extracted from bulk sediment beneath the minerogenic sediment unit is 1.1 ± 0.2 ka (Briner, unpublished data). Because radiocarbon ages on humic acids from elsewhere on Baffin Island are typically ~300 years older than macrofossil-based ages (Wolfe et al., 2004), the glacier likely reached its near-LIA extent after ~1200 AD.

To summarize, lake sediments from Donard Lake indicate that alpine glaciers on Baffin Island may have initialized Neoglacial expansion as early as ~6 ka. Most moraine records indicate near-LIA positions as early as 3.5 ka. Subsequent pre-LIA glacier maxima were attained during several intervals (~2.3 ka, ~500 AD and ~900 AD) prior to the last millennium. During the last millennium, glacier expansion recorded at widespread sites were centered ~1300 AD and 1450–1600 AD. The most extensive position of alpine glaciers was generally between 1650 and 1900 AD. Although the Barnes Ice Cap was retreating throughout the Holocene, its lichenometrically constrained record of stillstands/advances during overall retreat reveals a similar chronology, as summarized in Andrews and Barnett (1979).

5. Ongoing glacier change

Although little glacier monitoring is being conducted on Baffin Island, all available evidence indicates that glaciers and ice caps are melting at rapid rates. The geomorphic record and comparison of glaciers in recent LANDSAT imagery with earlier imagery clearly demonstrate widespread retreat of alpine glaciers throughout Baffin Island. For example, by comparing prominent LIA trimlines and moraines with modern ice limits defined from satellite imagery, Paul and Kääb (2005) show that 225 glaciers in a study area on Cumberland Peninsula have decreased on average 11% in area over the 20th century. By plotting the areal extent of small northern plateau ice caps (Fig. 11) over 50 yr of available imagery (aerial photographs and satellite imagery), Anderson et al. (2008) show that all such ice caps will disappear within the next few decades. Repeated airborne laser surveys in 1995 and 2000 document pronounced thinning on both the Barnes and Penny ice caps (Abdalati et al., 2004). The Barnes Ice Cap (5671 km^2) thinned even at its highest elevations, and is estimated to be decreasing by 3.5 km^3 of ice per year; although the alpine Penny Ice Cap (7335 km^2) slightly thickened at its highest elevations in the 5-year interval, it is estimated to be decreasing by 1.1 km^3 of ice per year (Abdalati et al., 2004). Abdalati et al. (2004) applied the empirical relationship of thickness change vs. elevation determined on the Penny Ice Cap to estimate the total volume of ice melt from all ice caps and glaciers in the eastern Baffin Island mountains. The additional glacierized area of 21,171 km^2 adds an additional decrease of ice volume on Baffin Island of 6.43 $km^3\,yr^{-1}$. In total, the amount of ice volume decrease on Baffin Island accounts for 45% of the ice volume decrease in the entire eastern Canadian Arctic, and contributed 0.029 $mm\,yr^{-1}$ to sea-level during the 5-year survey

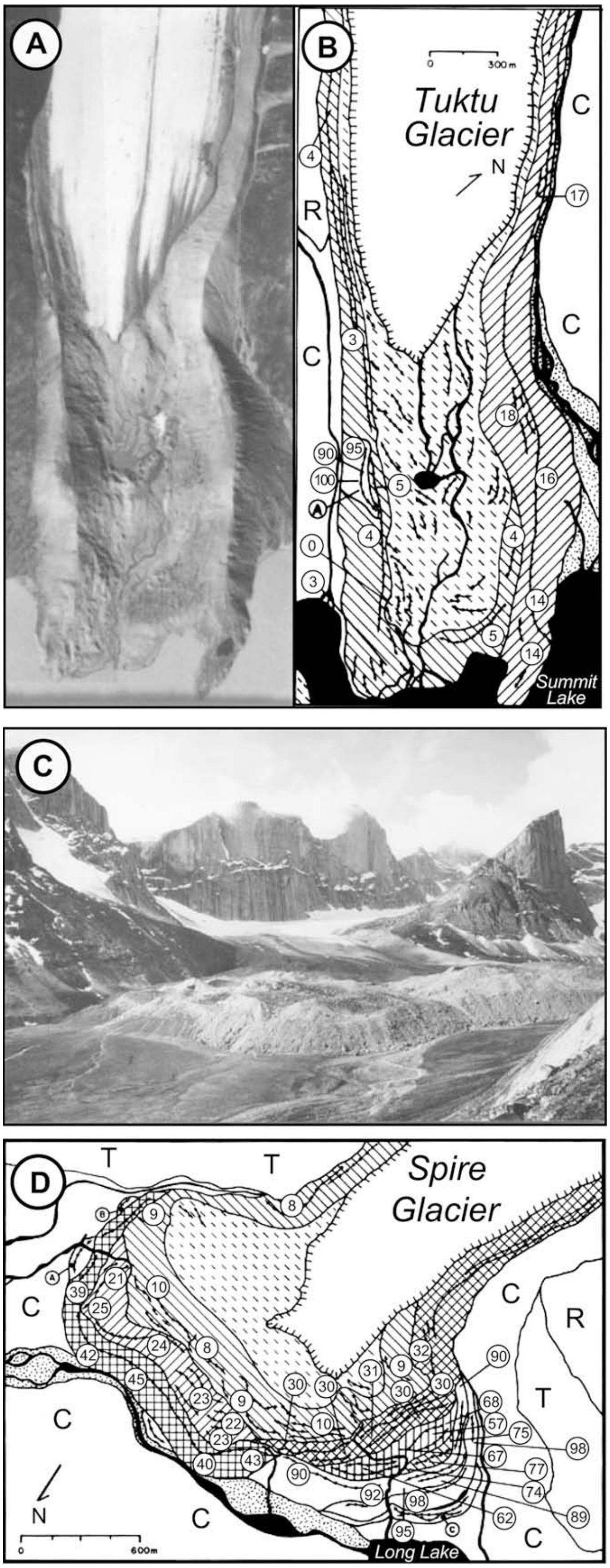

Fig. 7. (A) Vertical aerial photograph of Tuktu Glacier (66° 37.3′ N, 65° 14′ W), extending into south end of Summit Lake in Pangnirtung Pass (Fig. 1; 1959, originally 1:60,000 scale; reproduced with the permission of Natural Resources Canada 2008, courtesy of the National Air Photo Library). (B) Map of Neoglacial lateral moraines flanking Tuktu Glacier shown by lines with dots, along with glacial drift sheets indicated by different hachured patterns and maximum *Rhizocarpon geographicum s.l.* diameters (in mm) within circles. Location of soil profile in ice-cored moraine indicating early Neoglacial age shown with letter A in circle at left side of figure. Other labels include bedrock (R), talus (T), and colluvium (C). Outer left lateral moraine crest with 100 mm maximum diameters of *Rhizocarpon geographicum s.l.* suggest a Neoglacial advance ~3.3 ka. (C) Photograph looking south at moraines fronting Spire Glacier (66° 26′ N, 64° 48′ W) near head of Kingnait Fiord (Fig. 1). Large, steep, high ice-cored Neoglacial moraines overlie small, vegetated moraines, presumably Cockburn age, as seen at left. Spire Peak in the far background exhibits ~600 m relief. (D) Map with Neoglacial moraines fronting Spire Glacier shown by lines with dots, along with glacial drift sheets indicated by different hachured patterns and maximum *Rhizocarpon geographicum s.l.* diameters (in mm) within circles. Location of soil profile in low, non-ice-cored moraine indicating pre-Neoglacial age shown with letter A in circle at left side of figure. Other labels include bedrock (R), talus (T), and colluvium (C). Modified from Davis (1985).

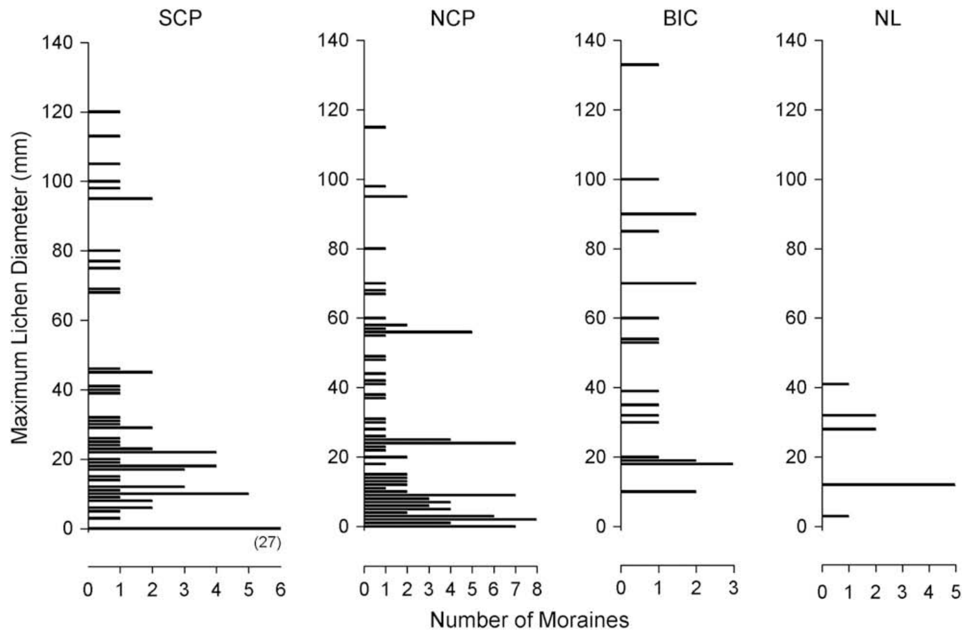

Fig. 8. Correlation chart for Neoglacial moraine records from three areas of Baffin Island and one area in the Torngat Mountains of northern Labrador. SCP, southern Cumberland Peninsula (Davis, 1985); NCP, northern Cumberland Peninsula (Carrara and Andrews, 1972; Mears, 1972; Miller, 1973b); BIC, Barnes Ice Cap (Andrews and Barnett, 1979); NL, northern Labrador (McCoy, 1983).

period (Abdalati et al., 2004). Somewhat surprising is that the Barnes Ice Cap has been steadily decreasing in volume despite relatively mild, and in some cases cool, summer temperatures over the study period (Jacobs et al., 1993, 1996). In addition, although the Barnes Ice Cap has been generally retreating throughout the 20th century, there were several decades in the middle of the 20th century with cooling summer temperature trends (Jacobs et al., 1993, 1996). Thus, the melting of the ice cap is not directly attributable to decadal-scale climate conditions, but rather is part of an ongoing response to a longer-term deglaciation. In any case, if current warmth is sustained or increases, mass loss of ice caps and glaciers on Baffin Island will continue to be pronounced.

6. Discussion and conclusions

Although much remains unclear about latest Pleistocene and Holocene glaciation of Baffin Island, the broad trends and timing of events are generally known. Initial recession of the LIS led to the most distal sites on Baffin Island (Cumberland Peninsula; northeastern coastal lowlands) to become ice free by ~14 ka. At two sites that were deglaciated prior to the Younger Dryas, alpine glaciers were larger during the LIA than any other time interval since late Pleistocene deglaciation, including during the Younger Dyras. Thus, alpine glaciers seem to have been smaller during the Younger Dryas than during the LIA, a surprising finding that has been found elsewhere in the North Atlantic (e.g., Mangerud and Landvik, 2007). It is intriguing that cooling during the LIA apparently led to more extensive alpine glaciation than during the Younger Dryas, which was likely a cold perturbation of greater magnitude. One explanation is that although the Younger Dryas interval was cold, it was also dry (Alley et al., 1993), thus, snowline lowered less than during the LIA when precipitation was closer to today's value.

Following the rapid retreat of the LIS in fiords and sounds around Baffin Island between 12 and 10 ka, the LIS retreated at slower rates, interrupted by advances, throughout the Holocene. A major decrease in areal extent occurred ~7 ka as the western sector of the Foxe Dome collapsed in Foxe Basin (Fig. 9). In contrast to the gradual retreat of the LIS throughout the Holocene, alpine glaciers re-grew in the middle–late Holocene from reduced extents during the early Holocene thermal maximum. Pleistocene ice resides at the base of the alpine Penny Ice Cap on Cumberland Peninsula (Fisher et al., 1998), thus at least some alpine glaciers on Baffin Island survived the thermal maximum. However, it remains unclear how much ice persisted through this interval that was up to 5 °C warmer in some locations (Briner et al., 2006a; Axford et al., in press).

Abundant evidence exists for ice sheet advances during overall retreat in the Holocene. Most widely acknowledged in the literature are datable ice limits at fiord heads along eastern Baffin Island within the Cockburn Substage. Calibrating the radiocarbon ages reported in Andrews and Ives (1978) suggests that the so-called Cockburn moraines represent multiple advances between 9.5 and 8.5 ka. There are several ice limits in fiord heads dated to ~8 ka, and younger ice limits dated to 7–6 ka. Once ice margins become land-based and are no longer in contact with fossiliferous marine deposits, potential for dating decreases. Nonetheless, existing chronologies reveal that the Barnes Ice Cap steadily decreased to its present extent, leaving behind a distinguishable moraine record spanning the middle and late Holocene. Thus, it is difficult to gauge whether ice sheet advances during the Cockburn Substage are more significant than other advances during the Holocene. Moraines of the Cockburn Substage are traceable across hundreds of kilometers. Indeed, these ice limits were once thought to represent the maximum position during the late Wisconsinan. On the other hand, focus on this interval may be disproportionate because the deposits are more datable than both younger and older ice limits. In any case, tighter chronologic control and more detailed comparison to high-resolution climate records are needed to assess more fully the significance of the early Holocene ice sheet advances.

Neoglaciation began as early as 6 ka, as demonstrated at Donard Lake on Cumberland Peninsula. That Caribou Glacier reached to within 1 km of its LIA margin as early as ~6 ka suggests that Holocene glacier changes on Cumberland Peninsula were of low amplitude, and that perhaps many alpine glaciers in addition to the high-elevation Penny Ice Cap (Fisher et al., 1998) persisted throughout the Holocene. In contrast, more widespread evidence

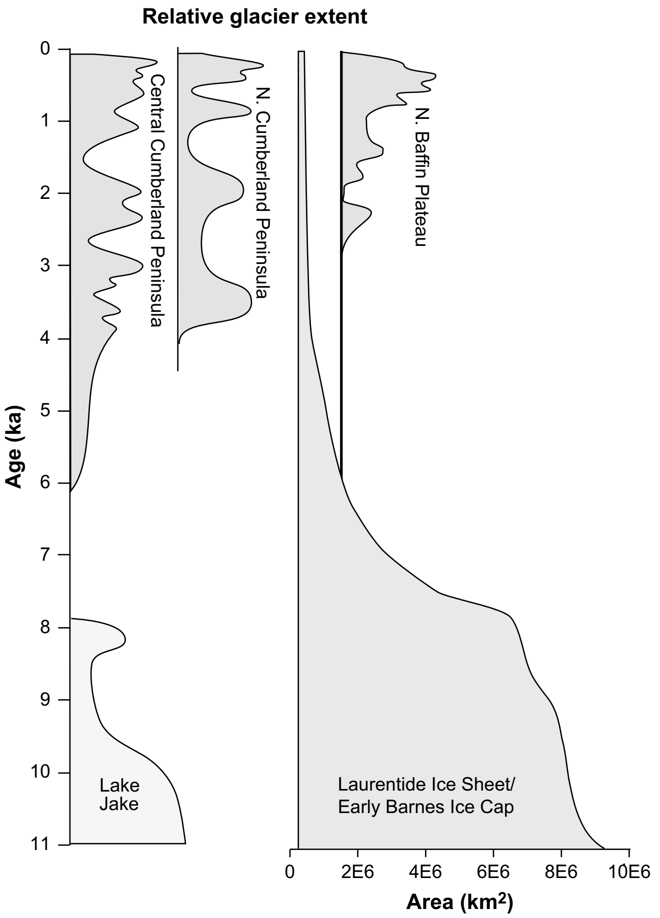

Fig. 9. Time–distance diagrams for glacier extent on Baffin Island through the Holocene. The curve for the LIS/Early Barnes Ice Cap is from digitizing the area of ice over Baffin Island/Foxe Basin at 18 time increments from 11.0 to 1.0 ka from Dyke et al. (2003). The boundary of the Baffin Island/Foxe Basin sector prior to the collapse between 8.4 and 7.8 ka is defined as the extent of ice at 7.8 ka (Cumberland Peninsula was not included in the LIS glaciation curve). Central and Northern Cumberland Peninsula curves are from Davis (1985) and Miller (1973a,b), respectively, and were redrafted according to approximate time in calendar years. Lake Jake and North Baffin Plateau curves based on information in Miller et al. (2005) and Anderson et al. (2008), respectively.

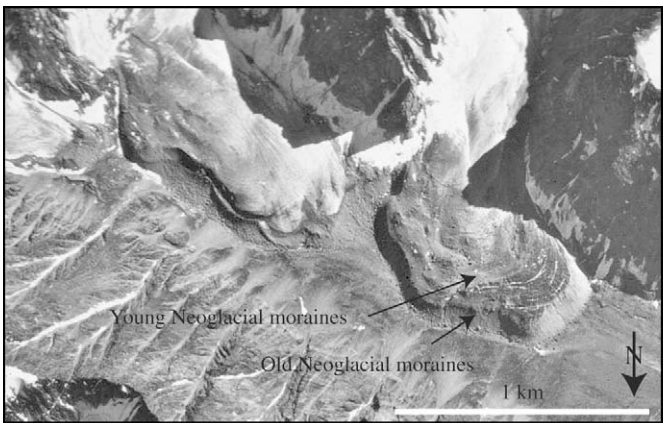

Fig. 10. Vertical aerial photograph showing ice-cored moraines (67° 38.5′ N, 65° 2′ W) on northern Cumberland Peninsula (ca 1960; originally 1:60,000 scale; reproduced with the permission of Natural Resources Canada 2008, courtesy of the National Air Photo Library). Outer moraine crests deposited by the glacier at right have lichenometric ages of ~3.2 ka, whereas inner moraines have lichenometric ages of <400 years (modified from Miller, 1973b).

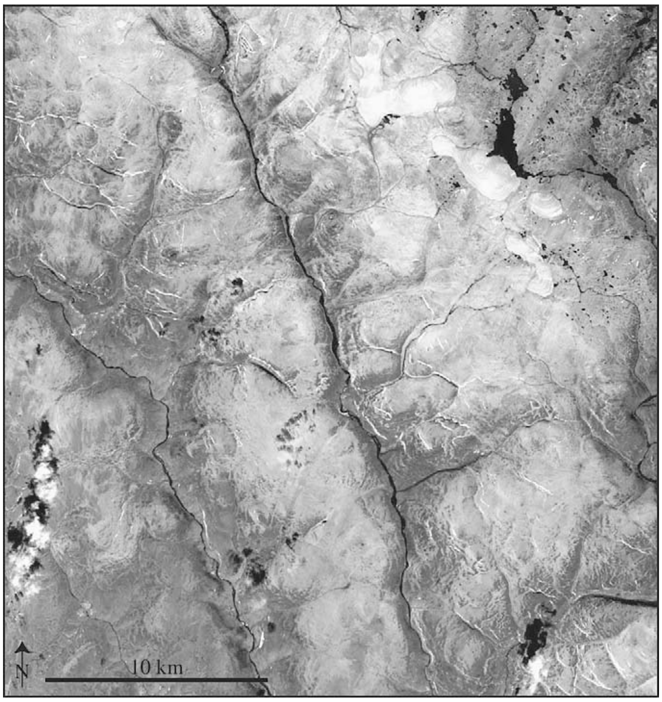

Fig. 11. Region of northern Baffin Island plateau landscape (Fig. 1) showing strong tonal variation that represents mature lichen cover of low-lying valleys and bare uplands that lack mature lichen cover (2000 AD LANDSAT image). Small remnants of once much larger plateau ice caps can be seen on the right side of the image (see Anderson et al., 2008).

from across Baffin Island suggests that most alpine glaciers did not significantly expand until ~3.5 ka. The apparent great magnitude of re-growth and 20th century retreat of ice caps on the northern Baffin plateau also strongly contrasts with the behavior of Caribou Glacier, whose terminus has not changed by more than 1 km in over 6000 years. The LIA seems to have been consistently the time period during which alpine glaciers reached their maximum extents during Neoglaciation. However, because pre-LIA moraine segments are often preserved just outboard of LIA moraine crests, glaciers were likely not dramatically larger during the LIA than during earlier Neoglacial maxima in most settings.

What is currently known about the climate history of Baffin Island is summarized in Kaufman et al. (2004) and Miller et al. (2005); however, several new records since then have used lacustrine-based climate proxies to quantify summer temperature changes in the Holocene (Briner et al., 2006a; Michelutti et al., 2007; Thomas et al., 2008; Axford et al., in press; Thomas and Briner, in press). Although little is known about precipitation on Baffin Island during the Holocene, several proxies record paleotemperature. Cooling followed a period between ~11 and 7 ka that was at least several degrees warmer than today. The timing of cooling corresponds well with the earliest evidence of Neoglaciation ~6 ka. However, most of the lacustrine-based climate proxy records show gradual cooling from ~6 to 7 ka until the LIA. Thus, a large amplitude climate change in the ~3.5 to ~2.5 ka interval that corresponds with the geologic evidence of pronounced Neoglaciation seems not to be recorded in paleoclimate proxy records. Perhaps the existing records lack the resolution to reveal such a climate change. Or, pronounced glacier expansion simply may be an artifact of the gradual lowering of snowline over high-elevation plateaus. In this case, accumulation zones would dramatically increase in area and lead to the pronounced expansion of alpine glaciers even under gradual climate change. In any case, both the glacier and climate proxy data recorded in lacustrine sediments and ice cores (e.g., Fisher et al., 1998; Miller et al., 2005; Anderson et al., 2008) depict the LIA as the coldest interval on Baffin Island during the Holocene. Glacier snowline and climate proxies suggest that the amplitude of LIA-20th century summer warming was 1–1.5 °C (e.g., Locke and Locke, 1977; Thomas and Briner, in press). Continued warming, as has been projected (ACIA, 2005; IPCC, 2007), will result in the continuation of the dramatic retreat of alpine glaciers witnessed in the 20th century.

Future research should focus on several important issues regarding the glaciation of Baffin Island that remain most uncertain. Foremost is the need to determine the sensitivity of Arctic glaciers and ice caps to climate change. Because projected warming by 2100 AD is of a magnitude comparable to the Holocene thermal maximum, improved reconstructions of glaciers during this interval should be targeted. Although moraine records tend to be incomplete, continuous proxy records from proglacial lakes may have potential to assess glacier change through the Holocene. In addition, mechanisms of rapid climate change need to be better understood. The pronounced expansion of glaciers in discrete pulses leading up to the LIA and rapid retreat of glaciers throughout the 20th century (e.g., Anderson et al., 2008), despite stable or slightly cooling summer temperatures spanning the middle 20th century, hints at non-linear responses to climate forcing mechanisms. Finally, in what fashion the significant amount of glacier ice (and sea-level equivalent) that is stored in the eastern Canadian Arctic responds to climate change, and the linkages with other Earth systems (snow cover, sea ice extent, ecosystems), needs to be better elucidated.

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