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Research paper

Using *in situ* cosmogenic ^{10}Be, ^{14}C, and ^{26}Al to decipher the history of polythermal ice sheets on Baffin Island, Arctic Canada

Jason P. Briner [a],[*], Nathaniel A. Lifton [b], Gifford H. Miller [c], Kurt Refsnider [d], Rebecca Anderson [e], Robert Finkel [f]

[a] University at Buffalo, Department of Geology, 411 Cooke Hall, Buffalo, NY 14260, USA
[b] Purdue University, Departments of Earth and Atmospheric Sciences and Physics, West Lafayette, IN 47907, USA
[c] INSTAAR and Department of Geological Sciences, University of Colorado, Boulder, CO 80303, USA
[d] Prescott College, 220 Grove Avenue, Prescott, AZ 86301, USA
[e] Alliance for Climate Education, 360 22nd Street, Oakland, CA 94612, USA
[f] Department of Earth and Planetary Sciences, University of California, Berkeley, CA 94720, USA

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ABSTRACT

Constraining the timing of past ice-sheet change is important for assessing the cryospheric expression of climate change and improving our understanding of ice sheet dynamics. Geochronology used to construct past ice-sheet reconstructions, however, can be ineffective in polar environments where ice sheets were polythermal and left varying imprints on landscapes. Cosmogenic-nuclide exposure dating, for example, is especially hampered by the lack of ice-sheet erosion and resultant cosmogenic nuclide inheritance. Here, we apply *in situ* cosmogenic ^{10}Be, ^{14}C and ^{26}Al methods to decipher various elements of the Laurentide Ice Sheet history of north-central Baffin Island. A clearly defined erosion boundary across the landscape reveals the transition in basal ice-sheet conditions as ice flow became channelized into northern Baffin Island fiords. ^{10}Be and ^{26}Al concentrations indicate that the boundary represents a juxtaposition of sliding, erosive ice and cold-bedded ice that preserved ancient bedrock that has not been significantly impacted by the ice sheet in perhaps one to two million years. We combine ^{10}Be measurements from ice-sculpted bedrock with measurements of *in situ* ^{14}C, which has no inheritance due to its quick decay during ice-sheet cover, to determine the local timing of deglaciation. The average ^{10}Be and *in situ* ^{14}C ages for upland deglaciation in north-central Baffin Island are 7.7 ± 0.9 and 8.4 ± 1.4 ka, respectively. Finally, *in situ* ^{14}C measurements from surfaces being uncovered by present-day retreat of small ice caps mantling uplands within the study area have concentrations too low to be compatible with continuous post-glacial exposure. These samples require shielding by ice for a significant portion of the Holocene, and more burial than during the Little Ice Age alone. Simple exposure-burial modeling suggests that 2400–2900 yr of total ice cover during Neoglaciation is required to explain measured *in situ* ^{14}C inventories. Combined, multiple cosmogenic nuclides with varying half-lives can be used to decipher many aspects of the history in landscapes occupied by polythermal ice sheets.

1. Introduction

Cosmogenic-nuclide exposure dating has emerged as a premiere tool for reconstructing glacier and ice sheet change (Balco, 2011). Users who first applied cosmogenic-nuclide exposure dating in high latitudes quickly realized that pervasive cosmogenic-nuclide inheritance (cosmogenic isotopes accumulated during prior periods of exposure; henceforth termed “inheritance”) is a significant complication (e.g., Brook et al., 1993, 1996; Marsella et al., 2000). Yet, several workers used inheritance to their advantage by making inferences about ice-sheet erosion, and hence, the nature of past ice-sheet basal thermal regimes and dynamics (e.g., Fabel et al., 2002; Stroeven et al., 2002; Briner et al., 2003, 2005; Marquette et al., 2004; Staiger et al., 2006). The disequilibrium of ^{26}Al and ^{10}Be in rock samples with inheritance was also used to model the history of ice-sheet occupation over long timescales (100 ky to My; e.g., Bierman et al., 1999; Stroeven et al., 2002; Sugden et al., 2005; Li et al., 2008). Combined, these relatively recent applications of cosmogenic nuclides in polar landscapes

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[*] Corresponding author. Tel./Fax: +1 716 645 6800.
E-mail address: jbriner@buffalo.edu (J.P. Briner).

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have contributed to tremendous progress in our knowledge of ice-sheet history and dynamics.

Despite this progress, there remain limitations to using cosmogenic-nuclide exposure dating in landscapes covered by polythermal ice sheets. First, it is difficult to date the last deglaciation in these landscapes, although avoiding bedrock and dating erratics has proven successful in some instances (Stone et al., 2003; Håkansson et al., 2007; Goehring et al., 2008; Stroeven et al., 2011). However, terrain with pervasive inheritance commonly yields erratics with inheritance, and thus cosmogenic-nuclide exposure ages of erratics do not always chronicle the timing of deglaciation. Second, given the relatively long half-lives of ^{26}Al and ^{10}Be ($\sim$0.7 and 1.4 My, respectively), their ratio is unable to resolve whether burial occurred within the last glaciation (post-Marine Isotope Stage 5e) or earlier (Steig et al., 1998; Bierman et al., 1999; Fabel and Harbor, 1999). Thus, without an independent source of information about the presence or absence of past ice cover, ^{26}Al and ^{10}Be measurements alone may not reveal whether or not a landscape was occupied by non-erosive ice during the last glaciation.

In situ ^{14}C, with its relatively short half-life (5730 years), enables one to circumvent the issues discussed above (Lifton et al., 2001). For example, while cover by non-erosive ice may not remove ^{10}Be or ^{26}Al from a landscape surface, *in situ* ^{14}C will decay to background levels when a surface is occupied by non-erosive ice for intervals longer than $\sim$30 ky. Therefore, the occupation of landscapes by non-erosive ice can be detected simply with measurements of *in situ* ^{14}C. Miller et al. (2006) used *in situ* ^{14}C measurements to determine deglaciation ages for summits in eastern Baffin Island and paired these *in situ* ^{14}C ages with ^{26}Al and ^{10}Be measurements to place additional constraints on long-term ice sheet erosion. Following Miller et al. (2006), Anderson et al. (2008) measured *in situ* ^{14}C in rock surfaces emerging from beneath small ice caps on uplands in north-central Baffin Island. The *in situ* ^{14}C inventory, influenced by the shielding of late Holocene ice-cap occupation, allowed Anderson et al. (2008) to constrain the duration of late Holocene ice-cap cover. Here, we combine *in situ* ^{14}C inventories with measurements of ^{10}Be and ^{26}Al to decipher the history of a landscape occupied by the polythermal northeastern Laurentide Ice Sheet (LIS). We add seven new *in situ* ^{14}C measurements to six measurements published by Anderson et al. (2008) and combine these with 16 new ^{10}Be and 2 new ^{26}Al measurements. With this dataset we evaluate a major landscape boundary that separates non-eroded from ice-sculpted terrain in northern Baffin Island, which in turn allows us to assess the long-term pattern of LIS erosion. Furthermore, we provide additional age constraints on the timing of deglaciation and subsequent re-growth of small ice caps during the Holocene.

2. Study area

North-central Baffin Island consists of a broad, low relief upland that ranges in elevation from $\sim$500–$\sim$800 m above sea level (asl) and hosts several small extant ice caps (Fig. 1). The upland is dissected by high-relief fiords along its northern and eastern margins, and gradually decreases in elevation into the Foxe Basin to the south and west. A visually well-defined landscape boundary separates terrain with obvious signs of glacial scouring to the north from terrain that lacks evidence of glacial scouring to the south (Fig. 2). The glacially scoured landscape consists of uneven terrain with exposed bedrock and with relief of up to 100s of meters and the presence of lakes; the non-scoured terrain has much lower relief, is covered by till, has rare outcrops of bedrock and the occasional meltwater channel, and has felsenmeer on mountain summits. Bedrock in this terrain is not glacially striated or polished, and includes various degrees of weathering of both bedrock outcrops and autochthonous blockfield (Fig. 3). The “erosion boundary” separating the two landscape zones runs irregularly across the northern side of the central Baffin uplands, and in our field area it separates the upland from the heads of valleys and fiords to the north. The elevation of the boundary varies and it does not seem to coincide with significant breaks in slope. The boundary is sharp in most locations, where ice-sculpted bedrock with straie appear a few 10s of meters from highly weathered bedrock; in some locations the transition from ice-sculpted bedrock to highly weathered bedrock spans hundreds of meters.

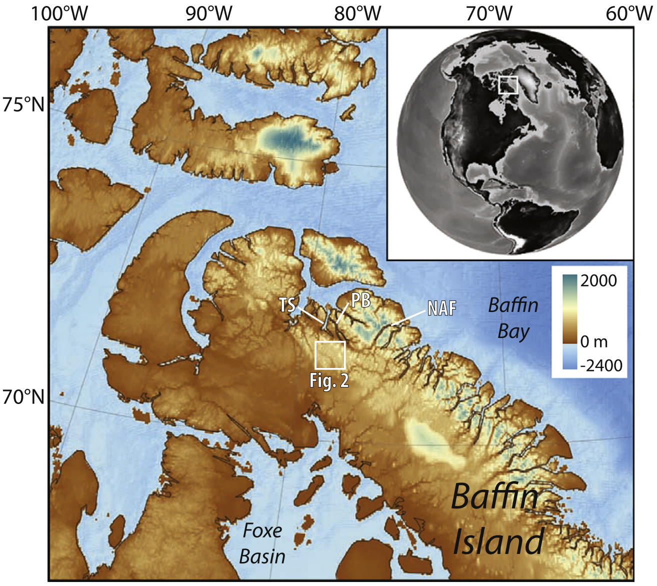

Fig. 1. Northern Baffin Island showing study area and context in northeastern North America (inset); TS = Tay Sound, PB = Paquet Bay, NAF = North Arm Fiord.

The glacial history of north-central Baffin Island has been the focus of a number of papers spanning the last half-century (e.g. Ives and Andrews, 1963; Falconer, 1966; Andrews and Barnett, 1979; Little et al., 2004; Utting et al., 2008; Refsnider and Miller, 2010). The LIS flowed northward and eastward across the plateau during the last glaciation, feeding ice streams that terminated in northern Baffin Bay (Dyke et al., 2003; De Angelis and Kleman, 2007). Staiger et al. (2006) measured ^{10}Be and ^{26}Al in till samples from northern Baffin Island and discovered a significant amount of cosmogenic nuclide inheritance in most of the high-elevation samples. This led Staiger et al. (2006) to show definitively that much of the upland escaped significant glacial scouring, as was previously hypothesized (Sugden, 1978; Andrews et al., 1985). Refsnider and Miller (2010) further support the antiquity of the north-central Baffin Island landscape by combining the cosmogenic radionuclide concentrations from Staiger et al. (2006) with till geochemistry measurements (Dredge, 2004; Utting et al., 2008) and burial/exposure modeling. The results reveal that till deposited on some parts of the north-Baffin landscape has been relatively immobile since 1.9–1.2 Ma, indicating at least minimally-erosive ice sheet conditions since that time.

Following the last glacial maximum, the LIS retreated from the continental shelf in northern Baffin Bay, through fiords, and eventually across north-central Baffin Island. Although recent research has focused on the timing of fiord deglaciation on southern (Marsella et al., 2000; Kaplan et al., 2001), eastern (Miller et al., 2002; Briner et al., 2007, 2009) and western Baffin Island (Dyke, 2008), there have not yet been any published detailed studies on the deglaciation of north-central Baffin Island. The deglaciation of eastern Baffin Island fiords typically occurred between 10 and 8 ka, followed by readvances in fiord heads $\sim$8 ka (Andrews and Ives,