Cosmogenic radionuclides from fiord landscapes support differential erosion by overriding ice sheets

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ABSTRACT

The interpretation of differentially weathered mountainous areas along the fringes of Pleistocene ice sheets is fundamental for determining ice-sheet behavior and thickness during the last glaciation. Two existing interpretations are either that highly weathered uplands remained as nunataks while freshly eroded troughs held outlet glaciers during the last glaciation or that uplands and lowlands were equally covered by ice, but that it was differentially erosive as a function of its spatially variable basal thermal regime. Cosmogenic radionuclide measurements from 33 bedrock samples and 27 upland erratics from differentially weathered fiord landscapes on northeastern Baffin Island shed light on Laurentide Ice Sheet (LIS) dynamics and thickness. Tors on weathered upland surfaces have minimum ^{10}Be ages between ca. 50 and ca. 170 ka (*n* = 12), whereas the majority of erratics perched in the uplands range from ca. 10 to ca. 13 ka (*n* = 14), indicating that the whole landscape was glaciated during the Last Glacial Maximum (LGM). Glacially sculpted bedrock (*n* = 7) near sea level reflects the age of deglaciation, and increasing amounts of isotopic inheritance in high-elevation sculpted bedrock (*n* = 3), higher elevation intermediately weathered bedrock (*n* = 11), and highest elevation intensely weathered bedrock reflect the weakening of erosive power as the ice sheet transitioned from fiords to interfiord plateaus.

Cold-based glaciation of uplands was contemporaneous with warm-based glaciation in fiords, suggesting the presence of ice streams. ^{26}Al and ^{10}Be concentrations measured in 10 bedrock samples indicate that tors have experienced a complex history of alternating periods of exposure, burial, and limited glacial erosion. The paired isotope data reveal a minimum duration of burial of 80–500 k.y., presumably by cold-based ice, and indicate that the tors have an age of at least 150–580 ka. These data, together with other reconstructions of ice streams in the eastern Canadian Arctic, suggest that ice streams were a large influence on northeastern LIS dynamics throughout the late Quaternary. Thus, the northeastern LIS was sensitively tied with North Atlantic thermohaline circulation and abrupt climate change.

Keywords: fjord, cosmogenic radionuclide, Laurentide Ice Sheet, weathering zone, glacier erosion, basal thermal regime.

INTRODUCTION

Ice sheets play an important role in landscape evolution and climate change. The role of ice streams and the variability in basal thermal regime in contemporary ice-sheet dynamics are becoming increasingly better understood (Hughes et al., 1985; Bentley, 1987; Fahnestock et al., 2001; Joughin et al., 1999). However, although we have an unobstructed view of the surface of contemporary ice sheets, their obscured beds make information regarding basal conditions difficult to obtain. On the other hand, the beds of former ice sheets are accessible and reveal a picture of highly dynamic behavior of past ice sheets. For example, the reconstruction of low-gradient outlet glaciers (e.g., Clark, 1994; Patterson, 1998), ice streaming (e.g., Stokes and Clark, 2001, 2002; Boulton et al., 2001; Kaplan et al., 2001), and frozen-bedded conditions (e.g., Dyke, 1993; Kleman, 1994; Dredge, 2000; Kleman and Hättestrand, 1999) in formerly glaciated regions, along with widespread ice-rafted debris in adjacent ocean basins (e.g., Bond et al., 1992; Andrews and Tedesco, 1992; Andrews et al., 1998) point toward Pleistocene ice sheets that have all the dynamic components of their modern-day counterparts. These reconstructions also add a longer-term perspective on contemporary ice-sheet dynamics, allowing for the rates and time scales of various processes to be addressed.

Although reconstructions of Pleistocene ice sheets have changed in recent decades from single-domed, stable ice masses to multi-domed, relatively dynamic systems, gaps in our understanding remain. For example, our understanding of the number and locations of former ice streams (Stokes and Clark, 2001), details of the spatial patterns of frozen- and warm-bedded conditions (Kleman and Hättestrand, 1999), and links between these and ice-sheet erosion (Hall and Glasser, 2003) has improved greatly in recent years, but is far from comprehensive. In addition, the reconstruction of former ice-sheet margins during the Last Glacial Maximum (LGM), especially in marine realms, is still evolving in many locations (Landvik et al., 1998; England, 1998; Mangerud et al., 1999). In some areas, ice extent and thickness debates revolve around the interpretation of differentially weathered fiord landscapes, which are common along the former eastern Laurentide, Greenland, British, and western Fennoscandian ice sheets.

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While some have supported the contention that highly weathered uplands persisted as nunataks during the last glaciation (e.g., Ives, 1978; Nesje and Dahl, 1990; Steig et al., 1998; Ballantyne, 1998; Rae et al., 2004), others have suggested that different "weathering zones" (Pheasant and Andrews, 1973; Boyer and Pheasant, 1974) represent different basal thermal regimes, hence, differential glacial erosion of an overriding LGM ice sheet (Sugden, 1977; Sugden and Watts, 1977; Hall and Sugden, 1987; Bierman et al., 2001; Hall and Glasser, 2003; André, 2004; Marquette et al., 2004).

Baffin Island, eastern Canadian Arctic (Fig. 1), was occupied by the northeastern sector of the Laurentide Ice Sheet (LIS) and is a region rich with prior weathering zone and ice extent research (Pheasant and Andrews, 1973; Boyer and Pheasant, 1974; Ives, 1975; Sugden, 1977; Sugden and Watts, 1977; Ives, 1978; Steig et al., 1998; Marsella et al., 2000; Bierman et al., 2001; Wolfe et al., 2001; Kaplan et al., 2001; Miller et al., 2002; Briner et al., 2003). This sector of the LIS is important because it is thought to have discharged icebergs into Baffin Bay and the North Atlantic during the last glaciation (e.g., Andrews and Tedesco, 1992; Andrews et al., 1998), altering global thermohaline circulation and playing a central role in abrupt climate change. Hudson Strait, a major artery for ice discharge into the North Atlantic, had a large, fast-flowing ice stream during the LGM (e.g., Andrews and MacLean, 2003), as did Cumberland Sound, the second largest marine trough on southern Baffin Island (Kaplan et al., 2001; Fig. 1). Lake density studies (Sugden, 1978; Andrews et al., 1985a) and erratic trains (Tippett, 1985; Dyke and Morris, 1988) have revealed the probable locations of former ice streams in Lancaster Sound and Home Bay, northern and central Baffin Island, respectively (Fig. 1). Thus, the available evidence suggests that ice streams existed in the major sounds of Baffin Island. However, ice-sheet dynamics in fiord landscapes are less well known.

We use cosmogenic radionuclides (cosmogenic radionuclides) to assess weathering zones and their implications for ice dynamics in fiord landscapes of the Clyde Region (Fig. 1) on northeastern Baffin Island. This paper expands on Briner et al. (2003), which presented initial cosmogenic exposure ages for three freshly weathered erratics perched on two highly weathered tors. Here, we provide cosmogenic exposure ages for 27 erratics, along with cosmogenic radionuclide concentrations for 33 bedrock samples from three different weathering zones. We conclude that differentially weathered landscapes do not represent different durations of ice cover or ice-sheet thickness, but rather demarcate zones of differential glacier erosion by overriding ice sheets. We suggest that ice occupying fiords was erosive and warm-based, whereas ice that covered terrain adjacent to fiords was much less erosive and likely frozen to its bed, supporting Sugden (1977). This finding implies that ice streams drained through the fiord landscape of northeastern Baffin Island, and, by inference, ice streams may have existed in differentially weathered fiord landscapes along the fringes of other Pleistocene ice sheets.

BACKGROUND AND SETTING

Differentially Weathered Landscapes

The description and interpretation of differentially weathered fiord landscapes has been a major component of Pleistocene ice sheet research for over a century (e.g., Blytt, 1876; Daly, 1902; Coleman, 1920; Odell, 1933; Flint, 1943; Ives, 1957, 1978; Dahl, 1966; Nesje and Dahl, 1990; Rea et al., 1998; Ballantyne et al., 1998; Marquette et al., 2004). Fiord terrain along formerly glaciated areas is composed of highly weathered uplands that contain block fields (felsenmeer) and tors, and valleys with fresh glacial features and freshly exposed bedrock. Along mountainous eastern Canada, these landscapes have been divided into three altitude-distinct weathering zones: weathering zones 1, 2, and 3 (WZ I, WZ II, and WZ III; Fig. 2; Pheasant and Andrews, 1973; Boyer and Pheasant, 1974). WZ I is the highest and presumably oldest, consisting of tors and block fields with advanced weathering characteristics. WZ III is the lowest and presumably youngest, consisting of freshly exposed glaciated bedrock on lower fiord walls and on islands within fiords. WZ II lies at intermediate altitudes, and consists of bedrock with intermediate weathering characteristics. In many places, well-defined moraines lie at the WZIII/WZII boundary.

There are two major groups of interpretations of differentially weathered fiord landscapes. The first is that highly weathered landscapes were nunataks during recent glaciation (Fernald, 1925; Linton, 1950; Dahl, 1966; Ives, 1966), and thus provide information on vertical ice-sheet extent during the LGM (e.g., Ives, 1957, 1978; Boyer and Pheasant, 1974; Nesje and Dahl, 1990; Ballantyne, 1997; Rae et al., 2004). In this scenario, glaciers that filled fiords were outlet glaciers (bounded by bedrock). The main lines of evidence supporting this notion include the seaward dip of, and presence of moraines along, weathering zone boundaries (mimicking ice margins), the general lack of glacial features in WZ I, and the belief that an overriding ice sheet would remove tors and other delicate, highly weathered features. The second interpretation is that weathering zones represent the differential modification by a topographically overriding ice sheet (e.g., Sugden, 1977). The differential modification stems from a varying basal thermal regime, such that highly weathered uplands were less modified beneath frozen-bedded ice than less-weathered landscapes in fiords and valleys, which were highly modified by sliding-basal conditions; this concept is the basis for Sugden's (1978) model of "selective linear erosion." In this scenario, fiord glaciers would have been a type of ice stream, or a fast-flowing glacier at least partly bounded by ice. This interpretation is mainly supported by the existence of fresh glacial erratics in WZ I, and in some cases glacially modified tors (e.g., Sugden and Watts, 1977; Hall and Glasser, 2003; André, 2004).

At present, the majority of evidence points toward the latter interpretation: in most settings, highly weathered landscapes reveal information about past subglacial thermal regimes and not about ice extent. Several supporting factors have now been well documented. Preglacial landscapes are preserved beneath the interiors of Pleistocene ice sheets (e.g., Hall and Sugden, 1987; Dyke, 1993; Kleman, 1994; Kleman and Hättestrand, 1999; Dredge, 2000; Hättestrand and Stroeven, 2002; Stroeven et al., 2002; Hall and Glasser, 2003). Evidence is mounting that weathered landscapes along many ice-sheet fringe areas were covered during the LGM (e.g., Bierman et al., 2001; Davis et al., 2002; Briner et al., 2003; Marquette et al., 2004), weathered bedrock has been preserved beneath modern cold-based glaciers (e.g., Rea et al., 1996, 1998), and tors within highly weathered uplands have been noted to be variably modified by cold-based ice (e.g., André, 2004).

Study Area

Our study area includes Clyde, Inugsuin, and McBeth fiords and the intervening landscapes (Fig. 1). Bedrock in the study area consists of a mixture of Archean layered monzogranite, granodiorite, and tonalite gneiss, and Proterozoic banded migmatite (Jackson et al., 1984). The fiords in our study area are over 100 km long and completely traverse the mountain range of the eastern Canadian Rim, spanning from the interior Baffin Island plateau (at 500–600 m above sea level [asl]) to broad coastal lowlands (Fig. 2). Although the core of the mountain range contains peaks over 1500 m asl, the area mainly consists of high, rolling plateaus that are generally glaciated above ~1000 m asl. Uplands are lower and unglaciated along the outer fiord reaches, where the plateau surfaces range from

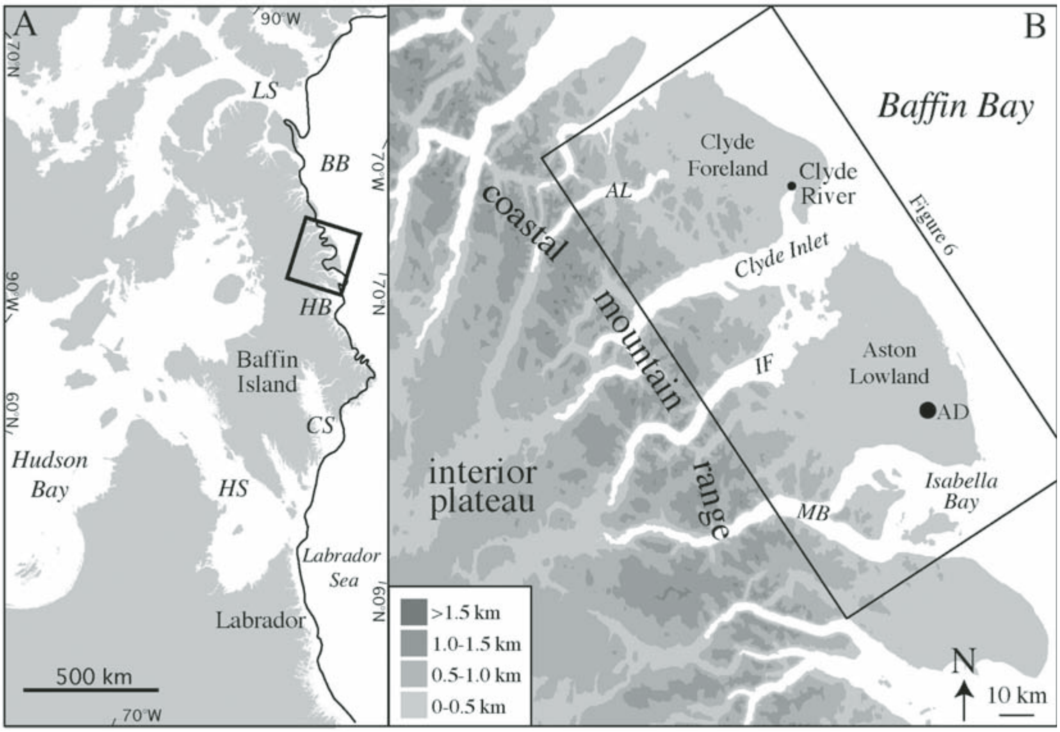

Figure 1. (A) Eastern Canadian Arctic showing Baffin Island, study area (box), and the limit of the Laurentide Ice Sheet during the Last Glacial Maximum (from Dyke et al., 2002). LS—Lancaster Sound; BB—Baffin Bay; HB—Home Bay; CS—Cumberland Sound; HS—Hudson Strait. (B) Elevation map of the Clyde region, northeastern Baffin Island, showing major physiographic features. AL—Ayr Lake; IF—Inugsuin Fiord; AD—Aston Delta; MB—McBeth Fiord.

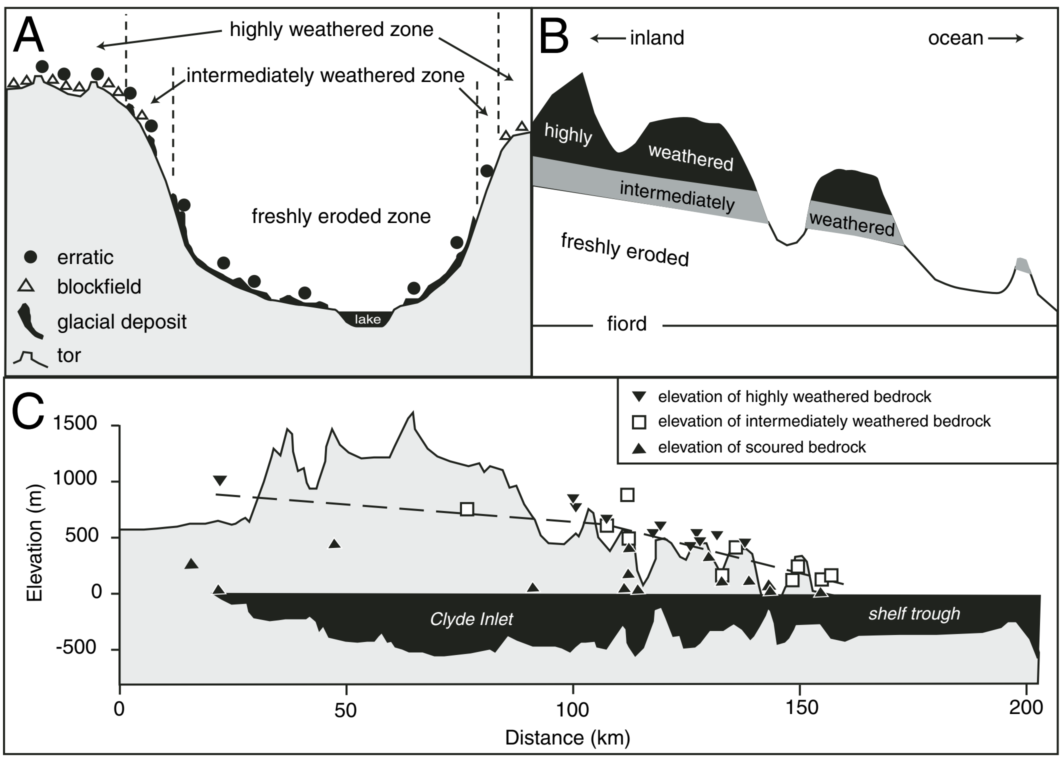

Figure 2. (A) Topographic cross section of a typical glacial valley showing different weathering zones and associated geomorphic features (modified from Ives, 1978). (B) Along-fiord topographic profile showing the three distinct seaward-dipping weathering profiles typical of Baffin Island fiord landscapes (modified from Boyer and Pheasant, 1974). (C) Profile of Clyde Inlet bathymetry and bordering mountain topography; bedrock weathering characteristics from Clyde Inlet and neighboring fiords are plotted to show the elevation and seaward dip of weathering zones. The dashed line separates WZ I from WZ III and is the approximate location of WZ II.

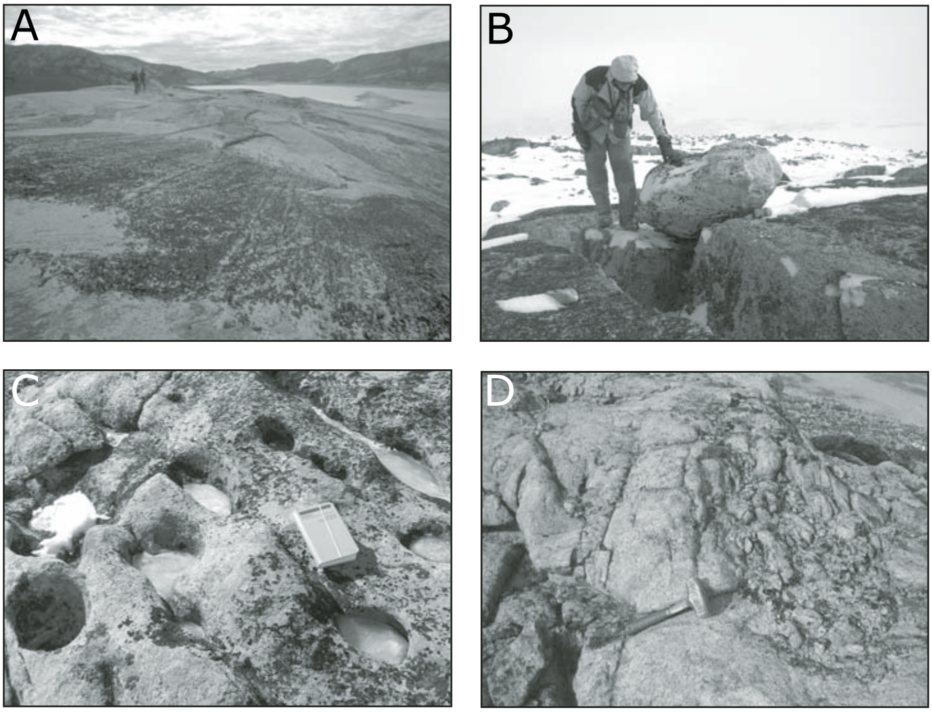

Figure 3. Photographs of bedrock from different weathering zones. (A) Sculpted bedrock island in outer Clyde Inlet. (B) Erratic perched on intermediately weathered bedrock. (C) Tor surface with weathering pits. (D) Tor surface with grusified character.

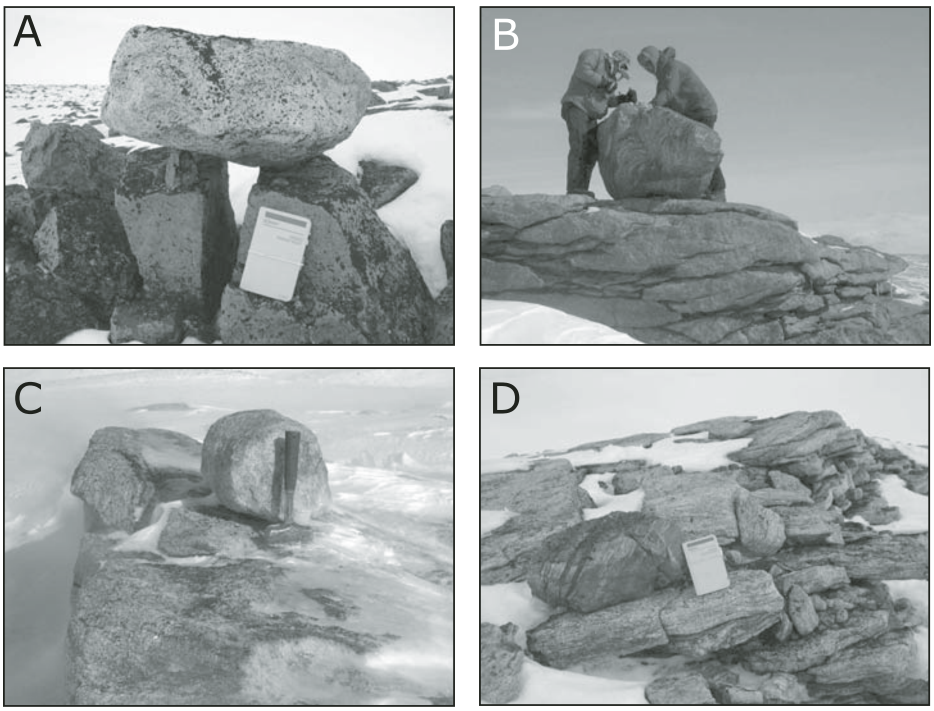

Figure 4. Photographs of erratics. (A) Erratic boulder perched on WZ II bedrock at ~200 m above sea level (asl) on an island in Isabella Bay. (B–D) Erratics perched on WZ I bedrock on the north side of outer Clyde Inlet at 688, 426, and 580 m asl, respectively.

~400–700 m asl. The Clyde Foreland and Aston Lowland are broad coastal lowlands punctuated by numerous hills, reaching ~100–200 m asl near the coast and increasing to >500 m asl farther inland (Fig. 1).

Like much of northeastern Baffin Island, landscapes in the study area are differentially weathered (Fig. 2; e.g., Boyer and Pheasant, 1974; Ives, 1978). Islands in the fiords are polished and striated, and the lower fiord walls contain suites of sharp-crested lateral moraines, categorizing these landscapes into WZ III (Fig. 3; Boyer and Pheasant, 1974). Upland surfaces, in contrast, are mantled by block fields, with tors on most summits. The block fields, or felsenmeer, are composed of boulders, some with advanced weathering characteristics. Similarly, the tors exhibit large weathering pits (up to 60 cm in diameter and 30 cm deep), deep differential weathering, and quartz veins that stand up to 10 cm in relief, placing these landscapes into WZ I (Fig. 3; Boyer and Pheasant, 1974). Where not vertical, upper fiord slopes are typically a mix of block field and scattered till, and where bedrock is present, it lacks both tors and a freshly scoured morphology (Fig. 3). Upper fiord slopes are devoid of lateral moraines but contain sparse lateral meltwater channels. We categorize these landscapes as WZ II (Boyer and Pheasant, 1974), and thus define landscapes as such if they lack both striated or polished surfaces and advanced weathering characteristics, such as tors and weathering pits.

Boundaries between weathering zones in the study area dip seaward, like elsewhere on Baffin Island (Boyer and Pheasant, 1974; Ives, 1978; Fig. 2). For example, ~30 km inland from the mouth of Clyde Inlet, sculpted bedrock exists up to ~450 m asl (Fig. 2), whereas near the mouth of the fiord, tors exist between ~360 and 400 m asl, and intermediately weathered bedrock exists as low as ~100 m asl. Along outermost McBeth Fiord, sculpted bedrock is found on low summits (~60 m asl), but on nearby summits at 125 m asl, the bedrock exhibits intermediate weathering characteristics.

Erratics are widespread in the study area, including the uplands, a finding noted 30 yr ago by Ives (1975). Although upland (WZ I and II) erratics (Fig. 4) are most abundant near fiords, they can still be found many tens of kilometers inland. The erratics are fresher than the bedrock and block fields upon which they are perched, lacking pits and exhibiting minimal grain-to-grain relief. In most cases, these perched blocks are locally erratic lithologies, and are not tabular like underlying bedrock, but are subrounded, and sometimes faceted (Fig. 4). We interpret the perched erratics to have been deposited by the LIS versus local ice carapaces, because many erratics lie on local summits. Local ice would presumably flow radially off the summits, and therefore would not carry erratics to these sites. Furthermore, the source material for transport by local glaciers would be tabular blocks derived from weathered bedrock and block fields; the perched erratics often differ in shape, lithology, and weathering characteristics from surrounding materials.

The continental shelf along northeastern Baffin Island is dissected by a series of overdeepened troughs that connect the deep fiords to the abyss of central Baffin Bay (Løken and Hodgson, 1971; Gilbert, 1982; MacLean, 1985). These troughs range in depth, but are shallower than their fiord counterparts, and are 100–200 m deeper than the adjacent shelf. Outboard of Clyde Inlet is the Clyde trough (Løken and Hodgson, 1971). The Clyde trough is ~250 m deep, as opposed to the adjacent shelf at ~100 m depth, whereas the mouth of Clyde Inlet is ~300–400 m deep. Similar to other troughs along northeastern Baffin Island, lateral shoals fringing the Clyde trough are interpreted as subaqueous moraines (Løken, 1973; cf. Gilbert, 1982).

The glacial history of the Clyde region, first investigated in the 1960s and 1970s (e.g., Løken, 1966; Miller et al., 1977; Mode, 1980), has been the focus of recent studies (Briner, 2003; Davis et al., 2002; Briner et al., 2003, 2005). The earlier group of studies showed that the lowlands in this area contained a rich record of glacial landforms, a thick and comprehensive sequence of glacial deposits in wave-cut exposures, and abundant raised marine features. More recently, cosmogenic exposure dating has been utilized to directly date glacial deposits on the coastal lowlands, providing an updated chronology for their most recent glaciation. Exposure ages of >100 erratics from across the Clyde Foreland (Fig. 1) indicate that the landscape was mostly, if not completely, covered by ice during the LGM (Briner et al., 2005); outside of a few locations, the foreland was mostly covered by cold-based ice. The 80 m asl emerged Aston Delta (Fig. 1), dated to older than 54 ka using in situ bivalves (Løken, 1966), along with an associated assemblage of marine features and meltwater channels spread across ~100 km of the Aston Lowland, was similarly overrun by nonerosive ice during the LGM (Davis et al., 2002). Thus, Laurentide ice, mostly cold-based, crossed both the Clyde Foreland and Aston Lowland during the LGM.

COSMOGENIC RADIONUCLIDES

We measured ^{10}Be cosmogenic radionuclide concentrations in 27 erratics (boulders and cobbles) perched in the uplands (WZ I and II) and in 33 bedrock samples from the three weathering zones. In addition, we measured ^{26}Al in three of the same erratics and in 23 of the same bedrock samples. Because the region experiences low amounts of precipitation (mean annual precipitation = 200 mm yr^{-1} and average snow density = 0.1 g cm^{-3} reported at Clyde River [Fig. 1]; http://www.climate.weatheroffice.ec.gc.ca/) and persistent high winds, surfaces in the study area experience minimal snow cover. We collected small boulders and cobbles perched directly on bedrock and tor surfaces in May, when snow depth is at a maximum. In some cases, we collected samples from perched erratic boulders, perched atop erratic cobbles, while in other cases we collected whole cobbles resting directly on bedrock surfaces. We sampled the uppermost surfaces and considered surface geometry, sample height, potential surface erosion, sample thickness, and topographic shielding in the age calculations.

Samples were prepared at the University of Colorado Cosmogenic Isotope Laboratory (CUCIL) following procedures modified from Kohl and Nishiizumi (1992) and Bierman and Caffee (2001). A subset of samples was prepared at the University of Vermont using the same procedures for interlaboratory comparison. A duplicate sample prepared at both laboratories yielded identical ^{10}Be concentrations (TM1-01-1; Table 1). Exposure ages were calculated using ^{10}Be and ^{26}Al production rates of 5.1 and 31.1 atoms g^{-1} yr^{-1}, respectively (Stone, 2000; Gosse and Stone, 2001). Site-specific production rates were corrected for elevation (Lal, 1991; Stone, 2000) and sample thickness. Because these samples are from high latitude (~70°N), radionuclide production rates are not influenced by changes in the geomagnetic field. The ages reported here are not corrected for atmospheric pressure anomalies, where average low pressure over Baffin Bay may increase the production rates by ~2% (Stone, 2000). Where both ^{10}Be and ^{26}Al concentrations were measured, an uncertainty weighted average exposure age was determined. Ages reported here include only the one standard deviation accelerator mass spectrometry (AMS) measurement uncertainty. We report all ages without accounting for surface erosion and snow shielding, which are likely negligible in this arid environment.

The calculation of surface exposure ages from cosmogenic radionuclide concentrations in glacial landforms relies on several assumptions. These include a lack of radionuclides from prior periods of exposure (termed inheritance, which requires more than ~2 m of erosion in bedrock surfaces; Briner and Swanson, 1998; Davis et al., 1999), a single period of exposure, and minimal postglacial subaerial surface erosion. Most erratics and scoured bedrock samples meet these assumptions, however, intensely weathered bedrock surfaces violate some of these assumptions and thus yield minimum exposure durations (Lal, 1991; Gosse and Phillips, 2001).

We consider burial by weakly or nonerodding ice sheets to be a more important factor that needs to be considered when interpreting our radionuclide concentrations. In locations where ice does not erode, or only partially erodes its bed, there is an inherited concentration of cosmogenic radionuclides upon deglaciation from prior periods of exposure (e.g., Briner and Swanson, 1998; Davis et al., 1999; Colgan et al., 2002). In these locations, cosmogenic radionuclides are a result of exposure since deglaciation, in addition to an inherited component minus what was removed via glacial erosion. In this complex case, such as for WZ II surfaces, the total concentration still provides a minimum exposure duration. Long periods of shielding by nonerosive ice, e.g., more than ~100 k.y., can be detected via the faster decay of ^{26}Al (700 k.y. half-life) with respect to ^{10}Be (1.5 m.y. half-life: Lal, 1991; Bierman et al., 1999; Fabel and Harbor, 1999; Fabel et al., 2002). In this case, such as in WZ I samples, paired isotopic data can provide minimum burial and exposure durations, and an overall minimum surface (upper ~2 m) age.

RESULTS

Upland Erratic Exposure Ages

Cosmogenic exposure ages were calculated for 27 erratic samples collected from weathered upland (WZ I and II) bedrock surfaces (Table 2; Fig. 5). Sample elevations range from 688 m asl ~50 km from the coast of Baffin Bay, to ~118 m asl adjacent to the coast (Table 2). Half of the population ranges between 9.5 and 12.7 ka (n = 14) and the majority of the others range between 14.8 and 23.1 ka (n = 9). One erratic is anomalously young at ca. 7 ka and three are anomalously old, with ages of ca. 28, ca. 40, and ca. 100 ka. There is no spatial relationship of the ages; for example, two erratics from each of three different tors have one age that lies between ca. 10 and 12 ka and the other between ca. 18 and 20 ka (Fig. 6). Although most erratics are from near fiords (Fig. 6), erratics from three upland areas >10 km inland have ages that fall within the 10–13 ka age cluster (Table 2; Fig. 6).

Bedrock Radionuclide Concentrations

A total of 16 individual ^{10}Be and ^{26}Al measurements were made from 10 samples from

TABLE 1. SAMPLE LOCATION INFORMATION AND COSMOGENIC EXPOSURE AGES OF BEDROCK

Sample	Site number on Figure 6	Latitude (°N)	Longitude (°W)	Elevation (m asl)	^{10}Be (10^5 atoms g^{-1})*	^{26}Al (10^5 atoms g^{-1})*	^{10}Be age (ka)†	^{26}Al age (ka)†	[^{26}Al]/[^{10}Be]
Freshly eroded bedrock (weathering zone III)									
CAM02-31	1	70°10.352′	68°27.911′	345	2.03 ± 0.06	11.58 ± 0.36	28.4 ± 0.8	26.8 ± 0.8	5.72 ± 0.24
IBP02-2a	2	70°12.978′	69°05.145′	372	2.46 ± 0.07	ND	34.0 ± 0.9	ND	ND
IBP02-3a	2	70°13.125′	69°05.693′	177	1.52 ± 0.05	ND	25.4 ± 0.8	ND	ND
BDF-00-2x	3	70°16.500′	68°58.403′	36	0.49 ± 0.02	3.93 ± 0.21	9.4 ± 0.4	12.5 ± 0.7	7.49 ± 0.53
AI1-00-1	4	70°17.926′	68°30.901′	80	0.71 ± 0.04	4.48 ± 0.55	12.9 ± 0.7	13.4 ± 1.7	6.34 ± 0.85
AI1-00-3	4	70°17.981′	68°30.928′	87	ND	3.59 ± 0.16	ND	11.0 ± 0.5	ND
CR03-28	5	69°44.077′	67°06.463′	62	0.88 ± 0.03	ND	16.2 ± 0.5	ND	ND
CR03-15	6	69°43.323′	67°58.439′	46	0.57 ± 0.02	3.46 ± 0.19	10.8 ± 0.4	10.9 ± 0.6	6.12 ± 0.43
CR03-22	7	69°30.708′	67°30.807′	145	0.71 ± 0.03	4.20 ± 0.17	12.1 ± 0.4	11.7 ± 0.5	5.88 ± 0.31
CR03-18	8	69°36.860′	68°26.070′	36	0.48 ± 0.03	3.45 ± 0.22	9.3 ± 0.5	10.9 ± 0.7	7.15 ± 0.60
Intermediately weathered bedrock (weathering zone II)									
TM1-00-1	9	70°17.913′	69°08.822′	605	1.99 ± 0.07	12.00 ± 0.41	22.1 ± 0.7	21.9 ± 0.7	6.03 ± 0.29
TM1-00-1x	9	70°17.913′	69°08.822′	605	1.99 ± 0.05	ND	21.9 ± 0.5	ND	ND
TM2-01-1	9	70°17.720′	69°07.543′	608	2.15 ± 0.09	ND	23.7 ± 1.0	ND	ND
IBP02-1a	2	70°12.731′	69°4.360′	501	2.67 ± 0.09	15.81 ± 1.44	32.5 ± 1.1	31.8 ± 0.8	5.91 ± 0.24
AL02-11	10	70°29.307′	69°39.505′	862	5.64 ± 0.13	34.13 ± 1.14	50.0 ± 1.2	50.3 ± 1.7	6.05 ± 0.25
CF02-80	11	70°46.784′	69°12.966′	150	2.76 ± 0.07	ND	47.5 ± 1.2	ND	ND
CF02-179	12	70°26.339′	68°31.118′	231	1.99 ± 0.05	11.78 ± 0.45	31.5 ± 0.8	30.8 ± 1.2	5.93 ± 0.27
CR03-13	5	69°44.339′	67°08.836′	125	1.93 ± 0.05	12.92 ± 0.48	34.2 ± 0.9	37.8 ± 1.4	6.68 ± 0.30
CAM02-6	13	70°15.656′	68°14.623′	107	3.94 ± 0.13	21.54 ± 0.57	72.2 ± 2.3	65.7 ± 1.8	5.46 ± 0.23
CR03-26	14	69°31.575′	67°39.002′	400	2.13 ± 0.07	13.28 ± 0.56	28.6 ± 0.9	29.5 ± 1.2	6.22 ± 0.33
CAD02-1	15	69°53.532′	67°37.435′	124	4.56 ± 0.11	ND	81.1 ± 2.0	ND	ND
CAD02-2	15	69°53.532′	67°37.436′	124	1.89 ± 0.05	ND	33.2 ± 0.9	ND	ND
Highly weathered bedrock (weathering zone I)									
SIV7-00-1	16	70°20.001′	68°48.044′	465	4.90 ± 0.14	27.02 ± 0.77	63.1 ± 5.6	57.3 ± 5.9	5.51 ± 0.23
BDFT2-01-2	17	70°19.817′	68°57.845′	523	8.44 ± 0.28	42.61 ± 1.14	102.3 ± 3.4	86.3 ± 2.3	5.05 ± 0.22
BDFT3-01-2	17	70°19.195′	68°55.515′	582	7.38 ± 0.18	38.30 ± 0.79	84.4 ± 2.0	73.0 ± 1.5	5.19 ± 0.17
SIV8-01-1	16	70°20.115′	68°46.798′	503	5.54 ± 0.13	32.10 ± 1.55	67.5 ± 1.6	65.6 ± 1.8	5.79 ± 0.21
SIV9-01-1	16	70°20.365	68°46.418′	520	6.34 ± 0.15	35.02 ± 0.94	76.6 ± 1.8	70.6 ± 1.9	5.52 ± 0.20
CF02-114	16	70°21.212′	68°48.671′	426	4.96 ± 0.12	24.20 ± 0.59	66.8 ± 1.6	54.0 ± 1.3	4.88 ± 0.17
AL5-01-1	18	70°31.206	68°02.076′	497	6.75 ± 0.16	ND	83.5 ± 2.0	ND	ND
PT01-03	15	69°53.164′	67°37.465′	125	7.77 ± 0.19	42.5 ± 1.2	139.5 ± 3.4	128.9 ± 3.7	5.46 ± 0.21
CAM02-22	19	70°00.732′	68°02.382′	396	9.67 ± 0.23	ND	132.9 ± 3.3	ND	ND
CR03-88	-	69°52.893′	70°36.107′	952	5.71 ± 0.15	30.89 ± 1.18	47.8 ± 1.2	42.8 ± 1.3	5.41 ± 0.25
CR03-16	20	69°35.755′	67°13.407	766	17.28 ± 0.57	85.25 ± 5.09	171.9 ± 5.8	143.4 ± 8.6	4.93 ± 0.39
CR03-17	20	69°35.240′	67°15.441′	739	11.80 ± 0.30	61.75 ± 1.66	116.8 ± 3.0	102.6 ± 2.8	5.23 ± 0.19

Note: m asl—m above sea level; ND—no data.

*Radionuclide concentrations reported here are not scaled for elevation, topographic shielding (which was negligible in all cases), and sample depth (corrections range between 1.00 and 0.97). Concentrations were scaled for elevation according to Lal (1991) with muogenic scaling according to Stone (2000).

†Calculated with ^{10}Be and ^{26}Al production rates of 5.1 atoms g^{-1} yr^{-1} and 31.1 atoms g^{-1} yr^{-1}, respectively (Stone, 2000).

WZ III bedrock, 19 measurements from 11 samples from WZ II bedrock, and 22 measurements from 12 WZ I bedrock samples (Table 1; Fig. 5). The samples are from plateau summits bordering Clyde Inlet, the Ayr Lake trough, and McBeth Fiord; from the Aston Lowland; and from bedrock islands in Clyde Inlet and McBeth Fiord (Fig. 6). The samples yield deglacial exposure ages from WZ III, minimum ages between ca. 20 and ca. 80 ka for WZ II, and minimum ages between ca. 50 and ca. 170 ka for WZ I (Table 2; Fig. 5).

Of the 10 striated bedrock samples in WZ III, seven have concentrations that equate to exposure ages between ca. 10 and ca. 16 ka. Three samples have pre-LGM exposure age equivalents that range between ca. 25 and ca. 34 ka. Elevations for the seven younger ages range between 36 and 145 m asl, whereas elevations for the older three range between 177 and 372 m asl.[1] The ^{26}Al/^{10}Be ratios calculated for six samples is within one standard deviation of the equilibrium ratio of ~6 or higher (Table 2; Fig. 7).

The 11 nonstriated bedrock samples that lack the advanced weathering characteristics of WZ I were placed into WZ II and have radionuclide abundances that equate to minimum exposure ages between ca. 22 and ca. 81 ka. Near the outer coast, WZ II is lower than farther inland (Fig. 2); sample elevations range between ~107 m asl at the outer coast to ~862 m asl ~30 km inland. All but one of the ^{26}Al/^{10}Be ratios for WZ II samples lie near the equilibrium ratio of ~6 (Table 2; Fig. 7).

The 12 samples collected from WZ I bedrock have radionuclide concentrations that yield minimum exposure ages between ca. 48 and ca. 173 ka. The samples are from uplands that range between ~125 m asl at the outer coast to ~766 m asl ~30 km inland. Ten WZ I samples have ^{26}Al/^{10}Be ratios that fall below the equilibrium ratio, ranging between ~4.9 and

[1]GSA Data Repository item 2006046, Figure DR1: scatter plot of exposure age versus elevation, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

TABLE 2. SAMPLE LOCATION INFORMATION AND COSMOGENIC EXPOSURE AGES OF ERRATICS

Sample	Site number on Figure 6	Latitude (°N)	Longitude (°W)	Elevation (m asl)	^{10}Be (10^5 atoms g^{-1})*	^{26}Al (10^5 atoms g^{-1})*	^{10}Be age (ka)†	^{26}Al age (ka)†	Weighted mean age and 1 S.D. uncertainty (ka)
Outer Clyde Inlet region									
SIV7-00-2	16	70°19.993′	68°48.015′	462	0.85 ± 0.04	6.34 ± 0.24	10.8 ± 1.1	13.2 ± 2.1	11.3 ± 1.7
TM2-01-2	9	70°17.720′	69°07.543′	608	0.87 ± 0.03	ND	9.5 ± 0.3	ND	ND
BDFT2-01-1	17	70°19.817′	68°57.845′	523	0.83 ± 0.02	ND	9.8 ± 0.3	ND	ND
BDFT3-01-1	17	70°19.195′	68°55.515′	582	1.13 ± 0.05	ND	12.7 ± 0.5	ND	ND
SIV7-01-3	16	70°20.000′	68°48.042′	464	0.81 ± 0.02	5.33 ± 0.30	10.1 ± 0.3	11.0 ± 0.4	10.5 ± 0.7
SIV8-01-2	16	70°20.115′	68°46.747′	504	1.46 ± 0.05	7.97 ± 0.44	17.5 ± 0.6	15.8 ± 0.6	16.7 ± 1.2
SIV9-01-2	16	70°20.366	68°46.418′	520	1.30 ± 0.03	ND	15.5 ± 0.4	ND	ND
CF02-30	12	70°25.241′	68°28.789′	430	0.93 ± 0.04	ND	12.0 ± 0.5	ND	ND
CF02-31	12	70°25.537′	68°29.675′	380	1.29 ± 0.03	ND	17.5 ± 0.5	ND	ND
CF02-115	16	70°21.212′	68°48.671′	426	3.04 ± 0.08	ND	39.8 ± 1.0	ND	ND
CF02-117	21	70°22.292′	68°50.760′	543	1.74 ± 0.04	ND	20.3 ± 0.5	ND	ND
CF02-118	21	70°22.292′	68°50.760′	543	0.88 ± 0.03	ND	10.3 ± 0.3	ND	ND
CF02-120	21	70°23.157′	68°50.980′	610	1.13 ± 0.03	ND	12.3 ± 0.4	ND	ND
CF02-119	21	70°23.157′	68°50.980′	610	1.69 ± 0.05	ND	18.6 ± 0.5	ND	ND
CF02-134	9	70°18.463′	69°07.258′	620	1.05 ± 0.05	ND	11.4 ± 0.5	ND	ND
CF02-138	9	70 °18.829′	69°08.661′	688	1.14 ± 0.03	ND	11.6 ± 0.3	ND	ND
CF02-142	22	70°23.620′	69°14.311′	632	1.57 ± 0.05	ND	16.9 ± 0.5	ND	ND
CF02-143	22	70°23.997′	69°14.783′	670	1.86 ± 0.06	ND	19.3 ± 0.7	ND	ND
IBP02-1b	2	70°12.731′	69°4.360′	501	2.33 ± 0.06	ND	28.3 ± 0.7	ND	ND
Aston Lowland–McBeth Fiord region									
CAM02-19	23	69 57.654′	67 47.694′	350	6.93 ± 0.40	ND	99.8 ± 5.8	ND	ND
CAM02-11	24	70 10.791′	68 10.952′	362	1.67 ± 0.04	ND	23.1 ± 0.6	ND	ND
CAM02-16	25	69 57.645′	68 23.847′	537	0.84 ± 0.03	ND	9.8 ± 0.4	ND	ND
CAM02-24	19	70 00.709′	68 02.177′	389	1.10 ± 0.04	ND	14.8 ± 0.5	ND	ND
CAM02-14	26	70 02.169′	68 20.375′	463	0.83 ± 0.03	ND	10.4 ± 0.3	ND	ND
CAM02-17	25	69 57.632′	68 23.789′	530	0.98 ± 0.05	ND	11.9 ± 0.6	ND	ND
CR03-14	5	69°44.399′	67°08.723′	118	0.66 ± 0.04	ND	11.5 ± 0.6	ND	ND
CR03-24	14	69°31.566′	67°39.039′	391	0.52 ± 0.03	ND	6.9 ± 0.4	ND	ND

Note: m asl—m above sea level; ND—no data; S.D.—standard deviation.

*Radionuclide concentrations reported here are not scaled for elevation, topographic shielding (which was negligible in all cases), and sample depth (corrections range between 1.00 and 0.97). Concentrations were scaled for elevation according to Lal (1991) with muogenic scaling according to Stone (2000).

†Calculated with ^{10}Be and ^{26}Al production rates of 5.1 atoms g^{-1} yr^{-1} and 31.1 atoms g^{-1} yr^{-1}, respectively (Stone, 2000).

§Uncertainty weighted mean = 1/SQRT[Σ(1/S.D.^{2})].

~5.8, indicating significant periods of burial (Table 1; Fig. 7).

SIGNIFICANCE OF UPLAND ERRATICS

Upland erratics from mountainous areas along ice-sheet fringes have been noted in the literature throughout the last century (e.g., Odell, 1933; Ives, 1975; Sugden and Watts, 1977), but only recently has cosmogenic exposure dating allowed for an unequivocal interpretation (e.g., Briner et al., 2003; Marquette et al., 2004). The significance of perched erratics with deglacial ages is twofold: first, it indicates that interfiord uplands were covered by ice during the last glaciation, and second, it demonstrates the existence of nonerosive ice and its preservation of delicate features.

The largest mode in the erratic exposure age distribution is centered between 10 and 13 ka (Fig. 5), and is interpreted to represent the final deglaciation of the uplands. This timing of upland deglaciation overlaps with the final deglaciation from the adjacent lowlands (Davis et al., 2002; Briner et al., 2005), indicating regional deglaciation during this interval. The one young age is interpreted as having rolled, an anomalous snow-cover history, or an analytical problem. The three oldest erratics are outliers that do not form clusters, and are thus interpreted as either being deposited during the LGM and containing isotopic inheritance or, like the bedrock upon which they are perched, were preserved beneath overriding nonerosive ice. The cluster of ages between 15 and 23 ka requires further explanation. One possibility is that these erratics were deposited between 10 and 13 ka, but they contain isotopic inheritance. However, because the ages fall into distinct modes, an alternative explanation may also be possible. The erratics may have been deposited during a short-lived fluctuation of the ice margin during the LGM but prior to a final advance and retreat phase that ended 10–13 ka. Alternatively, the erratics could have been deposited prior to, and shielded during, the LGM advance, and thus represent a glacial episode ~5–10 k.y. before the onset of the LGM (cf. Marsella et al., 2000). A combination of these scenarios is also possible. Nonetheless, the major cluster of erratic ages between 10 and 13 ka, similar to dozens of exposure ages that delimit deglaciation of the adjacent lowlands, indicates that nonerosive Laurentide ice covered the uplands of the Clyde region during the LGM.

INTERPRETATION OF BEDROCK RADIONUCLIDE CONCENTRATIONS

The striated bedrock surfaces of WZ III have not been subaerially eroded significantly since deglaciation. Thus, we interpret their ages as indicating the timing of deglaciation from the outer fiords. Six of the 10 ages are between 10.1 and 13.2 ka, overlapping with the timing of deglaciation estimated from both the ages of upland erratics (above) and erratic ages from the adjacent lowlands (Davis et al., 2002; Briner et al., 2005). Four ages predate this interval; one has a slightly older age of ca. 16 ka from 62 m asl in outermost Isabella Bay (Fig. 6), and three have ages that predate the LGM. The three pre-LGM samples are also from the three highest elevations of any striated bedrock sample. Because their cosmogenic radionuclide concentrations equate to pre-LGM exposure ages, they must contain isotopic inheritance from a period of exposure prior to the LGM. These data indicate that while the LIS was sliding during part or all of the LGM at these sites, glacier erosion was less effective (less than ~2 m of erosion) relative to lower elevation sites (greater than ~2 m of erosion) within WZ III.

Bedrock of WZ II is not striated, indicating that it was not covered by sliding ice during the LGM, yet erratics found throughout the uplands indicate that it and WZ I were covered by ice during that time. The cosmogenic radionuclide abundances in WZ II bedrock indicate that less than ~2 m of glacial erosion took place during the LGM. Subtracting the duration of exposure since deglaciation, these surfaces have periods of prior exposure ranging from ~10 to ~70 k.y. However, because there may have been some glacial erosion during the LGM that would have removed radionuclides, these are minimum estimates for prior exposure. Furthermore, the $^{26}Al/^{10}Be$ ratios of all but one of the WZ II bedrock samples do not suggest prolonged (>100 k.y.) periods of burial, indicating that their pre-LGM concentrations were in equilibrium. Today's WZ II surfaces may have been below thin (<2 m) bedrock slabs that were removed during the LGM by a cold-based ice process documented by Atkins et al. (2002) and André (2004).

WZ I bedrock yields the highest cosmogenic radionuclide concentrations in the landscape, equating to minimum ^{10}Be exposure ages between ca. 48 and ca. 172 ka. The antiquity of these surfaces clearly indicates that the overriding LGM ice sheet did not erode them significantly. The $^{26}Al/^{10}Be$ ratios of 10 WZ I samples provide reasonable constraints for the history of WZ I bedrock, implying a minimum duration of burial that ranges between 80 and 500 k.y. (Table 3; cf. Bierman et al., 1999; Fabel and Harbor, 1999; Marquette et al., 2004). Taking into account the last 12 k.y. of exposure, the ratios indicate minimum tor surface ages of 150–580 ka, suggesting that these surfaces have not experienced greater than ~2 m of glacial erosion in at least this amount of time.

There are several possible histories for these uplands that could explain their radionuclide concentrations. The uplands of eastern Baffin Island were likely exposed for millions of years before the onset of Quaternary glaciations (Andrews, 1979; Funder et al., 1985), and the terrain would have developed deep weathering profiles (e.g., Hall, 1985; Locke, 1979; Rea et al., 1996) with high concentrations of radionuclides. To reduce the sample's nuclide concentrations to what is measured today requires glacial erosion that amounted to ~2 m or more since the onset of Quaternary glaciation. One plausible scenario is that the uplands were glacially eroded >2 m during the early Quaternary, and were buried and exposed intermittently since then following the pacing of ice age cyclicity. A second scenario that could yield the measured cosmogenic radionuclide abundances and ratios is one that involves radionuclide-saturated uplands that have remained as nunataks until the LGM, when they experienced some amount of glacial erosion that was less than ~2 m. We favor the

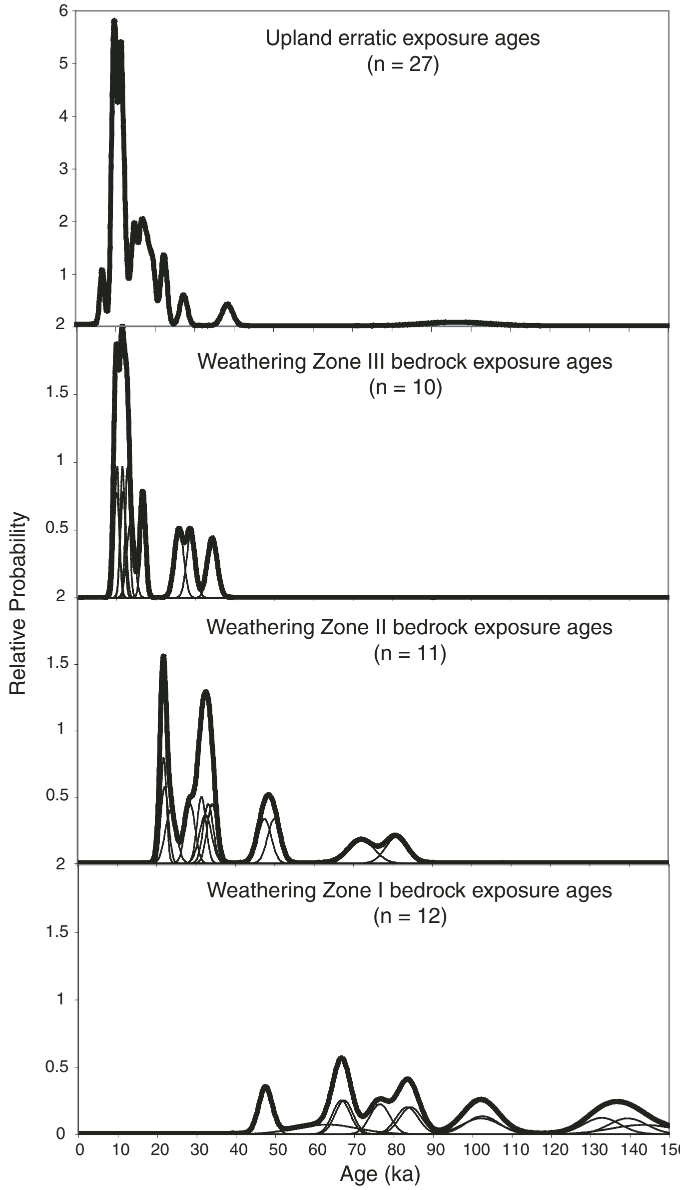

Figure 5. Relative probability plots of cosmogenic exposure ages for upland erratics and for bedrock from the three weathering zones that characterize the fiord landscapes of northeastern Baffin Island. Bold lines are summed probability; thin lines are relative probability for individual samples; curves for individual erratic samples are left out for clarity (probability calculations modified from Lowell, 1995).

first scenario, however, because the uplands have weathering pits several tens of cm deep, which probably took many tens of thousands of years to form (e.g., Hall, 1985; Locke, 1979). Therefore, there is no reasonable scenario that would allow the LIS to have eroded the tors significantly during the last glacial cycle. Our data require that the tors are at least several hundred thousand years old, indicating that the uplands have only been slightly modified in the middle and late Quaternary, despite probably having been repeatedly covered by the LIS.

DIFFERENTIAL ICE-SHEET EROSION AND WEATHERING ZONES

The cosmogenic radionuclide data and relative weathering characteristics of the three weathering zones provide empirical data that link subglacial thermal regime to glacial erosion (cf. Hall and Glasser, 2003). Because all weathering zones were glaciated during the LGM, as indicated by deglacial-age erratics found in all zones, we link weathering zone to ice-sheet erosion style. Although our data distinguish distinct erosional zones with unique characteristics, we note that they probably represent a gradual transition, albeit over a short distance, between warm-based fiord ice and cold-based plateau ice.

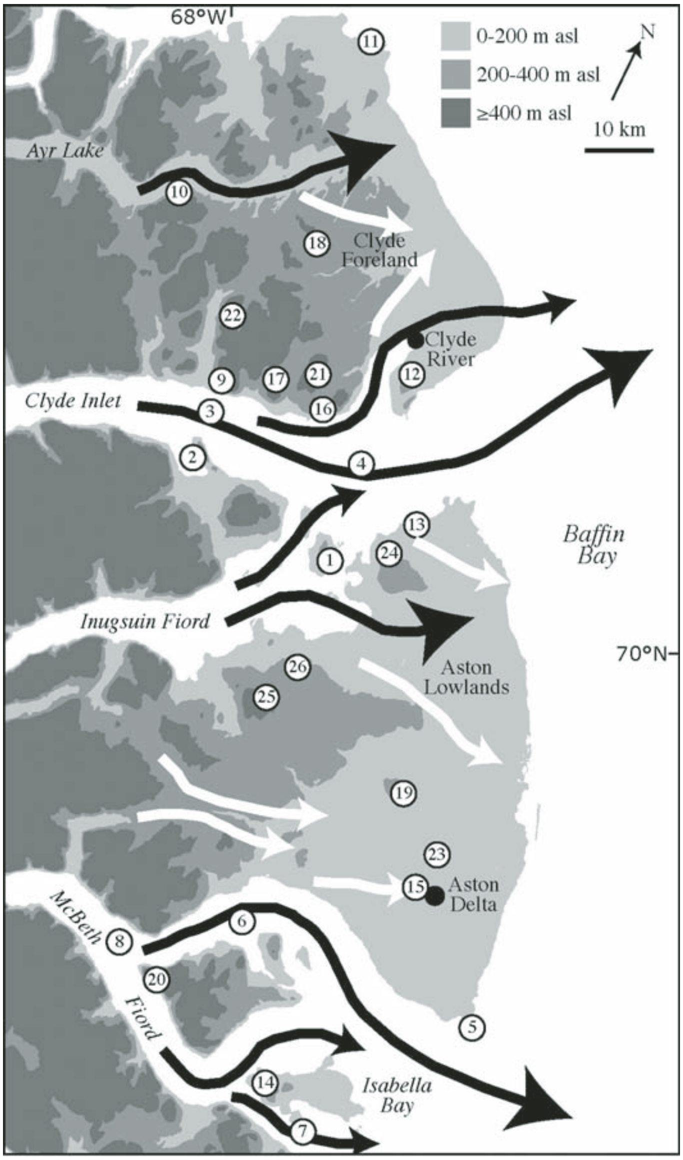

Figure 6. Shaded elevation map of study area showing sample locations (numbers refer to data listed in Tables 1 and 2) and major ice pathways (black arrows depict fast-flowing warm-based ice and white arrows depict the pathway of cold-based ice across the coastal lowlands). Map area shown in Figure 1; asl—above sea level.

Weathering Zone III—Erosive, Warm-Based Ice

Along the main axis of Baffin Island fiords, ice slid rapidly over its bed, accomplishing significant glacial erosion (>~2 m). Ice occupying the fiord was warm-based for several reasons. First, ice overrode unfrozen, unconsolidated fiord sediments (e.g., Gilbert, 1982) with relatively low shear strength, promoting sliding. Second, ice converging into fiords had a relatively high velocity, helping to maintain a warm base via strain heating. Third, unlike ice on plateaus, ice in fiords was thick and insulative.

Pre-LGM exposure ages at our highest three sites in WZ III indicate that glacial erosion amounted to less than ~2 m during the LGM. We interpret this inefficient erosion as representing ice that began to slow down, cool, and lose erosive power. In addition, erosive ice would have occupied the uppermost elevations of WZ III for shorter durations than at lower elevations.

Weathering Zone II—Inefficiently Erosive Ice

These surfaces pre-date the LGM, and have not been buried for >~100 k.y. (based on $^{26}Al/^{10}Be$ equilibrium) without being glacially eroded >~2 m. Because WZ II concentrations are lower and ratios higher than in WZ I, it is most likely that WZ II surfaces experienced >~2 m of glacial erosion within the last glacial cycle, but before the LGM. If any erosion took place during the LGM, it may have been via plucking (e.g., Atkins et al., 2002) or some other "uncapping" mechanism (e.g., André, 2004) that did not leave behind a fresh, abraded surface. In general, because WZ II separates warm-based ice at lower elevations from cold-based ice at higher elevations, the thermal regime likely fluctuates as ice thickens and thins, and WZ II may occasionally experience abrasion.

Weathering Zone I—Nonerosive, Cold-Based Ice

Based on paired radionuclide data, these surfaces are at least ca. 150 to ca. 580 ka, and have experienced ~80 to ~500 k.y. of accumulated burial. At some point since the onset of Quaternary glaciation, these surfaces have been glacially eroded >~2 m, and the variability in radionuclide abundances among the sampled tors likely reflects the distinct history, that is, glacial modification (cf. André, 2004), of each sampled tor. Because of the lack of striae and the predominance of weathering pits and other characteristics of long-term subaerial exposure, WZ I was probably not glacially modified during the LGM, but rather was protected by slow-moving, frozen-bedded ice.

This overall pattern of cold-based ice occupying uplands and warm-based ice occupying fiords, with two narrow transitional zones in between, suggests that topographically controlled fiord-type ice streams, or isbræ (Truffer and Echelmeyer, 2003), occupied northeastern Baffin Island fiords during the LGM. The accepted definition of an ice stream, "a region in a grounded ice sheet in which the ice flows much faster than in the regions on either side" (Patterson, 1994), fits the description of the flow regime reported here, although in cross section, there may be more bedrock in contact with the fiord glaciers than cold-based ice. These fiord ice streams would have been bounded by shear zones (incorporating WZ II) that separated slow-moving from fast-moving ice. The seaward dip of weathering (and erosion) zone boundaries likely arises from ice-sheet thinning as: (1) the LIS approached its margin, and (2) glaciers became unconstrained as they flowed from the deep central fiords to the less topographically constrained fiord mouths. This ice thinning was accompanied by a decrease in velocity as the ice margin spread out and experienced divergent flow. These factors resulted in cooler ice that, once out of the axis of flow, became cold-based. We suggest that these processes resulted in weathering (or erosion) zones existing at lower elevations near the coast as opposed to farther inland. Finally, the moraines that are commonly found at the WZIII/WZII boundary were most likely formed during deglaciation (Briner et al., 2005).

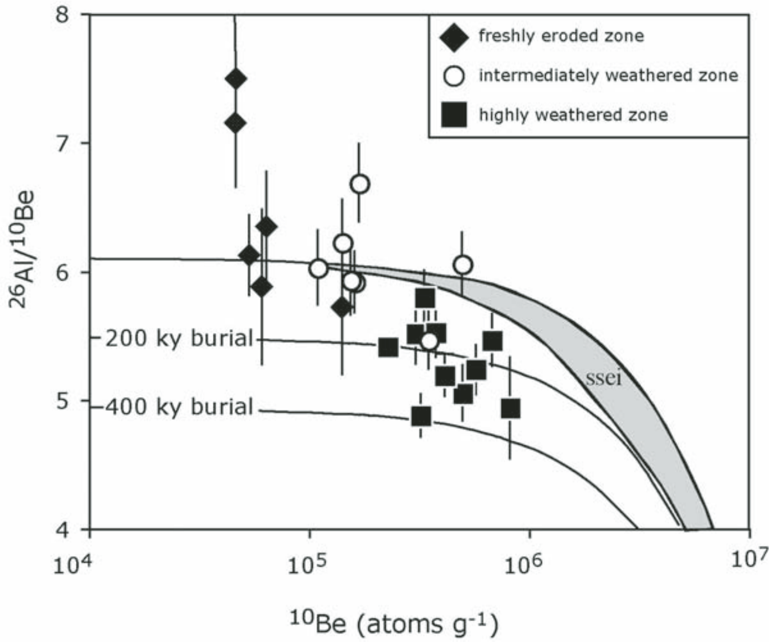

Figure 7. Plot of ^{26}Al/^{10}Be ratio versus ^{10}Be concentration for bedrock samples from freshly eroded (*n* = 6), intermediately weathered (*n* = 7), and highly weathered (*n* = 10) bedrock. The horizontal line at ~6.1 that leads into the shaded zone (the steady-state erosion island; Lal, 1991) is where samples will plot if burial (where ^{26}Al decays differentially with respect to ^{10}Be; see text) is below detection. The 200 and 400 k.y. burial isochrons are shown. Samples that plot above the steady-state erosion island have analytical problems relating to Al (cf. Brook et al., 1996; Gosse and Phillips, 2001).

TABLE 3. EXPOSURE/BURIAL CONSTRAINTS FOR WEATHERING ZONE I BEDROCK

Sample	Exposure prior to burial (k.y.)	Burial (k.y.)	Exposure since deglaciation (k.y.)	Total history (k.y.)
SIV7-00-1	54	216	12	282
CF02-114	66	504	12	582
BDFT2-01-2	104	357	12	473
BDFT3-01-2	82	325	12	419
SIV8-01-1	57	84	12	153
SIV9-01-1	69	191	12	272
PT01-03	135	163	12	310
CR03-17	117	271	12	400
CR03-16	182	347	12	541
CR03-88	41	267	9	317

Note: Calculations follow Bierman et al. (1999). ^{10}Be and ^{26}Al half lives are 1.5 m.y. and 0.7 m.y., respectively. ^{10}Be and ^{26}Al production rates are 5.1 atoms g^{-1} yr^{-1} and 31.1 atoms g^{-1} yr^{-1}, respectively.

IMPLICATIONS

Ice Dynamics of Study Area

The minimum vertical extent of ice during the LGM can be constrained using the altitude of upland erratics, and approximate ice-surface profiles can be reconstructed using lateral moraines deposited during deglaciation. Ice-surface gradients along outer Clyde Inlet were ~13 m km^{-1} (similar to Jakobshavns Isbræ, eastern Greenland; Echelmeyer et al., 1991) during deglaciation, yielding a basal shear stress of ~0.4 bar. The low gradient and basal shear stress support warm-based, sliding ice (Patterson, 1994), and place the LGM terminus at least ~30 km across the continental shelf, most of the way down the 50-km-long Clyde trough (Fig. 2). In contrast, the ice-surface gradients of the outlet glacier on the Clyde Foreland emanating from the Ayr Lake trough (Fig. 6), a land-based margin, are ~40 m km^{-1} and yield a basal shear stress of ~1 bar. The contrast of basal shear stresses is consistent with the different substrates that the ice overrode (unconsolidated, unfrozen fiord sediments versus frozen drift and bedrock).

Combining the mapped basal thermal regime pattern of the fiord landscape with the adjacent lowlands gives a more complete depiction of the dynamics of the northeastern LIS. The low-lying Clyde and Aston Lowlands were dominated by frozen-bedded ice during the LGM (Davis et al.,

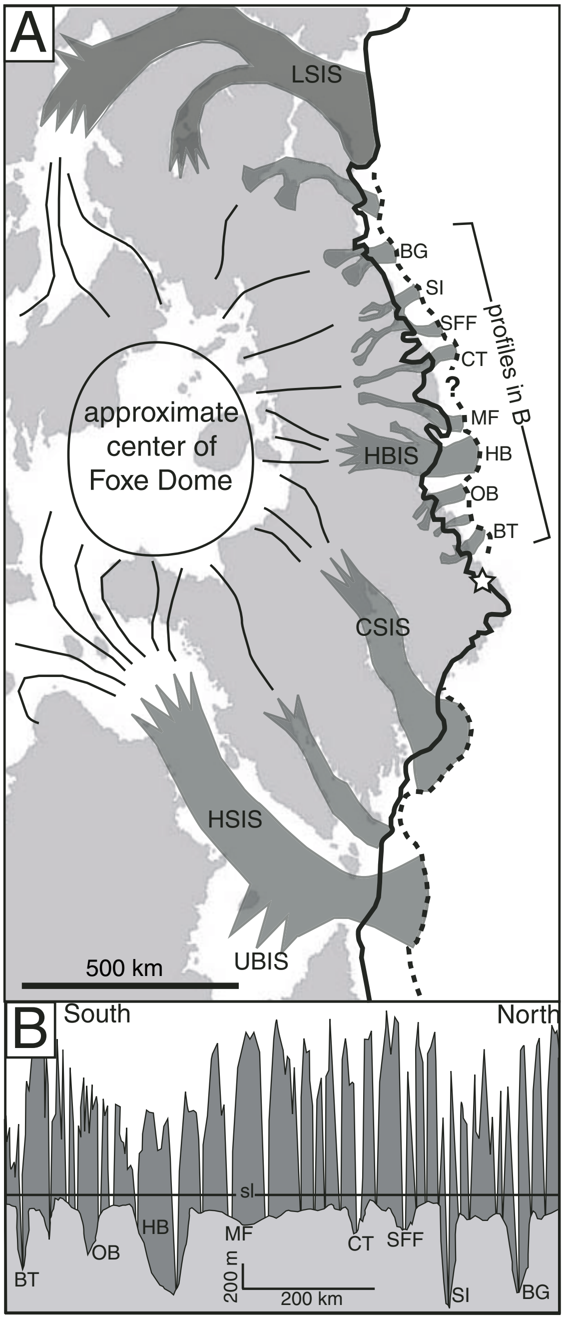

Figure 8. (A) Northeastern sector (Foxe Dome) of the Laurentide Ice Sheet showing schematic ice flow (thin black lines), ice streams (gray areas), most recently compiled Last Glacial Maximum ice limit (thick black line) from Dyke et al. (2002), and proposed modifications of the ice limit (dashed lines) where ice streams are thought to have terminated at the continental shelf break in the south (Andrews and MacLean, 2003) and in the northeast (this study). The star represents the location where LGM nunataks apparently existed (Steig et al., 1998). Reconstructed ice streams are labeled: LSIS—Lancaster Sound ice stream (Dyke and Morris, 1988; Andrews et al., 1985b); HBIS—Home Bay ice stream (Tippet, 1985); CSIS—Cumberland Sound ice stream (Jennings et al., 1996; Kaplan et al., 2001); HSIS—Hudson Strait ice stream (e.g., Andrews and MacLean, 2003); UBIS—Ungava Bay ice streams (Jansson et al., 2003). Overdeepened shelf troughs are labeled: BG—Buchan Gulf; SI—Scott Inlet; SFF—Sam Ford Fiord; CT—Clyde trough; MF—McBeth Fiord; HB—Home Bay; OB—Okoa Bay; BT—Broughton trough. (B) Cross sections of height of land for fiord troughs/intertrough areas and continental shelf (south to north) with major troughs labeled.

2002; Briner et al., 2005). Landscapes on either side of the fiords, including both the uplands and coastal lowlands, appear to have been covered by cold-based ice and experienced only slight modification during the LGM. The great antiquity of the upland surfaces indicates that they have been only slightly modified throughout the entire Quaternary, suggesting long-term selective erosion in the Clyde region (Sugden, 1978). Evidence of selective erosion continues offshore to the shelf troughs. There are fewer shelf troughs than fiords, however, implying further organization as ice streams approach the ice-sheet margin (Fig. 8).

Northeastern Laurentide Ice-Sheet Dynamics

Baffin Island holds a rich geomorphic record of LIS bed conditions and is the site of the classical studies by Sugden (1977, 1978) that linked glacial landscapes to ice-sheet dynamics. Our data on the ice dynamics in fiord landscapes, which make up a significant portion of the northeastern Laurentide's bed, help to improve the picture of LIS behavior during the LGM.

With an increased understanding of modern ice streams, arguably the most dynamic components of contemporary ice sheets, comes the recognition that ice streams likely played a large role in Pleistocene ice-sheet configuration and stability (e.g., Denton and Hughes, 1981; Andrews et al., 1985b; Hughes et al., 1985; Alley and MacAyeal, 1994; Stokes and Clark, 2001). Several studies have depicted ice streams along the periphery of the LIS, especially in deep marine channels like Hudson Strait and Cumberland Sound (Denton and Hughes, 1981; Laymon, 1992; Kaplan et al., 2001; Andrews and MacLean, 2003; Jansson et al., 2003). Although it is commonly difficult to reconstruct the former locations of ice streams (Stokes and Clark, 2001), several lines of evidence point toward past ice stream behavior. "Boothia-type" erratic dispersal trains (e.g., Dyke and Morris, 1988; Tippet, 1985) and mega-scale lineation patterns (e.g., Clark and Stokes, 2001; Stokes and Clark, 2002, 2003), similar to those associated with modern ice streams (e.g., Anderson et al., 2001), suggest the locations of LIS ice streams.

Numerous overdeepened troughs that crosscut the continental shelves all along the eastern seaboard of Canada (e.g., Løken and Hodgson, 1971; Gilbert, 1982), including northeastern Baffin Island (Fig. 8), provide additional evidence that erosion was concentrated in select, distal reaches of the LIS. Shelf troughs, which have long been interpreted to be glacial in origin (Shepard, 1931; Holtedahl, 1958; Hughes et al., 1977; Hughes et al., 1985; Bougamont and

Tulaczyk, 2003), likely indicate the former locations of ice streams. In Antarctica, overdeepened shelf troughs outboard of modern ice streams, floored with mega-scale lineations, are direct evidence for linking shelf troughs with paleo–ice streams (Hughes et al., 1977; Canals et al., 2000; Anderson et al., 2001, Anderson et al., 2002). Overdeepened shelf troughs along western Norway are interpreted to have contained ice streams along the fringe of the Fennoscandian ice sheet during the LGM (Sejrup et al., 1998; Landvik et al., 1998; Ottesen et al., 2005; Boulton et al., 2001; Landvik et al., 2005). The troughs on the continental shelf of northeastern Baffin Island (Fig. 8) further support the argument that the northeastern sector of the LIS was drained by a system of ice streams.

The preservation of pre-glacial (uneroded) terrain in ice-sheet interiors beneath cold-based ice is becoming increasingly documented. Islands in the Canadian High Arctic that contain preglacial weathered bedrock and block fields (e.g., Dyke, 1993) and weathered terrain in central regions of the Fennoscandian ice sheet (e.g., Kleman, 1994) represent persistent cold-based zones within ice-sheet interiors. Indeed, the locations of block fields (e.g., Dredge, 2000) and ribbed moraines (e.g., Kleman and Hättestrand, 1999) demonstrate that large portions of the LIS interior were likely cold-bedded. Unconsolidated Tertiary sediments preserved adjacent to the Barnes Ice Cap (Andrews et al., 1972; Morgan et al., 1993) require minimal erosion beneath portions of the Foxe Dome. In addition, lake density studies (Sugden, 1978; Andrews et al., 1985b) reveal that central Baffin Island above ~250 m asl remained unscoured by the LIS. Recent studies focusing on the bed of the Fennoscandian ice sheet have provided a more detailed spatial pattern of former bed conditions (e.g., Kleman and Stroeven, 1997; Hättestrand and Stroeven, 2002), and cosmogenic exposure dating has confirmed the locations of nonerosive ice (Fabel et al., 2002; Stroeven et al., 2002). Cosmogenic ^{10}Be and ^{26}Al isotope concentration data indicate that some repeatedly glaciated sites have experienced negligible glacial erosion over the entire Quaternary (e.g., Stroeven et al., 2002). Finally, cosmogenic radionuclide data from the southern margin of the LIS also support thin and cold-based marginal ice that accomplishes little erosion (Colgan et al., 2002).

Differentially Weathered Fiord Landscapes

Differentially weathered fiord landscapes have been the focus of recent study elsewhere. On southern Baffin Island, Bierman et al. (1999) obtained disequilibrated ^{26}Al/^{10}Be data on upland tors within WZ I, and concluded that the tors experienced periods of exposure and burial that require 500–700 k.y. of surface history. In Labrador, home to long-standing weathering zone research (e.g., Daly, 1902; Coleman, 1920; Odell, 1933; Ives, 1957, 1978), weathering zones have recently been interpreted to represent differential ice-sheet erosion, based on soil weathering and cosmogenic radionuclides. Marquette et al. (2004) collected ^{26}Al and ^{10}Be data from bedrock in all three weathering zones, as well as for five erratics from WZ II (their "intermediate zone") and one erratic from WZ I (their "felsenmeer zone"). The erratics have deglaciation ages (11–13 ka), whereas bedrock ages are typically ca. 80–150 ka in the felsenmeer zone, ca. 20 to ca. 90 ka in the intermediate zone, and 12 ka in the Saglek zone (WZ III). In addition, ^{26}Al/^{10}Be data suggest that the upland sites record burial by cold-based ice (Marquette et al., 2004). On the other hand, Steig et al. (1998) make a convincing case for upland refugia during the LGM on northern Cumberland Peninsula, eastern Baffin Island (Fig. 8). Although upland erratics were not dated, they used cosmogenic exposure dating on a flight of lateral moraines, the highest of which has two exposure ages that predate the LGM. Thus, although in many locations it appears that interfiord uplands were covered by glacial ice during the LGM, some areas on Baffin Island may have remained ice-free. We speculate that Cumberland Peninsula typically gets covered by thinner continental ice because of its isolation by the large Cumberland Sound and Home Bay troughs (Fig. 8), which help to efficiently discharge continental ice before it reaches the peninsula.

In the UK, differentially weathered landscapes were the impetus for Sugden's (1978) original ideas on the selectivity of glacial erosion and its link with ice-sheet thermal regimes. At present, however, the classical debate on how to interpret differentially weathered glacial landscapes continues. For example, some believe that the dip of the scoured bedrock/felsenmeer boundary along a flow line is enough evidence to support that it represents the maximum thickness of LGM ice (e.g., Ballantyne 1997, 1998, 1999; Ballantyne et al., 1998; Rae et al., 2004). However, others have documented glacial modification of surfaces above this trimline in some areas (e.g., Glasser and Hall, 1997; Hall and Glasser, 2003). Published cosmogenic exposure ages that apply to this debate from the UK thus far have been predominately on bedrock, and support relative weathering studies (Ballantyne et al., 1998; Stone et al., 1998), but do not directly address upland glacierization.

Differentially weathered fiord landscapes along the western Fennoscandian ice sheet have also received attention (e.g., Dahl, 1966; Nesje and Dahl, 1990; Sollid and Sorbel, 1994). Cosmogenic radionuclide data from four bedrock samples acquired by Brook et al. (1996) along a transect from valley to plateau support the break in weathering zones; however, no erratics were dated in their study. Along western Svalbard, Landvik et al. (2005) have reconciled seemingly conflicting data by calling for highly dynamic ice that overrode but preserved pre-LGM deposits. They reconstruct LGM ice streams that filled fiords and shelf troughs to the shelf edge, and depict less dynamically active ice along interfiord/trough areas. On the other hand, Landvik et al. (2003) provide exposure ages for four erratics perched among upland block fields at the northwestern tip of Svalbard that leave open the possibility that upland surfaces there remained ice-free during the LGM.

In many locations, cosmogenic radionuclide and other data have recently supported relative weathering studies by confirming that interfiord upland surfaces are indeed older than those in fiords and valleys. On the other hand, the majority of these studies have not explicitly addressed whether uplands were covered by cold-based ice or not, as has been hypothesized for many decades. By focusing future efforts on cosmogenic exposure dating of erratics perched on upland surfaces, we can address ice-sheet thickness histories for many locations along former ice-sheet margins.

CONCLUSIONS

Our results shed light on the weathering zone debate and support Sugden's (1978) notion that differentially weathered glacial landscapes can be the result of selective erosion by overriding ice sheets, and do not, a priori, provide information about the maximum thickness of ice sheets during the LGM. Cosmogenic radionuclide measurements from differentially weathered bedrock can provide insights into general surface ages and burial history, but interpretations of weathering zone surfaces are not complete until it is known whether the entire landscape was covered by LGM ice or not. Only careful field investigations (e.g., Hall and Glasser, 2003; André, 2004; Marquette et al., 2004) or cosmogenic exposure dating of erratics can address upland glacial histories. Moreover, because upland surfaces have high cosmogenic radionuclide concentrations, they can yield misleadingly old exposure ages when they are the sources for erratics. Thus, many erratics may need to be dated before making interpretations about ice cover history.

The reconstruction of ice streams filling northeastern Baffin Island fiords, combined with previously depicted ice streams in the larger sounds and straits of the Baffin region (e.g., Stokes and Clark, 2001), point toward an ice-stream–dominated northeastern LIS with sharp gradients in basal thermal regime. These topographically controlled ice streams would have been stable in space and time as long as ice discharge remained sufficiently high. However, several ice streams, including the Hudson Strait ice stream, thought to be the major supplier of Heinrich Event debris (Andrews and MacLean, 2003), are known to have "binged and purged" (MacAyeal, 1993) throughout the last glaciation. Thus, the northeastern LIS was highly dynamic and sensitively tied with North Atlantic thermohaline circulation and abrupt climate change. However, the behavior of different sectors of the LIS during binge and purge cycles is as yet unknown. Other problems, such as the location and behavior of ice stream onset zones and the origin and development of fiords, also remain unresolved.

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