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Deglaciation and ice shelf development at the northeast margin of the Laurentide Ice Sheet during the Younger Dryas chronozone

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Core 2011804-0010 from easternmost Lancaster Sound provides important insights into deglacial timing and style at the marine margin of the NE Laurentide Ice Sheet. Spanning 13.2-11.0 cal. ka BP and investigated for ice-rafted debris (IRD), foraminifera, biogenic silica, and total organic carbon, the stratigraphy comprises a lithofacies progression from proximal grounding line and sub-ice shelf environments to open glacimarine deposition; a sequence similar to deposits from Antarctic ice shelves. These results are the first marine evidence of a former ice shelf in the eastern Northwest Passage and are consistent with a preceding phase of ice-streaming in eastern Lancaster Sound. Initial glacial float-off and retreat occurred >13.2 cal. ka BP, followed by formation of an extensive deglacial ice shelf during the Younger Dryas which acted to stabilise the retreating margin of the NE LIS until 12.5 cal. ka BP. IRD analyses of sub-ice shelf facies indicate initial high input from source areas on northern Baffin Island delivered to Lancaster Sound by a tributary ice stream in Admiralty Inlet. After ice shelf break-up, Bylot Island became the dominant source area. Foraminifera are dominated by characteristic ice-proximal glacimarine benthics (*Cassidulina reniforme*, *Elphidium excavatum* f. *clavata*), complemented by advected Atlantic water (*Cassidulina neoteretis*, *Neogloboquadrina pachyderma*) and enhanced current indicators (*Lobatula lobatula*). The biostratigraphy further supports the ice shelf model, with advection of sparse faunas beneath the ice shelf, followed by increased productivity under open water glacimarine conditions. The absence of Holocene sediments in the core suggests that the uppermost deposits were removed, most likely due to mass transport resulting from the site's proximity to modern tidewater glacier margins. Collectively, this study presents important new constraints on the deglacial behaviour of the NE Laurentide Ice Sheet, with implications for past ice sheet stability, ice-rafted sediment delivery, and ice-ocean interactions in this complex archipelago setting.

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The Northwest Passage (NWP), comprising the main east-west channels of the Canadian Arctic Archipelago (CAA), represents a major oceanic gateway that accounts for ~25% of all modern Atlantic-Arctic Ocean exchange (Aagaard & Carmack 1989; Kliem & Greenberg 2003). During the Last Glacial Maximum (LGM, 26.5-19.0 cal. ka BP; Clark *et al.* 2009) grounded glacial ice of the confluent Laurentide and Innuitian ice sheets (Dyke 2004) blocked the main channels of the CAA (MacLean *et al.* 2010; Lakeman & England 2012, 2013; Bennett *et al.* 2013, 2014). Coalescent Innuitian and Greenland ice sheets further resulted in the closure of Nares Strait (cf. Zreda *et al.* 1999), thereby eliminating Arctic Ocean water inflow into Baffin Bay (Knudsen *et al.* 2008; Jennings *et al.* 2011).

Whilst early deglaciation of the CAA was marked by time-transgressive marine-based ice retreat along Parry Channel (the main axis of the NWP Fig. 1), the unsuturing of the Laurentide and Innuitian ice sheets (LIS and IIS respectively) remains poorly constrained, although it was probably completed by 10.5 cal. ka BP (Pieńkowski *et al.* 2014). The opening of Nares Strait remains debated, but likely occurred after 9.8 cal. ka BP (Zreda *et al.* 1999, but see England *et al.* 2006 vs. Mudie *et al.* 2006 for conflicting interpretations). The progressive retreat of marine-based glaciers from the CAA permitted the establishment of oceanic throughflow under conditions of elevated glacioisostatic sea-level and enhanced meltwater efflux (Dyke *et al.* 1991) resulting in greater Atlantic water inflow and increased biological productivity during the early Holocene (Pieńkowski 2015). These factors, combined with climate forcing, imply a highly variable Late Pleistocene to early Holocene CAA environment (Bradley 1990; Dyke *et al.* 1991; Kaufman *et al.* 2004). Although terrestrial and marine records for the eastern CAA spanning Termination I are relatively well constrained and in broad agreement (e.g. Dyke *et al.* 1991; Dyke 1999; Dyke & Hooper 2001; Pieńkowski *et al.* 2014), some localities are marked by late Quaternary histories that stand in contrast to the emerging regional model. For example, the glacial and deglacial history of Bylot Island (Klassen 1981, 1985, 1993), at the critical intersection of Lancaster

Sound and Baffin Bay (Fig. 1), suggests only limited MIS 2 (Late Wisconsinan) glaciation, contrary to evidence for extensive grounded glacial ice in Lancaster Sound (MacLean *et al.* 2010, 2017; Li *et al*. 2011; Bennett *et al.* 2014) and on adjacent Baffin Island (Dyke 2000; Dyke & Hooper 2001; Briner *et al.* 2003, 2009).

In order to elucidate the timing and style of ice sheet retreat during the last deglaciation at the junction of the NWP and Baffin Bay, and to clarify the palaeoenvironmental and oceanographic conditions during this interval of rapid change, we present new results from a high-resolution marine sediment core from outer Lancaster Sound (Fig. 1). The available stratigraphy permits a multiproxy (sedimentology, litho- and biostratigraphy, biogeochemistry) reconstruction of glacimarine and palaeoceanographic conditions. In this complex archipelago setting, determining the chronology and mechanisms of deglaciation carries major implications for understanding the relationship between sea-level, climate, and ice sheet retreat and the establishment of oceanic through-flow between Arctic and Atlantic oceans. This is particularly germane when considering interactions between grounded ice margins, floating ice shelves and sea ice; and the timing of Laurentide and Innuitian ice sheet unsuturing. This study carries with it implications for understanding the deglacial behaviour of the LIS, including the development of deglacial ice shelves and is also relevant to understanding past rates and patterns of seafloor sedimentation, which constitute fundamental variables in regional geohazard risk assessments.

Regional setting

Lancaster Sound (Fig. 1), at the eastern entrance to the NWP (Pharand 1984) is a large marine channel some 90 km wide and in some places >1000 m deep at its confluence with Baffin Bay (Canadian

Hydrographic Service 1984, 1985a, b; Jakobsson *et al.* 2012), with water depths in Lancaster Sound dropping rapidly offshore from typically steep coastlines to ~800 m in <5 km. The sound is bounded to the north by Devon Island (including tidewater margins of the Devon Icecap) and to the south by Baffin and Bylot islands (with tidewater glacial margins on northern Bylot). Large channels and fjord systems feed into Lancaster Sound from the south, such as Prince Regent, Admiralty, and Navy Board inlets – the former two serving as major conduits for grounded Laurentide ice export during the Wisconsinan glaciation and earliest Holocene (Dyke & Hooper 2001; MacLean *et al.* 2010; Margold *et al.* 2015a, b).

Stratigraphic and geomorphic evidence obtained during limited geophysical surveys indicates that Lancaster Sound was occupied by a grounded ice stream that extended into northern Baffin Bay during the LGM (MacLean *et al.* 2010, 2013, 2017; Li *et al.* 2011; Bennett *et al.* 2013, 2014). This is supported by terrestrial data from Devon Island (Dyke 1999) and Borden and Brodeur peninsulas on northwest Baffin Island (Dyke 2000; Dyke & Hooper 2001). Regional calibrated radiocarbon chronologies indicate coastal deglaciation between 12 and 10.5 cal. ka BP (Dyke 1999; Dyke & Hooper 2001; Pieńkowski *et al.* 2014), with the elevation of marine limit declining eastwards from ~60 to ~22 m a.s.l. on Brodeur and Borden peninsulas (Baffin Island) and from ~55 to 37 m a.s.l. on southeast Devon Island (Dyke 1999; Dyke & Hooper 2001). Deglacial marine limits have not been defined for the steep coast of southeast Devon Island east of Croker Bay, though the area has been undergoing submergence since the late Holocene (Dyke 1999; Taylor & Frobel 2006). To the south, the proposed glacial chronology of Bylot Island (Klassen 1981, 1985, 1993; Klassen & Fisher 1988) stands in contrast to adjacent Lancaster Sound coastlines. The extent of LGM glaciation on Bylot Island remains unclear, with some reconstructions suggesting limited interior glaciation with an ice-shelf occupying eastern Lancaster Sound (Dyke & Prest 1987; Dyke 1999). Evidence for an eastern CAA response to Younger Dryas (YD) cooling is limited. On eastern Baffin Island, any YD advances have been overprinted by more extensive 8.2 ka BP event and

Little Ice Age moraines (Miller *et al*. 2005; Briner *et al.* 2009; Young *et al.* 2012) with only limited YD responses evidenced from eastern Greenland (Funder & Hansen 1996; Kelly & Lowell 2009). While aridity is invoked for a limited Younger Dryas response in Greenland and Baffin Island (Cuffey & Clow 1997; Miller *et al.* 2005; Young *et al.* 2012), offshore records suggest elevated IRD output and ice discharge from the eastern CAA and Baffin Bay during this time (e.g. Andrews *et al.* 1996, 2012). A major readvance into Lancaster Sound of the Prince Regent Inlet ice stream draining LIS ice from the Gulf of Boothia (Niessen *et al*. 2009; MacLean *et al.* 2017) may also date to the Younger Dryas (Pieńkowski *et al.* 2014), though age control remains poor.

The development of modern oceanographic circulation in eastern Lancaster Sound is closely tied to the deglacial and glacioisostatic adjustment history of the broader Parry Channel and Baffin Bay regions, although chronologically well-constrained marine sediment cores from Parry Channel that span the critical period of immediate deglaciation are limited in number (MacLean *et al*. 1989; Pieńkowski *et al*. 2012, 2013, 2014). An east-west connection through Barrow Strait and Lancaster Sound was achieved 11-10.5 cal. ka BP with final unsuturing of Innuitian and Laurentide ice between Prince of Wales and Bathurst islands across the shallow Lowther-Young islands sill (Dyke 1993, 1999; Pieńkowski *et al.* 2014). Greater water depths in Barrow Strait and Lancaster Sound upon deglaciation appear to have resulted in enhanced Atlantic origin water penetration into the CAA relative to present (Pieńkowski 2015). The final deglaciation of Nares Strait ~8 cal. ka BP (~7.5 ^{14}C ka BP; England 1999; England *et al.* 2006; <9.8 cal. ka BP in Zreda *et al.* 1999) also opened an additional connection from the Arctic Ocean into Baffin Bay, increasing Arctic Ocean cold water contributions to the southward-flowing Baffin Current (Knudsen *et al.* 2008). The modern regional oceanography (effectively established by ~6 cal. ka BP; Pieńkowski *et al.* 2013, 2014) is characterised by net outflow of cold and relatively fresh Arctic Ocean Surface Water (<300 m depth) from the marine channels of the CAA towards Baffin Bay due to steric sea-level differences (Ingram & Prinsenberg 1998). Westward-moving Baffin Bay Atlantic Water (300-1200 m; Atlantic Ocean origin) enters Parry Channel below eastward-moving Arctic Ocean Surface Water, progressively mixing westwards (Jones & Coote 1980; Coote & Jones 1982). Water exiting eastern Parry Channel contributes to the southward flowing Baffin Current, along with Arctic Ocean water via Nares Strait and the return flow of the West Greenland Current (Tang *et al.* 2004). Seasonal sea-ice dominates both Lancaster Sound (first-year landfast ice) and Baffin Bay (pack ice), with much of this ice, along with icebergs from Greenland, being exported southward by the Baffin Current (Tang *et al.* 2004).

Geologically, eastern Lancaster Sound is underlain by Cretaceous to Plio-Pleistocene units associated with Cretaceous rifting and subsequent deposition (Daae & Rutgers 1975; MacLean *et al.* 1990; Li *et al.* 2011). Bylot Island to the immediate south and southeast of the core site is underlain by predominantly Archean to Palaeoproterozoic crystalline basement rocks of the Canadian Shield (Fig. 2). Mesoproterozic carbonates and clastic sedimentary rocks outcrop along the island's north and west coasts, bisected by Neoproterozoic mafic dykes (Scott & de Kemp 1998, 1999). Cretaceous to Neogene sedimentary rocks also occur along the southwestern and northern coasts (Miall *et al.* 1980; Csank *et al.* 2013). Adjacent Borden peninsula exhibits a complex assemblage of Palaeozoic to Mesoproterozoic carbonate and clastic sequences, Palaeoproterozoic granites, and Archean intrusive igneous and metamorphic units. Farther west, Brodeur Peninsula is almost entirely underlain by Silurian and Ordovician carbonates (Scott & de Kemp 1998, 1999). Archean gneiss and metavolcanics occur across north-central Baffin Island south of Eclipse Sound (Scott & de Kemp 1998, 1999; Jackson 2000) and include characteristic iron ore deposits of the Mary River Iron Formation (Johns & Young 2006). Together, these bedrock sequences represent potential source areas for erratics found on the north coast of Bylot Island and IRD (ice-rafted debris) in eastern Lancaster Sound, providing important evidence for ice transport trajectories.

Material and methods

Core material

A trigger weight – piston core pair (2011804-0010TWC and 2011804-0010PC, henceforth “0010TWC” and “0010PC” respectively; Fig 3) were collected in Lancaster Sound, 8.3 km north of Cape Hay, Bylot Island (Nunavut, Canada; 73°48.50’ N 80°00.54’ W) by the Geological Survey of Canada-Atlantic (GSC-A) during an ArcticNet cruise aboard the CCGS Amundsen (October 2011). The cores, retrieved from 837 m water depth, measure 275.5 cm (0010PC) and 175 cm (0010TWC), and show a similar sedimentary succession. The lithostratigraphy mainly consists of sandy diamictic mud to coarse sand, interspersed with laminated clays (Fig. 3).

Chronostratigraphy

Six benthic foraminiferal radiocarbon dates constrain 0010PC and 0010TWC (3 dates each; Table 1). Molluscs present in the cores were not dated as they were identified as deposit-feeding genera (*Yoldiella* spp., *Nucula* spp.) prone to enhanced, non-linear age effects (England *et al.* 2012; Pieńkowski *et al.* 2014). Due to the paucity of datable materials, a situation common in deglacial sequences (e.g. Ó Cofaigh *et al*. 2001), and likely marked changes in ice-proximal glacimarine deposition rates (Cowan *et al*. 1997; Gilbert *et al*. 2002), an age-depth model was not constructed for this record.

Radiocarbon dates were calibrated in CALIB 7.1 (Stuiver *et al.* 2016), using the MARINE13 calibration curve (Reimer *et al.* 2013). Calibration, however, is complicated by uncertainties in time-variable marine reservoir age (MRA) and appropriate correction values (ΔR). Whilst Coulthard *et al.* (2010) report a ΔR term of 220±20 ^{14}C years for the study region, this value is only applicable for the late Holocene with Arctic Ocean – Baffin Bay surface water exchange and Atlantic water downwelling in the Labrador Sea. Values on northern Labrador Sea gorgonian corals (ΔR = 132±23 ^{14}C a; Sherwood *et al*. 2008) are also only applicable during the late Holocene.

Given the site location at the intersection of Lancaster Sound and Baffin Bay and growing evidence for significant intrusion of Atlantic Water into Parry Channel from the east prior to final Laurentide and Innuitian Ice Sheet severance (Pieńkowski *et al.* 2014), a northern or northwestern North Atlantic ΔR estimate may be appropriate. The majority of North Atlantic studies attempt to estimate Bølling-Preboreal surface water MRAs for the eastern North Atlantic and Norwegian Sea (e.g. Bard *et al.* 1994; Austin *et al*. 1995; Voelker *et al*. 1998; Bondevik *et al*. 2001, 2006; Björck *et al*. 2003) or Iceland region (e.g. Eiricksson *et al*. 2004; Thornalley *et al*. 2011). Cao *et al*. (2007), however, use coldwater corals (co-dated using ^{230}Th/^{234}U/^{238}U and ^{14}C) from Orphan Knoll, east of Labrador, supplemented by a database of previously published North Atlantic values to derive northwestern North Atlantic MRAs (Bølling-Allerød 380±140 ^{14}C a; Younger Dryas 590±130 ^{14}C a; Preboreal 270±20 ^{14}C a). When compared against the INTCAL13 and MARINE13 curves of Reimer *et al*. (2013), the MRAs of Cao *et al.* (2007) yield a ΔR value of effectively zero for the Allerød, a Younger Dryas ΔR of 185±140 ^{14}C a, and 75±25 ^{14}C a during the Preboreal. These calculated values closely resemble the molluscan ΔR for Holocene West Greenland (-10±80 ^{14}C a; McNeely *et al*. 2006; Furze *et al*. 2014), suggesting that Atlantic water in the West Greenland Current is the primary source of marine carbonate in the northern Baffin Bay region. Thus, the ΔR values derived from Cao *et al*. (2007) are used here to calibrate the 2011804-0010 foraminiferal ^{14}C dates. We recognise that under deglacial conditions of reduced ocean ventilation and the admixture of isotopically “old” meltwater from retreating ice margins, the time-appropriate ΔR may have been larger, compounded by the downwelling of surface waters during seasonal sea-ice formation and brine expulsion. Nevertheless, given the dominance of Atlantic water circulation in Baffin Bay and the absence of Arctic Ocean to Atlantic through-flow at this time, we consider NW Atlantic ΔR values as broadly appropriate calibration terms. Irrespective of whether derived time-appropriate NW Atlantic values or the late Holocene ΔR term proposed by Coulthard *et al.* (2010) are used, the chronology still straddles the YD chronozone.

Ice-rafted debris (IRD) and micropalaeontology

For IRD and micropalaeontological analyses (sample interval 10 cm and 10-20 cm, respectively), samples were wet weighed, wet sieved at 63 µm with de-ionised water, oven-dried at 45 °C, and finally dry weighed.

The >250 µm fraction was examined for IRD under stereo microscopy, with the 250-2000 µm and >2000 µm components being examined separately for IRD species to facilitate ease of examination. A minimum of 300 IRD grains were picked from the 250-2000 µm fraction, and all grains were picked from the >2000 µm fraction. Grains were assigned to a total of 30 IRD categories adapted from Bischof & Darby (1999) and Esteves (2012; Table S1), based on lithic/mineral content and roundness. Both relative (%) and absolute abundances (grains per dry gram of sample = grains g^{-1}) were calculated for each IRD species. Results present the combined 250-2000 µm and >2000 µm fractions.

For micropalaeontological analyses, all sample residues >63-2000 µm were re-combined and systematically picked for foraminifera (benthic, planktic) and other calcareous microfossils (ostracods, juvenile molluscs) under stereo microscopy. In most cases, all calcareous microfossils were picked from each sample to achieve a statistically valid count (≥300 benthic foraminifera). Foraminiferal identification followed Vilks (1969, 1989), Feyling-Hanssen *et al.* (1971), and Knudsen & Seidenkrantz

(1994). Absolute abundances of microfossils are given as individuals per dry gram of sediment (ind. g^{-1}); benthic foraminiferal data are also presented as relative abundances (%).

To facilitate the identification of stratigraphic zonation within the IRD and foraminiferal data, respective datasets were run through a stratigraphically constrained cluster analysis (UPGMA, Euclidean similarity index) in PAST 3.11 (Hammer *et al.* 2001).

Biogeochemistry

For analyses of total organic carbon (TOC), samples (5-10 g of sediment; 15-20 cm interval) were oven-dried at 60-65 °C. Carbonates and inorganic carbon were removed by applying 10% hydrochloric acid to powdered sediments. Samples were then rinsed with de-ionised water, oven-dried, and combusted in a high-frequency induction furnace. TOC was measured at GSC-A using a LECO WR-112 Wide Range Carbon Determinator; values being reported as weight percentages. For biogenic silica (BioSil) assay (1-2 g sediment; 0010PC: 20 cm interval; 0010TWC: 40 cm), sediments were dried at 45 °C, ground, and subsequently analysed at the Pacific Centre for Isotopic and Geochemical Research (University of British Columbia), following Mortlock & Froelich (1989). BioSil content is reported as %Si(opal).

Geomorphic mapping

To provide a regional context for ice-flow trajectories (Fig. 2), glacial lineations (including mega-scale glacial lineations, drumlinoid ridges, crag and tail features and streamlined bedrock) and probable grounding zone wedges and moraines were mapped within the main channels of the eastern CAA using ArcticNet 30 kHz multibeam ecosounder data acquired by the CCGS Amundsen icebreaker and publically available online from the University of New Brunswick Ocean Mapping Group (http://www.omg.unb.ca/). Mapping criteria and definitions of bedforms followed MacLean *et al.* (2010, and references therein) and Dowdeswell *et al*. (2016, and articles therein), with iceberg scours and plough marks being excluded. Mapping was supplemented by previously published bedform maps for parts of the study area (De Angelis & Kleman 2005, 2007; MacLean *et al.* 2010, 2014, 2016, 2017; Li *et al.* 2011; Bennett *et al.* 2014, 2016). It should be noted that multibeam coverage for much of the study area is limited, especially in Prince Regent and Admiralty inlets and central Lancaster Sound, thus the nonappearance of mapped bedforms in many areas reflects this lack of data rather than an actual absence of glacial bedforms.

Results

Five basic units (Fig. 3) are identified from the record based on lithostratigraphic variations (colour, grain size/texture, structure), correlating well with IRD clusters (Fig. 4). All five units are present in 0010PC, with the uppermost three units identifiable in the shorter 0010TWC. Litho- and chrono-stratigraphic correlation between the two components is excellent (Fig. 3). Deposits <11.0 cal. ka BP are, however, missing from the top of both cores due to erosion or non-deposition.

Lithostratigraphy and IRD

The lower stratigraphy constitutes massive to faintly stratified sandy muds of units 5 and 3 (Fig. 3), marked by frequent granules, pebbles, and rounded clay clasts, which are interrupted by a prominent soft red clay (Unit 4). This clay horizon is notable for the absence of any granules or pebbles, an abrupt contact with Unit 5 below, and marked millimetre-scale laminae of alternating red and grey colour. Bracketing foraminiferal ^{14}C ages from units 5 and 3 indicate that Unit 4 was deposited between 13.2 and 12.5 cal. ka BP (Table 1; Fig. 3). Upcore, Unit 3 is truncated by a coarse well-sorted sandy bed representing the basal horizon of Unit 2: a stratified to laminated upward-fining sandy silty clay with infrequent granules and pebbles, that also dates to 12.5 cal. ka BP. Distinct large-scale (2-20 cm) couplets of upward-fining silty sand to clay characterise Unit 2, with couplets demonstrating a finer internal structure of horizontal and cross-stratified clay and silt laminae. Minor erosional surfaces and scours are present at the base of some of the thicker (10-20 cm) couplets. The laminations of Unit 2 are generally horizontal, but are interspersed with prominent cross-bedded fine sand lenses. Lonestones are common and exhibit deformation of underlying laminae, suggesting an interpretation as ice-rafted dropstones. The uppermost lithostratigraphic unit (Unit 1) is a massive to faintly stratified mottled stony mud, yielding foraminiferal ages of 11.0 and 11.1 cal. ka BP (Table 1).

Typical IRD (>250 µm) contributions by weight through the record are around 20%, but reach 77% at the contact between units 2 and 3; 0% contributions are apparent in Unit 4 and near the base of Unit 2 (Fig. 4). A total of 30 IRD species was identified in core 2011804-0010 (Table S1). Highest IRD concentrations are found within Unit 3, reaching >1800 grains g^{-1} dry sed. Relative abundances of IRD species show little variation through the record (Fig. 4), being dominated by angular clear quartz. Minor IRD species including iron-stained quartz, mafic minerals, and gneiss/granite maintain relatively low (typically <10%) background abundances throughout the record. Some subsidiary species, however, do show minor variances in relative abundance upcore. Limestone (particularly creamy-yellow) and calcareous sandstone abundances marginally decrease from the middle of Unit 3 upwards, whilst total feldspar and rounded and angular coloured quartz exhibit minor increases upcore from the same stratigraphic interval. Other minor variations include elevated gneiss/granite abundances in Unit 2

(0010PC) and a decrease in chlorite (as inclusions within quartz) from the middle of Unit 3 upwards. This general pattern of low amplitude IRD variability is punctuated by the red Unit 4 (Sample 250-252 cm 0010PC), which contains almost no IRD (>250 µm); the rare grains present being dominantly limestone and angular clear quartz (Fig. 4).

Biostratigraphy and biogeochemistry

Foraminifera are present throughout the record (Fig. 5), with benthics outnumbering planktics (Fig. 6). Benthic abundances range from <1 to >50 ind. g^{-1}, with maximum abundances occurring in Unit 1. Benthic faunas (Fig. 5; Table S2) are dominated by two species: *Cassidulina reniforme* and *Elphidium excavatum* f. *clavata* (Fig. 5). *Cassidulina neoteretis* is prominent towards the base of the sequence (units 5, 4, lower 3), whereas *Astrononion gallowayi* and *Lobatula lobatula* appear towards the core top (units 1 and 2). Other benthic foraminifera appearing throughout the record include *Stainforthia feylingi*, *Stetsonia horvathi*, *Islandiella norcrossi* and *Islandiella helenae*. *Neogloboquadrina pachyderma*, the exclusive planktic foraminiferal species, is present sporadically throughout the record (Fig. 5), as are juvenile bivalve molluscs (*Nucula* spp.) and ostracods (Fig. 6; Table S2). Bivalves of the family Yoldiidae (fragments and paired juvenile valves), some with intact periostraca, occur towards the base of Unit 1.

BioSil fluctuates between 0.7 and 2.2%, with slightly elevated values notable in Unit 5 and the top of Unit 1 (Fig. 6). It should be noted that none of the samples analysed for BioSil were above the laboratory error of 5%. Whereas this is suggestive of little productivity within the sequence, the presence of foraminifera in all of the investigated samples implies sufficient primary productivity to maintain such a community at, or adjacent to, the study site. Furthermore, TOC, though low and variable throughout the record (0.8-2.7%; Fig. 6), supports the notion of some bioproductivity.

Discussion

Interpretations

Both 2011804-0010TWC and -0010PC show a near identical litho- and chrono-stratigraphic progression, recording glacimarine sedimentation from 13.2 to 11.0 cal. ka BP over a ~3 m interval, and permitting a detailed reconstruction of sedimentary and environmental conditions from immediate deglaciation to the earliest Holocene.

The crudely stratified, waterlain gravelly, foraminifera-bearing muds of units 5 and 3 represent an ice-proximal environment with high sedimentation rates and clastic rainout from melting glacial ice. Similar sediments have been described throughout the marine channels of the CAA, constituting the lower units of piston- and trigger weight cores (e.g. Andrews *et al.* 1991; Bennett *et al.* 2014; Pieńkowski *et al.* 2014) and identified as acoustically unstratified, unsorted sediments in seismic profiles (MacLean *et al.* 1989, 2010, 2017; Niessen *et al.* 2010). The typically low shear strength of this regionally extensive deposit, coupled with its crude stratification and apparent *in situ* sparse glacimarine foraminiferal faunas (Pieńkowski *et al.* 2012, 2014; this study), suggests that this diamictic facies does not represent subglacial deposition beneath grounded ice. As a diagnostic characteristic, shear strength alone can be problematic. Deformation tills (*sensu* Benn & Evans 1996) associated with deforming beds and mega-scale glacial lineations (MSGLs) beneath palaeo-ice streams and modern Antarctic streaming glaciers (Evans *et al.* 2005; Ó Cofaigh *et al.* 2005, 2007; King *et al.* 2009; Reinardy *et al.* 2011; Spagnolo *et al.* 2016) can exhibit shear strengths <20 kPa (Kamb 1991; Humphrey *et al.* 1993). However, foraminiferal faunas yielding post-LGM ages indicate that units 5 and 3 record rapid rainout from debris-rich floating glacial ice associated with a quickly retreating and destabilizing marine margin (*sensu* Pieńkowski *et al.*

2014). Similar waterlain rain-out diamictons and proximal sub-ice shelf deposits have been reported from Antarctic, Barents Sea, and NW European continental shelves (e.g. Domack *et al.* 1998; Evans & Pudsey 2002; Knight 2006; Murdmaa *et al.* 2006; Christ *et al.* 2015; Peters *et al.* 2015). The age of 13.2 cal. ka BP at the base of Unit 5 (Table 1; Fig. 3) provides a minimum limiting age for deglaciation and lift-off of grounded glacial ice in eastern Lancaster Sound, with rapid westward retreat (runaway grounding line conditions) to a previously identified grounding zone wedge (GZW) north of Brodeur Peninsula (Fig. 2; Bennett *et al.* 2014) 140 km to the west. Rapid ice stream lift-off and retreat, punctuated by still-stands and GZW construction has been described from numerous locations in Antarctica (e.g. Dowdeswell *et al.* 2008; Ó Cofaigh *et al.* 2008; Dowdeswell & Fugelli 2012) and the Barents Sea (e.g. Winsborrow *et al.* 2010; Andreassen *et al.* 2014; Bjarnadóttir *et al.* 2014). Well-preserved MSGLs and drumlinoid ridges marked by a thin sediment cover and extensive lateral continuity (Maclean *et al.* 2015), and an absence of iceberg scours in Lancaster Sound adjacent to northern Bylot Island (Fig. 2; Bennett *et al.* 2013, 2014; MacLean *et al.* 2017) further confirm rapid ice stream lift-off and retreat.

Unit 4 represents an interruption in coarse clastic sedimentation sometime between 13.2 and 12.5 cal. ka BP. The effective absence of clastic material >250 µm in this unit (Fig. 4) indicates a temporary cessation in ice-rafted deposition with continued suspended sediment supply under quiescent ice-proximal conditions (units 5 and 3). The contrasting red colouration of this unit (Munsell colour 2.5YR 4/2; Fig. 3) suggests a markedly different sediment source. The apparent laminations at the base of Unit 4 are primarily colour variations from grey (2.5Y 6/2) to red (2.5YR 4/2) reflecting alternation in sediment sources rather than pronounced grain size changes. The sparse foraminifera within this unit (2.5 ind. g^{-1}) are consistent with advection from a more ice-distal source (Domack *et al.* 1998; Evans & Pudsey 2002; Post *et al.* 2007, 2014; Riddle *et al.* 2007). Collectively, these data are considered indicative of glacimarine suspended sediment deposition beneath a cover of pervasive debris-limited floating ice that prevented IRD and lonestone advection/deposition and biological productivity. Ice-distal conditions can be discounted given the low microfossil abundances and diversities (Korsun & Hald 1998; Jaeger & Nittrouer 1999; Knudsen *et al.* 2008). Similar units, interrupting coarse diamictic rain-out facies, have been described from ice-proximal deglacial sequences elsewhere within the CAA and suggest time-transgressive ice-shelf occupation or pervasive landfast sea-ice occurrence (Pieńkowski *et al.* 2012).

The ice-proximal environment of the basal unit 5 to 3 sequence is superseded by the stratified to laminated silty clays and cross-bedded sand lenses of the more ice-distal Unit 2. Available ^{14}C foraminiferal ages suggest rapid deposition around 12.5 cal. ka BP (Table 1, Fig. 3). The lithostratigraphy supports an interpretation of continued suspended sediment deposition and reduced iceberg rafting consistent with increasingly ice-distal conditions (Powell 1984; Powell & Molina 1989; Gilbert *et al.* 2003; Pieńkowski *et al.* 2014), punctuated by turbidity current erosion and deposition (Dowdeswell & Murray 1990; Gilbert 1990; Cowan *et al.* 1997; Ó Cofaigh & Dowdeswell 2001) from the channel margin to the south. Foraminiferal faunas in Unit 2 (Figs 5, 6) suggest less fluctuating environmental conditions compared to units 3 and 5, with an increase in typical Arctic continental shelf species (*I. norcrossi*; *I. helenae*) and appearance of taxa indicative of ample food supply to the benthos (*Buccella frigida*; Jennings *et al.* 2004). The minor changes in IRD species seen from the top of Unit 3 into Unit 2 (Fig. 4) indicate the increasing importance of local sediment sources. In particular, decreasing amounts of carbonates and increasing abundances of feldspar, rounded and angular coloured quartz, and granite clasts are consistent with the increasing importance of sediment delivery from the tidewater margins of glaciers D181 and D183. These glaciers drain ice from the northwestern Byam Martin Mountains into a bay west of Cape Hay (13 km south of the core site; Figs 1, 7). Primary bedrock types within the glacial catchment (Fig. 2) – Palaeoproterozoic migmatites, Mesoproterozoic quartz arenites and feldspar sandstones (Scott & de Kemp 1998) – are in keeping with the shift seen in the IRD (Fig. 4).

This overall ice-distal character marked by the increased contribution of locally sourced IRD continues upwards into Unit 1. However, unlike Unit 2, Unit 1 (~12.5 to 11.0 cal. ka BP) is a massive, mottled mud with frequent dispersed small granules and pebbles. Foraminiferal faunas in Unit 1 are similar to Unit 2, but with taxa indicative of increased current activity (*A. gallowayi*, *L. lobatula*; Murray 2006). Higher planktic to benthic (P:B) ratios in units 1 and 2 may also indicate increasing oceanic (versus glacimarine continental shelf) conditions (Vilks 1974; Darling *et al.* 2007; Xiao *et al.* 2014). Mottling, absence of laminations, and the increased BioSil values and higher diversity benthic foraminifera (Fig. 6) suggest greater biological productivity in both surface waters and benthos. Though still influenced by ice-rafting, Unit 1 resembles typical early Holocene marine sequences encountered elsewhere in Parry Channel characterised by low sedimentation rates and bioturbation (MacLean *et al.* 1989, 2010; Pieńkowski *et al.* 2012, 2013, 2014). Notably, the uppermost sediments of the record were deposited during the very early Holocene (11.0 and 11.1 cal. ka BP, 0010TWC and 0010PC respectively), indicating that the majority of Holocene marine sedimentation is absent from the record. The similarity in upper ages from both cores precludes the possibility that Holocene sediments were lost during recovery.

Lancaster Sound glacial extent and deglacial chronology

The lowermost ^{14}C age from core 2011804-0010 (Unit 5) dates ice-proximal glacimarine conditions consistent with rapid sedimentation from a destabilizing ice margin (rain-out diamicton; Powell 1984; Alley *et al.* 1989; McKay *et al.* 2009) and float-off of grounded glacial ice ~13.2 cal. ka BP or earlier. Similar to other piston core records from Parry Channel, subglacial sediments with high shear strengths (e.g. subglacial traction till *sensu* Evans *et al.* 2006) were not recovered due to limited core penetration of diamictic materials. Post-LGM glacimarine foraminiferal faunas recovered from units 5 to 3 argue against a subglacial interpretation for the basal diamictons in this study.

Denton & Hughes (1981; based on Blake 1970, 1975) depict a grounded outlet glacier in Lancaster Sound draining the LIS during the LGM, whereas Dyke & Prest (1987) and Dyke (1999, 2004) portray a floating ice tongue in eastern Parry Channel during this time. The latter reconstructions are based on the interpretation of the poorly defined Button drift on northern Bylot Island (Fig. 7; Klassen 1985, 1993; Klassen & Fisher 1988) as being deposited by a Late Wisconsinan ice shelf (Dyke 1999). Subsequent work by Li *et al.* (2011), however, has provided evidence for grounded glacial ice extending eastwards from Lancaster Sound repeatedly through the Quaternary (including the LGM), to a trough mouth fan in northern Baffin Bay (maximum water depths ~1300 m). This is further supported by MacLean *et al.* (2017) and Bennett *et al.* (2013, 2014) (and this study) who identify highly attenuated drumlins and glacial lineations in outer Lancaster Sound (Fig. 2), the preservation and strategic position of which is consistent with an LGM to deglacial age. Additional supporting evidence for a grounded LGM ice stream in eastern Lancaster Sound is provided by studies of IRD provenance in central Baffin Bay (Simon *et al.* 2014) and mapped and modelled ice sheet dynamics for the late Wisconsinan LIS (Margold *et al.* 2015a, b; Stokes *et al.* 2016).

The primary arguments for floating LGM ice in outer Lancaster Sound are founded on the interpretation of locally derived and extra-local glacial deposits and landforms on northern Bylot Island (Klassen 1981, 1985, 1993; Klassen & Fisher 1988). Tills containing abundant erratic material and detrital shell fragments have been recorded up to 500 m a.s.l. on the northern Bylot Island coast (Fig. 7; Hodgson & Haselton 1974; Klassen 1993) and assigned to the “foreign Eclipse Glaciation” pre-dating 43 ka BP by Klassen & Fisher (1988) and Klassen (1993). The altitudinal extent of these tills, which contain erratics likely sourced from Prince Regent and Admiralty inlets, indicates an ice thickness of ≥1500 m. Such ice thicknesses are too great to have been supported by a floating contiguous ice shelf given water depths of 800-1000 m and the necessary freeboard required to emplace till up to elevations ~500 m a.s.l. (Hodgson & Haselton 1974; Klassen 1993). The tills thus record a former glaciation that supported a grounded ice stream in outer Lancaster Sound, although their correlation to the Early Wisconsinan can be questioned. Klassen (1993) based the age of the Eclipse Glaciation on problematic molluscan amino acid ratios (i.e. overlapping and poorly-defined ratios for mollusc fossils considered to span more than 50 ka) and scant radiocarbon dating in Eclipse Sound, with north coast deposits being assigned an Eclipse age primarily by correlation. Furthermore, the Button drift, which contains erratics (up to 40 m a.s.l.; Fig. 7) from non-Bylot Island source areas was also correlated to a pre-LGM glaciation based solely on limited amino acid ratios (Klassen 1993). Indeed, Klassen (1993) reported that there is no direct stratigraphic evidence for the Button drift on the north coast of the island. Consequently, the chronostratigraphic relationship between the Eclipse moraines and the Button drift remains uncertain. Based on our new results of Late Wisconsinan deglacial environments, we propose that the Eclipse moraines were deposited by a grounded ice stream in Lancaster Sound following the LGM and that the Button drift was deposited subsequently by a deglacial ice shelf. Our reinterpretation is consistent with recent marine data from outer Lancaster Sound (Li *et al.* 2011; Bennett *et al.* 2013, 2014; MacLean *et al.* 2017), and more broadly with the re-assessment of the extent of LGM glaciation across the CAA (e.g. England *et al.* 2009; Lakeman & England 2012, 2013; Batchelor *et al.* 2012, 2014; Briner *et al.* 2003, 2009). Thus, the basal age of ~13.2 cal. ka BP from core 2011804-0010 PC constitutes a minimum-limiting age for emplacement of the Eclipse moraines and subsequent deglaciation, and also a maximum-limiting age for a deglacial ice-shelf that may have deposited the Button drift.

Origin of "red" sediments

The prominent red massive to laminated clay bed (Unit 4) interrupts otherwise massive to crudely stratified stony muds (units 3 and 5) in 2011804-0010. Dated to between 13.2 and 12.5 cal. ka BP, the unit is interpreted as a cessation in clastic debris ice-rafting beneath a cover of pervasive floating ice. All investigated GSC-A piston cores from Parry Channel (Table 2; Fig. 2) that sample immediate deglacial condition show similar lonestone-free (or -limited) clay intervals within typical ice-proximal diamictic stony muds and stratified rainout tills, although the prominent red colouration of this unit appears confined to eastern Lancaster Sound.

The petrological and mineralogical signature of this unit, and thus its provenance, is difficult to determine in 2011804-0010 due to the near absence of IRD-sized grains. Nevertheless, characteristically red lithologies outcrop on northern Baffin and Bylot islands that may represent potential sources (Fig. 2). In the Adams and Strathcona sounds of Borden Peninsula (Admiralty Inlet), red to purple quartz arenites of the Nauyat Formation (Jackson & Ianelli 1981) and red shales of the Society Cliffs Formation (Lemon & Blackadar 1963a, b) form prominent exposures. Southwest near Agu Bay (Gulf of Boothia/Prince Regent Inlet), red sandstones of the Nyeboe and Agu Bay formations (Blackader 1958, 1970; Chandler 1988) represent another potential source, as do outcrops of the Mary River Iron Formation (Jackson 2000; Johns & Young 2006) in central northern Baffin Island. Klassen (1981, 1993) noted that erratic clasts in tills (Eclipse drift) on northern Bylot Island are derived entirely of material originating on northern Baffin Island and potentially Foxe Basin, with an absence of lithologies from sources to the north or west of Bylot Island. Importantly, erratic clasts of red-brown volcanic rocks of the Nauyat Formation from Adams Sound occur only on the north coast of Bylot Island (Klassen 1993), with rare clasts of the Mary River Iron Formation occurring only on the island's southernmost coast. The Arctic Bay area (Nauyat and Arctic Bay formations) thus represents the most likely source for the fine-grained red sediments encountered in Unit 4 and throughout eastern Lancaster Sound, though direct geochemical investigation is required to confirm this hypothesis. Notably, piston cores containing early Holocene red-brown muds and sandy turbidites markedly similar in colour to Unit 4 (and its equivalents) in Lancaster Sound have been recovered from Strathcona Sound (Table 2, Fig. 2; Lewis *et al.* 1977a) supporting an Arctic Bay area sediment source. The IRD signature of the underlying Unit 5 is also consistent with a general Admiralty Inlet origin. Englacial transport of Adams/Strathcona Sound source debris into Lancaster Sound and subsequent suspended sediment deposition from meltwater plumes is implied based on chronology, distribution of cores containing red clays, and ice flow indicators. Unit 4 was deposited prior to 12.5 cal. ka BP, whilst Strathcona and Adams sounds did not become ice-free until 9.9-9.3 cal. ka BP (≈9.5-9.0 ^{14}C ka BP, Dyke & Hooper 2001). MSGLs, streamlined drumlins, and crag and tail features in Admiralty Inlet confirm ice flowed northwards into Lancaster Sound (Fig. 2) before turning eastwards to contribute to the southern component of the Lancaster Sound Ice Stream (De Angelis & Kleman 2005, 2007; MacLean *et al.* 2014, 2017; this study). Similar ice-moulded bedforms and striae indicative of erosion by eastward flowing Lancaster Sound ice have been described from the northern Bylot Island coast near Cape Hay (Klassen 1993). Taken together, it is inferred that the red clay was deposited by sediment plumes emanating from a deglacial trunk glacier in Admiralty Inlet, most likely from a potential grounding line on a prominent bedrock sill at the northern end of the inlet, and constituting a valuable regional marker horizon.

Deglacial ice shelf development

Lithofacies models for Antarctic ice shelves (e.g. Powell 1984; Domack & Harris 1998; Evans & Pudsey 2002; McKay *et al.* 2009; Christ *et al.* 2015) show remarkable similarities to deglacial NWP records, including 2011804-0010. In Antarctic records, rapid ice margin retreat, grounded ice lift-off and sub-ice shelf cavity development result in coarse muddy rainout from debris-rich basal ice immediately adjacent to the grounding line (Alley *et al.* 1989; Domack & Harris 1998; Christoffersen *et al.* 2010). This is succeeded by massive to laminated fine-grained sedimentation with limited/no lonestone deposition beneath debris-free ice above the main sub-ice shelf cavity (Domack & Harris 1998; McKay *et al.* 2009; Christ *et al.* 2015; Muto *et al.* 2016), followed by a subsequent return to coarse sedimentation and IRD deposition at the calving line associated with iceberg mobility and the delivery of en-/supra-ice shelf debris (Evans & Ó Cofaigh 2003; Gilbert & Domack 2003). Such lithofacies progressions are also reflected in microfossil abundances and assemblage changes, as demonstrated from Antarctica (e.g. Domack *et al.* 1995; McKay *et al.* 2009; Kilfeather *et al.* 2011) and the Atlantic margin of the Late Devensian/Midlandian British-Irish Ice Sheet (Peters *et al.* 2015). Foraminiferal abundances and productivity proxies (e.g. TOC, $\delta^{13}C_{org}$) typically decrease from *in situ* deposition at calving margins to sparse populations in sub-ice shelf cavity deposits (Domack *et al.* 1995; Kilfeather *et al.* 2011; Christ *et al.* 2015) advected from beyond the calving line by tidal and thermohaline pumping (e.g. O'Brien *et al.* 1999; Hemer *et al.* 2007). Although established benthic ecosystems have been documented from beneath stable, centennially-persistent Antarctic ice shelves characterised by the basal freeze-on of marine water (Post *et al.* 2007, 2014; Riddle *et al.* 2007; Rose *et al.* 2015), the development of *in situ* communities below short-lived ice shelves in rapidly deglaciating systems (cf. Knight 2006; Peters *et al.* 2015) is considered unlikely due to the ephemeral, variable, and stressed nature of these environments.

This picture of Antarctic sub-ice shelf deposition accords well with 2011804-0010 (Fig. 8). The massive pebbly sandy mud of Unit 5 is interpreted as initial sub-ice shelf cavity rainout from debris-rich basal ice. The till pellets present in this unit may be the product of sediment shearing under streaming glacial ice and re-deposited during ice shelf lift-off (Goldschmidt *et al.* 1992; Domack *et al.* 1998; Cowan *et al.* 2012). High current velocities due to thermohaline and tidal pumping within a confined sub-ice shelf cavity near the grounding line (Harris & O'Brien 1998; O'Brien *et al.* 1999; Hemer *et al.* 2007) are suggested by the presence of *L. lobatula* (Hansen & Knudsen 1995; Korsun & Hald 1998; Murray 2006). *C. neoteretis* (Seidenkrantz 1995; Jennings *et al.* 2004) and high planktic:benthic ratios (Fig. 6) further indicate the advection of foraminifera beneath the ice shelf to the grounding line from a distal (though contemporaneous) seasonally open Atlantic source. Following Antarctic models, Unit 4 is considered a low energy "Null Zone" (*sensu* Domack & Harris 1988) sub-ice shelf cavity deposit (Fig. 8), with low absolute foraminiferal numbers and diversities (Domack *et al.* 1995; Christ *et al.* 2015) and no coarse ice-rafted sedimentation.

Correspondingly, Unit 3 represents ice shelf break-up deposition associated with iceberg rafting and delivery of englacial and supraglacial debris, consistent with ^{14}C dates indicative of rapid sedimentation. This is supported by highly variable absolute foraminiferal abundances and P:B ratios (Fig. 6), with assemblages marked by both glacimarine (*C. reniforme*, *E. excavatum* f. *clavata*; Korsun & Hald 1998) and seasonally open Arctic shelf species (*I. norcrossi*, *I. helenae*; Vilks 1969, 1989). In this environment, fluctuating clastic sedimentation rates resulting in variable foraminiferal absolute abundances are to be expected (Evans & Ó Cofaigh 2003; Gilbert & Domack 2003). Although Kilfeather *et al.* (2011) and Pudsey & Evans (2001) suggest ice shelf calving line facies should show greater diversity in IRD species compared to sub-ice shelf units due to deposition by icebergs from multiple sources, the IRD spectra of units 5 and 3 are near identical (though a greater diversity is seen towards the top of Unit 3; Fig. 4). However, in rapidly ablating or collapsing ice shelf settings supplied by fast-flowing ice streams, pronounced supraglacial sediment loads can develop, derived from originally sub- and englacial sources (Glasser *et al.* 2006, 2014; Nicholls *et al.* 2012). Thus the lithologies of ice shelf break-up deposits may strongly resemble those from subglacial and sub-ice shelf rainout facies (Reinardy *et al.* 2009), particularly in the case of topographically constrained ice streams/shelves surrounded by steep

531 slopes and tributary glaciers, as was the case at the core site with ice exiting Admiralty and Navy Board inlets.

Furthermore, an interpretation of former deglacial ice shelf conditions (prior to 12.5 cal. ka BP) off northwest Bylot Island (Fig. 8) is supported by the recognition of the Button drift along the island's north coast as an ice shelf moraine (Fig. 7; Klassen 1981, 1985, 1993; Dyke 1999). Our interpretation is further strengthened if, in light of evidence for grounded LGM ice in eastern Lancaster Sound (Li *et al.* 2011; Bennett *et al.* 2013, 2014; MacLean *et al.* 2017), Button drift ice shelf deposits are assigned to a post-LGM regional deglaciation.

Previously, similar deglacial stratigraphies from elsewhere in Parry Channel have been interpreted as suspended sediment deposition beneath pervasive landfast sea-ice succeeding clastic ice-proximal sedimentation under seasonally open water conditions (Pieńkowski *et al.* 2012, 2013, 2014). However, a climatologic or dynamic model for ice-proximal pervasive sea ice formation in a rapidly deglaciating setting could be problematic, given conditions of elevated meltwater discharge maintaining sea ice mobility (*sensu* Dyke *et al.* 1997). The close correspondence between lithofacies models for Antarctic ice shelves and the stratigraphy seen in 2011804-0010, coupled with a reassessment of the chronology of moraine emplacement on northern Bylot Island, lends credence to proposals for a deglacial ice shelf in eastern Lancaster Sound rather than pervasive sea ice. However, it should be noted that elsewhere in the CAA, the lack of ice shelf moraines similar to those formed by the former Viscount Melville Sound Ice Shelf (Hodgson & Vincent 1984; Hodgson 1994; England *et al.* 2009; Furze *et al.* 2013, 2016) and adjacent to several modern Antarctic ice shelves (Fitzsimons 1997; Hjort *et al.* 2001; Smith *et al.* 2006; Roberts *et al.* 2008; Fitzsimons *et al.* 2012), cannot be considered as evidence for deglacial ice shelf absence. We suggest that rapid ice stream float-off and ice shelf formation within the main channels of the CAA, whilst grounded glacial ice still occupies adjacent islands to some distance offshore (e.g. Pieńkowski *et al.* 2014), would likely not result in significant ice shelf moraine formation (Fig. 8). Only in settings where coasts have been deglaciated prior to ice shelf formation and/or where an ice shelf actively advances and grounds along an ice-free coast (e.g. Viscount Melville Sound; Hodgson & Vincent 1984; Hodgson 1994) may ice shelf moraines be constructed (Fig. 8).

Radiocarbon dates from the immediate ice shelf lift-off and calving margin facies (units 5 and 3 respectively) suggest that the ice shelf persisted for some 1000 cal. a, formation immediately predating the onset of the YD chronozone (12 890-11 650 cal. a BP; Stuiver *et al*. 1995) and ice shelf collapse occurring towards the end of the YD. While the record does not permit a distinction to be made between a purely dynamic versus climatic cause for ice shelf development, cold dry conditions evidenced in the eastern Canadian Arctic during the YD (Cuffey & Clew 1997; Miller *et al.* 2005; Briner *et al.* 2009; Young *et al.* 2012) would have likely favoured ice shelf stability. Enhanced sea-ice occupancy and rigid mélange formation would limit calving and help buttress the floating glacial ice margin (Khazendar *et al.* 2009; Amundson *et al.* 2010; Moon *et al.* 2015). The extent to which the formation and persistence of the deglacial eastern Lancaster Sound ice shelf was a direct consequence of presumed Younger Dryas cooling must, however, at this stage remain speculative.

Palaeoceanography

The appearance of *C. neoteretis* and *N. pachyderma* (Figs 5, 6) in the early part of the record argues for the advection of cool, Atlantic-derived water (Vilks 1974) over the core site and beneath the proposed ice shelf. These species are currently absent from the central CAA (Vilks 1969, 1974, 1989), only observed in adjacent offshore waters of Atlantic origin in Baffin Bay, the Beaufort Sea, and the Arctic Ocean (Vilks 1974; Darling *et al.* 2007; Xiao *et al.* 2014). The presence of these taxa implies productive conditions in northern Baffin Bay/eastern Lancaster Sound ~13.2-12.5 cal. ka BP close to the ice shelf covered core site. The delivery of such an Atlantic water mass can only occur via Davis Strait and central Baffin Bay given that Parry Channel, Nares Strait, and the Queen Elizabeth Islands remain blocked by grounded glacial ice at this time (Zreda *et al.* 1999; Dyke 2004; England *et al.* 2006; Pieńkowski *et al.* 2014). This is supported by palaeoenvironmental data from northern Baffin Bay indicative of extensive, but seasonal, sea ice (Knudsen *et al.* 2008) and delivery of Lancaster Sound-derived IRD to central Baffin Bay implying mobile ice conditions (Simon *et al.* 2014). The hypothesis that dense sea ice persisted in Baffin Bay into the Early Holocene (Gibb *et al*. 2015) is not supported by our results, or those of Knudsen *et al*. (2008).

Above the postulated ice shelf facies (units 5-3), foraminiferal assemblages of units 1 and 2 imply more favourable environmental conditions, remaining glacimarine, but seasonally open, in character. Despite low TOC and BioSil values (Fig. 6), these glacimarine ecosystems were sufficiently productive to support a benthic community, including ostracods and molluscs. The appearance of *A. gallowayi* and *L. lobatula* in tandem with increasing *N. pachyderma* towards the core tops suggests increased bottom current velocities (Murray 2006) and enhanced Atlantic water advection (Darling *et al.* 2007; Xiao *et al.* 2014) over the core site. Chronologically (~11.0 cal. ka BP), this interval broadly coincides with increases in planktic foraminifera described from the central CAA which are linked to the final separation of Laurentide and Innuitian ice across the Lowther-Young islands sill and the establishment of oceanic throughflow along Parry Channel, as well as the onset of postglacial sedimentation (Pieńkowski *et al.* 2012, 2013, 2014). The extent to which the trends in foraminiferal assemblages at the top of 2011804-0010 represent these region-wide events must remain speculative, given the absence of Holocene sediments in the record and the currently limited direct chronological control on LIS-IIS unsuturing, oceanographic re-organization, and ecosystem response in the CAA.

Absence of Holocene sediments

Although both core components sample the same stratigraphic interval (0010PC extending deeper), sediments younger than ~11.0 cal. ka BP are absent. Significant loss of sediment during coring is unlikely given the remarkably similar recovery in both cores. The absence of Holocene sediments (<11.0 cal. ka BP) is thus attributed to either non-deposition or erosion/slope failure. Significant non-deposition at this location, only 14.5 km distal to the Holocene tidewater calving margin of glacier D181 (Fig. 7) and subject to ongoing iceberg rafting and suspended sediment plume sedimentation (Dowdeswell *et al.* 2007; Van Wychen *et al.* 2015), is unlikely. Nevertheless, foraminiferal assemblages in Unit 1 do imply increasing current velocities, and thus the potential for winnowing, towards the core top. Whilst current winnowing and lag deposits are encountered throughout Parry Channel, frequently resulting in poor piston- and gravity-core penetration (e.g. Falconer 1977; Lewis *et al.* 1977b), such lags are not evident in 2011804-0010. Consequently, slope failure and mass movement remain the most likely explanations for the apparent missing Holocene at this location. Notably, nearby core (core 40, Fig. 2, Table 2) and seismic data from a similar bathymetric setting indicate mass transport deposition. The core site (837 m depth) 14.5 km north of the calving margin of glacier D181 (Fig. 7) is ideally configured to favour mass failure and rapid downslope sediment transfer potentially linked to a Little Ice Age (LIA) ice advance during which the glacier (and formerly confluent D183) achieved its postglacial maximum (Klassen 1993; Dowdeswell *et al.* 2007). A recent (LIA or later?) failure is consistent with the absence of any mid to late Holocene sediments in 2011804-0010. Given that Lancaster Sound and Baffin Bay have been seismically active during the Holocene (Stein 1979; Basham *et al.* 1997; Bent 2002; Adams & Halchuk 2003), a seismic origin for slope failure and mass transport deposition offshore of northern Bylot Island may also be a possibility. An expansion of the currently limited seismic and multibeam resolution and data coverage for the study region, and further litho- and chronostratigraphic investigation may permit the confident identification of past sediment mass failures and the frequency of their occurrence.

Conclusions

Core 2011804-0010 provides an important high-resolution perspective on the marine based deglaciation of the CAA, in a region previously occupied by a major ice stream of the Laurentide and Innuitian ice sheets. The core suite records the transition from ice-proximal to ice-distal conditions in detail and, for the first time, permits a chronologically-constrained examination of deglacial ice shelf development and marine dynamics in the eastern Canadian Arctic. Such a high resolution deglacial record is unique within the context of the Canadian Arctic and allows the application of ice shelf lithofacies models previously defined from Antarctic settings to reconstruct rapid ice shelf formation and collapse in an important oceanographic gateway connecting the Arctic and Atlantic oceans.

The following conclusions and implications can be drawn from the analysis of this record:

- Grounded ice forming a major eastward-flowing ice stream likely occupied eastern Lancaster Sound at LGM up to deglaciation, draining the confluent margins of the Innuitian and Laurentide ice sheets. This is contrary to existing interpretations of the terrestrial surficial geology and geomorphology of Bylot Island (Klassen 1985, 1993; Klassen & Fisher 1988), but in agreement with recent marine reconstructions (MacLean *et al.* 2010, 2017; Li *et al.* 2011; Bennett *et al.* 2013, 2014; Margold *et al.* 2015a, b).
- As demonstrated by both IRD (this study) and terrestrial tills (Hodgson & Haselton 1974; Klassen 1985, 1993), ice and sediment on the southern margin of the Lancaster Sound ice stream was

derived from northern Baffin Island sources, in particular Admiralty Inlet. This suggests pronounced lateral flow partitioning within the trunk glacier from LGM to deglaciation.

- Deglaciation occurred prior to ~13.2 cal. ka BP, marked by the destabilisation, lift-off, and rapid grounding line retreat of the Lancaster Sound trunk glacier and the development of an extensive ice shelf potentially emplacing moraines on northern Bylot Island. This is supported by lithofacies models and foraminiferal assemblages indicative of sub-ice shelf and calving line deposition. The ice shelf persisted in eastern Lancaster Sound through much of the Younger Dryas chronozone until ~12.5 cal. ka BP. Our results are consistent with the rapid deglaciation of eastern Parry Channel previously described by Pieńkowski *et al.* (2014).
- Ice shelf collapse and transition from ice-proximal to -distal conditions are marked by a characteristic lithofacies progression and subtle shifts in IRD species from northern Baffin Island and Admiralty Inlet sources to northwest Bylot Island sources. This is consistent with the increasing importance of local glacially-driven sediment delivery. Foraminiferal faunas from this interval are marked by a mix of glacimarine and Arctic shelf species, and taxa indicative of elevated productivity and current velocity.
- Atlantic watermass inflow evidenced by *N. pachyderma* and *C. neoteretis* early in the record implies seasonally open conditions in Baffin Bay prior to 12.5 cal. ka BP given that other marine routes remain occupied by grounded glacial ice during this time. This contrasts with proposals that central Baffin Bay was covered by severe sea ice until 7.4 cal. ka BP (*sensu* Gibb *et al.* 2015).
- The absence of Holocene sediments in the core record (and mass transport deposits in GSC-A core 2008029-0053PC) provides evidence of sediment failures at the core site. Increased sedimentation from nearby glaciers associated with Little Ice Age advance and/or recent seismic activity are potential triggers for these failures, which must be better understood in order to improve regional geohazard risk assessments.

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List of Figures and Tables

Fig. 1. Map of study area and core location. A. Location of study area within Parry Channel and the Canadian Arctic Archipelago. Parry Channel constitutes the main axis of the Northwest Passage, from Baffin Bay westwards through Lancaster Sound, Barrow Strait, Viscount Melville Sound, and M'Clure Strait to the Arctic Ocean. Is = Island; Sd = Sound; St = Strait; PoW Is = Prince of Wales Island; Som Is = Somerset Island. B. Core location and bathymetry of Lancaster Sound and adjacent channels. Bathymetry (m) derived from GEBCO database (http://www.gebco.net/).

Fig. 2. Geological Survey of Canada piston core locations, bedrock geology, and subglacial bedforms mapped from the floors of the main channels. Information on cores is given in Table 2. Cores extending to deglaciation and possessing red-coloured glacimarine clay units are marked by red-filled circles. Those extending to deglaciation but lacking red-coloured deposits are marked by grey-filled circles whilst cores with insufficient recovery to prove deglacial deposits are marked with white-filled circles. Cores for which no materials have been archived and no detailed records exist (primarily cruise no. 76025) are not shown. Note that multibeam coverage for the region is limited, particularly in Prince Regent and Admiralty inlets and western Lancaster Sound, such that the distribution of mapped bedforms in this area does not fully encompass their true occurrence. Bedrock geology simplified from Scott & de Kemp (1998) and Harrison *et al.* (2015a, b, c).

Fig. 3. Lithostratigraphy of core 2011804-0010 trigger weight and piston components, including X-radiographs, calibrated radiocarbon dates and unit descriptions and interpretations. For details on radiocarbon dates, see Table 1.

Fig. 4. Results of IRD (>250 µm) analyses on core 2011804-0010, showing selected mineral and lithic species. A full list of all IRD categories is given in Table S1.

Fig. 5. Relative abundances of selected benthic foraminiferal species from core 2011804-0010. For a full list of species found, see Table S2.

Fig. 6. Microfossil absolute abundances, planktonic to benthic foraminiferal ratios, and biogeochemical results for core 2011804-0010.

Fig. 7. Northern Bylot Island showing core locations, modern glacial margins, and distribution of Eclipse drift and moraines and Button moraines (after Klassen 1993).

Fig. 8. Conceptual model of deglaciation and ice shelf formation along the Lancaster Sound coast of Bylot Island. Illustrations depict cross-sectional views of Lancaster Sound, with the Bylot coast on the left. Differences in deglacial style between northwest and northeast Bylot Island are shown, indicating formation of Eclipse and Button drift.

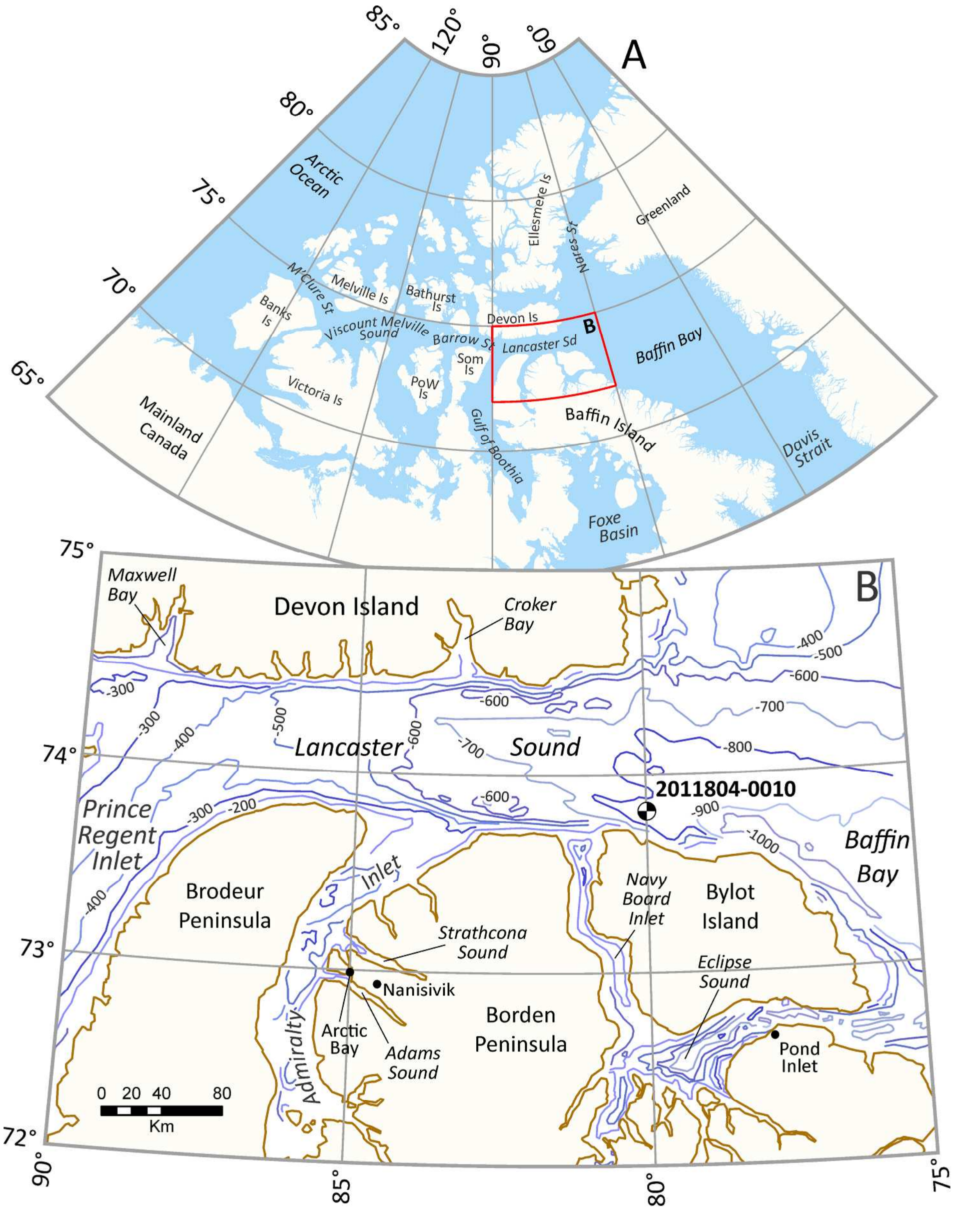
Table 1. Radiocarbon dates from GSC-A core 2011804-0010 piston and trigger-weight components. All dates measured at the National Ocean Sciences Accelerator Mass Spectrometry Laboratory at the Woods Hole Oceanographic Institution, Commonwealth of Massachusetts, United States of America. Ages calibrated using CALIB 7.1 (Stuiver *et al.* 2016) using the MARINE13 calibration curve (Reimer *et al.* 2013). ΔR values are those derived from Cao *et al.* (2007) using Reimer *et al.* (2013) for the Allerød[1] (zero), Younger Dryas[2] (185±140 ^{14}C a), and Preboreal[3] (75±25 ^{14}C a).

Table 2. List of GSC-A piston cores from the eastern CAA detailing the presence or absence of red-coloured glacimarine clay beds within core stratigraphies. Core positions are shown in Fig. 2. Natural Resources Canada (NRCan) publically accessible on-line Expedition Database (http://ed.gdr.nrcan.gc.ca) used as the primary source for records of Geological Survey of Canada piston cores taken from the study area. GSC Cruise reports and previously published records are cited where they are available for specific cruises or cores. Core recovery <50 cm listed as No/limited recovery. [1] Core identified in Pieńkowski *et al.* (2012) using old GSC core numbering convention as 86027-144. [2] Core identified in Pieńkowski *et al.* (2014) using old GSC core numbering convention as 86027-154. [3] Core identified in Ledu *et al.* (2010) using CASES no. as 2004804-009.

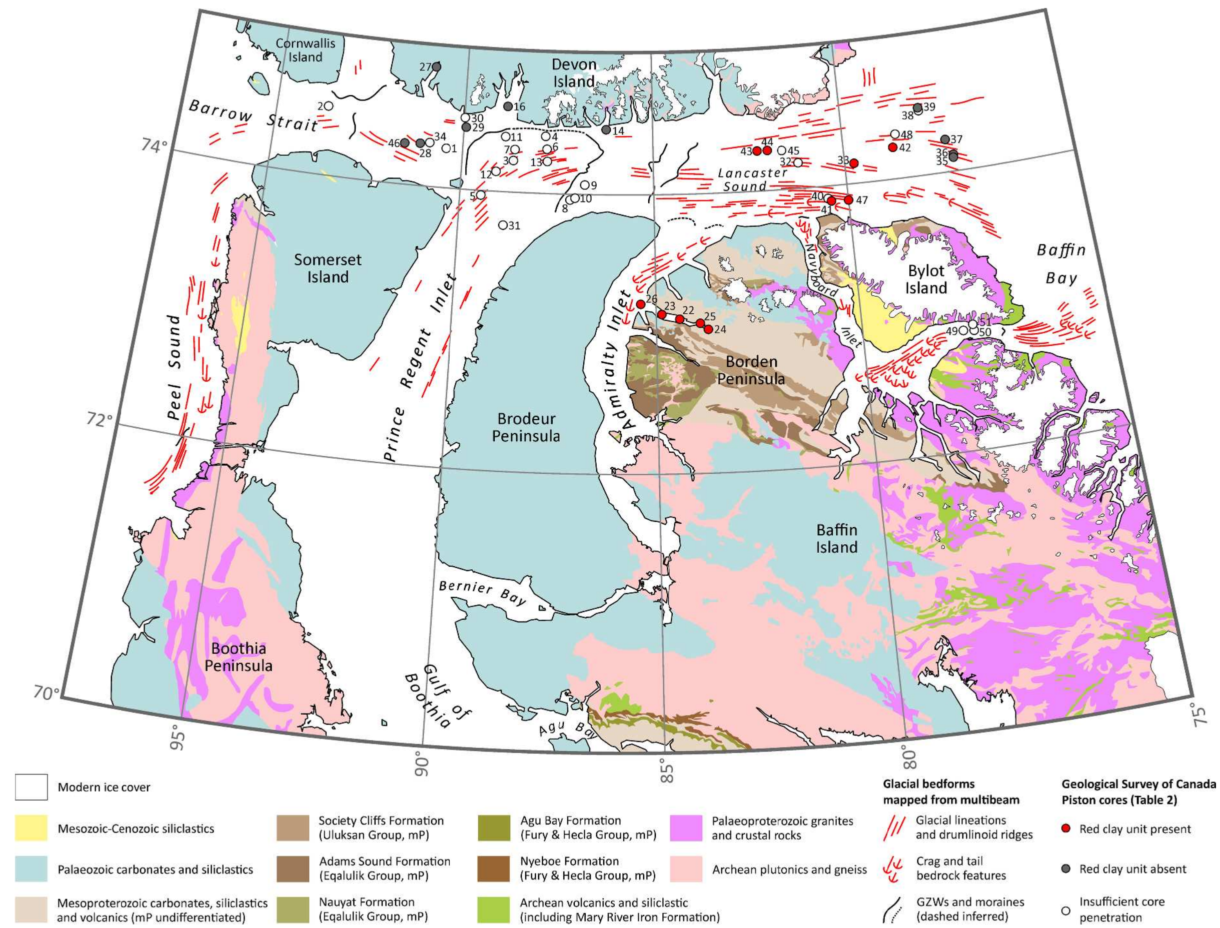
Supporting Information

Table S1. List of IRD (>250 µm) species categories used in this study.

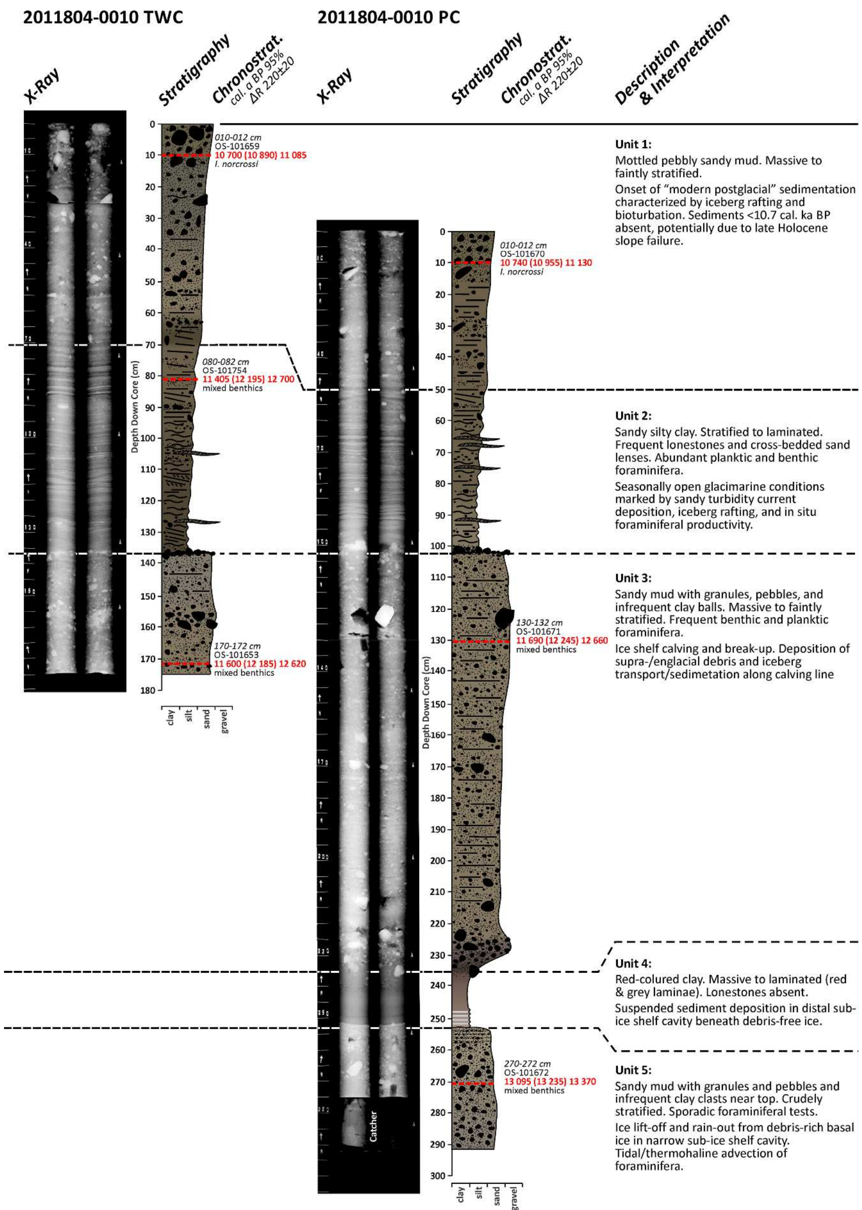
Table S2. List of taxonomic names of micro- and macro-fossils found in core 2011804-0010. Species names follow WoRMS Editorial Board (2016).



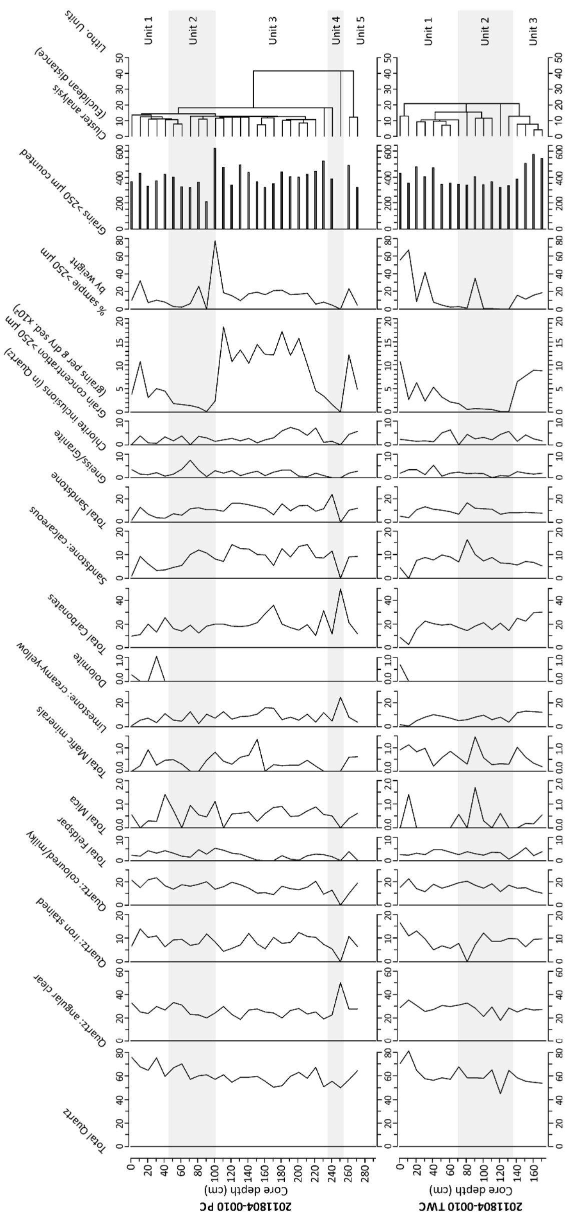
226x287mm (300 x 300 DPI)



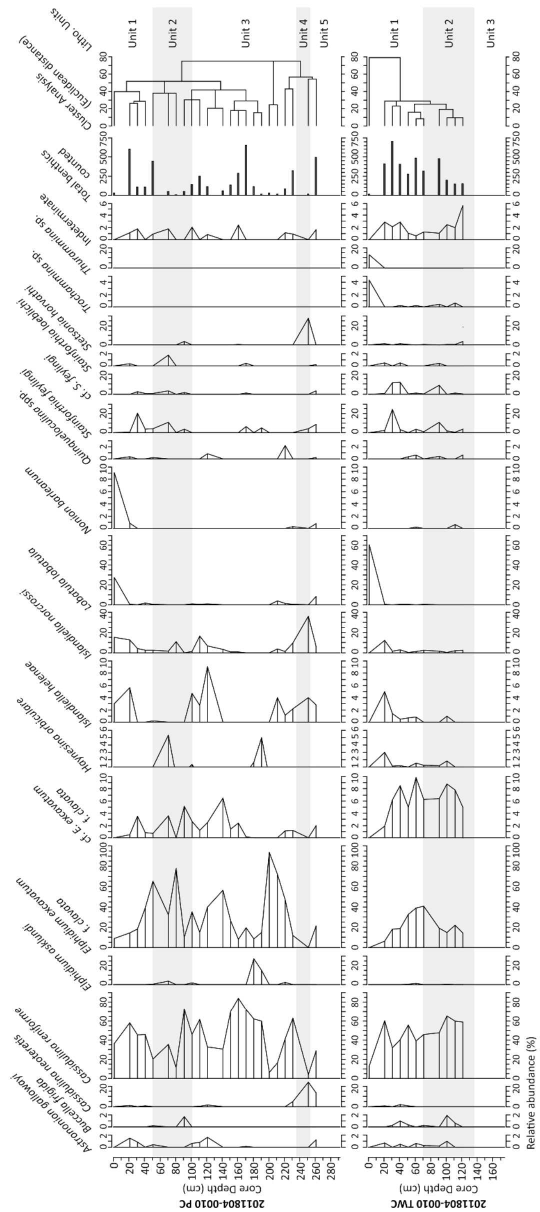
257x337mm (300 x 300 DPI)



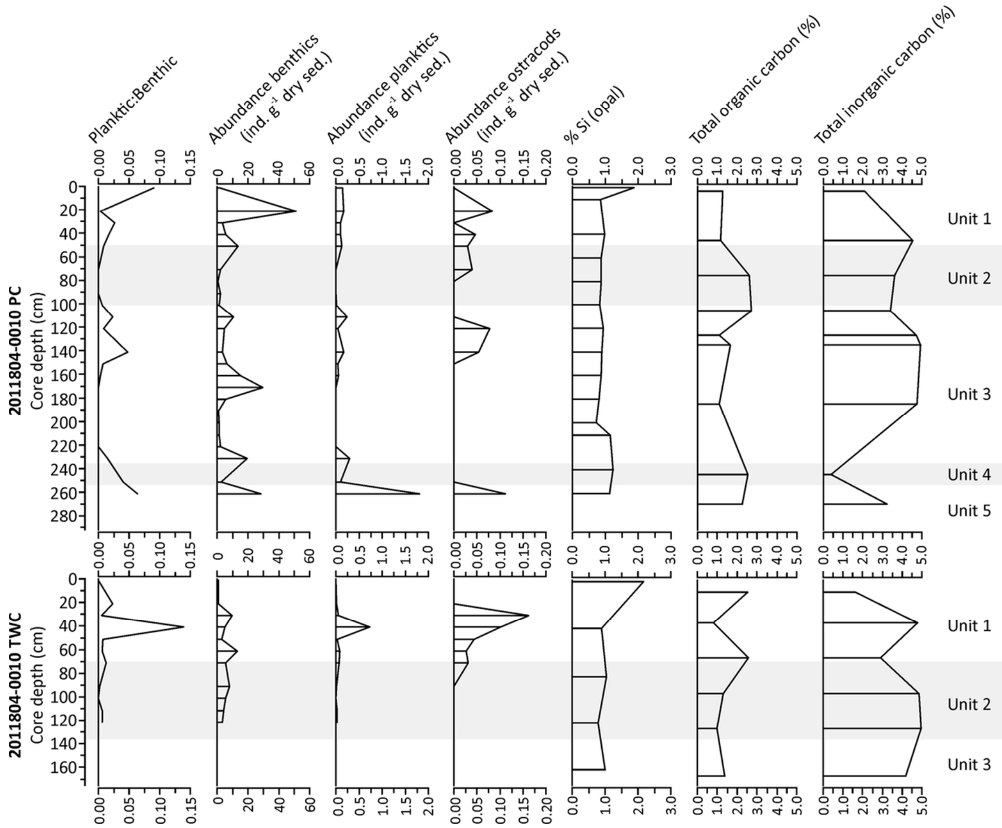
258x362mm (300 x 300 DPI)



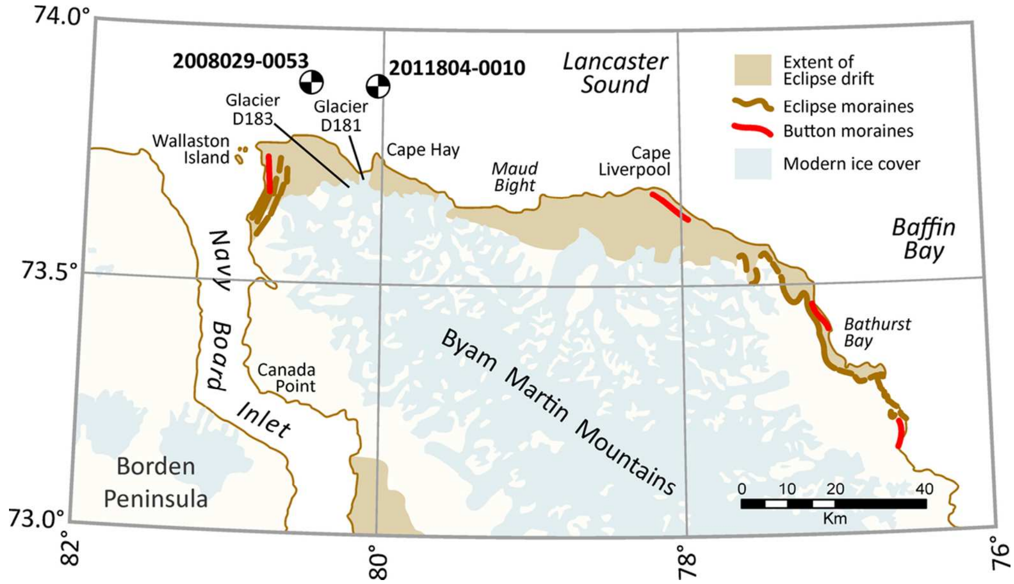
289x621mm (300 x 300 DPI)



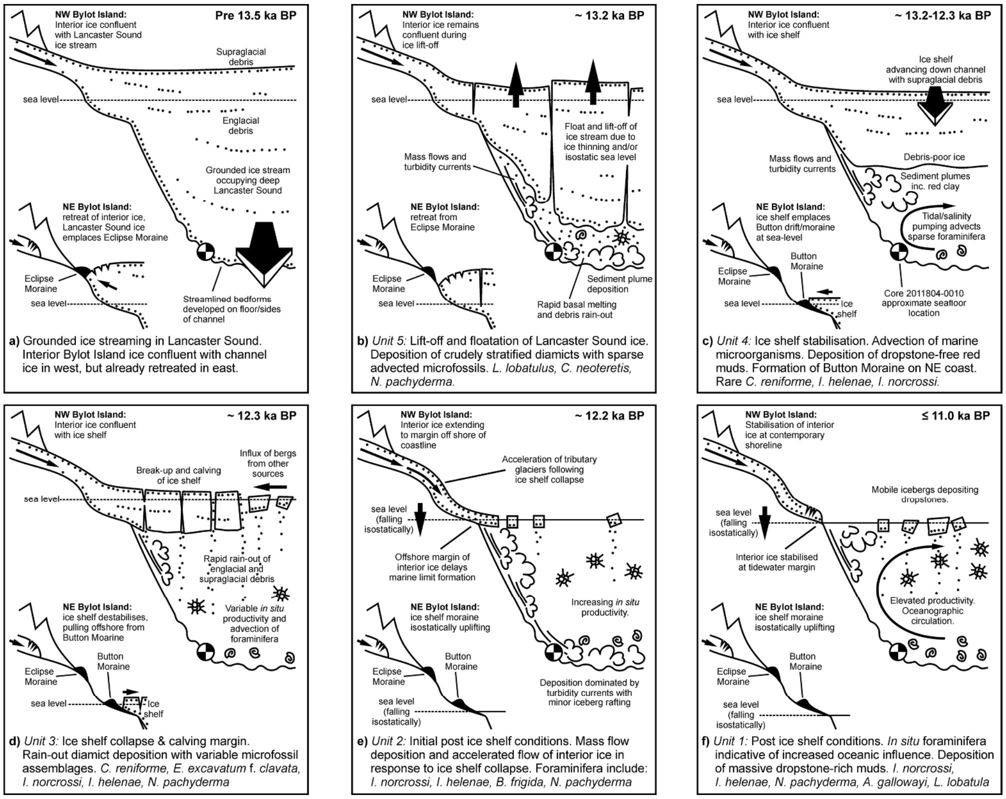
309x704mm (300 x 300 DPI)



112x91mm (300 x 300 DPI)



104x60mm (300 x 300 DPI)



a) Grounded ice streaming in Lancaster Sound. Interior Bylot Island ice confluent with channel ice in west, but already retreated in east.

b) *Unit 5:* Lift-off and floatation of Lancaster Sound ice. Deposition of crudely stratified diamicts with sparse advected microfossils. *L. lobatulus, C. neoteretis, N. pachyderma.*

c) *Unit 4:* Ice shelf stabilisation. Advection of marine microorganisms. Deposition of dropstone-free red muds. Formation of Button Moraine on NE coast. Rare *C. reniforme, I. helenae, I. norcrossi.*

d) *Unit 3:* Ice shelf collapse & calving margin. Rain-out diamict deposition with variable microfossil assemblages. *C. reniforme, E. excavatum* f. *clavata, I. norcrossi, I. helenae, N. pachyderma*

e) *Unit 2:* Initial post ice shelf conditions. Mass flow deposition and accelerated flow of interior ice in response to ice shelf collapse. Foraminifera include: *I. norcrossi, I. helenae, B. frigida, N. pachyderma*

f) *Unit 1:* Post ice shelf conditions. *In situ* foraminifera indicative of increased oceanic influence. Deposition of massive dropstone-rich muds. *I. norcrossi, I. helenae, N. pachyderma, A. gallowayi, L. lobatula*

237x187mm (300 x 300 DPI)

Core depth (cm)	Laboratory code	Dated material	Conventional ^{14}C date ±1σ (^{14}C a BP)	$\delta^{13}C$ (‰)	Calibrated age range (95.4%) (cal a BP)	Median probability age (cal a BP)
2011804-0010 Piston core						
010 - 012	OS-101670	Foraminifera: *Islandiella norcrossi*	10 050±60	-1.05	10 740 - 11 130	10 955 [3]
130 - 132	OS-101671	Foraminifera: mixed benthics	11 000±75	-2.84	11 690 - 12 660	12 245 [2]
270 - 272	OS-101672	Foraminifera: mixed benthics	11 750±65	-0.72	13 095 - 13 370	13 235 [1]
2011804-0010 Trigger weight						
010 - 012	OS-101659	Foraminifera: *Islandiella norcrossi*	10 000±55	-1.43	10 700 - 11 085	10 890 [3]
080 - 082	OS-101754	Foraminifera: mixed benthics	11 000±170	-1.80	11 405 - 12 700	12 195 [2]
170 - 172	OS-101653	Foraminifera: mixed benthics	10 950±60	-1.60	11 600 - 12 620	12 185 [2]

Number (in Fig. 2)	GSC core identifier Cruise no.	Core no. (old)	Length (cm)	Location	Latitude
1	69050	0813 (871)	0	E. Barrow Strait	74°16'48.00"N
2	69050	5438 (913)	0	E. Barrow Strait	74°28'00.01"N
3	73014	1085 (028)	61	W. Lancaster Sound	74°09'00.00"N
4	73014	1090 (051)	7	W. Lancaster Sound	74°19'18.01"N
5	73014	7819 (015)	0	W. Lancaster Sound	73°55'30.00"N
6	73014	7822 (017)	26	W. Lancaster Sound	74°15'00.00"N
7	73014	7831 (027)	11	W. Lancaster Sound	74°13'59.99"N
8	73014	7836 (031)	9	W. Lancaster Sound	73°57'42.01"N
9	73014	7840 (034)	89	W. Lancaster Sound	74°01'00.01"N
10	73014	7849 (044)	12	W. Lancaster Sound	73°58'12.00"N
11	73014	7854 (049)	37	W. Lancaster Sound	74°15'47.99"N
12	73014	7857 (051)	80	W. Lancaster Sound	74°05'30.01"N
13	73014	7861 (055)	42	W. Lancaster Sound	74°10'00.01"N
14	74026PHASE2	1326 (090)	259	W. Lancaster Sound	74°26'48.01"N
16	76025	1230 (003)	594	Maxwell Bay	74°35'35.99"N
17	76025	1232 (006)	498	Radstock Bay	74°47'31.20"N
18	76025	1242 (013)	100	Radstock Bay	74°42'42.01"N
19	76025	1245 (014)	135	Radstock Bay	74°40'19.81"N
20	76025	1249 (015)	597	E. Barrow Strait	74°17'12.01"N
21	76025	1265 (024)	221	E. Barrow Strait	74°11'12.01"N
22	76025	1288 (035)	921	Strathcona Sound	73°04'36.01"N
23	76025	1295 (040)	401	Strathcona Sound	73°07'30.00"N
24	76025	1296 (041)	424	Strathcona Sound	72°58'59.99"N
25	76025	1301 (042)	718	Strathcona Sound	73°01'43.79"N
26	76025	1304 (043)	397	Strathcona Sound	73°11'21.01"N
27	76025	1579 (009)	460	Radstock Bay	74°51'47.99"N
28	86027	9682 (144)[1]	438	E. Barrow Strait	74°15'33.59"N
29	86027	9696 (154)[2]	621	E. Barrow Strait	74°22'00.59"N
30	86027	9706 (159)	375	E. Barrow Strait	74°26'33.00"N
31	86027	9709 (162)	137	N. Prince Regent Inlet	73°43'59.99"N
32	2004804	0050[3]	593	E. Lancaster Sound	74°11'06.00"N
33	2005804	0003	577	E. Lancaster Sound	74°03'23.52"N
34	2005804	0004	697	E. Barrow Strait	74°16'09.30"N
35	2008029	0046	465	E. Lancaster Sound	74°01'23.79"N
36	2008029	0049	609	E. Lancaster Sound	74°01'34.24"N
37	2008029	0050	322	E. Lancaster Sound	74°06'44.43"N
38	2008029	0051	78	E. Lancaster Sound	74°18'26.26"N
39	2008029	0052	489	E. Lancaster Sound	74°18'25.47"N
40	2008029	0053	480	E. Lancaster Sound	73°50'25.98"N
41	2008029	0054	250	E. Lancaster Sound	73°50'20.30"N

42	2008029	0057	1092	E. Lancaster Sound	74°05'31.33"N
43	2008029	0059	734	E. Lancaster Sound	74°15'34.64"N
44	2008029	0061	451	E. Lancaster Sound	74°15'29.55"N
45	2008029	0062	76	E. Lancaster Sound	74°15'09.11"N
46	2011804	0007	391	E. Barrow Strait	74°15'01.80"N
47	2011804	0010	308	E. Lancaster Sound	73°48'30.00"N
48	2011804	0012	317	E. Lancaster Sound	74°11'37.20"N
49	2013029	0065	817	Pond Inlet	72°48'53.61"N
50	2013029	0066	630	Pond Inlet	72°50'56.75"N
51	2013029	0067	1087	Pond Inlet	72°48'56.04"N

Longitude	Desription
90°16'59.99"W	No archieved core, no detailed records
93°31'59.99"W	No archieved core, no detailed records
88°37'59.99"W	No archieved core, no detailed records
87°58'12.00"W	No archieved core, no detailed records
89°20'60.00"W	No archieved core, no detailed records
87°57'00.00"W	No archieved core, no detailed records
88°37'59.99"W	No archieved core, no detailed records
87°19'00.01"W	No archieved core, no detailed records
86°55'00.01"W	No archieved core, no detailed records
87°13'00.01"W	No archieved core, no detailed records
88°54'00.00"W	No archieved core, no detailed records
88°55'05.99"W	No archieved core, no detailed records
87°57'29.99"W	No archieved core, no detailed records
86°19'05.99"W	No red clay unit observed
88°49'59.88"W	Lewis *et al.* (1977) report no red unit
90°55'12.00"W	No archieved core, no detailed records
91°06'18.00"W	No archieved core, no detailed records
90°57'06.01"W	No archieved core, no detailed records
91°04'05.99"W	No archieved core, no detailed records
88°45'24.01"W	No archieved core, no detailed records
84°21'47.99"W	Red-brown muds and sandy turbidites
84°51'29.99"W	Red-brown muds and sandy turbidites
83°39'36.00"W	Red-brown muds and sandy turbidites
83°48'06.01"W	Red-brown muds and sandy turbidites
85°22'59.99"W	Red-brown muds and sandy turbidites
90°49'23.99"W	No red clay unit observed
91°14'12.59"W	Dropstone-free laminated unit, no red unit
89°51'15.59"W	Dropstone-free laminated unit, no red unit
89°52'30.00"W	Not proving ice-proximal glacimarine
88°44'12.01"W	Not proving ice-proximal glacimarine
81°12'36.00"W	Not proving ice-proximal glacimarine
79°51'11.64"W	Reddish-pink laminated mud, 560cm, 586cm
91°05'02.88"W	Not proving ice-proximal glacimarine
77°06'58.31"W	No red clay unit observed
77°07'30.95"W	No red clay unit observed
77°24'03.00"W	No red clay unit observed
78°01'12.14"W	Not proving ice-proximal glacimarine
78°01'10.56"W	No red clay unit observed
80°23'40.49"W	Not proving ice-proximal glacimarine
80°18'43.48"W	Red Clay no/few dropstones, 50 cm

78°43'05.37"W	Red Clay no/few dropstones, 925 cm
82°23'02.94"W	Red Clay no/few dropstones, 250 cm
82°13'49.27"W	Red Clay no/few dropstones, 50 cm
81°38'05.44"W	Not proving ice-proximal glacimarine
91°33'54.60"W	Dropstone-free laminated unit, no red unit
80°00'32.40"W	Red Clay no/few dropstones, 230 cm
78°38'33.60"W	Not proving ice-proximal glacimarine
77°40'35.99"W	Not proving ice-proximal glacimarine
77°26'29.56"W	Not proving ice-proximal glacimarine
77°25'34.63"W	Not proving ice-proximal glacimarine

Information source
NRCan Expedition Database, Reiniger & Latremouille (1987)
NRCan Expedition Database, Reiniger & Latremouille (1987)
NRCan Expedition Database, Williams (1973)
NRCan Expedition Database, Williams (1973)
NRCan Expedition Database, Williams (1973)
NRCan Expedition Database, Williams (1973)
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NRCan Expedition Database, Williams (1973)
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NRCan Expedition Database, Lewis *et al.* (1977a)
NRCan Expedition Database, Lewis *et al.* (1977a)
NRCan Expedition Database, Lewis *et al.* (1977a)
Observed by authors, Lewis *et al.* (1977a)
Observed by authors, Lewis *et al.* (1977a)
Observed by authors, Lewis *et al.* (1977a)
Observed by authors, Lewis *et al.* (1977a)
Observed by authors, Lewis *et al.* (1977a)
Observed by authors, Lewis *et al.* (1977a)
Observed by authors, Pieńkowski *et al* . (2012), MacLean (1986)
Observed by authors, Pieńkowski *et al.* (2014), MacLean (1986)
MacLean (1986)
MacLean (1986)
Ledu et al. (2010), Bennett *et al.* (2008a)
Observed by authors, Bennett *et al.* (2008b)
Ledu et al. (2010), Bennett *et al.* (2008b)
Observed by authors
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NRCan Expedition Database
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Observed by authors
Observed by authors
NRCan Expedition Database

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This study, Observed by authors
Observed by authors

Observed by authors
Observed by authors
Observed by authors

Table S1. List of IRD (>250 µm) species categories used in this study.

Quartz minerals
Rounded, Clear
Rounded, Milky
Rounded, Iron Stained
Angular, Clear
Angular, Milky
Angular, Iron Stained
Rounded, Coloured
Angular, Coloured

Feldspar minerals
Pink
Grey
White

Mica minerals
Biotite/Phlogopite (dark)
Muscovite (light)
Other

Mafic minerals

Carbonate lithics
Limestone, White
Limestone, Grey
Limestone, Brown
Limestone, Creamy-yellow
Dolomite

Sandstone lithics
Calcareous Sandstone
Sandstone

Metamorphic/Igneous lithics
Schist
Phylite
Gneiss/Granite
Gabbro
Diorite

Other Lithics

Other Mineral Grains

Chlorite inclusions in Quartz

Table S2. List of taxonomic names of micro- and macro-fossils found in core 2011804-0010. Species names follow WoRMS Editorial Board (2016).

Benthic foraminifera

Astrononion gallowayi Loeblich & Tappan, 1953

Buccella frigida (Cushman, 1921)

Cassidulina neoteretis Seidenkrantz, 1995

Cassidulina reniforme Nørvangi, 1945

Lobatula lobatula (Walker & Jacob, 1798)

Dentalina sp.

Discorbis sp.

Elphidium excavatum f. *clavata* Cushman, 1930

cf. *Elphidium excavatum* f. *clavata*

Elphidium spp.

cf. *Elphidium albiumbilicatum*

Elphidium asklundi Brotzen, 1943

Elphidium frigidum

cf. *Elphidium margaritaceum*

Cribroelphidium subarcticum (Cushman, 1944)

Epistominella arctica Green, 1959

Epistominella exigua (Brady, 1884)

Fissurina spp.

Globocassidulina crassa (d'Orbigny, 1839)

Haynesina germanica (Ehrenberg, 1840)

Haynesina orbiculare (Brady, 1881)

Laryngosigma hyalascidia Loeblich & Tappan, 1953

Islandiella helenae Feyling-Hanssen & Buzas, 1976

Islandiella norcrossi (Cushman, 1933)

Melonis barleeanus (Williamson, 1858)

Oolina spp.

Oridorsalis tenerus (Brady, 1884)

Parafissurina spp.

cf. *Protelphidium anglicum*

Pullenia bulloides (d'Orbigny, 1846)

Quinqueloculina spp.

Robertinoides charlottensis (Cushman, 1925)

Stetsonia horvathi Green, 1959

Fursenkoina acuta (d'Orbigny, 1846)

Fursenkoina complanata (Egger, 1893)

Stainforthia feylingi Knudsen & Seidenkrantz, 1994

cf. *Stainforthia feylingi*

Stainforthia loeblichi Feyling-Hanssen, 1954

Spiroculina sp.

Rhizammina sp.

Textularia sp.

Trochammina sp.

Thurammina sp.

Triloculina trihedra Loeblich & Tappan, 1953

Triloculina tricarinata d'Orbigny, 1826

Valvulineria arctica Green, 1959

Planktic foraminifera

Neogloboquadrina pachyderma (Ehrenberg, 1861)

Ostracods

Cytheropteron paralatissimum Swain, 1963

Cytheropteron pseudomontrosiense Whatley & Masson, 1979

Clithrocytheridea sorbyana (Jones, 1857) Schweyer, 1949

Molluscs

Nucula spp.

Yoldiella spp