Spring 1-1-2010

Holocene Reconstruction of the West Greenland Current and the Greenland Ice Sheet Margin Near Disko Bay Using Foraminiferal Assemblages

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HOLOCENE RECONSTRUCTION OF THE WEST GREENLAND CURRENT AND THE GREENLAND ICE SHEET MARGIN NEAR DISKO BAY USING FORAMINIFERAL ASSEMBLAGES

by

MARIAH E. WALTON

B.S. Chemical Engineering, Rose-Hulman Institute of Technology, 2008

A thesis submitted to the Faculty of the Graduate School of the University of Colorado in partial fulfillment of the requirement for the degree of Master of Science Department of Atmospheric and Oceanic Sciences 2010
This thesis entitled:
Holocene Reconstruction of the West Greenland Current and the Greenland Ice Sheet Margin near Disko Bay
using Foraminiferal Assemblages
written by Mariah E. Walton
has been approved for the Department of Atmospheric and Oceanic Sciences

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Date________________

The final copy of this thesis has been examined by the signatories, and we find that both the content and the form meet acceptable presentation standards of scholarly work in the above mentioned discipline.
Abstract

Walton, Mariah E. (M.Sc., Atmospheric and Oceanic Sciences)
Holocene Reconstruction of the West Greenland Current and the Greenland Ice Sheet Margin near Disko Bay using Foraminiferal Assemblages
Thesis directed by Dr. Anne E. Jennings

The Greenland Ice Sheet (GIS) is currently thinning and retreating. One of the focal points for present research on the GIS retreat is Jakobshavn Isbrae, the largest ice stream on Greenland’s West coast, which is retreating today at least in part from ocean subsurface warming (Holland et al., 2008). Jakobshavn is located in Disko Bay, which receives warm Atlantic water from the West Greenland Current (WGC). In this thesis we present multi-proxy data from four marine sediment cores that help constrain the timing of ice retreat from the continental shelf and identify the marine conditions that accompanied retreat. Cores were taken in a transect from the outer shelf to near the mouth of the isfjord and basal dates give minimum ages of retreat from the shelf, bay sill, and near shore of 11.4, 11.1, and 9.4 cal kyr BP.

Foraminiferal assemblages and IRD counts from these cores indicate that after rapid retreat of the GIS the region experience cold conditions. Atlantic waters, via the WGC were first felt in the bay sometime between 9 and 8.5 cal. kyr BP. A possible strengthening of the WGC is seen on the shelf at approximately 7.0 cal. kyr BP. A dramatic rise in calcareous fauna from 6.2 – 3.5 cal. kyr BP in all cores indicates a significant shift in the strength of the WGC, with warm, Atlantic associated fauna gradually increasing and peaking at 4.5 cal. kyr BP. This coincides with the believed retreat of Jakobshavn Isbrae behind its present margin (Weidick et al., 1990), suggesting that subsurface warming may have played a part in the ice stream’s retreat in the past. The region returns to colder conditions by 3.0 cal. kyr BP, indicating a late onset of neoglacial cooling.
Acknowledgements

The work presented here was part of a large collaborative effort, and as such there are many individuals and parties that deserve recognition for its completion. If I have omitted anyone unknowingly I apologize now, though I like to think that I have at the very least acknowledged you in the references included throughout this thesis.

First and foremost I must recognize the sources of funding for my research. The vast majority of this work was funded by NSF grant OPP-0713755. Several radiocarbon dates and the pending $^{18}$O samples were paid by a Beverly Sears grant which was based on donations by Scott and Carol Winston. Many of the radiocarbon dates referred to herein were obtained by Matthias Moros (Baltic Research Institute), Jerry Lloyd (Durham University), Colm Ó Cofaigh (University), and Aoibheann Kilfeather (Durham University), and were paid for by NERC and other agencies.

The marine sediment cores analyzed in this these came from many sources. The 070HU cores were obtained in collaboration with Anne de Vernal at the University of Quebec. Core 343300 was obtained in partnership with Matthias Moros and Jerry Lloyd. Cores VC05 and VC20 were obtained in collaboration with Colm O Cofaigh and Aoibheann Kilfeather at Durham University (UK), and the British Antarctic Survey and Geologic Survey. I extend special thanks to Colm and Aoibheann for allowing me to join the JR175 cruise, as it truly was a wonderful opportunity—as a student, researcher, artist, and outdoor enthusiast! Thanks also go to Joseph Ortiz (Kent University) for performing the color reflectance analysis for the 070HU cores.

I would like to give particular thanks to several colleagues at the Institute for Arctic and Alpine Research. Thanks got to John Andrews for performing the mineralogical analysis of all samples and for his mentorship in OxCal, MSVP, and general paleo-Greenland knowledge. Thanks also go to Wendy Roth for showing me the ins and outs of the sediment lab, and helping with lab equipment and logistics. Thank you Ursula Quillman for being a supportive office mate and a spring-board for ideas. Christina Sheldon—you were a phenomenal microscope lab-mate, and your work with the surface samples within Baffin Bay was extremely useful. Lastly but most significantly I would like to thank Anne Jennings for being my adviser and mentor in this project. You were my guide in a new field and your leadership allowed me to accomplish a great deal in a short amount of time.
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I. Setting the Scene

Purpose

With the retreat of many Greenland glaciers today, it is increasingly important to improve our understanding of the local and regional interplay between ice sheets and ocean currents to anticipate changes we may expect in the future. This project uses foraminiferal assemblages from four marine sediment cores to examine changes in the West Greenland Current and ocean conditions from 12.5 to 3 thousand years ago in a large embayment in West Greenland. Specific goals are to identify periods of increased West Greenland Current inflow, to identify possible ranges for the Holocene Thermal Maximum and Neoglacial, and to identify the rough position of the ice margin through time based on faunal changes.

Disko Bay and the West Greenland Current

Disko Bay is located on Greenland’s West coast and is known primarily for Jakobshavn Isbrae, West Greenland’s largest ice stream, which funnels 50 km$^3$ of glacial ice into the bay each year. Jakobshavn is of particular interest today because of its accelerating calving rate. In the 1980’s, the ice stream was losing only roughly 14 km$^3$ of ice per year (Andersen, 1981), and the Isbrae was even thickening in the early 1990’s, according to yearly NASA airborne surveys (Holland et al., 2008). Since 1997, however, Jakobshavn has shown rapid thinning and retreat, with the calving rate doubling by 2000, and its 15 km floating ice tongue disintegrating shortly thereafter. This thinning and retreat was at first attributed to rising air temperatures and increased surface melting, which migrates to the ice base and lubricates the ice-bedrock interface (Joughin, 2006). However a more recent study has traced the initial thinning and subsequent retreat to a
sudden increase in subsurface ocean temperatures (Holland et al., 2008). Hydrographic surveys taken for fisheries and spanning Greenland’s West Coast show a distinct increase in the average temperature of 150 to 600 m waters beginning in 1997 and persisting today (Holland et al., 2008).

The source of this warm water is the West Greenland Current (WGC), which forms at the southern tip of Greenland and extends along its West coast (Fig I.1). The WGC is formed by the merging of the East Greenland Current, which brings cold, low-saline water from East Greenland and the Arctic Ocean, and the Irminger Current, which is the northernmost extension of the Gulf Stream and is comprised of warm, saline Atlantic waters. The WGC is initially highly stratified, with East Greenland Current (EGC) water overlaying the deeper Irminger Current (IC) water. By the time these waters reach Disko Bay, however, they are no longer distinct water masses, and a warm water core is the most pronounced signal of the current. The WGC substantially weakens as it moves north, which has lead to different classifications of this warm core.

Ribergaard (2005) defines Pure Irminger Water as having a temperature of ~ 4.5°C and salinity >34.95, found from Cape Farewell to Cape Desolation, and Modified Irminger Water (T>3.5°C, 34.95>S>34.88) found from Cape Desolation to Fylla Bank. North of Fylla Bank, WGC is less dominated by Irminger water, and hence the warm core has no specific designation. A series of hydrographic transects taken in 2005 (Ribergaard, 2005) clearly show this progressive weakening of the current signal as it moves North (Fig I.2 & I.3). The strength of the WGC on the continental shelf is determined by the overall strength of the IC and EGC components as well as the strength of eddies in the Labrador Sea, which are advected from the Norwegian Seas and enhance water mass mixing in the Labrador Basin (Jakobson et al., 2003).

The WGC splits prior to and at the Davis Strait—which is the sill between Baffin Bay and the Labrador Sea—with one arm joining the Labrador Current and the other continuing North along the West Greenland coast (Cuny et al., 2002). Much of the Irminger Water rounding the tip of Greenland follows bottom topography as it enters the Labrador Sea, which leads to a gradual decrease in the transport of Irminger Water along West Greenland as it moves north. This decrease was quantified by Myers et al. (2007), who estimated multi-decadal mean Irminger Water transports (pure and intermediate combined) of 3.8-4.9 Sv (1 Sv = 10^6 m^3/s) at Cape Farewell, decreasing to 1.0-0.8 Sv by Paaimut (see Fig I.2&3 for location). Myers et al. used hydrographic transects that did not extend through the Davis Strait, but it is clear that the bulk of Irminger water is incorporated into the Labrador sea prior to it reaching Davis Strait. In a follow up paper Myers et al. (2009) quantified the EGC component by looking at freshwater transports at the same transect locations used by Ribergaard (2005), up to and including Sisimiut. Mean freshwater transports increased from 45.8±11.8 mSv
Figure I.2. Salinity profiles of hydrographic transects from Greenland’s west coast, taken from Ribergaard (2005). Profiles were stretched to have the same length scale.

Figure I.3. Potential temperature profiles of hydrographic transects from Greenland’s west coast, taken from Ribergaard (2005). Profiles were stretched to have the same length scale.
at Cape Farewell to 54.4±22.4 mSv at Cape Desolation due to influx from local glaciers, then decreased to 44.7±12.3 mSv at Paamiut. Unlike Irminger Water, which is gradually deflected with rising bottom topography, freshwater transport dramatically drops only at Davis Strait itself (from 29.5±11.0 mSv at Fylla Bank to 2.9±5.4 mSv at Maniitsoq, and 0.03±5.4 mSv at Sisimiut), supporting its presence as primarily a surface current.

The Ribergaard transect on the Disko shelf (Aasiaat) indicates that the reduced WGC that reaches Disko Bay is still vertically stratified, with warmer waters residing in Disko’s bathymetric trough (Outer Egedesminde Dyb) than on the high shelf. This depth dependence is significant when looking for WGC penetration into Disko Bay, as a high sill (250m) separates the bay from the shelf. One deep channel (750m) breaks up this sill, and it is primarily through this channel, called (Inner) Egedesminde Dyb (Fig I.4), that WGC enters the bay (Andersen, 1981). Detailed transects taken in 1980 (Andersen, 1981) indicate that WGC is felt most strongly in the southern half of the bay, and only slowly moves N, eventually re-circulating to exit the bay both through the Vaigat and just S of Godhavn (also called Qeqertarsuaq). The more recent Ribergaard transects (2005) are taken at too few points within the bay to speculate on circulation but do confirm warmer water moving through the break in the sill, as well as warm deep water (3°C) close to the mouth of Jakobshavn’s inner channel.

IC flow peaks in autumn and winter, and the EGC peaks in spring and early summer (Ribergaard, 2005). Northward transport at Davis Strait peaks in November and has a minimum in late spring (Myers et al., 2009), following IC flow. As shown above, EGC transport when it rounds the tip of Greenland is two orders of
magnitude smaller than IC transport. In Disko Bay the WGC is felt as a somewhat seasonal signal, with the strongest influence present from July through September. This is likely due to local, seasonal changes in ocean stratification caused by sea ice formation in the winter (Andersen, 1981). Over longer times scales, WGC strength has been most strongly correlated with the North Atlantic Oscillation.

**North Atlantic Oscillation and its relation to the WGC**

![Sea Level Pressure](image)

The North Atlantic Oscillation (NAO) refers to the oscillation of air mass between the Arctic and the subtropical Atlantic (Hurrell et al., 2003). The name was given by Sir Gilbert Walker, who was one of the first to give it concerted study. As described in his seminal work (1924): “it is generally recognized that an accentuated pressure difference between the Azores and Iceland in autumn and winter is associated with a strong circulation of winds in the Atlantic, a strong Gulf Stream, high temperatures in winter and spring in Scandinavia and the east coast of the United States, and with lower temperatures in the east coast of Canada and the west of Greenland”.

The Northern Hemisphere Sea Level Pressure (SLP) mean state is shown in Figure I.5. The NAO swings between two extremes of this initial state, particularly during the winter, when pressure gradients are more pronounced. During an extreme positive NAO, the low pressure system centered near Iceland is intensified, with lower than mean SLPs, and the high pressure over the subtropics is intensified as well, creating a stronger than mean pressure gradient between...
the Arctic and mid latitudes. During a negative NAO, these pressure centers are weaker than the mean state, creating a weaker pressure gradient. The NAO is typically characterized by looking at the pressure difference between the two pressure centers. There are two indices commonly used today—the first (Rogers, 1984) uses the SLP difference between Akureyri, Iceland and Ponta Delgada, Azores; the second (Hurrell, 1995) uses Stykkisholmur, Iceland and Lisbon, Portugal, its creation motivated by a longer temporal record.

The effect of these pressure differences on regional climate and ocean circulation is profound, and surprisingly well summarized by Walker, above (1924). We focus here on the ocean, and while the full mechanism for NAO-ocean response is not yet known, in general the ocean is most strongly affected by NAO driven changes in wind stress and convection. As air moves counter clockwise around a Low and clockwise around a High in the Northern Hemisphere, a +NAO leads to stronger west winds by intensifying the pressure gradient in the North Atlantic (Hurrell, 1995). It also promotes an enhanced storm track in the Nordic Seas, which strengthens and alters the location of ocean convection in that region. A –NAO by contrast has weaker winds and a strong storm track off the northeast U.S. coast (Hurrell and Dickson, 2003). A +NAO is also associated with increased convection in the Labrador Sea and a colder North Atlantic due to enhanced NW winds. A –NAO is associated with the opposite (Hurrell and Dickson, 2003). These responses are not guaranteed, however, and depend on the exact position of the low pressure system centered over Iceland. For example 1993-1995 and 1999-2000 were two periods of strong +NAO, but different Labrador Sea convection. In 1993 convection was strong, but in 1999 the Iceland Low shifted east, leading to slightly weaker NW winds, even some S winds, and correspondingly weaker Labrador Sea convection (Hurrell and Dickson, 2003). Using two points as the basis for the NAO index unfortunately does not capture the exact position of each pressure system, and as a result can sometimes be misleading when characterizing Labrador Sea convection and WGC strength.

Curry and McCartney (2001) created the equivalent of an ocean NAO index by looking at the Potential Energy Anomaly between the subtropical gyre and the subpolar gyre, as calculated from station data at Bermuda and in the Labrador Basin. They found, at 95% confidence, that this ocean index follows the Hurrell NAO index with a 1-2 yr lag (correlation of r=0.58). This lag is likely due to the natural delay in ocean response to surface forcing and to the ‘memory’ of the ocean. Here the subpolar gyre feeds the Labrador basin, where convection sends any signal felt near Labrador to the mid depths, where it moves into the subtropical gyre—all of which takes time. Curry and McCartney used a weighting function on the NAO index to simulate an idealized ocean response (where each year back in time gets progressively less weight at any
one time step). It is through this weighting that they obtained their estimate of a 1 to 2 year lag. Joyce et al. (2000) similarly correlated the northward transport of subtropical mode water and the position of the Gulf Stream to the NAO index by looking at potential vorticity at Bermuda. They found that a positive NAO is associated with a southeast shift in the mean location of the gulf stream and increased transport of subtropical mode water.

Interestingly, when looking at the Irminger Current and WGC, Myers et al. (2007) found an anti-correlation between ocean transport and the NAO index. A –NAO was associated with more transport of Irminger mode and intermediate water, with a one year lag (correlation of $r=0.42-0.52$ depending on data set). This observation (Fig I.6), based on data from 1950-2005, is supported by Ribergaard (2005), who saw in his transects that the recent trend toward a negative NAO corresponded to enhanced IC flow. Sea ice extents in Baffin Bay and the Labrador Sea also correlate with the NAO index, a –NAO leading to decreased total ice area—in the same year and with a one year lag ($r=0.57$ and $r=0.43$ respectively) (Heide-Jørgensen, 2007). The ice margin is influenced by the strength and position of the WGC, so this sea ice relationship further supports the association of a strong WGC/IC arm with a –NAO.

Figure I.6. NAO index taken from Hurrell et al (2003) and Irminger water volume transport at Cape Desolation estimated from hydro- graphic transect data, taken from Myers et al (2007).
Significant climactic periods of the Holocene

Up to this point we have discussed only the modern situation in Disko Bay and the North Atlantic. The paleoceanographic work presented within this thesis spans the last 13,000 years, however, which makes a brief review of significant climactic changes during that time pertinent. Deglaciation from the Last Glacial Maximum (LGM) was neither swift nor steady. The Bolling-Allerod (BA) was the initial period of significant warming instigating deglaciation, but it was quickly followed by the Younger Dryas (YD), during which temperatures returned to glacial levels. The resumption of rising temperatures marked the beginning of the Holocene. Climactic ‘events’ show much variation in timing depending on location, due to the complex interplay between forcings and earth processes. This applies to the YD and BA, but a good estimate for their timing in West Greenland can be found using the GISP2 ice core record (Fig I.7).

The cause for the rapid switch between cold and warm climates is tied to ocean circulation, but the exact trigger and mechanism remains unclear. We know comparatively little of the ocean circulation during glacial periods, as the only method we have for reconstructing circulation is through water mass tracers. These consist primarily of $^{14}$C, $^{13}$C, $^{18}$O, and trace nutrients (Cd, Mg, Ca), all of which can be measured from foraminifera, which directly incorporate them into their calcite shells from sea water as they grow (Gupta 1999). Phytoplankton productivity at the ocean surface creates a steady state of enriched $^{13}$C (relative to $^{12}$C), and low nutrient concentrations at the surface, as phytoplankton preferentially uptake $^{12}$C from dissolved inorganic
carbon in sea water (Bradley, 1999). Measuring $^{13}$C and nutrient concentrations at different depths worldwide reveals regions of depletion and enrichment that can indicate which basins are sources and sinks for these tracers.

From these tracers we have been able to infer that North Atlantic Deepwater (NADW) was shoaled in the north Atlantic at the Last Glacial Maximum (LGM) sinking to a depth of 1500-2500m, as opposed to the present 2500-3500m (Marchitto and Broecker 2006). It is also likely that the Southern Ocean played a much larger role in deep water formation than it does today (Rosenthal et al. 1997). Problems arise, however, when some of these tracers conflict. They are often assumed to be conservative, but $^{13}$C/$^{12}$C ratios are greatly influenced by air-sea exchange, and fractionation into the calcite shells they are sampled from. Furthermore, both $^{13}$C and nutrients are affected by changes in local productivity (Broecker and Maier-Reimer 1992, Lynch-Stieglitz et al. 1995). Resolving the problems with ocean tracers has lead in some part to an increasing number of modeling studies, which generally focus on identifying stable modes of the ocean, and the trigger that initiates glacial-interglacial transformations.

There is no strong consensus on which of these models is more probable, but one which is increasingly supported by tracer reconstructions was put forth by Keeling and Stephens (2001). They hypothesized that the ocean has two modes—a stable interglacial mode, and an unstable glacial mode. The glacial mode is fundamentally different from the modern circulation in that the salinity gradient in the ocean is reversed, with the Southern Ocean more saline than the North Atlantic. This reversal is achieved by cooling the deep ocean to near freezing; the Southern Ocean would become saline due to brine rejection when sea ice forms. Ordinarily this would self-regulate, as the sinking induced from brine rejection would enhance mixing, which inhibits sea ice formation. If the entire water column were to cool to near freezing, however, this is not the case, and we would be left with a salty, near freezing deep ocean. A study by Adkins, McIntyre and Schrag (2002) using $^{18}$O to reconstruct deep ocean salinity and temperature supports the possible existence of this cold, saline body. The Keeling and Stephens modeled glacial state is incredibly sensitive to changes in the freshwater budget, because a change in the formation of Antarctic Intermediate Water (AAIW) would impact NADW formation. If AAIW was too dense (due to insufficient influx of freshwater at the Antarctic Polar Front), NADW would be unable to sink beneath it. Similarly, a strong freshwater influx into the North Atlantic would result in shoaled or halted NADW formation temporarily, but increased convection after the event, as saline waters would build up while deep water formation was halted. This bi-stable mode could explain the Dansgaard-Oeschger oscillations of the last glacial period and the bi-polar seesaw.
The time frame for this study includes part of the Younger Dryas, which is why it is important to be aware of the potentially significant changes the ocean underwent as it switched from glacial to interglacial modes. The transition out of the Younger Dryas at the very least involved a change in North Atlantic convection and deep water circulation, which likely induced changes in the WGC.

The Bolling Allerod and Younger Dryas were periods of dramatic climate change, and the Holocene looks quite stable by contrast. Yet when placed in context of a smaller time frame, there has been considerable change over the last 11.5 calibrated thousand years before present (cal. kyr BP). The broadest of these changes was a gradual warming and then a gradual cooling of the Northern Hemisphere, leading to higher than present temperatures in the Arctic in the early Holocene and steadily decreasing temperatures until the present (ignoring recent global warming). The period of increased temperatures is termed the Holocene Thermal Maximum (HTM), while the period of reduced temperatures is referred to as the Neoglacial Period (NG). At their most basic level these phases were driven by changes in solar insolation caused by the precession of the Earth with respect to the Sun (Kaufman et al., 2004). The Arctic had relatively high summer insolation during the early Holocene 11.5-8.5 cal. kyr BP (Fig I.7), and decreasing summer insolation through the mid and late Holocene. This primary forcing is confounded, however, by physical processes—particularly those influenced by the Laurentide Ice Sheet (LIS), which covered most of Northern Canada during the last glacial period.

The LIS affected global and local energy balances through changes in surface cover, as rock and vegetation supplanted ice during deglaciation, dramatically changing albedo. It also affected the air-sea heat budget through meltwater flux into the ocean and cold air advection as air passed over the ice sheet (Kaufman et al., 2004). The LIS retreated roughly linearly from 16.8 to 8 cal kyr BP with two significant exceptions. During the Younger Dryas retreat slowed, and was interspersed by multiple small readvances in certain locations. The second exception was from roughly 8.4 to 8.2 cal kyr BP, when retreat sped up as ice was cleared from Hudson Bay (Dyke, 2004). This second excursion is linked to the rapid drainage of the glacial lake Ojibway, which covered parts of Northern Ontario and Quebec, and the remnant of Lake Agassiz, which occupied the region west of Hudson Bay. By 8 cal kyr BP less than 10% of the LGM extent of the LIS remained. By 6.8 cal kyr BP, the Foxe dome collapsed, leaving only remnant ice caps on the Melville Peninsula, Southampton Island, and most significantly on Baffin Island, where ice has persists in the Penny Ice Cap even in the present day (Dyke, 2004).
A comprehensive study of all robust proxy data available in the Western Arctic (0-180°) (Kaufman et al., 2004) found that sites near or formerly under the LIS presented a delayed and varied HTM (7-3 cal. kyr BP), while temperature reconstructions in NE Russia and in the high N Arctic tended to mirror solar insolation. The cooling influence of the LIS appears to have suppressed rising temperatures in its vicinity until its disintegration. As the timing of the HTM is defined by the high temperatures it manifests rather than the forcing that drives it, we would expect sites within the sphere of influence of the LIS to experience a ‘delayed’ HTM because temperatures could not peak until the ice was gone. The onset of the HTM in proxy records hence seems to occur in two waves: one at the beginning of the Holocene, aligning with peak solar insolation, and one following the main collapse of Laurentide ice sheet at 8.4-8.2 cal kyr BP (Fig I.8).

Determining exactly how and to what extend the LIS influenced any one site, particularly distal sites, is somewhat speculative. Other factors, such as atmospheric circulation, may have had an equally significant influence on regional temperatures. For example, tree line in the Russian Arctic heavily correlates with the Southern extent of the Polar front, and proxy records of tree pollen in Beringia indicate that the polar front was further South in the early Holocene (Kaufman et al., 2004).

Rising in prominence in North Atlantic paleoclimate studies is the hypothesis that atmospheric circulation may have shifted between positive or negative NAO-like modes through time, which could also explain some of the variation in HTM onset (Kaufman et al., 2004; Frechette and de Vernal, 2009). The NAO is highly variable today, and does not appear to prefer one mode more than the other on anything more than a decadal

Fig I.8. Frequency-longitude diagram of HTM onset, taken from Kaufman et al. (2004).
time scale. To look at a longer time period, the NAO index has been correlated to ice core records of selectively chosen cores with mixed results. Appenzeller et al. (1998) made a proxy NAO index using normalized accumulation rates in the NASA-U ice core, which they correlated to the 130 yr instrument record such that over a third of the variance could be explained by a linear relationship between the proxy and actual record. The NASA-U is a short core (extending to ~350 years BP) taken in W Greenland, just N of the Umanak and Disko fjord system. A similar correlation was achieved by Vinther et al. (2006) by using principle component analysis with the stable isotopes of five ice cores taken from the Southern half of Greenland. However both of these studies smoothed out high frequency noise and linear trends so that variance was the only basis of correlation. A shift in the dominant mode would therefore be difficult to find were these studies extended to a millennial time scale, as a permanent switch could manifest as a linear transition. Furthermore, ice cores must be in specific locations to experience significant changes in precipitation (and therefore accumulation) due to changes in the NAO. West Greenland and Baffin Island generally experience wetter conditions during a negative NAO and dryer during a positive NAO, whereas central Greenland is less affected (Hurrell, 1995). Precipitation is also largely orographically driven, making it highly dependent on local topography.

Of the Arctic ice cores that extend through the full Holocene, the Penny core—taken at the Penny ice cap on Baffin Island—is probably the most appropriate for looking at the NAO, but accumulation data is not available at a high enough resolution to compare to present day NAO records. A statistical comparison between the Penny core and NAO is hence not presently available. Qualitatively, however, possible climate shifts can still be seen by looking at the Penny oxygen isotope and dust/particle records (Fig I.9).

The $\delta^{18}O$ record, representing temperature, indicates a fairly early and long HTM from 9.5-5 cal. kyr BP, in line with changes in August insolation. Beginning at 6 cal. kyr BP, a gradual increase in dust concentration and a decrease in temperatures could indicate a shift to a different NAO state. A shift to a more +NAO would strengthen NW winds, which carry cold, polar air into the region, and could increase dust transport from the plateau on Western Baffin Island to the Penny ice cap, but dust can be transported from much further locations, so the change seen in the Penny core does not preclude the possibility of a shift to a –NAO dominant mode. The increase in dust at 6 cal. kyr BP could also, be attributed to the final disappearance of the LIS, which would have freshly exposed large areas of ground susceptible to erosion. Further, the decline in temperature could merely by the onset of Neoglacial cooling. These explanations would neither support nor disprove a shift in atmospheric circulation, rather indicating that other forcings proved more influential at this site.
Looking for a NAO signal is useful in as much as it may allow us to attribute observed changes in temperatures and other climate parameters to a single mechanism. It can become counterproductive, however, when past observations are themselves insufficient to constrain timing and identify co-incident trends. This conversation will be revisited when discussing the paleoceanographic interpretation of our results.

Returning to the long term climate record, the Holocene Thermal Maximum was followed by Neoglacial cooling, forced by decreasing summer insolation. The timing of the NG is similarly staggered to the HTM, as sites with a late HTM tended to have a similarly late NG (Kaufman et al., 2004). In the Penny Ice cap $^{18}$O record, temperatures begin to decrease at approximately 6 cal. kyr BP, but temperatures do not significantly drop until closer to 4.5 cal. kyr BP. The NG is thought to have a very late onset in Disko Bay. Transported subfossils at the ice margin indicate that Jakobshavn Isbrae had retreated to behind its present position from 4 – 5 cal. kyr BP, suggesting warm temperatures at that time (Weidick et al., 1990). These ages do not support a readvance indicative of cooling until at least 3.2 cal. kyr BP, and probably closer to 2.7 cal. kyr BP. This discussion will also be revisited when discussing the paleoceanographic interpretation of our results.

Figure I.9. Penny $^{18}$O (Fisher et al., 2003) and dust record (Zdanowicz 1998), the GISP2 temperature reconstruction (Alley, 2004), and summer insolation, taken from Kaufman et al (2004). The Penny $^{18}$O record presented is a 25yr running mean. The dust record presented uses a variable running mean, due to increasing resolution upcore (0-1.0 cal. kyr bP = 31 pt running mean, 1.0-2.5 = 19pt, 2.5 – 6.0 = 11pt, beyond 6.0 cal. kyr BP the data is not smoothed).
The importance of Jakobshavn Isbrae today, and by extension Disko Bay and the West Greenland was presented in the first section, but our main focus is on how the ice stream and the WGC behaved through the Holocene. To justify such a topic, we must first, at a basic level, verify the importance of Jakobshavn Isbrae on a millennial time scale and the presence of ice in Disko Bay during more than the recent past. A bathymetric trough mouth fan is present on the Disko slope and is a strong indicator of ice extent. Many trough mouth fans have been found in the Arctic along formerly and currently glaciated margins, and in the Norwegian Seas they have been accepted as the depocenters of ice stream frontal sediments (Vorren and Laberg, 1997). The Disko fan, and the neighboring Umanak trough mouth fan (North of Disko) are the largest fans on Greenland’s West Coast. Single beam (TOPAS) profiles of the shelf edge in Disko reveal debris flows consistent with other trough mouth fans and with remobilized glacially deposited sediments in general (O’Cofaigh, 2009). The Disko trough mouth fan and accompanying debris flows therefore strongly support a GIS ice margin that extended to near the edge of the continental shelf at the LGM. High resolution swath bathymetry within the bay and on the shelf shows scoured bedrock features further indicative of ice movement (Weidick and Bennike, 2007). Many marine sediment cores within this span additionally contain fine, gray, glacial silt, and in some cases possible till (Kilfeather et al., 2010). Till is defined as ice deposited sediments (Foster, 1982) but usually takes the form of consolidated diamicton (which is poorly sorted sediment, meaning grains and rocks of different sizes found together). There is therefore substantial evidence for an extensive glacial ice presence in the Disko region in the past; we focus our research on a less well constrained subject—that is how fast and in what manner did the ice retreat, and what was the ocean doing at that time?

Foraminifera as Proxies

Foraminifera (forams) are single celled protists that build shells, called tests, out of calcium carbonate or particulates and grains. Forams are broadly classified by where in the water column they live, and what they construct their tests of. Planktic forams live in the mixed layer of the ocean, benthic forams live on the ocean floor. Calcareous forams have tests made of calcite (CaCO₃), agglutinated (or arenaceous) forams have tests made of detrital particles cemented together. This study focuses on benthic species, primarily because very few planktic specimens were found in the cores examined, but also because bottom dwelling fauna should reveal more about the WGC.
Foraminifera have existed since at least the Cambrian (>500 million yrs ago) and while they have undergone some evolutionary changes since then, they are have not changed over the period of our record. Their long geological record and their sensitivity to changes in water conditions make them a good proxy for past ocean and lake environments. Approximately 10,000 species are believed to exist today, and of those only about 50 are planktic (Gupta 1999). Benthic forams have a life span ranging from a few weeks to a year, and in some larger taxa up to five years (Lee and Anderson, 1991). Species living in highly seasonal environments such as the arctic often reproduce only annually, and as reproduction effectively terminates an individual, the lifespan of these species is approximately one year. The sensitivity of forams to changes in ecological parameters such as temperature, salinity, and nutrients is highly species dependent. Benthic forams can live on top of the sea floor (epifaunal) or burrow several centimeters beneath the sediment-water interface (infaunal), and some species are adapted to specific environments (like glacier margins).

Forams are used as proxies either by chemical/isotopic analysis of their tests, or by looking at how an assemblage—the relative distribution of different species—changes through time at a set location. We focus on assemblages in this thesis, as chemical analysis is only possible for calcareous specimens and our cores show poor preservation of calcareous tests through extensive time periods, particularly on the shelf.

Assemblage work is most meaningful when the specific environmental tolerances of all species present is known. This is often not possible due to insufficient observation of modern distributions, but extensive effort has been made in the Arctic, Baffin Bay, and specifically in Disko bay, to identify and characterize the dominant fauna present (Rytter, 2002; Schafer and Cole, 1988; Schroder-Adams et al., 1990; Sheldon, 2010; Lloyd, 2006; to name just a few). Our knowledge of foraminifera within Disko Bay has been greatly expanded by Jeremy Lloyd (Durham University, UK), who has championed the ‘wet’ method of counting foraminifera in West Greenland. Drying samples can cause agglutinated tests to shrivel and break apart, leading to undercounting of agglutinate fauna, which can be particularly problematic in regions with low abundance of calcareous fauna. With the wet method samples are never dried. Sediment is wet sieved and forams are counted in solution. Using the wet method, Lloyd and others have been able to use agglutinated faunal shifts to identify changes in climate and WGC strength within Disko Bay through the Holocene (Lloyd et al., 2005, 2007; Perner, 2010; McCarthy, 2010, Jennings et al, 2010).

The wet method is particularly necessary in this study due to the poor preservation of calcareous tests on the outer Disko shelf. Baffin Bay is known to have high dissolution rates, particularly at depth. Near Davis Strait, tests are subject to 100% dissolution below 900m, and intense dissolution below 600m, probably as a result of glacial activity.
result of high CO\textsubscript{2} content in the water and low temperatures (Asku, 1983). These conditions are influenced by changes in bottom waters and by processes at the sea surface. With bottom waters, cold water can hold more CO\textsubscript{2}, so relatively warm Atlantic sourced bottom waters can promote better preservation by reducing dissolve CO\textsubscript{2} concentrations, and by importing carbonate. Reduced sea ice conditions can increase dissolved CO\textsubscript{2} concentrations by exposing more surface water to atmospheric exchange, and by increasing planktonic productivity, which increases carbon flux to the sea floor, and subsequently oxidizes organic matter in sediments (de Vernal \textit{et al.}, 1992).

The Disko region is generally more ice free than the bulk of Baffin Bay due to the WGC. Yet any increase in surface CO\textsubscript{2} absorption this might promote is compensated by the warmer bottom waters the WGC introduces. Modern benthic assemblages within the bay are almost uniformly dominated by agglutinated fauna, with noticeably corroded calcareous tests (Lloyd, 2006). They are not, however, subject to the complete dissolution seen in Baffin Bay, and many sites have significant calcareous fauna. Another study, incorporating assemblages on the Disko shelf (Sheldon, 2010), showed similarly agglutinated dominated fauna throughout the Disko area, with extremely low calcareous abundances in two of the three shelf cores examined (<5\% of total assemblage). The exception to the poor preservation observed in modern assemblages in Disko is found in deglacial sequences with high sedimentation rates, where the rapid burial of tests appears to protect them to some extent from dissolution (Lloyd \textit{et al.}, 2007).

Preservation differences between the inner bay and shelf could be an artifact of depth—shelf cores are typically taken within Outer Egedesminde Dyb, at depths >400m—but they may also be a result of differences in surface productivity. Mean monthly climatologies from 1972-1994 reveal that inner Disko Bay typically has some persistent light sea ice cover even in the summer, while much of the shelf is ice free by June (Tang \textit{et al.}, 2004). Further, the upper water column within the bay is highly stratified, which Nielsen and Hansen (1999) have shown inhibits the mixing of nitrate into surface waters, and in turn limits phytoplankton productivity. Nielsen and Hansen associated phytoplankton blooms mostly with strong winds that promoted upward mixing of nitrate from the pycnocline. It is possible that the inner bay has slightly better preservation than the shelf due to the reduction in surface activity imposed by sea ice and this surface stratification. Temperature differences between bottom waters on the shelf and within the bay are small, though the shelf appears to have a slightly stronger WGC presence (Ribergaard, 2005).

The magnitude of dissolution in our cores can be roughly quantified by the number of remnant test linings present, as was done by de Vernal \textit{et al.} (1992). These test linings are incorporated into shells when they are
constructed; many species secrete calcite in layers as they grow, and the organic lining separates these layers (Gupta, 1999). The linings are more resistant to dissolution than calcite and can in some cases be identified as belonging to a specific species (most notably Elphidium excavatum f. clavata) even after most or all calcite is gone.

*Species Preferences*

The prominent species in our cores and their potential or previously identified environmental preferences are discussed below:

*Elphidium excavatum* forma clavata and *Cassidulina reniforme* have been found in numerous arctic fjord systems, and are common together in ice proximal environments (Hald et al., 1994; Jennings and Helgadottir, 1994, Nagy, 1965). *E. excavatum* is the most frequent benthic species in shallow glaciomarine sediments of high latitudes and has been shown to prefer waters of low or fluctuating salinity, turbid waters (like those found at glacier termini), and cold, seasonally ice covered water (Hald et al., 1994, Corliss, 1991). However in some cases, such as in Cumberland Sound (Jennings, 1992) and E. Greenland (Jennings and Helgadottir, 1994), *C. reniforme* can be much more abundant in extreme ice-proximal environments than *E. excavatum*. *C. reniforme* prefers reduced salinity and cold bottom temperatures (Hald and Korsun, 1997), but is not always found with *E. excavatum*. It is important to note then that while *E. excavatum* and *C. reniforme* can be indicative of an ice proximal setting, they are also found in ice distal environments (Polyak and Solheim, 1994), which provide similarly cold, low saline environments. Alone, each of these species can indicate cold environments dominated by polar water.

*Islandiella norcrossi* is an epifaunal to shallow infaunal species that thrives on algal blooms at the ice edge. It is consequently associated with polar front productivity and seasonal ice cover (Korsun and Polyak, 1989; Steinsund, 1994; Rytter et al., 2002), particularly when found with Nonionellina labradorical. *I. norcrossi* has also been found in relatively saline and stable waters, however, and in some locations can signify increased Atlantic inflow (Korsun and Hald, 1998).

*Islandiella helenae* is similar to *I. norcrossi* in that it is associated with seasonal ice cover and seasonal productivity pulses (Wollenburg et al., 2004). Like *I. norcrossi* it is an arctic species and is not uncommon in ice distal settings (Jennings, 1992). *I. helenae* presents a puzzle in one of our cores (VC20), however, because it is the first species to appear and is present in a diamicton band that could be till. Nor is this an isolated
occurance—small quantities of *I. helenae* are seen at the bottom of VC05, 343300, and 070, and significant abundances are found at the bottom of two other shelf cores that have not yet been fully analyzed (VC02, VC15). Its presence is anomalous because *I. helenae* is not typically found in ice proximal settings, which we expect must have existed to deposit diamicton. *I. helenae* has been found in ice proximal settings in a few isolated studies—in British Columbia (Patterson and Kumar, 2002), and Northern Norway (Olsen, 2002), always coupled with *C. reniforme* and *E. excavatum*. It may be that our core was actually in an ice distal environment, or it could be that the shelf had a non typical ice presence, such as an ice shelf, that provided a cold, ice frontal environment without the turbidity associated with a glacier margin.

*Nonionellina labradorica* is strongly associated with high organic flux (Steinsund, 1994; Korsun et al., 1995, Rytter et al., 2002; Jennings et al., 2004), and is as a result linked to the polar front, similar to *I. norcrossi*. The source of this organic flux in Disko is not straight forward, as discussed previously, but *N. labradorica* is considered an Atlantic water indicator within the bay because a strong Irminger influence must be present for us to get preserved specimens despite high organic flux.

*Stainforthia feylingi* and *Spiroplectammina biformis* are considered polar water species in past Disko work (Lloyd, 2006), but *S. biformis* has also been found in glaciomarine environments (Korsun and Hald, 2000, Jennings and Helgadottir, 1994). In a recent study of glacier mouths in Alaska (Ullrich et al., 2009) a variant of *S. biformis* (identified as T. earlandi in their paper, but identified as Spiroplectammina sp. by us from samples sent in personal communication) is found with *E. excavatum* at the ice margin, and the authors speculated that it thrived due to its ability to survive high sedimentation rates. This conclusion is supported by the assemblages presented here and by Jennings et al (2010). *S. biformis* and *S. feylingi* are the first species to appear in high abundances at the bottom of all cores, and are frequently found with the glaciomarine species *E. excavatum* and *C. reniforme*. In the bottom of our cores they most likely indicate an ice distal environment, which gives high sedimentation rates, but enough productivity to support the high abundances. In non-glaciomarine settings, *S. biformis* reflects cold, arctic waters (Schafer and Cole, 1986). In the upper sections of our cores we interpret *S. biformis* and *S. feylingi* as cold, polar water indicators.

*Adercotryma glomerata* and *Portotrochammina bipolaris* are often linked to high latitude, Atlantic sourced waters. Hald and Korsun (1997) found a strong correlation between their dominance and distance from tidewater glaciers in Svalbard, indicating a preference for transformed Atlantic water. A similar relationship was found with *Recurvoides turbinatus*. 
Eggerella advena is a weakly classified species (in as much as it does not seem to strongly distinguish between Atlantic and polar water). Jerry Lloyd’s work indicates that it prefers shallow, fine-grained sediments and low-salinity water.

Reophax arctica, or Cuneata arctica is associated with cold, polar waters, and is today found in regions of Disko Bay more strongly influenced by polar water (Lloyd, 2006).

Textularia torquata, Reophax catenata and Reophax catella are not well understood and they show no strong preference for a specific environment. They are present in varying concentrations in all our cores, at nearly all depths, and while they show some response to zone transitions, their responses are not consistent. Efforts to identify preferences in other records are confounded by variations in species labeling and identification, particularly for the Reophax species. One of the results of this thesis was the separation and identification of these uniserial Reophax species. Our classification of R. catenata and catella are based on detailed descriptions and sketches by Hans Hoglund (Hofsten and Horstadius, 1947), a pioneer of foraminiferal studies in the Norwegian Sea. In other studies very similar Reophax species have been classified as Reophax gracilis (Lloyd 2006), and Reophax scotti (Scott et al., 2009). Lloyd performed a principle component correspondence analysis on his Disko assemblages that indicated R. gracilis (possibly R. catenata, R. catella, and/or R. scotti), along with R. arctica and T. earlandi, prefer clay and silt-rich sediments. This preference is somewhat supported by the presence of R. catenata at the bottom of several of our cores, when those sites would have been receiving an influx of glacial silt. Here we label them as indifferent or indeterminate in their environmental preferences, because they do not appear to distinguish between Atlantic and Polar water influences.

The modern assemblage in Disko Bay varies by location and water mass characteristics (Lloyd, 2006, Sheldon, 2010). Nearly all sites are dominated by A. glomerata and are mostly agglutinated, regardless of position. Sites that show the strongest WGC influence based on CTD casts also contained (at lower abundances) Cribrostomoides crassimargo, T. earlandi, R. fusiformis, R. pilulifer, and in some locations S. biformis. Some N. labradorica and Cassidulina neoteretis is also present. Sites with mixed WGC and polar water had higher abundances of P. bipolaris and T. torquata, and lower abundances of T. earlandi and R. fusiformis. Sites more strongly influenced by polar water are co-dominated by A. glomerata and S. biformis. They also contain greater frequencies of R. arctica and C. crassimargo, and lower frequencies of N. labradorical.
Radiocarbon Dating

The stratigraphic chronologies of the cores presented in this thesis are based entirely on radiocarbon dates, making a review of the theory behind such dating and the errors associated with it helpful. Radiocarbon dating is based on the exchange of protons and electrons between nitrogen atoms and carbon atoms. Neutrons are produced by cosmic rays entering the upper atmosphere (~15km), and these neutrons collide with nitrogen atoms to create $^{14}$C and hydrogen:

\[
^{14}_N + ^{1}_n \rightarrow ^{14}_C + ^{1}_H
\]

$^{14}$C radioactively decays through time back into nitrogen, releasing Beta particles and neutrinos:

\[
^{14}_C \rightarrow ^{14}_N + \beta + \text{neutrino}
\]

The amount of decay is time dependent, and can hence be used to determine the age of a sample. To do so, however, requires foreknowledge of the background levels of $^{14}$C present in the atmosphere and biosphere through time. The rate of $^{14}$C production and decay is generally in a state of equilibrium (meaning the level of background radiocarbon is roughly constant) but even variations of 2% are significant, particularly when dating older samples. To resolve this source of error, dendrochronology and uranium dating of corals has been used to independently date radiocarbon concentrations, producing a curve of background $^{14}$C that extends 40,000 yrs BP (Bradley, 1999). Ages are determined by measuring either the rate of $\beta$-particle emission, or, more recently, by measuring the concentrations of all carbon isotopes ($^{12}$C, $^{13}$C, $^{14}$C). Our samples were analyzed for $^{13}$C depletion, which can be correlated to radiocarbon enrichment/depletion.

From a measurement point of view radiocarbon dating is very straightforward, but the accuracy of the measured age is greatly impacted by fractionation effects and the movement of $^{14}$C through the biosphere. Plants and organisms (such as foraminifera and mollusks) uptake carbon dioxide, but the ratio of heavy to light CO$_2$ ($^{14}$C to $^{13}$C to $^{12}$C) that is absorbed is species and temperature dependent. Furthermore, the CO$_2$ reservoir they draw from will not always be representative of the global CO$_2$ reservoir (and hence the global $^{14}$C reservoir). Fractionation at the ocean surface during the absorption of CO$_2$ causes an enrichment in $^{14}$C in the water relative to the atmosphere, but ocean circulation prevents isotopic equilibration as old waters replace “new” waters at the surface (Bradley, 1999). As a result, even surface waters have an apparent age that is older than actual. This problem is compounded in bottom waters, as it can take over a thousand years for surface waters to circulate into the deep ocean. The reservoir age in Disko Bay is assumed to be approximately 400 yrs (subtracted from $^{14}$C age), but this correction may be off by several hundred years, and may not have
been constant through time. Too little is known about the reservoir time near Disko to assign a more accurate age, however, and 400 yrs is used for consistency with past work in the region.

An additional source of error for marine dates can result from partial dissolution of the material to be dated. Calcium carbonate is primarily secreted as either aragonite or calcite by shell-building marine organisms. Aragonite is somewhat unstable, and has been known to dissolve and re-precipitate as calcite, sometimes incorporating “new” carbonate from the environment into the calcite as it does so. The introduction of younger carbon into a shell can greatly affect measured age, even in trace amounts, which is why poorly preserved shells, and certain species of mollusks and gastropods, should be avoided when selecting samples to date. In Disko Bay, we are often limited in the material we have available to date, but where possible seaweed and whole, paired shells are used.
II. Methods

Overview

In order to better elucidate the behavior of Jakobshavn and the WGC through the Holocene, a transect of four cores (VC20, HU070, 343300, VC05) from the Disko area is presented here. Core locations are shown in figure II.1, and lengths are listed in table II.1. VC20 and VC05 were obtained on the 2009 JR175 cruise, which took place aboard the RRS James Clark Ross and was funded by the Natural Environment Research Council (NERC). The chief scientist was Colm O’Cofaigh (Durham University, UK). Core 343300 was obtained on the 2007 MSM 05/03 cruise of the R/V Maria S. Merian and was funded by the German Research Foundation (Deutsche Forschungsgemeinschaft--DFG) and the German Federal Ministry of Education and Research (Bundesministerium für Bildung und Forschung—BMBF). The chief Scientist was Jan Harff (Baltic Sea Research Institute). The 070 cores were obtained on the 2008 CCGS Hudson cruise, funded by the Geological Survey of Canada. The chief scientist was Calvin Campbell (Geological Survey of Canada). The 070 cores were later correlated and combined into one composite record (see Inter-correlation of 070PC and 070TC). Foraminiferal analysis for core 343300 was done by Anne Jennings, Shannon Smith, and the author. Foraminiferal analysis for the other cores was done by the author alone.

Table II.1. Core types.

Figure II.1. Map of study area with core sites. Open circle sites are from previous work in the area that are mentioned or discussed later in this thesis (Lloyd et al., 2004; Lloyd et al., 2007; McCarthy, 2010). Closed circle sites are for the cores analyzed in this thesis.
lengths and locations.

<table>
<thead>
<tr>
<th>Core</th>
<th>type</th>
<th>Length</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Water Depth</th>
<th>Cruise</th>
</tr>
</thead>
<tbody>
<tr>
<td>HU070</td>
<td>PC piston</td>
<td>2.5</td>
<td>68.22788°N</td>
<td>57.61746°W</td>
<td>444</td>
<td>HU2008-029</td>
</tr>
<tr>
<td>HU070</td>
<td>TC trigger</td>
<td>2.05</td>
<td>68.22788°N</td>
<td>57.61746°W</td>
<td>444</td>
<td>HU2008-029</td>
</tr>
<tr>
<td>HU070</td>
<td>composite</td>
<td>4.30</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>VC20</td>
<td>vibracore</td>
<td>5.39</td>
<td>68°12.06’N</td>
<td>57°45.38’W</td>
<td>424</td>
<td>JR175</td>
</tr>
<tr>
<td>343300</td>
<td>gravity</td>
<td>11.4</td>
<td>68°28.311’N</td>
<td>54°00.118’W</td>
<td>518</td>
<td>MSM 05/03</td>
</tr>
<tr>
<td>VC05</td>
<td>vibracore</td>
<td>5.87</td>
<td>69°09.6’N</td>
<td>51°31.63’W</td>
<td>389</td>
<td>JR175</td>
</tr>
</tbody>
</table>

Chronology for these cores comes from several radiocarbon dates (Table II.2). Dates for HU070, VC20, and VC05 were obtained in collaboration with Colm Ó Cofaigh and Aoibheann Killfeather at Durham University (UK) and were funded by NERC, NSF grant OOP-0713755, and by a Beverly Sears Grant awarded to the author. Dates for 343300 were obtained in collaboration with Matthias Moros and the Baltic Sea Research Institute, as well as Jerry Lloyd at Durham University. All calibrations were carried out using OxCal 4.1, with the IntCal 09 calibration curve and a reservoir correction of 400 yrs. Dates are cited with a two-sigma range, but unless otherwise noted will be reported at their mean value in calibrated years before present (cal. yr BP), or calibrated thousand years before present (cal. kyr BP).

Cores were sampled at roughly every 10 cm, with the exception of 343300, which from 0-400cm was analyzed at roughly a 5cm resolution. For 343300 this corresponds to a sample every 100 yrs or less through most of the core, with a few samples spaced every 200yrs. The resolution of HU070 is roughly 150 yrs from 11.9 – 8.5 cal kyr BP (428-184cm), 700 yrs from 8.5 – 3.5 (175-104cm), and 250 yrs from 3.5 – near present (95-5cm). More samples have been requested to improve this resolution in the low sedimentation rate regions, but they have not arrived in time to be included in this analysis. VC20 and VC05 are not yet fully dated, though several dates are pending. In both cases possible dates will be suggested based on similar features in better dated cores. VC20 in particular closely mirrors HU070 in assemblage changes, though VC20 appears to end earlier and extend further than HU070 (12.5 – 2 cal kyr BP probable time span compared to 11.8 – 1 cal kyr BP in HU070). The modern assemblage for HU070 has additionally been included using the top most box core sample from the same site (Sheldon, 2010). Box core samples were stained with Rose Bengal dye, which is taken into the cytoplasm of living foraminifers. Alive (dyed) specimens were counted as present day (0 cal kyr BP), while dead (undyed) specimens were assumed to be younger than the trigger core top but older than modern. For the sake of analysis they were assigned an age of 0.2 cal kyr BP (the trigger core is assumed to end at 1.0 cal kyr BP).
Table II.2. Radiocarbon dates from all core locations

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth in Core (cm)</th>
<th>Lab Code</th>
<th>Material</th>
<th>C14 age</th>
<th>95% probability range</th>
<th>Probable age</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>lower</td>
<td>upper</td>
<td>median</td>
</tr>
<tr>
<td>070TC</td>
<td>25-26</td>
<td>seaweed</td>
<td>pending</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>070TC</td>
<td>82-83</td>
<td>AA84708</td>
<td>bivalve</td>
<td>3197 ± 43</td>
<td>2863  3147</td>
<td>3003</td>
<td>3005</td>
</tr>
<tr>
<td>070TC</td>
<td>97-98</td>
<td>AA84709</td>
<td>bivalve</td>
<td>3252 ± 37</td>
<td>2951  3205</td>
<td>3082</td>
<td>3078</td>
</tr>
<tr>
<td>070TC</td>
<td>134-135</td>
<td>SUERC25670</td>
<td>seaweed</td>
<td>5913 ± 39</td>
<td>6274  6400</td>
<td>6311</td>
<td>6320</td>
</tr>
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<td>10910 ± 60</td>
<td>12221 12601</td>
<td>12437</td>
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Calibrated ages obtained using OxCal 4.1, with the IntCal 09 calibration curve and reservoir correction of 400 yrs.

* 14-15 cm date in 070PC was not used, as the upper 25cm of the piston core was disturbed.
** 169-170 cm date was rejected due to large error and age inversion relative to 188-189 cm.
All samples were split into two portions, one for foraminiferal and one for mineralogical analysis. Mineralogical splits were weighed, freeze dried, and subsampled for x-ray diffraction (XRD) analysis to determine mineralogy; the remainder was kept as a bulk dry sample. Foraminiferal samples were weighed, sieved at 63µm, and stored in a carbonate rich solution made from deionized water, alcohol, and baking soda. These samples were kept wet at all times but equivalent dry weights were calculated for each based on moisture content in mineralogical splits. Foraminifera were counted in each full sample, or a sub-split of each sample (depending on their concentration). Counting aimed for 300 specimens per sample, but this was not always possible at lower depths in the cores. Counts were converted to percentages based on total counts of combined agglutinated and calcareous tests, and included test linings. Samples with less than 95 specimens are highlighted in pink in all stratigraphic diagrams, as percentages from these levels are not statistically sound. Percentage based assemblages were divided into faunal zones by stratigraphically constrained cluster analysis using MultiVariate Statistical Package (Kovach Computing Services).

Ice rafted detritus (IRD) was counted in two size fractions, >500µm and >2mm. The 500 µm fraction was sieved from foram samples and individual grains were counted. The 2mm fraction was counted in 2cm intervals from X-ray photographs of the archive half of each core. The internal diameter for the JR175 cores (VC20 and VC05) is ~7cm, and that for 070 is ~10cm. Any differences in the magnitude of x-ray counted IRD between VC20 and VC05, which used the same corer, should therefore be actual, and not an artifact of differences in sediment volume between cores, but the same cannot be said for comparisons with 070.

IRD records were supplemented by counts of tephra shards within each foram sample. Tephra is a broad term for material that is formed or fragmented from volcanic processes and made airborne during volcanic eruptions (Bradley, 1999). The tephra we see in our cores is clear volcanic glass, broken into fragments of former bubbles. Tephra is atmospherically deposited on the GIS and ocean surface after major eruptions, particularly from Iceland. Disko Bay is too distal to a major volcanic source to receive much direct deposition of tephra, rather it appears that the tephra in our cores is melting out of the ice sheet (Jennings et al., 2006). This hypothesis is supported by previous work in East Greenland (Jennings et al., 2002), where tephra from Ash Zone 2 (55 radiocarbon kyr BP) was identified from electron microprobe analysis in several independently dated marine cores. This tephra correlated with 18O spikes indicative of meltwater events and could not have been deposited in the ocean before 14 cal kyr BP according to core age models. Our working hypothesis is that tephra is deposited directly on to the GIS from Icelandic (or other) eruptions, where it is stored and transported in the ice, and melts out as a form of IRD. Tephra in these cores coincides very well with other
IRD signals, and looks to represent fine grain IRD (63µm - 125µm). Geochemical analysis is in progress on tephra samples from our Greenland cores, as part of the Volcanoes in the Arctic SysTem (VAST) project. Note that tephra has not yet been counted for the upper half of 343300.

**Stratigraphy of core sites**

Core sites are generally selected based on bathymetric profiles. On JR175 this was done using a Kongsberg-Simrad EM120 multibeam swath bathymetry system and a TOPAS sub-bottom profiler (O Cofaigh 2009). Swath bathymetry generates a 3D image of the ocean floor, with a high enough resolution to distinguish bedrock features. TOPAS uses a single, high-powered ping to penetrate through sediment on the ocean floor. Significant sediment interfaces reflect the signal back, creating a vertical profile of sediment layers in one location. Debris flows and turbidites (sediment driven gravity flows) typically show as disturbed layers, or no layering at all, depending on the degree of reworking. An undisturbed chronological sequence will show up as unbroken layers, with no significant disturbance. Bedrock is clear as a much darker band, as is diamicton (potential till). An undisturbed core ending in till should contain an unbroken record of deglaciation, and indeed finding coring sites with till and shallow overlaying sediments was one of the primary goals of the JR175 cruise. The cores presented here appear to have relatively clean stratigraphic sequences, and were obtained with that intent. The bottom of VC20 may include till or a debris flow, as a significant section consists of diamicton—rocks and clasts of different sizes—and is tightly packed and consolidated.

**Coring methods**

Cores are taken using different methods, potentially introducing distortion to the records they represent. These methods are described here to ensure that each sequence presented is viewed within the context of its original sediment.

Cores VC20 and VC05 were acquired using a Vibracorer. A Vibracorer uses vibrational motion to better move through sand and other notoriously stubborn sediment layers. The core frame is gently lowered to the sea floor, ideally only minimally disturbing upper sediment, and then the central shaft is triggered from the surface to start vibrating and push through the mud. Resulting cores from JR175 were up to six meters long and should have little distortion. If the core impacts a sand or rock band, the sediment within and around that
interface can become more consolidated, but it is difficult to attribute such consolidation to coring versus genuine sediment properties.

Core 343300 was obtained with a giant gravity corer, which is essentially a tubular barrel with a weighted head. The corer is gravity driven, and consequently limited in length by friction. Gravity cores often show compression and can have some disturbance, as the force of its decent creates a pressure wave that obliterates the sediment-water interface (Griffiths and Thorpe, 1996). They have additionally been known to ‘bounce’ on the ocean floor, repenetrating the surface sediments and essentially resampling within the same core (Weaver and Schultheiss, 1990). Our core 343300, however, appears undisturbed and has actually provided our best faunal record.

070PC and 070TC were obtained using a paired trigger and piston core. Piston cores have a piston within the barrel that ‘sucks’ sediment upward as the corer impacts, allowing for much longer cores due to the reduction in friction (Griffiths and Thorpe, 1996). The trigger core is a gravity core that is attached to an arm of the piston core. The trigger and piston core are lowered to the sea floor together, but the trigger core, on a longer chain, impacts the sea floor first. Its impact ‘triggers’ the fall of the piston core by releasing a free fall cable above the piston core’s weighted top (Weaver and Schultheiss, 1990). The piston core impacts the sea floor with much greater force due to a significantly larger weight and due to this free fall line. Piston cores often lose the upper few meters of sediment, particularly if freefall lines are not the correct length, and can have unexpected distortion from ship movement that transmits to the corer (Weaver and Schultheiss, 1990). The trigger core typically captures the uppermost sediments, though not always the sediment-water interface. Trigger and piston cores are often paired together at one site to obtain a more complete sediment record, but they must be correlated to each other by comparison of the data they provide.

Inter-correlation of 070PC and 070TC

In the case of 070PC and 070TC, which were taken at the same site but represent different sediment depths, the cores had to be inter-correlated to yield a composite record. Shipboard measurements of bulk density and magnetic susceptibility (Fig II.2) show very little possible overlap between the cores—the trigger core having much lower densities and magnetic susceptibilities than all but the uppermost 25 cm of the piston core. Slanted stratification and density variations in x-rays indicate that some of the upper 25 cm of the piston
core could be disturbed, or the surface sediments bypassed. To better define the extent of overlap between 070PC and TC, we examined shipboard measurements of diffuse spectral reflectance of both cores (Figure II.3). The color reflectance reveals that there could be a possible overlap, but not more than 30 cm beyond the disturbed upper 25 cm. It was decided that the upper 25 cm of 070PC should be cropped and ignored, and the cores put end to end, with no overlap. In the resulting composite core, which will hereafter be referred to as 070, the undisturbed portion of the piston core (25 cm and below) begins at 208 cm composite depth. The total composite depth is therefore 430 cm, though the last foram sample is from the interval 427-428 cm.
box core sample used to obtain modern assemblages (described previously), is never prescribed a depth, only an age, as all 070 foraminiferal analysis was done using an age scale, not a depth scale.

The reflectance data used here was preliminary data collected by Joseph Ortiz (Kent State University) during the Hudson cruise using a Minolta CM-2600d ultraviolet/visible spectrophotometer (400-700 nm wavelength range, 10 nm resolution, 3mm spot size). Ortiz has since used a center-weighted derivative of the spectral data and Principal Component Analysis to isolate contributing mineral factors (from standards) within the sediments. This work (Ortiz, 2010) is very promising but is not presented here, as it was not what was originally used to assign our composite depth, and it uses a slightly different depth scale that includes the 070 box core.

Figure II.3. Spectral reflectance of 070TC and 070PC, put on end to yield a composite record. Composite depth is relative to the top of the trigger core. 208cm in TC = 208cm in composite = 25cm in PC.

Age models

Material for radiocarbon dating was rinsed with deionized water, but no other pretreatment was performed prior to shipping samples for testing. All radiocarbon dates are summarized in the table II.2. These dates provide a very well constrained age model for 343300, as presented by Jennings et al., 2010, and a reasonable age model for HU070. There are insufficient dates to give an age model for either VC20 or VC05.
With the HU070 core, it should be remembered that there is some additional uncertainty in the age model between the dates at 172cm and 326cm due to the inter-correlation of 070’s two component cores. We are in the process of obtaining more dates for 070 to better constrain the overlap, but the current age model should be sufficient when looking at large scale changes in assemblages. HU070 is assumed to end near 1000 cal. yr BP.

Figure II.4. Age models for HU070 and 343300 cores. Points indicate depths with radiocarbon dates. The red dotted line indicates assumed ages.
III. Results

[Graph showing results with various species and time periods]
Figure III.5. Ice rafted detritus in HU070 core. Tephra shards can be viewed as very fine IRD. Comparisons of course IRD abundance between cores should be made with caution, as core diameters (and therefore sediment volume) may have varied.

Figure III.6. Ice rafted detritus in VC20 core. Tephra shards can be viewed as very fine IRD. Comparisons of course IRD abundance between cores should be made with caution, as core diameters (and therefore sediment volume) may have varied.
Figure III.7. Tephra in 343300 core. Tephra has not yet been counted above 5.3 cal kyr BP.

Figure III.8. Ice rafted detritus in VC05 core. Tephra shards can be viewed as very fine IRD. Comparisons of course IRD abundance between cores should be made with caution, as core diameters (and therefore sediment volume) may have varied.
Foraminiferal Assemblages

**HU070 and VC20**

The shelf cores 070 and VC20 are located roughly 10km apart and show very similar assemblages and transitions. Zones I – IV in VC20 appear to directly correspond to zones I – IV in 070, and zone V in VC20 to zone VI in 070. Both cores begin with very low abundance of foraminifera in zone I, and a calcareous assemblage of *Cassidulina reniforme*, *Elphidium excavatum f. clavata*, and *Islandiella helenae*. This zone also has significant IRD, with the course fractions >500µm and >2mm peaking in both cores. In HU070 this section of the core is a sandy band; in VC20 it is mixed sand and diamict, with sand from the core bottom to roughly 525cm, diamict from 525–425cm, and sand from 425-350cm (Kilfeather *et al.*, 2010). VC20, with a basal date of 12.47 cal kyr BP, extends further back in time than 070, which likely extends no further than 11.8 cal kyr BP. This zone ends some time before 11.47 cal kyr BP (the lowest date in the core).

Zone II in both cores is dominated by *Stainforthia feylingi* and *Spiroplectammina biformis*, as well as *Islandiella helenae*, which is particularly prevalent in VC20. Also present are *C. reniforme* and *E. excavatum* in low abundance. *T. torquata* begins to increase towards the end of this zone as well. This zone has higher, though fluctuating total abundances, and marks the appearance of tephra in the cores, representing fine grain IRD. Interpolation between two dates in 070 sets the termination of zone II at approximately 10.9 cal kyr BP.

Zone III is dominated by *Reophax arctica* (also called *Cuneata arctica*) and *Spiroplectammina biformis* (each 30-40%), though the total abundance of *S. biformis* declines from the previous zone. *Textularia torquata* and *Reophax catenata* are also present in both cores (2-10% each). Some calcareous fauna can be seen at the beginning of this zone in the form of declining *I. helenae*, *C. reniforme*, and *E. excavatum*, but they are very low in abundance. More notable is the appearance of organic test linings, which could indicate increased dissolution, possibly due to a decrease in sedimentation rate. Tephra remains in high concentrations during this period as well. Dates in the 070 core are more widely spaced here, giving a greater range for the termination of this zone, but it ended some time before 8.36, and probably closer to 9 cal kyr BP.

Zone IV is still dominated by *R. arctica* and *S. biformis*, but *T. torquata* becomes more prominent, and *Adercotryma glomerata*, *Eggerella advena* and *Portotrochammina bipolaris* are seen in small percentages (2-8% each). In VC20 some *Islandiella norcrossi* and *E. excavatum* are additionally present. The lack of other calcareous fauna could be an indication of poor preservation rather than the true absence of other species, as
test linings are present in both cores. Some tephra is present in VC20 during this zone, and while we see almost no tephra in 070, we do see a small peak in >2mm IRD. The 070 age model constrains the termination of zone IV to between 7.12-7.83 cal kyr BP.

Zone V in 070 is only represented by 2 samples due to low sedimentation rates and low sampling resolution. Similar stratigraphy in VC20 could explain why this zone is not distinctive in VC20. Zone V is characterized by a drop in _R. arctica_ and _S. biformis_ and a simultaneous rise in _A. glomerata_ and _P. bipolaris_. _T. torquata_ is abundant, and many of the large agglutinated taxa appear: _R. fusiformis, R. subfusiformis, S. difflugiformis_. _Textularia earlandi, Portotrochammina weisneri_ and several species of trochammina also are first seen here. A date just above this zone puts its termination at some time before 6.32 cal kyr BP.

Zone VI in 070 (zone V in VC20) is distinctive in both cores for a sudden rise in test linings. The assemblage is similar to what began to emerge in Zone V, with no single dominant species, but a wide variety of abundant species. _A. glomerata, P. bipolaris, T. torquata, R. arctica_ and _R. catella_ are the most prominent, but _R. catenata_ and _S. difflugiformis_ are also significant.

Zone VII in 070 is based on the living and dead assemblages in the box core taken at that site. It reveals that the modern assemblage is heavily dominated by _S. difflugiformis_, though we also see _P. bipolaris, A. glomerata, R. subfusiformis_, and _T. torquata_. The dead assemblage (indicating the recent, but not present day assemblage), also contained _R. catella, R. catenata_, and _R. arctica_.

343300

343300 is the most well dated core in this transect, and can hopefully more closely constrain similar assemblage changes seen in the other cores. The bottom of 343300 (zone I) closely resembles zone II in 070 and VC20, with very low total abundance and a non diverse assemblage dominated by _S. feylingi_ and _S. biformis_. Unlike the shelf cores, we see more _I. norcrossi_ than _I. helenae_ at this site, as well as _R. catella_ and _catenata_. The distinction between _Reophax catenata, catella_ and _scotti_ was not consistently made throughout this core as several different people have contributed to its assemblage work. They have therefore been combined to avoid misrepresentation. This zone is constrained by a date at approximately 11.1 cal kyr BP.

Zone II here again closely resembles zone III in the shelf cores, with an assemblage dominated by _R. arctica_ and _S. biformis_, and with significant amounts of tephra. _S. feylingi_ is also dominant here, much more so
than on the shelf. *T. torquata* and *R. catenata/catella* are present both here and on the shelf, accounting for 5-20% of the assemblage each. The age model for 343300 puts the termination of this zone at 8.38 cal kyr BP.

Zone III, mirroring zones IV and V in 070, is characterized by a slight decline in the still dominant *R. arctica* and *S. biformis*, but an introduction of *A. glomerata*, *P. bipolaris*, several *Deuterammina* species, and an increase in *E. advena*. *Textularia torquata* still accounts for roughly 20% of the assemblage, showing little change from zone II. Tephra in peaks a second time at the end of this zone, after showing a marked decline at the end of zone II. The termination of zone III is constrained by a date at 6.24 cal kyr BP.

After this point, 343300 departs from the other cores in that it has a much higher resolution, and therefore reveals more subtle changes in climate, and more closely defines zone boundaries. Zones IV-VI in 343300, which look similar to zone V in 070, mark a shift to a more calcareous assemblage, with a large increase in test linings and preserved specimens. Most of this period is co-dominated by *E. excavatum*, *I. norcrossi* and *Nonionellina labradorica*, but other species rise and decline throughout. This general period terminates at roughly 2.5 cal kyr BP, but can be subdivided into three different zones based on changes in these other species. In zone IV (6.24-5.11 cal kyr BP), for example, *R. arctica*, *S. biformis*, and *C. reniforme* persist at 4-10% of the assemblage each, but they taper off in zone V (5.11-3.5 cal kyr BP), to be replaced with *P. bipolaris*, *E. advena*, and several *Trochamminacea*. In zone VI (3.5-2.5 cal kyr BP), *C. reniforme* peaks along with total foram abundance, and the minor taxa shift to include *B. pseudopunctata* and *S. feylingi*. *Globobulimina* is also present throughout zones IV-VI, indicating increased overall productivity.

In zone VII the assemblage becomes more evenly split between agglutinated and calcareous species, *P. bipolaris* dominating (20-40%), and the rest of the assemblage fairly evenly split between *E. excavatum*, *I. norcrossi*, *A. glomerata*, *E. advena*, *R. arctica*, *R. catenata* et al., *S. biformis*, and *T. torquata*. This period has the highest level of species diversity in the core, and spans from 2.5 to probably no sooner than 1 cal kyr BP.

**VC05**

VC05 is not yet fully dated, with only one lower date of 9.38 cal kyr BP. Assemblage changes however are in some cases distinct enough that we may draw comparisons to the other cores. Zones I and II in VC05 are very similar, and too low in abundance to justify much differentiation between the two. The assemblage is mostly calcareous, dominated by *S. feylingi* (40-60%) and *C. reniforme* (20%), but also containing *E. excavatum*, *S. biformis*, and *R. catenata*. Some tephra is present as well, indicating fine grained ice rafting.
In zone III, *S. feylingi* co-dominates the calcareous assemblage with *C. reniforme* and *E. excavatum*, and *I. norcrossi* has a significant presence as well. *R. catenata* and *S. biformis* remain the dominant agglutinated taxa, but we also see *A. glomerata*, *P. bipolaris*, and *R. arctica*. Tephra decreases to insignificant levels during this time. The dominance of calcareous fauna (versus agglutinated) in zones I-III may be a result of higher sedimentation rates, which can lead to better preservation.

The transition from zone III to zone IV signifies the most dramatic shift in the core. The assemblage changes from mostly calcareous to mostly agglutinated, with dominant species *A. glomerata*, *R. arctica*, *S. biformis* and *T. torquata*. *I. norcrossi* and *E. excavatum* persist in low abundance, but *S. feylingi* and *C. reniforme* are absent. The period is also marked by a sharp increase in tephra, and, shortly after the tephra, a smaller increase in >500µm and >2mm IRD.

Zone V in VC05 closely resembles zone V in HU070 and VC20 for its dramatic increase in test linings, as well as its assemblage of *A. glomerata*, *R. arctica*, and *T. torquata*. The primary difference between this zone and that on the shelf is that here *S. biformis* is still dominant, and we see less *P. bipolaris* than on the shelf. This zone is a period of high IRD in VC05, with large increases in both the >500µm and >2mm fractions, though a decrease in the fine grained (tephra).

Zone VI is represented by only one sample and may be the closest to modern assemblage we have at this site. The primarily agglutinated assemblage is dominated by *A. glomerata*, *S. difflugiformis*, and *S. biformis*.

**Paleoceanographic Setting**

In this thesis we look for how changes in the WGC current may have influenced ice retreat, but another significant driver of ice stream movement is bottom topography and sea level. Figure III.9 summarizes the prominent bathymetric features in and around Disko Bay that could have caused pinning points for the ice sheet, and lists all available limiting dates for ice retreat. Marine changes must be viewed with these physical constraints in mind.

The first potential pinning point is the Hellefisk moraine, which was deposited some time between 380-130 cal kyr BP (Gibbard *et al.*, 2005), indicating that the ice margin held at this point during a previous glacial stadial. The second pinning point is a basalt escarpment that shapes one side of Egedesminde Dyb. A long deglacial sequence in a marine core near the end of this escarpment (12.3 on the map) (McCarthy, 2010) would support the presence of a retreating ice margin at that location. Another potential pinning point is the high rise
at the mouth of the bay, with shallow island chains broken up by Egedesminde Dyb. The final pinning point is at the mouth of the inner Jakobshavn fjord—where a shallow sill (200m) could have stabilized the ice margin temporarily. The two other moraines indicated in Fig III.9 (the Marrait and Tasiussaq moraines) were dated by at roughly 9.5 cal kyr BP and 8.0 cal kyr BP respectively, as summarized by Weidick and Bennike (2007).

Relative Sea Level (RSL) on West Greenland was dependent on two competing factors—a rising ocean from melting ice sheets, and the isostatic uplift, or emergence of land from continually decreasing ice mass. In the Disko region, isostatic uplift largely overrode any rise in actual sea level, such that RSL was 50-120m higher at the beginning of the Holocene than at present. RSL curves for Disko Bay (Rasch and Jensen, 1997, Rasch, 2000) based on $^{14}$C dating and the presence of Eskimo and paleo-Eskimo historical sites give some indication of the timing of ice retreat from changes in the gradient of change in RSL. At 10 and at 7 cal. kyr BP Rasch (2000) interprets a reduction in this rate of change as possible standstills of ice. From 4 to 2 cal. kyr BP, RSL falls below present, and we see a change from a falling RSL to a rising RSL beginning at ~1 cal. kyr BP from submergence of land. Changes in RSL were not consistent across the bay, however—more dramatic changes were seen closer to the GIS then on Western Disko Island, and larger RSL gradient changes were seen on the south side of the bay than on the north.
Figure III.10 summarizes the fauna accompanying deglaciation as it is evident in these cores. Fauna have been roughly classified as indicators of Atlantic water or polar water to help elucidate what changes in bottom currents, if any, may have accompanied ice retreat. The ice margin in this chronology is loosely based on the dates shown in Fig III.9, taken from Weidick and Bennike (2007), and Long and Roberts (2003), who summarized decades of research by many people. It should not be taken as a definitive extent. The focus here is more on the marine influence than the precise location of the ice grounding line. The time steps were chosen based on the zones described above. The goal was to capture periods when all cores were well within a set zone, avoiding ages near zone shift.

**Paleoceanographic Interpretation**

The basal date in VC20 introduces a potential point of dispute in our record because it indicates the possibility of an ice sheet readvance during the Younger Dryas (O’Cofaigh et al., 2010). Two shells—one single valve *Nuculana pernula cotigera* at 524.5cm, and shell fragments at 477.5 are dated at 12.43 and 12.33 kyr BP respectively. Both of these shells are in a diamicton band that could be till (Kilfeather et al., 2010). If it is till, the shells could not have lived there during its deposition; the only way for the shells to have become embedded in the till is if they lived there previously and were picked up, pushed forward, or overridden by the ice as it advanced forward. If this diamicton is indeed till, we would expect it to be overlain by an ice proximal to ice distal glaciomarine sequence.

In VC20 we find low abundance *I. helenae* in the upper diamicton, which is gradually joined by *C. reniforme* and *E. excavatum*. As stated previously when discussing species preferences, this could be an ice proximal sequence of fauna, but it is more typical of an ice distal setting. One possible explanation is that the ice margin was present as an ice shelf, which would be more conducive to the ice marginal productivity preferred by *I. helenae*. This idea of an ice shelf has also been put forward by Weidick and Bennike (2007).

Another explanation is that the ice retreated rapid by calving as opposed to melting. What we consider ice proximal fauna are classified as such because they respond to the changes in turbidity, salinity, and temperature introduced by meltwater. The ice sheet may have rapidly calved in large chunks, which could have been transported away such that the margin was not characterized by meltwater plumes during ice retreat. The lack of a significant meltwater flux would explain the absence of traditional ice proximal fauna, as well as the absence of a thick deglacial sediment sequence. Further, the changes in sea level experienced on the outer
shelf—with global sea level rising, but local land uplifting—would likely have created an unstable environment favorable to a swift break up of ice.

More samples are being examined from the base of VC20, but at this time we do not have enough evidence to say with any certainty whether a readvance took place. The first panel of Figure III.10 sets the ice grounding line at its most likely stable, post break up position (according to Weidick and Bennike) should the fauna and diamicton in VC20 come from an ice shelf, or from rapid calving.
Beginning at 11.7 cal. kyr BP and moving forward in time, the shelf cores show ice distal fauna through 11.1 cal. kyr BP, despite their distance from the ice margin, which could have been as much as 200 km away. Rapid retreat off the continental shelf would have produced a massive volume of ice bergs, perhaps creating the moderately high sedimentation rates that *S. feylingi* and *S. biformis* seem to thrive in. The ice may also have halted at the basalt escarpment before Egedesminde Dyb for a time, promoting the long ice distal sequence in VC20 and HU070. Both shelf cores show a high influx of course IRD from 11.7 – 11.1 cal. kyr.
BP. The lack of *E. excavatum* and a greater dominance by *C. reniforme* at the 343300 site at 11.1 cal. kyr BP may indicate that the ice margin had by then retreated even further than shown here.

By 10.0 cal. kyr BP, and likely well before, mollusk dates on Disko Island indicate that the island periphery was ice free and the Vaigat was open. The basal date in VC05 indicates that the ice retreated E of this site some time before 9.38 cal. kyr BP, but the exact timing is unknown. The shelf and Egedesminde Dyb sites show similar, cold, arctic assemblages, and both sites are receiving significant fine fraction IRD in the form of tephra at this time.

At ~9.4 cal. kyr BP, after the ice retreats past VC05, that site shows an ice proximal to ice distal assemblage of *E. excavatum*, *C. reniforme* and *S. feylingi*. VC05 receives very little coarse IRD at this time, which could be a result of dilution by melt water and fine silt. This is supported by the presence of tephra in the core, of roughly the same magnitude as that seen on the shelf previously (40 shards/g). The outer sites are still dominated by arctic species, but the influx of tephra to the shelf has by this time declined, and is beginning to decline in Egedesminde Dyb. The cold assemblages found from 10-9.4 cal. kyr BP is in agreement with previous work that looked at two cores that sandwich VC05 and are indicated as open circles in Fig II.1 (Lloyd *et al.*, 2005).

This previous work (Lloyd *et al.*, 2005) also inferred the possible presence of transformed Atlantic water at this time based on the appearance of *I. norcrossi* and *R. turbinatus* in those cores. While VC05 does show a similar assemblage, with both *I. norcrossi* and *R. turbinatus* shortly after 9.4 cal. kyr BP, the shelf and 343300 sites do not have noteworthy amounts of either species, and do not show a significant faunal shift until closer to 9.0 cal. kyr BP on the shelf and 8.5 cal. kyr BP in Egedesminde Dybe. Lloyd *et al* (2005) also speculated that Jakobshavn Isbrae came to still stand some time around 9.0 cal kyr BP based on continuous ice proximal to ice distal fauna in their cores. VC05 is not dated well enough to support or refute this hypothesis. Young *et al* (2010) used radiogenic dating of moraines and minerogenic sediments in lake cores to show that in some places there was ice advance between 10 and 8 cal kyr BP. This conclusion is supported by the shift in the gradient of RSL change seen by Rasch (2000). The generally cold assemblages found in all cores at this time certainly would not preclude a standstill or advance.

By 8.3 cal. kyr BP (capturing the assemblage shift on both the shelf and within Egedesminde Dyb), a strengthening of the WGC is indicated by the appearance of *A. glomerata* and *P. bipolaris* in the outer cores, and a noticeable decline in the still dominant arctic species *R. arctica* and *S. biformis*. VC05 is not dated to this point, but similar stratigraphy in a neighboring core DA00-06 (Lloyd *et al.*, 2005) indicates that zone III in
VC05 does not end until roughly 7.5 cal. kyr BP. Based on this assumption, the inner bay is at this time still dominated by ice distal fauna, though WGC influence could be indicated by an increase in *I. norcrossi*, and the appearance of small but sustained frequencies of *A. glomerata*. The onset of this strengthened WGC is not entirely consistent between 343300 and 070, but is probably between 9 and 8.5 cal. kyr BP. The increased influence of WGC into the region may have been time-transgressive, impacting some sites before others, but it is difficult to determine whether this is the case with our current age models. 9 – 8.5 cal kyr BP is during the barren, low sedimentation section of 070 where the trigger and piston cores where aligned, and we have only assumed dates for VC20 and VC05.

By 7 cal. kyr BP, warmer, more Atlantic associated fauna are dominant or co-dominant at every site, and species diversity increases. This may indicate a further strengthening of the WGC, but could also be a result of weakened influence of the Laurentide Ice Sheet, or changes in sea ice. According to the Weidick (1990) reconstruction based on terminal moraines, Jakobshavn retreated to its deep inner channel after 7.8 cal. kyr BP, leaving a shallow lip to trap outflowing ice. A decrease in GIS ice within the bay could give the appearance of a stronger WGC. This explanation weakens, however, when looking at the shelf, which would be less influenced by a change in ice outflow, as most ice is channeled through the Vaigat. The LIS had almost entirely disintegrated by this point (Dyke, 2004), which could have affected the temperature and salinity signal of the WGC within the Labrador Sea, and consequently along West Greenland. These faunal changes on the shelf may therefore be reflecting not a genuine increase in the WGC so much as a reduction in glacial influence. In either case (a genuine strengthening or an apparent strengthening from less dilution), this warming marks the possible initial onset of the HTM on the shelf. Other work in the region shows different behavior at the ocean surface. A pollen reconstruction of continental temperatures from two lake cores (one in SW Greenland, one on Baffin Island), and dinocyst temperature reconstructions of 2 marine cores both show summer cooling at the surface from 7 cal. kyr BP to present, though they also show slight warming in the winter, creating a longer growing season (Frechette and de Vernal, 2009). The bottom warming seen in our cores therefore seems to diverge from the surface timing of the HTM.

Frechette and de Vernal (2009) also saw a strong thermal gradient between the E and W sides of Baffin Bay, which they thought could be explained by a shift to a more positive NAO-like state at ~7 cal kyr BP. Such a shift would be somewhat supported by the dust and temperature records of the Penny Ice cap. Yet a +NAO is associated with a weaker IC core in the WGC, which is not supported by our faunal data.
From roughly 6.2 to 2.5 cal. kyr BP the assemblage shifts on the shelf and in Egedesminde Dyb to one dominated by calcareous fauna, and/or by linings of calcareous foraminifera. VC05, while not well dated, shows a similar rise in calcareous linings from ~30 to 160cm, which we assume corresponds to this same time interval. The timing of the assemblage shift was confirmed by comparison to nearby cores DA06 and DA05 (Lloyd et al., 2004; Lloyd et al., 2007). In the presented reconstructions, we used fauna from DA06 and DA05 where possible to verify our analysis, but fauna taken from these cores is indicated by a question mark. The sudden increase in test linings and calcareous fauna seen in all cores by 5.5 cal. kyr BP (fig III.10) indicates an increase in carbonate and productivity in the region, which could be caused by an increase in Atlantic water inflow, by a decrease in organic flux from the surface (from increased sea ice extent).

Agglutinated fauna change more gradually during this period, with Atlantic associated fauna slowly supplanting arctic species from 6.2 to 4.5 cal. kyr BP, at which point all sites show warmer assemblages. This may indicate that the sudden increase in carbonate preservation at 6.2 cal. kyr BP was caused by changes in surface conditions rather than the WGC. But the discrepancy between agglutinated and calcareous fauna could also be a result of threshold responses. If the WGC strengthened, it may have initially changed the temperature comparatively little while introducing just enough carbonate for calcareous fauna to grow. A shift in WGC is certainly supported by the presence of *N. labradorica* and other productivity indicators. Interestingly this faunal shift coincides with the rise in dust concentration in the Penny Ice Cap record, which opens the possibility for an atmospherically driven change in WGC. No statistical analysis has yet been performed between our record and the Penny record, however, so at present this remains speculation.

By 4.5 cal. kyr BP the agglutinated fauna begin to mirror the calcareous fauna in showing a strengthened WGC. Arctic agglutinated fauna are supplanted by Atlantic associated agglutinated species (*P. bipolaris*), and *N. labradorica* is found in Southern Disko Bay (DA05). Recent reconstructions of temperature and salinity in several cores near Iceland (Quillman et al., 2010) show a gradual warming at the eastern periphery of the subpolar gyre, beginning at 5.4 cal kyr BP. Quillman *et al.* have tentatively interpreted this shift as a possible contraction of the subpolar gyre, and such a shift could have strengthened the Irminger arm of the WGC. Interestingly, this period in VC05 corresponds to peak coarse IRD (defining 4.5 cal kyr BP as ~ 120cm). This IRD may indicate increased melting of Jakobshavn, initiated by an increase in WGC strength and penetration into the bay. Weidick *et al.* (1990) believe that Jakobshavn retreated behind its present margin from 5 - 4 cal. kyr BP, based on fossil material pushed forward by the ice when it readvanced. Such a retreat would have been
felt at the mouth to the fjord as an increase in ice discharge. It is clear that some dramatic shift in the WGC took place from roughly 6.2 to 3.5 cal. kyr BP.

From approximately 3.5-2.5 cal. kyr BP the assemblage shifts again to a colder assemblage, indicating the onset of neoglacial cooling. This shift is particularly apparent in Egedesminde Dybe, where *C. reniforme*, *E. excavatum*, and *S. feylingi* suddenly rise, though some Atlantic species are still present (*N. labradorica*). The shelf cores lacks the sampling resolution to fully resolve this period, but the shelf does see an increase in *R. arctica* and a decline in *P. bipolaris*. 343300 ends before 1.0 cal. kyr BP, and HU070 is insufficiently dated above 3.0 cal. kyr BP to justify comparing assemblages beyond this age.

*Holocene Thermal Maximum*

When viewed as a period of peak temperatures, the Holocene Thermal Maximum may have ranged from 7 to 3.5 cal. kyr BP in Disko Bay and on the shelf. Atlantic fauna begins to increase on the shelf as early as 7.5 cal. kyr BP, but similar faunal changes at the mouth of the bay and within Disko Bugt itself are not seen until 6.2 cal. kyr BP and later. Peak temperatures are found from 5-3.5 cal. kyr BP, based on the abundances of *A. glomerata*, *P. bipolaris*, and *N. labradorica* in all cores. It is possible that the HTM was staggered (time-transgressive) across this transect due to the cooling influence of Jakobshavn Isbrae and GIS on the more inland sites, meaning the HTM had a much earlier onset on the shelf.

*Conclusions*

Foraminiferal assemblages and IRD counts indicate that the GIS rapidly retreated from the continental shelf and into inner Disko Bay at the start of the Holocene (12.5 – 10 cal. kyr BP). After extensive retreat the region was dominated by polar fauna, indicating cold conditions. Atlantic waters are first felt in the bay sometime between 9 and 8.5 cal. kyr BP, which is a later WGC onset than previously found (9.2 cal. kyr BP in Lloyd et al., 2005). A possible strengthening of the WGC on the shelf is detected in two cores, with an onset of approximately 7.0 cal. kyr BP, but this could be an apparent increase from weakening influence of the GIS and LIS. A dramatic rise in calcareous fauna from 6.2 – 3.5 cal. kyr BP in all cores indicates a much more dramatic shift and gradual strengthening of the WGC, with warm, Atlantic associated fauna peaking at 4.5 cal. kyr BP. This coincides with retreat of Jakobshavn Isbrae behind its present margin (Weidick et al., 1990),
suggesting that subsurface warming may have played a part in the ice stream’s retreat. The region appears to return to colder conditions by 3.0 cal. kyr BP, indicating a late onset of neoglacial cooling.

Work in Progress

In addition to continuing assemblage work in other cores, two projects are underway to advance our understanding of deglaciation in the Disko region. The first is a provenance study that will look at radiogenic isotopes of trace elements in four samples to help identify which fjord system was responsible for sediment deposition on the continental shelf. The second project is measuring oxygen isotopes in the tests of foraminifera in specific sections of the cores to identify possible melt water events.

The provenance study is partially completed at this time. It spans two fjord systems—Disko Bay, and the Umanak system, which is just north of Disko. The outlets to these embayments have comparable shelf fans, and Umanak is today home to the second largest ice source in West Greenland, Rink’s Ice Front. Less is known about deglaciation in Umanak relative to Disko. This provenance study will help define the sources of sediments in slope cores from these systems, which could indicate the path of ice out of these bays.

Provenance can be attributed using radiogenic isotopes of certain trace elements. We are looking at Neodymium (Nd), Strontium (Sr), Rubidium (Rb), and Samarium (Sm). Specifically, we are measuring the ratios of $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$, $^{87}\text{Rb}/^{86}\text{Sr}$, and $^{147}\text{Sm}/^{144}\text{Nd}$, and the total concentrations of Sm, Nd, Sr, and Rb. The principle behind this work is that $^{87}\text{Rb}$ naturally decays into $^{87}\text{Sr}$, and $^{147}\text{Sm}$ decays into $^{143}\text{Nd}$. The above ratios and concentrations will quantify this decay, which we can use to date the formation of and categorize the source of minerals in the sediment.

Samples were taken from the base of four cores (Fig 3.10): one at the mouths of Rinks Ice Front (VC39), at the mouth of Jakobshavn Isbrae (VC03), on the Umanak slope (VC46) and on the Disko shelf (VC20). The VC39 sample was from ice proximal sediments, and the others (VC03,46 and 20) were from possible till.

Each sample was sieved at three size fractions (<63μm, 63-125μm, >125μm) to avoid biasing samples with the coarse grains. Each of these fractions is ground to a fine powder and incinerated to burn off any organic material, then leached in 4N HCl. Samples are then dissolved in hydrofluoric, perchloric, and nitric acids and separated into a ‘spiked’ and ‘unspiked’ subsample. The ‘spiked’ sample is spiked with a series of standards with known concentrations of the elements of interest. By adding a set concentration of an element, we give ourselves a meter by which to gauge the magnitude of the original concentration in the sample. The spiked
and unspiked samples are then run through a series of pH dependent columns designed to capture those specific elements. The spiked and unspiked samples for each element will be analyzed at the Thermal Ionizing Mass Spectrometer (TIMS) Lab at the University of Colorado in Boulder using a Finnigan-MAT 6-collector solid source mass spectrometer. The TIMS directly ionizes a sample by passing a current through a metal filament. The ions are separated by mass and by mass to charge ratio, allowing for the distinction between different elements and different isotopes.

Currently we are running all <63µm samples on the TIMS. All >125µm samples have been spiked and partially separated into the pertinent elements. Nothing has yet been done with the middle fraction (63-125µm).

The second project is less advanced than the provenance study. This stable isotope study is based on the cycling of O$_{16}$ and O$_{18}$ between the atmosphere, ice, and oceans. O$_{16}$ preferentially fractionates into the atmosphere when water evaporates from the ocean surface, leaving water vapor in the air depleted in O$_{18}$. Globally speaking, most evaporation occurs in the tropics and water vapor is transported poleward. As it moves poleward, progressively more water vapor condenses out, falling as precipitation, and $^{18}$O preferentially fractionates into this condensate. By the time it reaches the poles, the remaining water vapor is very depleted in $^{18}$O relative to water vapor at the equator. This process of progressive isotopic lightening with latitude is called Rayleigh fractionation (Bradley, 1999). The depleted vapor is the source of precipitation on ice sheets, and while that precipitation is enriched in $^{18}$O relative to the polar air, it is still depleted relative to equatorial air, and to the ocean. Melt water from an ice sheet therefore has a lighter isotopic signature than sea water, which should allow us to pick up a melt signatures in our cores.

We can see this isotopic signal in the CaCO$_3$ tests of foraminifera because the carbonate they use to build their tests is ultimately derived from CO$_2$ complexing with water. To this end we are picking samples of $I$. 

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Fig III.11. Core locations for samples used in provenance study
helenae and I. norcrossi from continuous units of calcareous fauna in several cores. Islandiella was chosen because it has a well established $^{18}\text{O}$ calibration curve and is one of the more abundant families of calcareous fauna in our cores. Roughly 10 specimens are needed to measure $^{18}\text{O}/^{16}\text{O}$ ratios in a sample. Currently we have picked sufficient specimens only from the bottom of VC20, from 400-290cm, every 10cm (with the exception of 360-361cm, which was barren of both Islandiella species). We anticipate picking samples from sections of VC05, and possibly from HU070, if more samples arrive in time.
V. References


Alley, R. B. (2004). GISP2 Ice Core Temperature and Accumulation Data, IGBP PAGES/World Data Center for Paleoclimatology.


Ortiz, J. (2010). "Mineralogy of Labrador Sea and Baffin Bay sediment inferred from diffuse spectral reflectance." Arctic Workshop, Winter Park, CO, USA, Institute for Arctic and Alpine Research.


Quillmann, U., I. R. Hall, et al. (2010). "Mg/Ca-Based Sea Surface Temperature and Oxygen Isotopic Composition of Seawater During the Holocene from the Northern Iceland Basin." Arctic Workshop, Winter Park, CO, USA, Institute for Arctic and Alpine Research.


### IV. Appendix

**Species List**

<table>
<thead>
<tr>
<th>Calcareous</th>
<th>Agglutinated</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Angulogerina fluens</em> (Todd, 1947)</td>
<td><em>Adercotryma glomerata</em> (Brady, 1878)</td>
</tr>
<tr>
<td><em>Astronomion gallowayi</em> (Loeblich &amp; Tappan, 1953)</td>
<td><em>Ammotium cassis</em> (Parker, 1870)</td>
</tr>
<tr>
<td><em>Bolivina pseudopunctata</em> (Hoglund, 1947)</td>
<td><em>Ammotium salsum</em> (Cushman &amp; Bronnimann, 1948)</td>
</tr>
<tr>
<td><em>Buccella frigida</em> (Cushman, 1922)</td>
<td><em>Cribrostomoides crassimargo</em> (Norman, 1892)</td>
</tr>
<tr>
<td><em>Buccella tenerrima</em> (Bandy, 1950)</td>
<td><em>Cribrostomoides jeffreysi</em> (Williamson, 1858)</td>
</tr>
<tr>
<td><em>Cassidulina neoteretis</em> (Tappan, 1951)</td>
<td><em>Cyclolyra</em> sp</td>
</tr>
<tr>
<td><em>Cassidulina reniforme</em> (Norvang, 1945)</td>
<td><em>Deuterammina williamsoni</em> (Bronnimann &amp; Whittaker, 1988)</td>
</tr>
<tr>
<td><em>Cibicides lobatus</em> (Walker &amp; Jacob, 1798)</td>
<td><em>Deuterammina minuta</em> (Bronnimann &amp; Whittaker, 1988)</td>
</tr>
<tr>
<td><em>Dentalina baggi</em> (Galloway &amp; Wissler, 1927)</td>
<td><em>Deuterammina montagui</em> (Bronnimann &amp; Whittaker, 1988)</td>
</tr>
<tr>
<td><em>Dentalina sp.</em></td>
<td><em>Deuterammina ochracea</em> (Williamson, 1858, as Rosalina ochracea)</td>
</tr>
<tr>
<td><em>Elphidium albiumbilicatum</em> (Weiss, 1954)</td>
<td><em>Eggerella advena</em> (Cushman, 1922)</td>
</tr>
<tr>
<td><em>Epistominella sp.</em></td>
<td><em>Portatrochammina bipolaris</em> (Brady, 1881)</td>
</tr>
<tr>
<td><em>Globobulimina auriculata</em> (Bailey, 1894)</td>
<td><em>Recuvoridae turbinatus</em> (Brady, 1881)</td>
</tr>
<tr>
<td><em>Globobulimina turgida</em> (Bailey, 1851)</td>
<td><em>Reophax arctica</em> (Brady, 1881)</td>
</tr>
<tr>
<td><em>Islandiella helenae</em> (Feyling-Hanssen &amp; Buzas, 1976)</td>
<td><em>Reophax bilocularis</em> (Flint, 1899)</td>
</tr>
<tr>
<td><em>Islandiella islandica</em> (Norvang, 1945)</td>
<td><em>Reophax catella</em> (Hoglund, 1947)</td>
</tr>
<tr>
<td><em>Islandiella norcrossi</em> (Norvang, 1945)</td>
<td><em>Reophax catenata</em> (Hogland, 1947)</td>
</tr>
<tr>
<td><em>Lagena laevis</em> (Montagu, 1803)</td>
<td><em>Reophax fusiiformis</em> (Williamson, 1858)</td>
</tr>
<tr>
<td><em>Lagena sp.</em></td>
<td><em>Reophax dentiliniformis</em> (Brady, 1884)</td>
</tr>
<tr>
<td><em>Melonis barleeanus</em> (Wiliamson, 1858)</td>
<td><em>Reophax pilulifer</em> (Bandy, 1884)</td>
</tr>
<tr>
<td><em>Nonionella turgida</em> (Williamson, 1858)</td>
<td><em>Reophax scotti</em> (Chaster, 1892)</td>
</tr>
<tr>
<td><em>Nonionella auricula</em> (Heron-Allen &amp; Earland, 1930)</td>
<td><em>Reophax sp.</em></td>
</tr>
<tr>
<td><em>Nonionella labradorica</em> (Dawson, 1860)</td>
<td><em>Reophax subsfusiiformis</em> (Earland, 1933)</td>
</tr>
<tr>
<td><em>Nonionella iridea</em> (Heron-Allen &amp; Earland, 1932)</td>
<td><em>Saccammina diffugiformis</em> (Brady, 1879)</td>
</tr>
<tr>
<td><em>Pullenia bulboides</em> (d'Orbigny, 1826)</td>
<td><em>Silicosigmoilina groenlandica</em> (Cushman, 1933)</td>
</tr>
<tr>
<td><em>Pullenia ostolensis</em> (Feyling-Hanssen, 1954)</td>
<td><em>Spiroplectammina biforntis</em> (Parker &amp; Jones, 1865)</td>
</tr>
<tr>
<td><em>Quinqueloculina seminula</em> (Linne, 1758)</td>
<td><em>Textularia earlandi</em> (Phleger, 1952)</td>
</tr>
<tr>
<td><em>Rosalina sp.</em></td>
<td><em>Textularia torquata</em> (Parker, 1952)</td>
</tr>
<tr>
<td><em>Stainforthia concava</em> (Hoglund, 1947)</td>
<td><em>Trochammina sp.</em></td>
</tr>
<tr>
<td><em>Stainforthia feylingi</em> (Knudsen &amp; Seidenkrantz, 1994)</td>
<td><em>Trochammina squamata</em> (Jones &amp; Parker, 1860)</td>
</tr>
<tr>
<td><em>Triloculina trihedra</em> (Loeblich &amp; Tappan, 1953)</td>
<td><em>Trochammina sp.2</em></td>
</tr>
<tr>
<td></td>
<td>*sp.6 &quot;Mystery species&quot;</td>
</tr>
</tbody>
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