NEW EVIDENCE FOR ICE SHELF FLOW ACROSS THE ALASKA AND BEAUFORT MARGINS, ARCTIC OCEAN

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By
Jennifer L. Engels

Dissertation Committee:

Margo Edwards, Chairperson
Frederick Duennebier
Christopher Measures
Leonid Polyak
Jane Schoonmaker

We certify that we have read this dissertation and that, in our opinion, it is satisfactory in scope and quality as a dissertation for the degree of Doctor of Philosophy in Geology and Geophysics.

DISSERTATION COMMITTEE

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Chairperson

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ABSTRACT

The Arctic Ocean may act as a lynchpin for global climate change due to its unique physiography as a mediterranean sea located in polar latitudes. In our modern warming climate, debate over the bounds of natural versus anthropogenically-induced climate variability necessities a comprehensive understanding of Arctic ice extent and configuration over the last interglacial cycle. Longstanding controversy exists as to the volume, timing, and flow trajectories of ice in the Arctic Ocean during glacial maxima when continental ice sheets mantled circum-arctic landmasses. As a result of the Science Ice Exercise surveys of the Arctic Ocean in 1999, new evidence for ice grounding at depths down to 1000 m on the Lomonosov Ridge and 750 m on the Chukchi Borderland indicated the likelihood of large ice shelves flowing into the ocean from both the Barents/Kara Sea and the Canadian Arctic Archipelago or eastern Alaska. Sidescan imagery of ~14100 km^2 of seafloor along the Alaska and Beaufort margins in water depths from 250-2800 m maps a repetitive association of recognizable sub-glacially generated bedforms, ice carved-bathymetry, and ice-marginal turbidite gullies over a 640 km stretch of the margin between Point Barrow and the MacKenzie River delta. Glaciogenic bedforms occur across the surface of a flattened bathymetric bench or 'second shelf break' that is interpreted to have been formed by an ice shelf eroding the continental slope. The glacial geology of surrounding areas suggests that an ice shelf on the Alaska and Beaufort margins was likely to have been flowing from the mouths of overdeepened glacial troughs in the Canadian Arctic Archipelago westward and across the Chukchi Borderland due to an obstruction in the central Canadian basin. Evidence for an ice shelf along the Alaska and Beaufort margins supports an expanded interpretation of ice volume and extent during Pleistocene glacial periods. This has

far-reaching implications for Arctic climate studies, ocean circulation, sediment stratigraphy, and the stability of circum-Arctic continental ice masses. This dissertation provides compelling evidence for the provenance of the most recent major ice shelf transgression into the Arctic basin, and these data are critical for reconstructing Pleistocene ice history throughout the region.

LIST OF ABBREVIATIONS

ABM	Alaska/Beaufort margin
CAA	Canadian Arctic Archipelago
CB	Chukchi Borderland
dB	decibels
ERBE	Earth Radiation Budget Experiment
ESRI	Earth Systems Research Institute
FFT	Fast Fourier transform
GIS	Geographic information system
GPS	Global positioning system
IBCAO	International Bathymetric Chart of the Arctic Oceans
IPCC	Intergovernment Panel on Climate Change
INS	Inertial navigation system
IRD	Ice rafted debris
LGM	Last Glacial Maximum
LIS	Laurentide Ice Sheet
MSGL	Mega-scale glacial lineation
mwd	meters water depth
NADW	North Atlantic Deep Water
OIS	Oxygen Isotope Stage
SCAMP	Seafloor Characterization and Mapping Pods
SCICEX	Science Ice Exercises
SSBS	Sidescan Swath Bathymetric Sonar
THC	Thermohaline circulation
TMF	Trough mouth fan
WGS	World Geodetic System
YP	Yermak Plateau

INTRODUCTION

1.1 Why study ice in the Arctic?

The volume and distribution of ice in the Arctic region is important to many aspects of both modern and paleo-climate science. The freeze/thaw cycle of sea ice in the Artic Ocean impacts ocean circulation as far away as Antarctica and may cause major shifts in hemispheric weather patterns [Ruddiman, 2001]. In a warming climate, the Arctic is particularly susceptible to dramatic change because the majority of Arctic ice floats over water. Melting ice exposes open water that has low albedo (reflectivity) compared to ice or snow and therefore absorbs heat more efficiently, which in turn causes additional melting of sea ice. This self-perpetuating feedback mechanism can quickly result in reduced extent and thickness of sea ice due to melting, as well as large fluxes of fresh water out of the Arctic into the North Atlantic, as has been documented in the modern Arctic over the last three decades [IPCC, 2001]. Antarctica has less sea ice than the Arctic, and most of its ice volume is situated on land, making it less susceptible to the sea-ice – albedo feedback loop described above.

The total volume of Arctic ice is also determined by the way in which heat is redistributed over the Earth's surface. As measured at the top of the atmosphere, incoming shortwave radiation from solar insolation is concentrated at equatorial latitudes (Figure 1) [Pidwerny, 2004]. This energy is then transported to the poles via both oceanic and atmospheric processes [Merritts et al., 1997]. As a result, outgoing longwave radiation or sensible heat measured at the top of the atmosphere is more nearly equal at polar and equatorial latitudes (Figure 1). It is estimated that without this efficient meridional transfer of heat, equatorial regions could be up to 14˚C warmer than present, and polar regions could be as much as 25˚C colder [Pidwerny, 2004]. One of the primary mechanisms by which heat is redistributed in the Earth's oceans is via global thermohaline circulation (THC) (Figure 2) [Ruddiman, 2001].

Thermohaline circulation is driven by the sinking of cold, saline waters in the Norwegian-Greenland seas to create North Atlantic Deep Water (NADW). North Atlantic circulation is therefore very sensitive to perturbations in fresh water flux. Fresh water input to the North Atlantic from the Arctic may control large-scale heat and salt redistribution, regional and global THC, and thereby climatic conditions [e.g. Clark et al., 2002]. In a warming climate, a lens of ice-derived fresh water may cap the water column, causing the water mass structure to become stratified, and shifting the center of downwelling NADW to the south or causing THC to shut down entirely. Rahmstorf [1995] has shown that even relatively small (~0.01 Sv) and local perturbations in fresh water input to the North Atlantic may result in regional temperature changes of several degrees on timescales of only several years. Manabe and Stouffer, [1995] indicate that a complete breakdown of the THC and oceanic meridional heat transport would result in substantial climate change. In our warming modern climate, a great deal of debate centers on the threshold between natural vs. anthropogenically-influenced climate variability [IPCC, 2001]. By studying the extent and distribution of Arctic ice in the past and its impacts on fresh water flux and regional and global THC over time, we may be able to gain some insight into the bounds of natural climate change that will help us to understand the modern system.

1.2 Synopsis of the Arctic ice controversy: 2 scenarios

The modern Arctic Ocean region (Figure 3) is warm relative to Pleistocene glacial maxima, and the basins are covered by seasonally variable sea ice up to a maximum of only 10 m thick where concentrated by wind and currents [Jeffries, 1992]. Longstanding controversy exists as to the extent and continuity of ice in the Arctic Ocean during Pleistocene glacial periods when huge ice sheets existed at its periphery (Figure 4)[Dyke et al., 2002; Svendsen et al., 2004]. Based on theoretical $\delta^{18}O$ and ice mass balance considerations and comparisons to the setting of the West Antarctic Ice Sheet, Mercer [1970], Broecker [1975], and Keigwin [1982] hypothesized that vast and thick ice shelves could have formed in the Arctic Ocean, possibly coalescing to cap it entirely during glacial maxima. Building on the work of the CLIMAP group in the 1980s [CLIMAP, 1984], Grosswald and Hughes [1999] suggested that the entire Arctic Ocean was covered with a giant floating ice cap at the Last Glacial Maximum (LGM) that became grounded and pinned on topographic highs. Their hypothesis holds that the existence of a “Pan-Arctic” ice cap is the best way to reconcile evidence of abiotic conditions from the marine stratigraphic record, ice stream dynamics, and geophysical ice mass balance studies used to explain ice grounding traces around Arctic margins. Until now, this hypothesis has been mostly discarded by glacial geologists studying the Arctic Ocean and its margins who have taken a more conservative view of the LGM and previous ice ages, suggesting that they were probably similar to modern conditions, just with larger extents of year-round sea ice [e.g. Clark, 1990; Jeffries, 1992; Phillips and Grantz, 1997; Spielhagen et al., 1997].

The two contrasting scenarios described above have vastly different implications for both Artic science and climate science generally. The existence of continuous ice shelves and/or iceberg armadas in the Arctic Ocean would increase surface albedo and inhibit ocean-atmosphere heat and gas exchange as described in the previous section. The Bering Strait and Canadian Arctic Archipelago (CAA) channels would be closed restricting circulation [Goosse et al., 1997a, b; Forman et al., 2000] and the trajectories of floating ice and ice rafted debris (IRD) dispersal. The volume and timing of ice shelf growth could also complicate calculations of global sea level, for which significant questions remain, especially for ice ages preceding the LGM [Rohling et al., 1998]. Floating ice shelves contribute light isotopes to the $\delta^{18}O$-derived sea level curve in the same way as land-based ice, but without affecting sea level. For example, a 1000 m thick floating ice shelf in the Arctic Ocean generates a sea level bias of at least -40 m in a $\delta^{18}O$-based reconstruction [Williams et al., 1981]. For the LGM there is a ~20 m offset between sea level reconstructions based on marine $\delta^{18}O$ records and independent sea level data, possibly due to an ice shelf effect [Clark et al., 2001]. Such discrepancies could be larger for earlier glaciations when much less is known about sea level. Finally, an entirely ice-covered ocean might affect the stability of land-based ice sheets in the Northern Hemisphere via buttressing, as is documented for the West Antarctic Ice Sheet today [Anderson, 1999].

Table 1 – Oxygen isotope stage numbers and dates

Stage number	**Glacial/Interglacial**	**Approximate dates in years before present [Brigham-Grette, J., pers. comm., 2004]**
OIS 1	Interglacial	0-12,050
OIS 2	Last Glacial Maximum	12,050-24,110
OIS 3	Interglacial	24,110-58,960
OIS 4	Glacial	58,960-73,910
OIS 5a	Interglacial?	73,910-79,250
OIS 5b	Glacial?	79,250-99,380
OIS 5c	Interglacial?	99,380-110,790
OIS 5d	Glacial?	110,790-123,820
OIS 5e	Interglacial?	123,820-129,840
OIS 6	Glacial	129,840-150,000?

Recent marine evidence indicative of ice grounding on the seafloor from widely separated regions of the Arctic Ocean (Figure 3) suggests that larger and thicker ice volumes than previously thought may have existed in the basin during glacial Oxygen Isotope Stages (OIS) 6, 4, and at the LGM (Table 1). These locations include the Yermak Plateau (YP) where iceberg scours are documented down to 850 meters water depth (mwd) [Vogt et al., 1994; Kristoffersen et al., 2004]; the Kara and Barents Seas [Polyak et al., 1997] and Lomonosov Ridge [Jakobsson et al., 1999; Polyak et al., 2001] that show ice shelf grounding features in 630 mwd and 1000 mwd respectively; and the Chukchi Borderland (CB) [Polyak et al., 2001; Jakobsson et al., in press] and Alaska Margin [Engels et al., 2003] that both show ice shelf grounding features in 700-800 mwd and iceberg scours shallower than 400 mwd. In each of these locations extensive erosional surfaces plane the tops of submarine ridges and plateaus and form prominent angular unconformities with underlying stratified sediments (Table 2). The seafloor has diverse morphology that includes chaotically oriented to sub-parallel iceberg scours, sets of parallel flutes, and transverse ridges. The combination of these features indicates that both icebergs and coherent ice shelves grounded on the seafloor in the Arctic just as they did on glaciated continental margins around Antarctica [e.g., Anderson, 1999; Shipp et al., 1999]. These findings support the hypothesis that ice shelves several hundreds of meters thick to >1 km thick once existed in the Arctic Ocean, rivaling or even exceeding in size the largest modern ice shelves of West Antarctica.

Table 2 – Bedform type, location, and depth

Location	**Bedform style**	**Bedform depths**
Lomonosov Ridge	Flutes	> 980 m
Lomonosov Ridge	Iceberg scours	> 850 m
Yermak Plateau	Iceberg scours	> 850 m

Chukchi Borderland	Flutes	> 750 m
Chukchi Borderland	Iceberg scours	> 400 m
St. Anna Trough	Flutes	> 630 m

In some cases ice thickness and the directions of ice movement across the Arctic Ocean basins can be estimated from existing seafloor data [Polyak et al., 2001], but the overall configuration and trajectories of ice masses at specific times are far from being delineated. Identifying paleo-ice shelf boundaries and timing via the stratigraphic record is essential to differentiate between the two Arctic ice volume scenarios discussed above, as well as to understand mechanisms of ice formation and growth. The synchronicity of ice shelf advances on the American and Eurasian sides of the ocean, if established, could support the much-argued pan-Arctic ice shelf scenario [Mercer, 1970; Hughes et al., 1977; Grosswald and Hughes, 1999]. Alternatively, asynchronous ice shelf events would indicate a more complex pattern involving partial and variable ice shelf coverage of the ocean during different glacial periods. In addition to resolving the timing of ice shelf advances into the Arctic Ocean, it is important to detect their extent and trajectories to better understand the dynamics of the adjacent Laurentide and Eurasian ice sheets and the interaction between land-based and floating ice sheets, such as in West Antarctica today. The full extent of Arctic glacial incursions, whether they occurred simultaneously throughout the whole Arctic basin, or whether ice ages of differing intensities in disparate locations were caused by local variations in ocean and atmospheric conditions, all remain to be determined. By examining the glacial history of the Alaska and Beaufort margins of the Arctic Ocean, this dissertation helps to resolve some of the above outstanding issues.

1.3 What we know today: basin-wide glacial sediment distribution, bedforms, and age

Sediment cores recovered from widely separated areas across the Arctic Ocean generally display a cyclic sequence of grey, nearly abiotic, and brown, faunal-rich beds interpreted to represent a succession of Pleistocene glaciations and interglacial periods [e.g., Poore et al., 1993; Phillips and Grantz, 1997; Jakobsson et al., 2000b]. Glacial intervals have extremely low abundances or the complete absence of biogenic remains and very low sedimentation rates or even hiatuses, indicating the likelihood of very thick ice cover on top of the water column [Darby et al., 1997; Nørgaard-Pedersen et al., 1998; Polyak et al., 2004]. However, these glacial stratigraphic indicators in and of themselves are not sufficient to differentiate between the conflicting Artic ice scenarios described in the previous section. Where sediments from deep-sea areas have not clearly been affected by ice grounding it is not possible to make conclusive interpretations about whether ice cover consisted of thick sea ice only, or of a more extensive ice shelf.

High-resolution geophysical mapping of relatively elevated seafloor areas in the Arctic Ocean gives new evidence for ice grounding. The crests of ridges and plateaus are commonly flattened by erosion as indicated by a prominent unconformity with the underlying stratified sediments (Table 2). A striking example of the deepest documented erosion is at the crest of the Lomonosov Ridge that is truncated to water depths of almost 1000 m along a stretch of ~60 km at the very center of the Arctic Ocean (Figure 5). Sediments eroded from the Eurasian side of the ridge crest were dumped on the Amerasian side of the ridge, while the eroded surface shows multiple parallel, low-relief flutes [Jakobsson, 1999; Polyak et al., 2001]. Sets of similar flutes reaching tens of kilometers in length are also found on the CB, sometimes in combination with a system of nested transverse ridges (Figure 6) [Polyak et al., 2001]. Sediment core studies from fluted sites confirm the presence of a hiatus beneath a thin layer of recent sediments, commonly underlain by stiff, sub-glacially compacted sediments known as diamictons [Jakobsson et al., 2001; Polyak et al., 2003]. At least two generations of flutes with differing trends occur on the CB at different depth levels and with different post-erosional stratigraphies, indicating multiple erosional events. These deep, eroded, and fluted sites are interpreted to have formed when large masses of grounded ice transited across the seafloor surface, leaving wide fields of parallel bedforms in their wake. At water depths shallower than those of the fluted fields, ridges and plateaus are excavated by abundant, typically chaotic plowmarks that overprint pre-existing bedforms and are thought to have been excavated by the protruding keels of large icebergs. At some sites such as on the Lomonosov Ridge and Yermak Plateau in water depths of almost 900 m, plowmarks are sub-parallel indicating generally unidirectional movement of larger groups or armadas of icebergs.

It has been observed that bottom currents sometimes form long furrows on the seafloor called “contourites” that usually parallel continental margins in both polar and low-latitude settings [Hollister et al., 1974; Embley et al., 1980]. However, not all fluted sites documented in the Arctic have flute trends that parallel oceanic currents or the dominant topography. The combined presence at the ridge/plateau tops in close proximity to each other of extensive erosion, flutes, transverse ridges, and iceberg scours indicates that the linear flutes were almost certainly formed by the passage of grounded ice. Similar groups of features are documented on glaciated continental margins around Scandinavia [Solheim et al., 1990; Polyak et al., 1996], Canada [Josenhans and Zevenhuizen, 1990], and Antarctica [Anderson, 1999; Shipp et al., 1999, 2002]. Seafloor areas vacated by grounded ice commonly exhibit sub-glacial bedforms (flutes, drumlins, and transverse ridges) at the deepest depths, reaching 1200 m near West Antarctica [Canals et al., 2000], while iceberg scours occupy shallower sites. This distribution is consistent with the formation pattern of glaciogenic bedforms. Sub-glacial bedforms result from coherent masses of ice resting on and deforming the substratum. As the grounded ice mass begins to disintegrate and decouple from the seafloor, huge icebergs plow the substrate with chaotic or sub-parallel scours that overprint pre-existing bedforms in shallow waters.

In some areas of the Arctic, former grounded ice flow trajectories can be determined from stratigraphy and the orientation of seafloor bedforms. For example, the accumulation of eroded sediment on the Amerasian side of the Lomonosov Ridge (Figure 5) indicates that eroding ice was flowing basinward from the Barents-Kara Sea margin (Figure 3). This is consistent with the presence of large troughs traversing the Barents-Kara Sea margin and providing natural pathways for ice streams [Siegert and Dowdeswell, 2002; Jakobsson et al., 2003]. The prevailing orientation of flutes on the CB indicates that eastern Alaska and/or the western part of the CAA was the major source of ice eroding the borderland (Figures 3 and 6) [Polyak et al., 2001]. Similar to the Barents-Kara margin, the Canadian arctic margin is dissected by overdeepened troughs that were the likely outlets for ice streams (Figure 3). This is consistent with the orientation of major ice streams in the northern part of the Laurentide Ice Sheet (LIS) as indicated by geomorphic and modeling data [Dyke et al., 1992; Bigg and Wadley, 2001; Stokes and Clark, 2001]. The distribution and trajectories of ice discharged into the Arctic Ocean could differ from one glaciation to another as suggested by the two generations of flutes on the CB. Iceberg scours generally are not as useful for inferring ice flow trajectories since they are often chaotic and overlapping in their orientations. However, in some locations such as on the Lomonosov Ridge and Yermak Plateau in water depths <900 m, fields of sub-parallel iceberg scours indicate that large groups of icebergs were moving in a single direction in response to oceanic currents or other forcing [Polyak et al., 2001; Kristoffersen et al., 2004].

Although limited, sediment core data provide an initial chronostratigraphic framework for major glacial incursions into the Arctic Ocean (Figure 3). Data from the eroded portion of the Lomonosov Ridge indicate that this erosion occurred during OIS 6, ca. 150 ka BP (Table 1) [Jakobsson et al., 2001]. The age model used for this estimate is now widely accepted in the Arctic Ocean stratigraphic community [Backman et al., 2004; Spielhagen et al., 2004]. This age of erosion matches the reconstruction of a huge ice sheet in northern Eurasia during OIS 6 based on compelling glacial-geological evidence [Svendsen et al., 2004]. The stratigraphy of eroded sites on the CB is not as well understood, but there are indications that the major ice-grounding event there also took place during OIS 6 or a substage of OIS 5, with another glacial erosion event possibly occurring after OIS 5 [Polyak et al., 2003]. If this interpretation holds true, a likely scenario is that during OIS 6 very large masses of ice expanded into the Arctic Ocean both from the Eurasian and American sides (Figure 3). This raises the question of whether these ice masses coalesced into a thick ice shelf covering the entire Arctic Ocean. More complex configurations can also be imagined, with coherent ice shelves occupying only parts of the basin, especially if ice advances on various sides of the Arctic Ocean were not synchronous. The exact ages of maximal ice build-ups along the Arctic margins, especially for pre-LGM times, are not well understood in Eurasia or in North America. For the LGM, existing stratigraphic data provide no evidence of ice grounding at deep water depths, however the possibility of a thinner ice shelf or at least of very thick sea ice over the Amerasia Basin is indicated by a halt in biological production and a hiatus in sedimentary records [Poore et al., 1999; Polyak et al., 2004].

1.4 The Alaska/Beaufort margin and its importance to the Arctic ice controversy

An important area to clarify the sea ice vs. Pan-Arctic ice shelf controversy for the Arctic is the Alaska/Beaufort margin (ABM) of the U.S. and Canada (Figure 7), the northernmost location of the ice sheet that covered most of North America during the LGM from ~20,000-12,000 years ago [Dyke et al., 2002]. The ABM is situated between the glacially impacted CB and the glacial outlet troughs of the CAA. These troughs are the hypothesized source of the ice that grounded on the CB (Figure 3) [Polyak et al., 2001]. The ABM may therefore provide a link to understanding more about the location, trajectory, and scale of past ice volumes in the Arctic. Continental margins such as the ABM are likely locations for land-based ice sheet termination and grounding, ice shelf formation, sub-glacially derived sediment deposition and downslope landslides (turbidites), and water mass exchange triggered by the melting and freezing of ice (sub-ice brine rejection). In addition, submarine continental shelves and slopes in cold regions of the globe exert dynamic controls on important oceanic processes such as water mixing, upwelling, eddy formation, nutrient supply, open water (polynya) formation, water mass ventilation, and the location of boundary currents. Oceanographic data from the ABM show evidence for a seasonal westward-flowing subsurface boundary current derived from Pacific water inflow [Pickart, 2004] at depths of 100-150 m (Figure 3). The existence of deeper margin-parallel currents in this location would have to be related to the formation and transit of deep waters, such as at the Greenland margin, which is not observed in the Canada Basin today. It is even less likely that deep contourite-forming currents would have formed during glacial periods when the closure of the Bering Strait and CAA channels would have further restricted circulation in the Arctic Ocean [Goosse et al., 1997a, b; Forman et al., 2000].

The ABM extends from the base of the Chukchi Rise eastward to MacKenzie Bay at the mouth of the Amundsen Gulf in the CAA (Figures 3 and 7). This aseismic Atlantic-type continental margin has a continental shelf measuring on average 75 km wide that ranges in depths from 0-75 m. The shelf is characterized by gentle seaward dips with zones of down-to-the-north normal faults that are particularly prominent near the shelf break where some of them bound large and deep regional rotational slumps [Grantz et al., 1979]. The margin is bounded to the north by a steep, heavily canyoned continental slope that drops to 1000 mwd in the Beaufort Sea and Canadian Basin over an average distance of just 13 km. Stratigraphic studies of the region reveal that the North Beringian Marine Abrasion Platform is a broad, low relief plain that extends from the subaerial Arctic foothills of the Brooks Range to the continental shelf break [Dinter et al., 1990]. The whole length of the coastline up to 7 m above modern sea level is mantled by 10s-100 m of the unconsolidated clastic materials of the Gubik Formation. This enigmatic formation shows multiple successions of glaciomarine transgressive deposits dating from 3 Ma through the LGM that have puzzled researchers for years [Dinter et al., 1990]. Current stratigraphic studies hypothesize that deposition of glaciomarine sediments occurred when continental-scale ice in the CAA collapsed in OIS 5a, flooding the coastline and entraining IRD [Brigham-Grette et al., 2001]. However, the mapped formation shows little to no isostatic tilt along its length [Brigham-Grette, J., pers. comm. 2004], which is inconsistent with isostatic loading in a location so distal to the ABM as the CAA.

1.5 Science Ice Exercise (SCICEX) programs

Until recently, many portions of the Arctic Ocean had never been or were only sparsely surveyed due to the constraints of operating in a perennially ice-covered and climatically inhospitable ocean. Traditional Arctic surveying methods such as the use of floating ice islands, icebreakers, and aircraft suffer from restricted operations because they are subject to seasonal ice and weather conditions. In 1993 a nuclear-powered submarine was made available to the scientific community for a proof-of-concept Arctic research cruise. Based on the success of this venture the U.S. Navy invited researchers in the academic community to design and participate in a series of cruises entitled the SCience ICe EXercises (SCICEX) to the Arctic between 1995-1999 [Edwards and Coakley, 2003]. Research operations were conducted aboard Sturgeon-class submarines specifically designed to travel below and surface through sea ice (Figure 8). The advantages of submarine surveys include near-complete freedom of movement under the ice, a quiet and stable platform for collecting acoustic data, and the long survey times possible onboard submarines (up to 90 days submerged). The primary disadvantage of the submarine platform is poor navigational accuracy. Submarines navigate when submerged using inertial navigation systems corrected with Global Positioning Satellite (GPS) network fixes when the submarine surfaces. Navigational errors thus increase with time spent under the ice and are up to ~3 km in some surveyed areas based on data crossover points.

Several research programs were undertaken during the last of the SCICEX cruises conducted in 1999 aboard the *USS Hawkbill*. Surveys targeting the CB, Lomonosov Ridge, and Yermak Plateau (Figure 3) were anticipated to produce evidence of iceberg or ice shelf grounding that might clarify the Pleistocene Arctic ice controversy [Polyak and Edwards, pers.

comm., 2003]. Seafloor data from the CB and the Lomonosov Ridge were subsequently used to support an expanded interpretation of thick ice grounding in the basin during OIS 6 or 5 [Polyak et al., 2001]. The primary goal of the SCICEX-99 ABM survey was a physical oceanography study of the continental slope designed to characterize boundary currents and eddy structure (Figure 7). Fortunately, in accordance with the SCICEX-99 focus on underway data acquisition, the submarine's purpose-built 12 kHz Sidescan Swath Bathymetric Sonar (SSBS) collected data throughout the course of the six-day ABM margin survey. The seafloor image data were acquired in non-ideal conditions of constantly changing water depths to accommodate the water sampling goals of the program, but nonetheless cover a total area of ~14100 km^2. The ABM data remained unprocessed due to a lack of site-specific funding until the outset of this project.

1.6 Framework for the study: hypotheses

The location, areal extent, and depth range of the ABM data are ideally situated to capture any trace of ice passage along the margin from Alaska or Canada to the CB (Figures 3 and 7). Results of SCICEX data collection from 1998-99 as well as other icebreaker cruises to the Arctic in the early 1990s generated a large new dataset of glaciogenic seafloor features in widespread regions around the Arctic [Vogt et al., 1994; Polyak et al., 1997; Jakobsson, 1999; Polyak et al., 2001; Kristofferson et al., 2004]. However, a key component of the Arctic ice story still missing from these datasets is strong evidence for the provenance of the large ice masses that created seafloor features in water depths from ~400-1000 m. Certainly ice masses of the dimensions implied by the observed depths of ice impact should have left some trace of their genesis and flow trajectories. Based on the prevailing orientation of glaciogenic flutes on the

CB, Polyak et al. [2001] hypothesized that the glacial ice had to have come from eastern Alaska or the western CAA. Shallow seafloor depths down to ~40 m around the margins of the Canada basin are excavated by modern iceberg drafts that have obliterated any trace of older ice passage. To date, intermediate depths from 40-250 m were either unsurveyed or uninformative as to the provenance of the ice that left glaciogenic bedforms on the CB. Accordingly, my first hypothesis is that: ***(1) SCICEX data from the ABM will reveal evidence for the provenance of the ice that created the CB glaciogenic bedforms.***

The questions listed below were generated from a comprehensive review of the Arctic ice literature to structure my analyses and to develop hypotheses. During the course of my research, I used them as a guide for placing the bathymetric and sidescan results from the Alaska margin in the larger context of Pleistocene glacial ice ages throughout the Arctic basin.

1) Margin substrate information –

How thick are sediments below the Alaska margin? What is their composition? Would they produce fast basal sliding for an ice sheet/shelf? How strong and heavy are the sediments? How much "backstress" would be needed to erode those sediments and what minimum grounded ice size does this imply under what forcing conditions?

My second hypothesis is that: ***(2) if grounded ice transited across the ABM, it should have left behind sub-glacial sedimentary bedforms that I can recognize with my seafloor data.*** The above questions are largely addressed through a literature review, particularly the papers of Grantz et al. [1979], Dinter et al. [1990], Kayen and Lee [1991], and Brigham-Grette [2001] who have all done land-based and some seismic reflection work along the ABM. Their studies show that the margin is draped with up to 2700 m of marine and non-marine clay, silt, and sand sediments over bedrock and has been impacted by both faulting at the shelf break and gas hydrate mass wasting between 200-2000 mwd [Kayen and Lee, 1991]. Of principal interest to my study is that the margin is in fact covered with unlithified, easily erodible sediments. An ice mass sliding across these sediments should encounter little resistance at the seafloor interface and could successfully mold the substrate into recognizable sub-glacial bedforms. Additionally, an eroding ice mass could move quickly across the sediments, (relative to its speed across bedrock [Tulaczyk et al., 2001]), with minimal deformation and break-up.

2) Glacial setting and ice provenance -

How much of the continental shelves would be exposed at lower sea level during glacial periods OIS 2, 4, and 6 (Table 1)? How would isostatic changes affect the area during glacial intervals? What was the distribution of land-based ice during the period of interest? Would ice features on the ABM imply formation by big icebergs, ice shelves, or an ice sheet? How were ice sheets and shelves related on the American and Canadian Arctic margins?

My third hypothesis is that: ***(3) ice impacting the ABM was an extension of continental ice masses in Alaska or Canada as opposed to sea ice or icebergs, and therefore bedforms will show evidence for the cohesive structure of the ice and its origin on land.*** It is nearly impossible to sort out isostatic information for the Arctic prior to the LGM, but it is fair to assume that any large local ice mass would have caused subsidence of the underlying substrate and adjustments to regional sea level, as has been inferred for the LGM. Sea level in the Arctic is still not well constrained, particularly for glacial periods prior to the LGM, but has been delineated in some areas by the research of Rohling et al. [1998], Vincent [1992], Dyke et al. [1992; 2002], Clark et al., [2001], and Grosswald and Hughes [1999]. By most estimates, glacial sea levels in the Arctic were ~100-150 m lower than modern levels [Rohling et al., 1998]. However, even taking into account lowered glacial sea levels, the ice that created seafloor bedforms would still have been far too thick (up to 650 m thick on the CB) to have been generated by sea ice accumulation or icebergs. At the LGM there is little evidence to support large land-based ice sheets in northern Alaska, but during OIS 6, 5b, and 4 there is evidence to suggest thick continental ice in both the CAA [C. D. Clark, personal communication to L. Polyak, 2004], and Alaska [Brigham-Grette, 2001] as well as thick ice on the CB [Polyak et al., 2001]. Continental ice masses that propagated as far as the Arctic Ocean coastline would likely have produced marginal ice shelves just as they do in Antarctica today [Anderson, 1999] that could have grounded to produce glaciogenic bedforms.

3) Ice dimensions and flow characteristics –

What were the spatial dimensions, thickness, and flux of ice inferred from sea level calculations and ABM bedform dimensions? Is there evidence that bathymetry steered ice over the Alaska margin or that the ice created its own bathymetry? How do the sizes of ice features on the ABM compare to inferred former ice streams such as the M'Clintock Channel ice stream or Antarctic examples?

Hypothesis four: ***(4) the ice that impacted the ABM was derived from a grounded, flowing ice shelf that responded dynamically to changes in slope but was large enough to affect the bathymetry of the whole continental margin.*** The above questions are addressed directly in my data by examining the style, depth, distribution, and orientation of seafloor bedforms on the ABM (Figure 7). Correlations between bathymetry and bedform orientation are used to gauge bathymetric steerage and changes in the ice flow path along the margin. Analyses of ABM slope profiles highlight the impact of grounded ice on the local bathymetry. I compare the ABM features to other documented glaciogenic seafloor bedforms around the Arctic and to ice streams from the northern LIS margin [e.g. Stokes and Clark, 2001; Dyke et al., 1992; 2002; Vincent, 1992].

4) Ice trajectory and evolution –

Could ice transiting along the Alaska margin also have created bedforms on the CB? Do sediments found on the northern coast of Alaska relate to ice propagation along the ABM? Do ABM bedforms follow bathymetric contours or flow down slope? What controlled the overall position of the ice and the location of its margins?

Hypothesis five: ***(5) ABM ice flowed parallel to the margin and west to the CB due to an obstruction in the central Canadian basin.*** Two alternate hypotheses exist as to the reason for the preferred orientation of mapped CB glaciogenic bedforms (Figure 6). Grosswald and Hughes [1999] infer that a pan-Arctic ice cap in the central Canadian basin prevented CAA ice streams from penetrating into deeper waters, while Polyak et al., [2001] instead invoke thick glacial sea ice to block CAA ice stream flow. Relating ABM features to glaciogenic bedforms throughout the Canada basin will help to constrain their age, origin, and relationship to other glacial episodes. If bedforms on the ABM and CB (Figure 3) [Polyak et al., 2001] have similar depths, styles, and orientations, this might link them in both genesis and age. Bedforms on the CB have tentatively been dated to OIS 4-6. Several authors have noted glaciogenic clast material in the Flaxman Formation along the Alaska margin up to ~7 m above modern sea level that date to OIS 5a [Dinter et al., 1990; Brigham-Grette, 2001; Grantz et al., 1990]. It seems reasonable that glaciomarine deposits on the ABM coastline could be related to glaciogenic bedforms on the ABM and CB.

1.7 Chapter outline

Chapter 1 presents bedforms mapped in sidescan sonar imagery of the ABM continental slope at a range of water depths, and discusses their probable formation mechanisms. Chapter 2 focuses on the origin of a series of highly developed dendritic gully systems that initiate along the whole length of the margin deeper than 500 mwd. Chapter 3 characterizes the bathymetry of the margin using spectral analysis techniques to differentiate between areas eroded by different ice morphologies.

CHAPTER 1: Glaciogenic bedforms

2.1 Survey design, navigation, and methods

During the 1999 SCICEX cruise [Edwards and Coakley, 2003] ~2000 km of 12 kHz sidescan sonar and bathymetry data were collected by the nuclear submarine USS *Hawkbill* along a 680 km swath of the ABM between 140-160° W (Figures 3 and 7). The submarine mapped seafloor from 200-2800 m deep in a zigzagging pattern along the ABM to accommodate a water sampling program (Figure 7). Sidescan swath widths vary from 6 km at the shallowest depths to >20 km at the base of the continental slope. Sidescan sonar seafloor insonification angles exceed 80° in shallow water, emphasizing low-relief morphologic features at the outer swath edges. Sidescan data were iteratively processed to remove speckles, striping, nadir gaps, and other artifacts that reduce image quality using the Hawaii Mapping Research Group's BTYP program. Sidescan data were then gridded in 8 m cells yielding two-dimensional swath imagery for ~14100 km^2 of seafloor. Narrow-beam bathymetry data with resolutions of ~2% of water depth were stripped from the raw swath bathymetric data files every ~400 m along track and were then manually filtered to remove artifacts such as missed pings. Where the submarine traversed the continental slope along the outgoing west-east survey track, the bathymetry data were sub-sampled to create 19 slope profiles used for characterizing slope geometry. One-dimensional bathymetry data were used in this initial part of my analysis because the swath bathymetric data were not yet processed at that time (swath bathymetry data are presented in Chapter 3).

Geospatial reference for the ABM survey track locations was generated by the submarine's inertial navigation system (INS) and was then re-navigated using surface GPS fixes [Edwards, M., pers. comm., 2004]. The INS incurs errors that propagate and exacerbate with time spent under the ice and away from the last GPS reading. The ABM survey took 6 days and the submarine surfaced only at the beginning and end of the survey, resulting in relative positional navigation errors of ~3 km towards the end of the survey based on data crossover points. It is possible to shift sidescan sonar and bathymetry data swaths relative to each other to generate an apparent match between mapped seafloor features. This type of arbitrary data shifting is not ideal because it corrects the data in a relative sense but is not representative of absolute locational accuracy. As a result of the INS navigational issues described above it is not possible to pinpoint the exact localities of features of interest. However, a subsequent GPS-navigated survey of the ABM by the U.S. coastguard icebreaker Healy in 2003 intentionally replicated the SCICEX survey track over a short distance at the easternmost end of the survey. A comparison of geologic features seen in both the Healy and SCICEX data shows that at the halfway point of the SCICEX ABM survey, navigation had not yet deteriorated beyond a few 100s of meters (Figure 9).

The International Bathymetric Chart of the Arctic Ocean (IBCAO) [Jakobsson et al., 2000a] is the most comprehensive compilation dataset of Arctic bathymetry currently available and is generated from icebreaker and submarine data soundings contained in archives in the U.S., Canada, Denmark, Russia, Germany, and Iceland. The Arctic region still suffers from large data gaps and therefore the IBCAO dataset has an average resolution of only 2.5 km, though higher data densities exist for some areas. The IBCAO bathymetry grid is centered on the North Pole, uses a polar stereographic projection with the true scale at 75º N and the 1984 World Geodetic

System (WGS) datum. Co-registration of SCICEX and IBCAO data highlights data gaps in the IBCAO bathymetry while at the same time pointing to navigational offsets in the SCICEX data. However, there is generally good correspondence between the two datasets and I have used them in concert to characterize bathymetry along the ABM (Figure 7).

The goal of my analysis was to document the extent and distribution of any glacially generated bedforms in the ABM seafloor imagery. As a template for my investigations I employed the model of Stokes and Clark [2001] to identify the former location of ice streams. Indicators of paleo-ice stream bed location according to Stokes and Clark [2001] include: 1) characteristic shape and dimensions of the whole flow field (>20 km wide and >150 km long); 2) evidence of convergent ice flow patterns; 3) highly attenuated sub-glacial bedforms (length:width ratio >10:1); 4) Boothia-type erratic dispersal trains [Dyke and Morris, 1988]; 5) abrupt lateral margins (<2 km); 6) ice stream marginal moraines; 7) glaciotectonic and geotechnical evidence of pervasively deformed till; and 8) submarine accumulation of sediment into a trough mouth fan (TMF) or till delta. My analyses were accomplished via measurements from the sidescan images of bedform length, width, elongation ratio, transverse wavelength, and percentage bedform area per unit area using the Environmental Systems Research Institute's (ESRI) ArcMap Geographic Information Systems (GIS) program. This analysis is limited by the fact that sidescan sonar data are projected onto a flat surface, when in reality the seafloor surface is sloped, so measurements represent minimum values in all cases. I plotted the orientations of all seafloor bedforms documented in the sidescan imagery in rose diagrams (Figure 10). The indicators described above will help to determine whether the ice mass that created the bedforms was derived from a grounded ice stream or sheet [Stokes and Clark, 2001], icebergs and ice islands, or if in fact the features were instead formed by contour currents [Pichuring et al., 1989, Kuijpers et al., 2002; 2003].

2.2 Results

In the study area a repetitive association of seafloor bedforms, distinctive bathymetry, and mass wasting features is observed along the length of the margin. Eight separate regions of seafloor (Figure 7: A-H) totaling ~550 km^2 (~3% of survey area) show evidence of disturbed sediments in <250-700 mwd. In each locality the disturbed sediment zones occur at the shallowest and most landward point of the survey track on a prominent low-angle bathymetric bench (Figure 11a). The basinward terminations of the disturbed sediment zones and bathymetric bench are marked by deeply incised dendritic gullies that drain the continental slope (Chapter 2).

2.2.1 Bedforms

I group the observed seafloor disturbances into three distinct types. Type 1 bedforms are chaotically oriented overlapping scours <2 km in length and <25 m wide in water depths <500 m (Figure 12a). Type 2 bedforms are deeply incised, arcuate to nearly linear individual grooves up 5 km long, <50 m wide, and occur in <500 mwd (Figure 12b). In all locations where both type 1 scours and type 2 grooves are observed, the grooves overprint the chaotically oriented scours. Sidescan data show that negative-relief scour and groove furrows are bordered by positive-relief levees; total relief measured from the narrow-beam bathymetry is >10 m. Scours and grooves occur in fields of several hundred overlapping individual features with a range of trends 60-135º from north, but fields have a dominant overall fabric of ~95-100º and most trend obliquely down slope (Figure 7).

Type 3 bedforms, parallel flutes, are found between ~400-700 mwd. They reach >10 km in length and measure ~50 m wide yielding very high length:width ratios of up to 200:1 (Figure 12c). The true coverage and length of the type 3 flutes are underrepresented as they extend beyond sidescan swath edges. Flutes parallel isobaths trending 103-147º from north (average trend 114˚, Figure 10) and occur in fields that are 10->30 km wide to cover a minimum total area of 330 km^2. In area G (Figure 7) a bathymetric transect perpendicular to the lineation field shows a 5 km wide, >25 m deep trough that is floored by flutes. The total areal coverage of the type 3 fluted bedforms can be inferred from their distribution along the ABM (Figure 7). If flutes continue beyond mapped areas in similar water depths throughout the surveyed region, their along-margin extent totals ~680 km. The width of the widest field of flutes is >30 km, which indicates that the total inferred extent of the flutes falls within the characteristic flow field dimensions of paleo-ice streams outlined by Stokes and Clark [2001] (>20 km wide and >150 km long).

Four of the eight disturbed sediment regions (Figure 7: A, B, C, D) show type 1 scours, and they are often overprinted by type 2 grooves that are present in six regions (Figure 7: A, B, D, F, G, H). Type 3 flutes are concentrated in three regions (Figure 7: E, G, H) on the eastern half of the survey. The linearity and parallelism of flutes documented in the shallowest portions of areas G and H at the easternmost end of the survey begin to deteriorate with decreasing depths, causing them to grade progressively into type 2 grooves.

2.2.2 Bathymetric bench

Slope profiles created using the SCICEX narrow-beam bathymetry document the co-occurrence of the seafloor bedforms described above with the location of a flattened bathymetric bench that is 400-550 m deep and has an average slope of just 1.6º (Figure 11a). Bathymetric features on the bench surface are mostly subdued, but in fluted areas the bench exhibits undulations in the slope profiles with relief of ~25 m. The bench is also visible in IBCAO bathymetry for the Beaufort Sea (Figures 11b and 13) [Jakobsson, 2002]. The basinward margin of the bench is associated with the onset of dendritic gullies and a precipitous increase in slope to 13º. The slope width varies from 5->35 km, with the broadest portions mapped for ~100 km along the easternmost end of the surveyed area (Figure 13). The widest portion of the bench underlies areas G and H (Figure 7) where the most areally extensive examples of fluted terrain are documented.

2.2.3 Slope gullies

Throughout the survey area a well-developed series of gullies erodes the continental slope and the bathymetric bench described above in water depths >500-700 m (Figure 12a, b, d; Chapter 2). Gullies initiate in dendritic catchment areas up to 10 km across, then coalesce to form broad trunk channels >5 km wide in water depths >1000 m (Figure 12d). One mapped gully system exceeds 40 km in length from the southernmost catchment areas to the northernmost trunk channels where it continues out of the sidescan image (Figure 12d). Gullies and channels have higher backscatter returns than adjacent areas, suggesting a concentration of reflective materials in the channels compared to the surrounding sediments. Individual gully channels may be as narrow as <1 km, while broad canyon systems containing multiple gully channels are up to >100 km wide. Over 430 km of the returning east-west ABM survey path a total of 59 individual gullies are mapped, or one gully every 7 km. Continental slopes underlying the gully systems are steep: 10˚ between 500-600 mwd and 13˚ between 600-700 mwd.

2.3 Discussion

The three bedform types mapped on the ABM show a remarkable resemblance to glaciogenic features found on continental margins and oceanic plateaus in the Arctic and Antarctic [e.g., Davies et al., 1997; Anderson, 1999; Shipp et al., 1999; Polyak et al., 2001]. Chaotically oriented to sub-parallel scours and grooves (types 1 and 2; Figure 12a, b) are similar to iceberg plowmarks documented in regions of modern and past glaciation where more linear grooves result from persistent effects of winds or oceanic currents on icebergs [e.g., Barnes and Lien, 1988; Vogt et al., 1994]. Parallel flutes (type 3; Figure 12c) have been described in numerous glaciated regions and are interpreted to be sub-glacially generated flutes formed by an ice sheet or shelf sliding over the seafloor [Josenhans and Zevenhuizen, 1990; Polyak et al., 1997; Anderson, 1999; Shipp et al., 1999]. Many studies document a consistent pattern in the bathymetric distribution of glaciogenic bedforms: flutes are detected in the deepest parts of glacially eroded areas but are truncated and obliterated by iceberg scours in the shallows, while basinward continental slopes show mass-wasting features, with mudflows and turbidite gullies dominating low-angle and steep slopes, respectively [Shipp et al., 1999; Polyak et al., 2001; Ó Cofaigh et al., 2003; Dowdeswell et al., 2004]. In perfect agreement with the associations described in other studies, the ABM has iceberg scours at the shallowest depths in the study area

(<500 m), parallel flutes further downslope (400-700 m), and gullies originating at the lower edge of the glacially-disturbed areas.

Though the elongation and sedimentary sub-glacial character of the ABM flutes is adequate for them to be classified as mega-scale glacial lineations (MSGLs) [Clark, 1993], and their total areal extent places them in the size range of documented paleo-ice streams, I lack several lines of other sedimentary evidence to unequivocally classify them as being derived from an ice stream vs. an ice shelf. In many instances, however, ice shelves are the marine continuation of land-based ice streams exiting a confining trough, so many of the Stokes and Clark's [2001] ice stream indicators may also be useful for studying ice shelf configuration.

A peculiarity of the ABM is that the bedforms interpreted to be sub-glacially generated flutes formed by the passage of grounded ice (type 3 bedforms) parallel slope isobaths (Figures 10 and 12c) rather than traversing the continental shelf as in other glaciated areas. Similar-looking margin-parallel lineations on the Greenland continental slope are produced by contour currents [Kuijpers et al., 2002]. Oceanographic data from the ABM show evidence for a seasonal subsurface boundary current [Pickart, 2004], but it occurs at depths of 100-150 m, much shallower than the observed ABM flutes. The existence of deeper margin-parallel currents would have to be related to the formation and transit of deep waters, such as along the Greenland margin, and is not observed in the Canada Basin today. It is even less likely that deep currents would have formed during glacial periods when the closure of the Bering Strait and CAA channels further restricted circulation in the Arctic Ocean [Goosse et al., 1997a, b]. The absence of an oceanographic current capable of producing the linear bedforms on the ABM makes it unlikely that they are contourites. In addition, the characteristic association of sub-glacially generated flutes with iceberg scours above and turbidite gullies below is strong evidence for their glaciogenic origin.

Flute orientation (Figures 7 and 10) indicates that a cohesive, grounded ice mass moved parallel to the ABM at depths reaching ~700 m, rather than downhill across the continental slope as in other glaciated areas. As this ice mass slid across the erodible sedimentary slope, it might have carved the bathymetric bench that is coincident with the flutes and formed a 'second shelf break' along its lower edge (Figures 11a, b, and 13). Similar abnormally deep continental shelf breaks occur only on glaciated margins of the Arctic Ocean (Barents Sea and the Greenland/Canadian margin) [Jakobsson, 2002] and in Antarctica [Anderson, 1999]. The observed flute orientation along the ABM could indicate either west or east ice flow, but the glacial geology of adjacent regions suggests the ice source was the northwestern LIS margin. Fields of flutes with a similar orientation on the CB (Figures 3 and 6) show direct evidence for an east-west ice flow path in the form of blocks entrained in ice that left grooves in their wake as they were dragged across the seafloor (Figure 6). The broad, over-deepened inter-island troughs of the Amundsen Gulf and M'Clure Strait as well as the Mackenzie trough were likely outlets for large ice streams extending into the Canada Basin (Figure 3) [Blasco et al., 1990; Dyke et al., 1992; Stokes and Clark, 2001].

As ice exited the confining troughs of the CAA to enter the Canada basin, a floating ice cap filling the basin might have deflected ice flowlines parallel to the ABM [Grosswald and Hughes, 1999]. Alternatively, thickened glacial sea ice might have prevented ice streams from penetrating into the central basin [Polyak et al., 2001]. Propagating along the inferred path of least resistance parallel to the margin, this ice stream would have overrun the CB northwestwards, matching the orientation of flutes mapped in that region (Figures 3 and 6)

[Polyak et al., 2001]. The water depths of flutes on the ABM and the CB are similar, reaching almost 700 m and between 700-750 m, respectively. Assuming a glacial sea level lowering of 100-150 m [Rohling et al., 1998] in the Arctic Ocean, these depths of glaciogenic bedforms indicate ice thickness of 600-700 m. However, if ice shelves remained grounded long enough to continue to grow *in situ*, their total thickness could be greater than 600-700 m as they would no longer be floating in equilibrium with just 1/8-1/10 of their mass above water.

Iceberg scours were likely generated during deglaciation and the detachment of grounded ice from the seafloor, and possibly also during younger glacial event(s). The two types of scours/grooves observed suggest the possibility of two generations of iceberg activity, one during the collapse of an ice shelf that eroded the ABM and the other during a later episode of LIS expansion. Both types have a similar azimuth range (95-100˚ from north) (Figure 7) indicating that icebergs drifted clockwise along the ABM and impacted the slope at an oblique angle. This drift pattern is consistent with the distribution of IRD in Pleistocene sediments of the Canada Basin that were carried by the Beaufort Gyre [Phillips and Grantz, 2001] (Figure 3).

The age of maximal erosion on the CB has been estimated between early Wisconsinian to Illinoian, or OIS 4 to 6 [Polyak et al., 2003]. Because the orientation of glaciogenic flutes on the CB and ABM align spatially and have the same depth range (Figures 3, 6, and 12) I hypothesize that they were formed by the same glacial event. If a causal link exists between ice shelf encroachment on the ABM and coastal deposits with high contents of CAA clast material, I can narrow this age range using an estimate of 70-80 ka (OIS 5a) for the Flaxman Formation that mantles the coastline [Dinter et al. 1990]. Grosswald and Hughes [1999] hypothesize that the Flaxman Formation is a glacial till formed by the lateral margin of a transiting ice shelf, but geological data suggest a glaciomarine rather than sub-glacial origin for this deposit [Dinter et al., 1990]. I infer that ice shelf grounding on the continental shelf/slope isostatically depressed the ABM allowing for a glaciomarine transgression during ice shelf break-up (Figure 14). The glaciomarine transgression of the coastline deposited IRD in regions that are now up to 7 m above sea level due to post-glacial isostatic rebound effects. This interpretation is consistent with the lack of any mapped isostatic tilt in the Flaxman Formation deposits, which would indicate a local source of isostatic loading at the ABM, rather than a distal source such as the previously-invoked CAA center of glacial ice [Brigham-Grette et al., 2001]. Accordingly, the ice shelf grounding would have occurred during OIS 5b, the age of an inferred widespread glaciation in the Beringia region [Brigham-Grette et al., 2001]. A buried hummocky formation on the outer ABM shelf, tentatively interpreted as a barrier island [Dinter, 1985], might instead be a moraine that marks the landward limit of ice shelf grounding on the slope. A dedicated collection of high-resolution seismic reflection records and sediment cores is needed to delineate the boundaries of this grounding and to better establish its stratigraphic position.

2.4 Conclusions

The ABM is characterized by a repetitive association of glaciogenic seafloor bedforms and glacially sculpted bathymetry in water depths reaching 700 m, with ice-marginal turbidite gullies further downslope. At 400-700 mwd, broad fields of sub-glacially generated flutes cover a minimum seafloor area of 330 km^2, indicating widespread margin-parallel flow of a grounded, cohesive ice shelf or stream. Iceberg scours and grooves observed on the ABM in <500 mwd were probably generated during ice sheet break-up as well as later glacial event(s). Seafloor bedforms occur on the surface of a bathymetric bench that may have been formed by a large ice mass; the bench is truncated to the north by steep dendritic gullies produced by turbidite mass-wasting. ABM glaciogenic bedforms have depths, styles, and orientations similar to features discovered earlier on the CB and were probably formed during the same glacial episode(s). ABM bedforms link CB glaciogenic features and the over-deepened glacial outlet troughs of the CAA, and may have formed during an inferred widespread Beringian glaciation at OIS 5b. These findings provide compelling evidence for the provenance of the most recent major ice shelf transgression into the Arctic Basin and are critical for reconstructing Pleistocene ice history in the Arctic Ocean.

CHAPTER 2: Ice-proximal gullies

3.1 Introduction

It has been argued that a systematic relationship exists between glacial processes and continental slope morphology on high-latitude margins [Ó Cofaigh et al., 2002; O'Grady and Syvitski, 2002; Ó Cofaigh et al., 2003]. Glacial ice extending across continental shelves delivers large volumes of sub-glacial sediment to the upper continental slope [Alley et al., 1989] while some sediments remain entrained in the ice until it melts. In a widely accepted model, sub-glacial sediments may be remobilized down the continental slope by mass-flow processes, commonly debris flows, resulting in the development of broad, gently-sloping, fan-shaped sediment accumulations termed trough mouth fans (TMFs) [Ó Cofaigh et al., 2003]. TMFs have therefore been used in many studies to recognize the locations of former ice streams [e.g. Dowdeswell et al., 1996; Vorren and Laberg, 1997; Dowdeswell and Siegert, 1999]. However, as pointed out by Ó Cofaigh et al. [2003], several key questions remain concerning the nature of continental slope sedimentation at ice stream termini: (1) How variable is the relationship between ice streams and continental slope sedimentation deposits? I.e. do ice streams terminating at the shelf break always form TMFs? (2) What are the physical controls on the sedimentary architecture of continental slopes at ice stream termini? (3) How much variability is there in sediment delivery style to TMFs? I.e. do debris flows dominate?

In answer to the questions posed above, there are at least three well-documented examples of known glaciated continental margins that lack TMFs, but that instead have other recognizable and characteristic features. For example: (1) in the Ross Sea of Antarctica ice streams drained to the shelf edge during the LGM as inferred from the modern presence of glacially eroded troughs on the continental shelf. In this locality, however, there is no evidence for the traditional TMF but rather an extensive network of steep turbidite gullies that initiates at the shelf edge and rapidly deepens and coalesces to form larger channels [Ó Cofaigh et al., 2003]. (2) Similarly, in Marguerite Bay on the Antarctic Peninsula, streamlined sub-glacial bedforms in a cross-shelf trough indicate the former presence of a major ice stream draining to the continental slope. At the inferred location of the ice stream terminus the slope has an angle of 9˚ and is characterized by shallow dendritic gullies, while slopes to either side of the ice stream terminus are 10-12˚ and show deeper erosional gullies, but no TMF in either location. Ó Cofaigh et al. [2003] conclude that sediment delivery to the trough mouth in Marguerite Bay was moderate causing progradation of sediments that blanketed and masked erosional slope gullies and resulted in shallower (but still quite steep) slopes. To either side of the ice stream mouth sediment delivery was too low to mask gullies. As a result, erosional processes from turbidity flows sourced by downward-flowing sub-glacial meltwater along the ice edge dominated, creating steeper slopes and deeper turbidite gullies. (3) A third example from the East Greenland margin where LGM ice sheet sediment delivery to the shelf was inferred to be low also shows a network of gullies initiating on the upper continental slope and extending all the way to the abyssal plain, instead of the traditionally-recognized TMF. It is hypothesized that the East Greenland gullies formed in response to turbidity currents and cascading dense cold water formed by brine rejection during sea-ice formation along the shelf [Ó Cofaigh et al., 2002]. With reference to the questions from the first paragraph, the examples cited above give evidence that on formerly glaciated continental margins ice streams terminating at a shelf break do not always form TMFs. Moreover, in each of the localities described above, low glacial sediment delivery to the shelf break is inferred to control the sedimentary architecture of the margins. This low sediment delivery results in steep continental slopes and a predominance of characteristic turbidite gullies (instead of a TMF) that could not have been produced by debris flows.

A further refinement to the original TMF model comes from a recent study of the continental slope architecture of 66 regions around the Arctic Ocean [O'Grady and Syvitski, 2002]. O'Grady and Syvitski [2002] note that high sediment delivery from trough-mouth systems has been documented to lead to the formation of low-angle, debris-flow dominated fans [Elverhøi et al., 1998], while sediment-starved regions may be more prone to steep slopes and canyon development such as on the Norwegian Andøya margin [Taylor et al., 2000]. Using previously published studies of circum-Arctic ice extent and configuration, they attempt to determine whether the former ice regime along a given margin can be used to predict continental slope morphology. They compare the modern continental slope morphology of margins interpreted to have been impacted by primarily convergent, divergent, or no ice flow. IBCAO bathymetry [Jakobsson, 2000a] is used to characterize large-scale slope morphology, slope angle, and bathymetric roughness. Though the IBCAO grid has relatively low resolution (grid cell size 2.5 km), it is the only bathymetry grid that covers the entire Arctic Ocean at the same resolution and therefore serves as a useful comparative tool for their study. O'Grady and Syvitski [2002] conclude that the inferred method of glacial sediment delivery to the shelf edge and continental slope (via convergent or divergent ice or non-ice means) is very important to the resulting continental slope architecture. The specific method of glacial sediment delivery is key to determining the width and depth of the continental shelf, as well as the angle, shape, surface morphology, and number of canyons on the continental slope. They find that the mechanism of sediment delivery also influences the rate and patterns of sediment accumulation on the slope, the types of sediment dispersal that are present (e.g. buoyant plumes, icebergs, basal transport), and the origin of down-slope processes such as slope failures, debris flows, slumps, slides, and turbidity currents [O'Grady and Syvitski, 2002].

The ABM is highlighted in the O'Grady and Syvitski [2002] study but is incorrectly characterized as a non-glaciated, alpine-river dominated margin based on the work of Garrison and Becker [1976] and Dyke and Prest [1987]. They acknowledge that some authors infer a glaciated history for the region [Hughes and Hughes, 1994], but do not reference important previous work such as Grantz et al. [1979] and Polyak et al. [2001] that supports the possibility of thick ice on the margin. They therefore conclude that the rugged surface texture observed along the slope even in the low-resolution IBCAO bathymetry is related to slope failure from decomposing gas hydrates [Kayen and Lee, 1991]. My study uses high-resolution sidescan sonar and single beam bathymetry to refute O'Grady and Syvitski's [2002] initial assumption that the ABM was never glaciated. As a result, their findings with respect to the sedimentary architecture of the ABM are thrown into question, though their methodology may be appropriate for other Arctic margins that are better studied. At the same time, I characterize in detail a formerly glaciated continental slope that lacks a TMF, thereby broadening our knowledge of the range of glacially impacted continental margin configurations and contributing to our understanding of ice-marginal processes.

3.2 Methods

Bathymetric character and surface roughness along the ABM are evaluated using a combination of 12 kHz sidescan sonar imagery and narrow-beam bathymetric data collected by the submarine *USS Hawkbill*, as well as 2.5 km grid cell resolution IBCAO bathymetric contours. These datasets are combined in a GIS using the ESRI software ArcMap to co-register datasets and generate average slope profiles along the margin. A margin-parallel transect taken by the submarine on its returning east-west survey track followed the 200 fathom (366 m) contour (Figure 15). This transect is used to characterize gullies along the slope including their frequency, distribution, style, depth and width, depth of onset, and average distance from the shallowest part of the survey as represented by the 250 m contour. In one locality it is possible to observe almost the full length of an individual gully system (Figure 12d) and this example is used to infer gully dynamics and configuration elsewhere.

3.3 Results

The most striking feature of the ABM as observed in both the 12 kHz sidescan sonar and the narrow-beam bathymetry data is undoubtedly the abundance of dendritic gully systems along the whole length of the margin in water depths >~500 m (Figures 12a-d). Within the 430 km of the ABM covered by the returning east-west submarine track (Figure 15) there is on average one gully every 7 km (59 total). Gully systems are easily recognizable in sidescan sonar imagery due to the marked contrast in acoustic reflectivity between the bright gully channels and the surrounding dark sedimentary substrate (Figure 12d). This contrast in acoustic reflectivity suggests that the gully channels are filled with highly reflective hard materials such as coarse gravel lag deposits, and that finer sedimentary materials have been winnowed away or distributed further downstream. Gully channel depths measured vertically range from 80-770 m (average 307 m), and gully widths range from 0.8-18.1 km (average 5.5 km) (Figure 16).

Seventy-eight percent of the gullies initiate in water depths <700 m and 46% of the gullies initiate shallower than 600 mwd. Gullies are deep, with average width:depth ratios of 15:1, and occur in steep terrain, with average continental slope angles of 10˚ from 500-600 m and 13˚ from 600-700 m. The gully system whose length is mapped in Figure 12d measures in excess of 40 km long and covers depths ranging from 250->1500 mwd before continuing out of the sidescan swath image. The gully system in Figure 12d shows the dendritic catchment areas typical of the upper (<1000 mwd) parts of the gully systems (here ranging in width from 0.7-10 km). The thalwegs of the gullies in the dendritic catchment system then coalesce multiple times to form broad trunk channels in deeper water depths from ~1000-1500 m.

In an effort to better understand the physical controls on gully formation I analyzed their morphologies as well as their relationship to topographic features and landmarks. Gully walls generally are the same height on their east and west sides, with no trend of asymmetry with distance along the margin that would indicate preferential erosion due to factors such as different thicknesses of eroding ice or irregular currents. A striking attribute of the gully transect shown in Figure 16 is the flattened character of the shallowest portions of the margin substrate compared to the gullied terrain. At ~400-600 mwd the seafloor appears horizontally planed off with gullies downcutting through and dissecting this surface. This upper, flattened surface corresponds to the glacially eroded bathymetric bench described in Chapter 1 and has very low slope angles of only 1.6˚ as measured from 19 slope profiles (Figures 11a and 13). As seen in Figures 16 and 13, gully density is highest where the bathymetric bench is narrowest at ~148˚ W, and then decreases to the west. In addition, the character of the gullies changes from predominantly shallow, narrow gullies where the bathymetric bench is narrowest, to wider and deeper gullies and gullied canyons at the westernmost end of the survey where the bathymetric bench widens. Multiple individual gullies can be grouped into larger canyons where the walls of several successive gullies do not reach shallow enough depths to intersect the flattened bench surface (Figure 16). A prominent example of a gullied canyon is the ensemble of gullies indicated in Figures 16 and 17 that initiates on average 18 km north of the 250 m contour. Gullies initiate anywhere from 0.3-21.6 km (average 5.8 km) north and downslope of the shallowest depths measured in the survey area as represented by the 250 m contour. When the horizontal distance of canyon initiation away from the 250 m contour increases, the depth of canyon initiation deepens, and this is reflected statistically (at the 95% significance level, canyon initiation depth is highly correlated with average distance from the 250 m contour, $r = .81$). This provides confirmation that slope angles along the margin are relatively constant within the depth range of canyon initiation, as has also been measured from slope profile transects. Canyon depth and width increase nearly linearly with respect to each other (at the 95% significance level $r = .89$).

3.4 Discussion

The co-occurrence of the gullies described above with the glacially eroded 'second shelf break' of Chapter 1 (Figures 11a-b and 13), and their morphologic and contextual similarity to gully systems documented on glaciated continental margins around Antarctica (Figure 18) [Ó Cofaigh et al., 2003], East Greenland [Ó Cofaigh et al., 2002], the Gulf of Alaska [Carlson et al., 1990], and the Barents Sea [Vorren et al., 1988], are strong evidence for their glaciogenic origin. In each of the localities referenced above, ice is postulated to have extended to the continental shelf edge during glacial periods, releasing sub-glacially entrained sediments where the ice front separated from the seafloor, calved and broke apart [Vorren et al., 1988; Carlson et al., 1990; Ó Cofaigh et al., 2002; Ó Cofaigh et al., 2003]. Sediments were subsequently remobilized by downslope gravity flows, but the volume of sediments was too low in any of the above locations to form TMFs, and instead turbidite gullies were eroded into the upper continental slope. There is consensus that gullies are erosional features created during periods when sediment availability is too low to build TMFs, but authors differ as to their ultimate formation mechanism. Vorren et al. [1988] speculate that gullies in the Barents Sea formed in response to the interglacial wintertime sinking of cold, dense shelf water that winnowed sediments and formed concentrated sand and gravel lag deposits, similar to the highly reflective materials mapped in thalwegs along the ABM (Figure 12d). Carlson et al., [1990] infer that gullies in parts of the Gulf of Alaska were eroded by sediment gravity flows during Pleistocene lowstands of sea level, but that in other areas, for example at the mouths of cross-shelf ice streams, sediment delivery was high enough to bury slope gullies (but still not high enough to form a TMF). Ó Cofaigh et al. [2003] postulate that the gully formation characteristic of some glaciated continental slopes is governed by an interplay between three factors: (1) slope gradient and associated tectonic history of the margin, (2) geology of the adjoining continental shelf, and (3) the availability of sediment-laden sub-glacial meltwater for turbidity current generation.

I have used Ó Cofaigh et al.'s [2003] three gully-forming criteria above to analyze the ABM continental slope gullies. (1) Slope gradients along the ABM are higher than in any of the other locations described in the previous section, up to 13˚ between 600-700 mwd. Ó Cofaigh et al. [2003] argue that where slope gradients are sufficiently steep, the sub-glacial sediments available at the shelf edge may be transported directly to the abyssal plain before being re-deposited, bypassing the slope altogether. This results in a sediment-starved upper slope environment conducive to the formation of erosional gullies, regardless of the total amount of sediment available from glacial activity. Thus even though the ABM is known to be mantled with up to 2700 m of unconsolidated sediments [Grantz et al., 1979; Dinter et al., 1990; Kayen and Lee, 1991; Brigham-Grette et al., 2001], the continental slope in this area may simply be too steep to retain any sediments deposited off the shelf edge by ice. (2) The geology of the paleo-ABM continental shelf at 400-550 m described in Chapter 1 shows evidence for the west-east flow of extensive, cohesive, grounded ice parallel to the margin (Figures 12a-d). ABM ice-molded bedforms are elongate and suggest the fast ice flow typical of an ice stream [Stokes and Clark, 2001]. However, the direction of ice flow is isobath-parallel instead of transecting the continental shelf perpendicular to the shelf edge as in the ice stream localities described above. As a result, it appears that ice probably did not deliver sediments directly to the shelf edge. Some sediments may have arrived at the shelf edge at an oblique angle where pre-existing topographic variability along the shelf/slope break caused the ice shelf to float over open water, but the majority of the sediments were being moved parallel to and along the slope in the direction of dominant ice flow. Thus most sub-glacially entrained sediments may not have been released along the ABM margin itself, thereby starving the upper continental slope of sediments. (3) As in any fast flowing, marine based glacial setting, it is likely that there was abundant sub-glacial meltwater available to lubricate the flow interface of the ABM ice with its sedimentary substrate [Tulaczyk, 2001]. Dense, sediment-laden meltwater would quickly migrate to the areas of lowest pressure at the shelf break, flowing down the steep ABM slope and triggering erosive turbidites as it went. The depositional products of turbidity currents, known as Bouma sequences, are remarkable in the sedimentary record for their well-sorted deposits. The heaviest gravels and large clasts can only travel in the most competent and fast-moving core of the flow, while finer materials are distributed where the flow slows down: in overbank locations, at the distal ends of the flow, or in response to changes in slope. These criteria match well the sharp contrast in the distribution of reflective materials mapped in sidescan sonar imagery of the ABM (Figure 12d). The brightest acoustic returns are concentrated in the centers of gully channels and in the uppermost sections of the gully systems, consistent with the classic bedload distribution of a turbidite flow. The above discussion demonstrates that each of the three gully-forming criteria delineated by Ó Cofaigh et al. [2003] is maximized along the ABM to produce dramatic, ice-proximal, turbidite-derived gullies. Despite the glaciated history of the coastline, there is no evidence for a TMF, and instead continental slope architecture appears to be controlled by a low-sediment environment that results in steep slopes and a heavily dissected margin.

In their study of Arctic continental slope morphology, O'Grady and Syvitski [2002] attribute the rugged surface texture of the ABM as imaged by 2.5 km grid cell IBCAO bathymetry to massive slope failure caused by decomposing gas hydrates [Kayen and Lee, 1991]. Grantz et al., [1982; 1990] and Grantz and Dinter [1980] describe a 500 km long belt of very large-scale seafloor slumps and block glides 100-150 m thick that moved by shearing along their bases. The landward edge of this continental slope landslide belt corresponds with a gas hydrate zone imaged in geophysical records in water depths of 200-400 m [Kayen and Lee, 1991]. Kayen and Lee infer that decomposition of gas hydrates in response to Pleistocene sea level changes was sufficient to produce the belt of observed landslides. Both slump blocks and turbidite gullies co-exist along the continental slope. The resolution of the IBCAO bathymetry captures the topography of the landslide belt, but also images the largest of the mass wasting gullies. These two sets of features have very different morphologic character and should be difficult to confuse. The turbidite gullies provide compelling evidence of ice-marginal mass wasting processes that in an of themselves are an indicator of the formerly glaciated character of the margin, particularly when coupled with earlier reports of glaciogenic bedforms in the area [Grantz et al., 1979; Polyak et al., 2001]. However, despite the fact that O'Grady and Syvitski [2002] have incorrectly characterized the glacial history of this portion of the Arctic margin, their methodology in other areas may still prove useful to understanding the ABM ice regime.

In their study O'Grady and Syvitski [2002] characterize margins as having been impacted by fast-flowing convergent ice carrying high sediment loads, slower-moving divergent ice with lower sediment loads, or no ice. It is likely that ice on the ABM was fast-flowing (based on a predominance of highly attenuated bedforms), but in this location ice actually flowed parallel to the trend of the margin, thereby making the ice flow regime difficult to characterize in the context of their study (or in any comparative study; there is in fact no other location documented in the literature where ice of this scale is postulated to flow parallel to a margin). However, the defining features of the ABM margin according to the scheme of O'Grady and Syvitski [2002] include steep slopes, a narrow-medium width shelf, high numbers of canyons, and a shallow shelf break. This is a very close match to the defining characteristics of the Norwegian Andøya margin, an area that was impacted by slower-moving, divergent ice flow [Taylor et al., 2000]. Both the Andøya margin and the ABM have steep slopes, narrow continental shelves, and high numbers of canyons. The Andøya margin has a moderately deep continental shelf while the ABM shelf is classified as shallow, but if the ABM paleo-shelf at 400-550 m is used instead of the modern interglacial shelf, the shelf depths on the Andøya and ABM margins are analogous. It is likely that the sediment load available for deposition in each of these locations was also comparable. Along the Andøya margin, sediment transport is thought to have been low, both because of the slow-moving character of the ice, and also due to the lack of a source area

[Dowdeswell et al., 1996]. As described in the previous sections, sediment availability on the ABM was probably also quite low due to a combination of very steep slopes and the fact that ice was transporting sediments parallel to the margin rather than depositing them at the shelf edge. As a result, neither the Andøya margin nor the ABM shows evidence for a TMF, both areas are dominated by erosive processes that result in extensive canyoning and dissection of the continental slope, and yet both areas are known to have been glaciated. The Arctic-wide classification scheme of O'Grady and Syvitski [2002] may be too great an oversimplification of a complex series of processes and perhaps is not appropriate to the detailed study of any one margin. However, the concepts discussed provide a useful comparison of two formerly glaciated margins that improves our understanding of continental slope architecture along high latitude margins in areas where sediment availability is low.

3.5 Conclusions

The ABM is distinguished by rugged seafloor bathymetry north of and deeper than the 500 m contour along its entire length. The seafloor texture is created in part by a series of deep, steep dendritic gully systems that start at the paleo-shelf break and truncate the flattened, glacially eroded surface to the south. Gully systems initiate in branching catchment areas up to 18 km across that coalesce multiple times to form broad trunk channels in water depths >1500 m. Gullies are interpreted to have formed in response to sediment-laden sub-glacial meltwater cascading down the very steep continental slope and triggering turbidity currents that eroded deep channels on the slope before being re-deposited on the abyssal plain. Available glacial sediments therefore bypassed the upper continental slope on this margin and did not form a

TMF, despite strong evidence of both streaming ice and thick sediments on the shelf. The isobath-parallel trend of the ice flow in this location probably resulted in the majority of the sub-glacial sediments being deposited west of the ABM, contributing to the sediment-starved character of the margin. The similarity of the dissected ABM continental slope to formerly glaciated continental slopes at both poles, combined with evidence of glaciogenic bedforms along the paleo-shelf surface, are strong evidence for the glaciogenic origin of the gully systems draining the shelf.

CHAPTER 3: Bathymetry interpretation

4.1 Introduction

The Seafloor Characterization and Mapping Pod (SCAMP) system mounted on the submarine *USS Hawkbill* during the 1999 SCICEX cruise was equipped with a 12 kHz swath mapping sonar [Edwards and Coakley, 2003] that collected continuous seafloor depth and backscatter information along the length of the ABM. An acoustic pulse transmitted by the sonar array rebounds from the seafloor surface and is detected on its return by the SCAMP system. The returning acoustic information is interpreted differently for sidescan sonar than for bathymetric processing [Blackinton, 1986]. Sidescan processing is based on the strength of the acoustic return; bathymetry is based on the time delay between when parts of the same acoustic waveform are detected at each of two separate rows of transducers on the port and starboard sides of the survey instrument. For sidescan processing the strength of the acoustic return provides information on four combined factors: substrate type, seafloor slope, acoustic look-angle, and water column properties. Sidescan gives a good indication of fine-scale seafloor texture because substrate type is the most important of the factors contributing to the returning acoustic signal. Differing return signal strengths can therefore primarily be interpreted to identify locations of seafloor with varying acoustic impedance contrast. For example, fresh lava flows with high acoustic impedance return a very strong signal, while easily-penetrable soft sediments return a more attenuated signal. If the highest acoustic impedance barrier occurs below the actual seafloor surface, for example where a lava flow has been covered by sediments, sidescan sonar processing may still identify the lava flow as the dominant acoustic return depending on the thickness of the sediments. In contrast, bathymetric processing attempts to identify not the most powerful but the first recognizable seafloor return signature. Bathymetry processing very accurately identifies how far away in time (not angle/distance) the closest seafloor is. This procedure is automated by telling the system to identify the first acoustic return that is some set percentage of the most powerful acoustic return (e.g. 15%) as the seafloor surface. Based on how far away the seafloor is (and it is assumed to be directly under the sonar), the phase difference at the two arrays on each side of the acoustic instrument is used to compute bathymetry.

As a result of the differing processing techniques described above, sidescan and bathymetric data for a given area yield different yet complimentary seafloor information. For example, along the ABM, bathymetric data highlight old glaciogenic depressions or hills that have been covered with sediments deep enough to mask any sidescan signature of the feature. In contrast, sidescan processing highlights glaciogenic features whose seafloor amplitude is too shallow to be detected at the resolution of the bathymetry, but whose areal extent and impedance contrast are sufficient for them to be detected in sidescan imagery. Seafloor erosional glacial features formed by large ice masses are typically characterized by an overcompacted diamict layer at their base formed by the grinding, sliding pressure of the overlying ice mass [Anderson, 1999]. This stiff sediment layer is readily identifiable in sidescan sonar images due to its high impedance relative to the surrounding soft sediments, even when covered by a few centimeters of post-erosional pelagic sediments. Because bathymetry and sidescan data for the same area highlight different aspects of the same glacial terrain, I have processed the ABM swath bathymetry data in addition to the sidescan data to give greater insight into the genesis, distribution, and flow pattern of ice masses that impacted the margin. Additionally, swath bathymetric data yield information on the rugged character of the ABM turbidite gully systems which may impact boundary current flow and eddy generation along the length of the survey area.

4.2 Methods

Bathymetry processing is a multi-stage endeavor wherein the processor's judgment and geologic knowledge may significantly impact the final data product. All seafloor data acoustic returns are hand-edited to produce the best geologic match of seafloor depths to features mapped in the corresponding sidescan imagery. The bathymetric data are next gridded together taking into account location-specific water column information such as temperature and salinity that may affect the speed of the acoustic pulse and thereby the two-way travel time used to calculate seafloor depths. Gridded data are then mosaicked to form a map image of the area of interest at varying resolutions. Maximum bathymetry resolution with the SCAMP system is typically ~2% of water depth (>~5 m) along the ABM [Edwards, M., pers. comm., 2004]. It is possible create images at the full resolution of the data to zoom in on features of interest (Figure 19). The full resolution imagery captures well the signature of glaciogenic features in both the sidescan and the bathymetry data (Figure 19), but it is impractical to visualize the bathymetry at this resolution over large areas. As such, bathymetry data is most often displayed at less than the full resolution but covering larger seafloor areas, and this has the added benefit of averaging bathymetry values into coarser grid cells, which aids with noise reduction. The ABM seafloor covers too large an area to examine in detail at the full resolution of the bathymetry over its entirety, but lower resolution compilations smooth glacial features of interest and make it impossible to detect all but the largest glacial bedforms in the bathymetry (Figure 20). Thus, visual interpretation alone of bathymetric data at any resolution is inefficient and may result in data misinterpretations for a variety of reasons. As a result, depending on the skill of the processor, the display method, and the resolution of the maps being examined, visual interpretations of seafloor character may differ from person to person.

In order to overcome the practical constraints of visual bathymetric interpretation described above, and to quantify the seafloor bathymetric signature over large areas, I characterize the distribution and magnitude of the glacially-generated seafloor bedforms mapped in the sidescan imagery using spectral analysis techniques. The Fourier transform of any digital signal is the decomposition of the signal into sine and cosine waves of differing frequencies and amplitudes that when summed replicate the original signal. When correctly interpreted, these frequencies and amplitudes represent the wavelengths and strengths of repeating signals in the data, which in my study area may correspond to the spacing and severity of excavation of glaciogenic features on the seafloor. I perform a one-dimensional analysis, which means I use transects of the highest resolution seafloor data, rather than swaths of gridded seafloor data that would entail a more complex two-dimensional analysis. The raw bathymetric data (Figure 21) underwent several iterations in processing in order to prepare them for this analysis (Table 3).

Table 3 - Spectral analysis processing sequence

1. Isolate bottom detects in bathymetry data
2. Choose sample locations in areas where glacial features exist in the sidescan imagery
3. Choose samples where submarine speed was constant and away from turns
4. Project submarine track perpendicular to ice flow direction as inferred from sidescan data
5. Interpolate and resample data so that sample spacing and density is constant between areas
6. Remove slope trend and DC bias of seafloor depths
7. Filter data with high pass “boxcar” filter to remove long wavelength noise
8. Take fast Fourier transform of data in Microsoft Excel
9. Make power spectra to highlight wavelengths of interest
10. Scale amplitudes by sample size and plot as spectral estimates in dB

The highest resolution bathymetry were used for this analysis, of which 20 samples were chosen in areas where I observe glacial features in the sidescan imagery, and where the submarine was moving in a single direction at a constant speed since data quality may deteriorate around turns (Figure 22). The submarine track never crossed glaciogenic bedforms perpendicular to the direction of ice flow as inferred from sidescan imagery, resulting in erroneously large apparent spacing between features. I corrected the spacing by projecting each submarine track perpendicular to the trend of ice flow, and then resampling the data to generate consistent data densities for each sample area (Figure 23). The broad range of orientations of iceberg scours and grooves in areas A, B, and C make it impossible to reliably sample the bathymetry perpendicular to a single inferred direction of ice flow. As such, seafloor transects in areas A, B, and C were projected using the systematic relationship between glacial flute trends in sidescan imagery and the trend of bathymetric landmarks identified in areas E, G, and F (Figure 10). To avoid introducing systematic end artifacts into the data with sample replication, I removed the slope trend and direct current bias for each sample area. After testing multiple filter lengths, the projected, resampled, and detrended data were filtered using a high pass "boxcar" filter to remove long wavelength noise. These data were then subjected to a Fast Fourier Transform (FFT) [Smith, 2003] analysis in Microsoft Excel. The FFT output was plotted as power spectra to highlight the wavelengths of interest (Figure 24). I then produced spectral plots showing the amplitudes of seafloor irregularities at given wavelengths in decibels (dB) for each area (Figure 25). The sums of the amplitudes shown in the spectral plots are proxies for the total variance in seafloor height of a given area.

4.3 Results

The results of spectral analysis for 20 samples selected along the length of the survey area based on the criteria listed in Table 1 are shown in Figure 25, with their locations shown in Figure 22. In my study area, spectral analysis is most useful for determining the distance between repeated seafloor irregularities of the same amplitude. The amplitude of spectral estimates is an indication of the relative seafloor roughness at each repeated interval (wavelength). In Figure 25, colored amplitudes indicate the scale of roughness of the seafloor at the wavelengths indicated in meters on the y-axis. For example, in sample area C3, there is high amplitude roughness of the seafloor with repeated spacing of ~45 m perpendicular to the sample transect and parallel to the direction of ice flow, while in sample G2 the highest amplitude seafloor roughness within that area (but lower amplitude than in area C3) occurs with spacing of ~60 m. The apparent spacing between glacial flutes as measured from sidescan sonar maps is ~50 m. Sidescan measurements produce minimum values since acoustic imagery is projected onto a flat surface for display, though the smaller the spacing between features, the less the distortion has an effect. In Figure 25 the prominent bullseye pattern in the amplitudes centered around the 50 m wavelength that stretches the whole length of the margin matches the spacing of glacial flutes measured manually from the sidescan imagery (Chapter 2). Further information can be gained by looking at the relative strengths of the amplitude signals from east to west. In order from most to least cumulative roughness at the seafloor surface over the sample area are samples a2, b2a, c3, b2b, b1, h1b, g1, a1, g2, e1, h1a, h4c, h3c, h3b, h3a, h2a, h4b, h4a, h2b, and h1c. Samples clustered on the western end of the margin in areas A, B, and C (Figure 25) have the highest overall seafloor roughness, while samples on the eastern end of the survey in areas E, G, and H show more attenuated seafloor roughness. Depths in sample areas A, B, and C at the western end of the survey have a mean of 290 m with no depths >370 m, whereas depths in areas E, G, and H at the eastern end of the margin average 400 m with depths as great as 570 m.

Figure 10 shows the average strikes of glacial bedforms indicative of ice flow direction as measured from sidescan sonar imagery within each sample area, as well as the average trends of bathymetric landmarks such as the shelf/slope break for the same areas. In some locations it is possible to measure the trend of glaciogenic bedforms in the bathymetry, but over most of the study area larger-scale features such as the shelf/slope break are the most prominent morphologic landmarks. On a location-by-location basis the offset between the inferred ice flow direction and the trends of bathymetric features may be as great as 20˚, however the mean ice flow direction and bathymetry trend for all glaciated areas is nearly identical at 113˚ using the Rayleigh's test for trend significance at the 95% level [Swan and Sandilands, 1995]. If four areas where it is not possible to unequivocally determine the former ice flow direction from sidescan imagery are removed from the comparison (sample areas b1, b2, b2b, c3), the average trend of glacial bedforms as mapped from sidescan imagery is 114˚, and the average trend of bathymetric landmarks for the same locations is 113˚ (significant at the 95% level using Rayleigh's test for trend [Swan and Sandilands, 1995]).

4.4 Discussion

Bathymetric processing provides compelling corroboration of the impact of glacial ice on the ABM. At the broadest scales, swath bathymetric maps of glacially eroded areas known from sidescan imagery highlight the contours of the areally extensive, flattened bench or 'second shelf break' underlying glacial features the length of the margin (Figure 13), whose northern edge is truncated by deep turbidite gullies. The bench is widest where it underlies the best and most extensive examples of sub-glacially-generated fluted bedforms (Figures 12c and 13). It narrows to a minimum width near 148˚ W where the density of turbidite gullies is highest, then widens again to the west as it approaches Barrow Canyon (Figure 13). Rose diagrams comparing the trend of glacial flow directions inferred from sidescan maps with the trend of bathymetric landmarks across the surface of this bench (Figure 10) are so closely aligned as to suggest that the eroding ice mass actually carved the bench out of the continental slope. At full bathymetry resolution, imagery of matching bathymetry and sidescan swaths confirms the negative-relief character of the glaciogenic bedforms previously identified from sidescan maps (Figure 19). Glaciogenic features observed in sidescan imagery are therefore inferred in most instances to have been eroded into the substrate, certainly in areas where the seafloor was excavated by the descending keels of icebergs. In fluted terrain, it is possible that the weight of the overlying ice may have squeezed sediments up into levees where pre-existing basal roughness at the ice/substrate interface created void cavities, effectively molding the substrate rather than eroding it. Individual glaciogenic bedforms can measure >5 m deep as measured from swath bathymetry.

Spectral analysis contributes a quantitative and margin-wide elucidation of the impact of grounded ice on the continental slope. As seen in Figure 25, seafloor roughness the length of the margin has repeated spacing throughout of ~50 m. Closest to the inferred source of flowing ice and where I document the best examples of sub-glacial flutes on the eastern half of the survey, amplitudes of seafloor roughness in areas E, G and H are moderate and cover a narrow range of dominant wavelengths from ~45-100 m. On the westernmost half of the survey, samples from areas A, B, and C where iceberg scours and grooves erode the substrate show the strongest signals of seafloor excavation and a wider range of wavelengths from ~20-300 m. Spectral analysis appears to correctly identify this change in bedform character and inferred eroding agent over the 20 sample areas and along the length of the survey from east-west. In the same way that the bathymetric bench morphology may be the result of ice creating topography along the continental slope, the fact that measured seafloor roughness changes in concert with bedform type is more evidence that glaciogenic processes controlled the bathymetric signature of the seafloor on the ABM, at a variety of scales.

There are several factors that may contribute to the different character of seafloor roughness associated with each style of glacial bedform. It is probable that areas A, B, and C were impacted by the same cohesive ice mass that molded the seafloor in areas E, G, and H, as is indicated by the persistence of the bathymetric bench along the whole length of the margin (Figure 13). However, the chaotically oriented iceberg scours and grooves that characterize the western end of the margin are interpreted to post-date the passage of the larger ice shelf (see Chapter 1). When the ice shelf began to disintegrate into many large blocks, iceberg keels probably eroded and overprinted fluted bedforms in shallow water depths. As such, the younger groove and levee patterns of seafloor sediments at the western end of the survey would be crisper and deeper, not so attenuated or modified by subsequent deposition of sediments or the activity of currents as the older glacial flutes on the eastern end of the survey. In addition, iceberg keels have greater basal roughness than a large ice shelf, and would therefore be capable of creating deeper gouges in the sedimentary substrate than an ice shelf base. As an ice shelf transits across the seafloor over long distances, it is likely that any basal roughness will be eroded and equalized over broad areas to produce a more uniform ice base. The ice shelf base may mold or bulldoze the seafloor sediments, but would be unlikely to produce large, individual gouge and levee bedforms with high roughness amplitudes. Iceberg scours and grooves in sidescan imagery have more variable and less coherent spacing than the flutes observed in areas E, G and H, which would explain the wider range of dominant seafloor wavelengths on the western half of the survey detected with spectral analysis (Figure 25). It is also possible that the character of the sedimentary substrate changes fundamentally over the length of the survey. I am not able to evaluate the contribution of sediment type changes to seafloor roughness with my dataset, but the coincidence of the change in seafloor roughness with the change in glaciogenic bedform types suggests that erosion by ice was the dominant control on seafloor bathymetry along the ABM.

4.5 Conclusions

Visual and spectral analysis of swath bathymetric data for the ABM provides strong support for the paleo-ice flow interpretations of previous chapters. The coincident trends of inferred ice flow direction and bathymetric landmarks on the surface of the 'second shelf break' at 450-550 mwd support the hypothesis that a large ice mass carved the bench with its passage (Figure 10). Within the full resolution bathymetry data it is possible to discern the negative-relief signature typical of eroded glacial scours and grooves, while flutes may result from molding of seafloor sediments (Figure 19). A margin-wide one-dimensional spectral analysis of the highest resolution bathymetry data correctly identifies the transition from regularly-spaced, moderate amplitude fluted seafloor at the eastern end of the survey, to variably spaced and deeply grooved iceberg-eroded bathymetry at the western end of the survey (Figure 25). The change in seafloor roughness associated with different glaciogenic bedform styles is interpreted to result from differing erosive processes at the ice/substrate interface. I additionally infer that iceberg scours are younger and better preserved than the glacial flutes, which contributes to their seafloor roughness. This chapter demonstrates that spectral analysis may be an effective tool for quantifying seafloor bathymetric character over large distances. This has the potential to greatly streamline seafloor analysis methods in areas where it is not possible to visually examine all bathymetric data.

CONCLUSIONS: Implications and future work

5.1 Summary of findings

The ABM was surveyed using sidescan sonar and bathymetric mapping techniques over a region covering 640 km east of Point Barrow during the 1999 SCICEX cruise aboard the nuclear submarine *USS Hawkbill* (Figures 3 and 7). The margin is characterized throughout its length by a repetitive association of glaciogenic seafloor bedforms, glacially sculpted bathymetry, and ice-marginal turbidite gullies. In 400-700 mwd, broad fields of sub-glacially generated flutes >10 km long grouped in fields >30 km wide cover a minimum seafloor area of 330 km^2 and parallel bathymetric contours (Figure 12c). Margin-wide spectral analysis reveals that these flutes are regularly spaced every ~50 m and have moderate amplitudes ranging up to ~10 m deep based on slope profiles (Figure 25). Chaotically oriented iceberg scours and grooves detected in areas <500 m deep have variable spacing and severely turbate the seafloor (Figures 7a and b). Within swaths of full resolution bathymetry it is possible to discern the negative-relief, eroded signature of glaciogenic scours, and grooves, while flutes may form via molding of substrate sediments (Figure 19).

A flat bench or 'second shelf break' underlying the glaciogenic features identified from sidescan images is prominent in bathymetric maps of the whole margin (Figure 13). The parallel trends of bathymetric landmarks and ice flow directions inferred from glacial features on the surface of this bench (Figure 10) support the hypothesis that a large ice mass carved the bench with its passage. The bench surface is truncated north of the 500 m contour by a series of deep and steep dendritic gully systems that initiate at the paleo-shelf break (Figures 12d and 16).

Sediment-laden sub-glacial meltwater cascading down the steep continental slope triggered turbidity currents that eroded deep channels; sediments thereby bypassed the upper continental slope before being re-deposited on the abyssal plain. The margin-parallel trend of the ice flow in this location indicates that the majority of the sub-glacial sediments were likely deposited west of the ABM, making the ABM itself a sediment-starved margin.

I conclude that the pervasive bathymetric bench and >600 km-wide distribution of deep, sub-glacially generated flutes indicate margin-parallel flow of a grounded, cohesive ice sheet or stream 600-700 m thick (Figure 7). ABM flutes have depths, styles, and orientations similar to features on the CB, linking all features to the over-deepened glacial outlet troughs of the CAA (Figure 3). Ice sheet grounding on the continental shelf/slope may have isostatically depressed the ABM, probably during OIS 5b based on sediment cores from the CB which is the age of an inferred widespread glaciation in the Beringia region [Brigham-Grette et al., 2001]. Shallow seafloor areas were likely turbated by iceberg keels in the aftermath of ice sheet break-up or during later glacial event(s) and are therefore younger and better preserved than the glacial flutes. A glaciomarine transgression during ice sheet break-up might have deposited the CAA-sourced IRD-rich sediments that make up the areally extensive Flaxman Formation now found up to 7 m above sea level. These findings provide critical data for reconstructing the history of Pleistocene ice in the Arctic Ocean.

5.2 Implications of margin-parallel ice flow

There is little doubt based on my survey of seafloor bedforms along the ABM that they were formed by the erosive action of grounded ice. Based on a recent study of boundary currents in the area, it is unlikely that modern or paleo-current regimes could have created the bedforms observed [Pickart, 2004]. Kristoffersen et al. [2004] postulate that massive tabular icebergs calved from St. Anna Trough-sourced ice streams could have carved flutes at ~980 mwd on the Lomonosov Ridge (Figures 3 and 5) similar to the ones I document on the ABM, rather than a cohesive floating ice shelf as hypothesized by Polyak et al., [2001]. Even if this is true, the along-flow trajectory of the Lomonosov flutes is only ~10 km (limited by data coverage, Figure 5), while the physically separated but margin-wide flow-parallel distribution of ABM flutes the ice-carved 'second shelf break' implies a total ice stream length in excess of 600 km. It is unlikely that even icebergs of the dimensions documented around Antarctica would have the structural integrity or the momentum to ground the whole length of the ABM. The most striking sub-glacial flutes I catalogue occur down to 700 mwd indicating margin-parallel flow of a grounded ice mass 600-700 m thick (Figure 7). However, I know of no analogous modern or paleo ice sheets or streams documented at either pole that circle the periphery of an ocean basin rather than flowing into it. In other studies of ice streams [e.g. Stokes and Clark, 2001], ice draining a continental glacier flows across the continental shelf and deposits its ice and sediment load directly into the ocean basin. This is logically the most efficient ice flow path for at least two reasons. First, it is likely that the gradient from the central ice dome to the ice stream outlet at the ice/ocean interface is steepest when the ice flows perpendicular to the coastline. Second, the more quickly ice is attacked and calved away at the ice/ocean interface, the faster the ice stream can drain, accelerating flow along the shortest path to the ice stream outlet. If the ice does not take this most efficient course, it must be because the flow has been blocked or diverted by topographic barriers.

In the context of the modern ABM, it is hard to imagine what the topographic barrier preventing penetration of thick ice into the central Canada basin could be. Polyak et al., [2001] suggest that ice was propelled along the coastline and across the CB by the force of CAA/Eastern Alaska ice streams (Figure 3). However, based on ice mass balance considerations [Mercer, 1970], this ice would need to be buttressed on both sides of the stream to maintain its full thickness without calving and breaking apart on its northern, oceanward margin. Modern Arctic sea ice reaches thermodynamic equilibrium at 2.5-5 m [Weeks and Ackley, 1986], and may be concentrated and thickened by wind to >7 m north of the CAA [Bourke and Garrett, 1987]. There would be only a slight increase in equilibrium average sea ice thickness with colder temperatures during glacial extremes [Kristoffersen et al., 2004], though it is probable that the whole Canada basin would be capped by a near-continuous cover of sea ice (Figure 3). Continuous sea ice cover during glacials might briefly retard the calving rate along an ice sheet margin [Reeh et al., 2001] but it is unlikely that even wind-concentrated sea ice could buttress an advancing ice shelf whose thickness is more than two orders of magnitude greater than the sea ice [Kristoffersen et al., 2004].

An alternative mechanism to explain the circuitous flow path of ice around the margin of the Canada basin has been advanced in various forms by Mercer [1970], Hughes et al. [1977], and several papers by Grosswald and Hughes [e.g. 1999; 2002]. In these models, a glacial Arctic basin with reduced sea levels due to ice storage on land has only two-thirds of its modern surface area [Grosswald and Hughes, 1999]. The broad Arctic shelves are covered with continental glaciers that through calving eventually clog the diminished Arctic basin with massive icebergs that coalesce to form a cohesive floating ice shelf. The ice shelf may ground or become pinned on central basin topographic highs such as the Lomonosov Ridge, and also prevents land-sourced ice streams from penetrating directly into the central basins [Grosswald and Hughes, 1999]. While this theory provides an appealing explanation for the flow path of ABM ice between the hypothesized central Arctic floating ice shelf and land-based glaciers, it is not supported by many Arctic researchers due to dating inconsistencies and a general lack of corroborative subaerial evidence for an ice shelf/sheet combination of such massive proportions.

A scenario invoking Arctic ice volumes intermediate between a floating ice shelf and sea ice only might be the most plausible means to confine ice flow trajectories parallel to the ABM. For example, Kristoffersen et al., [2004] hypothesize that during deglaciations of the Barents/Kara Sea ice sheet, huge icebergs purged from the St. Anna Trough were entrained in thick glacial sea ice and then dragged across the crest of the Lomonosov Ridge by northward flowing Atlantic water advected through the Fram Strait (Figure 3). In their model, these Antarctic-type icebergs eroded the crest of the Lomonosov Ridge and were then deflected into the Amerasian Basin where I hypothesize they may have contributed to a pile-up of wind and current-driven ice along the CAA and ABM. The clockwise Beaufort Gyre in the Canada Basin combined with katabatic winds coming off the high Kara/Barents Sea ice dome could act to concentrate ice along the north coast of Alaska (Figure 26). This mix of very large and cohesive icebergs (at least 800 m thick and strong enough to remove ~50 m of sediments on the crest of the Lomonosov Ridge [Kristoffersen et al., 2004]) frozen into continuous sea ice cover might have sufficient strength and momentum to deflect the path of ice streams coming out of the mouths of the CAA troughs.

Kristoffersen et al., [2004] note at least 3 episodes when iceberg armadas may have eroded the crests of the Lomonosov Ridge and Yermak Plateau during OIS 16, 12, and 6. These dates are based on the paleoenvironmental record of ODP cores 910 [Flower, 1997], 909 and 912

[Hevroy et al., 1996] on the Yermak Plateau, and a piston core on the Lomonosov Ridge [Jakobsson et al., 2001] (Figure 3). Not all glacial maxima appear equally important in each area suggesting variations in their timing, extent, and severity by region. The glaciogenic features on the ABM are inferred to have been formed coeval with the CB features around OIS 5b or 70-80 ka. As such they are younger than the youngest dates of ice-impacted sediments on the Lomonosov Ridge. However, it seems conceivable that deglacials postdating OIS 6 could have generated iceberg armadas nearly comparable in size to prior surge events, but perhaps not quite thick enough to have scoured the deep crest of the Lomonosov Ridge at ~1000 mwd. Alternatively, the mega-berg purge trajectory could vary slightly from deglacial to deglacial allowing some very large icebergs to miss the shallowest portions of the Lomonosov Ridge but still collect in the Amerasian Basin.

If icebergs from the Barents/Kara ice sheet propagated as far across the basin as the CAA and ABM, there might be some IRD trace of their passage in seafloor sediments. Phillips and Grantz [2001] distinguish an Amerasian basin clast suite of dolostone and limestone sourced in the CAA, and a Eurasia basin suite consisting of siltstone, sandstone, and siliceous clasts probably from the Taymyr Peninsula (Figure 3). The suites appear neatly separated by the length of the Lomonosov Ridge which would seem to indicate that Eurasian-sourced ice rarely traverses the basin. However, the rainout pattern of IRD is thought to have an exponential distribution [Phillips and Grantz, 2001] implying that the majority of clasts would be deposited close to their source area. In addition, any icebergs scraping across the Lomonosov Ridge might have most of their basal clast load removed by erosion. A third hypothesis suggested by Kristoffersen et al. [2004] is that mega-bergs whose drafts were below the main influence of warm Atlantic waters might have their detrital rain-out delayed until they exited the basin into the North Atlantic.

There are clearly many outstanding questions concerning the timing and distribution of glacial ice accumulation in the Canada Basin that need to be addressed with further data collection. In the interim, my hypothesis invoking iceberg armadas trapped in glacial sea ice appears to encompass the physical constraints needed to confine ABM ice flow parallel to the margin, and is not inconsistent with land-based studies.

5.3 Further implications and future work

The Arctic basin plays a pivotal role in regulating and triggering global climate change via its connection with North Atlantic global thermohaline circulation [Ruddiman, 2001], and thus new information for the Arctic would also aid a broad range of scientific endeavors at lower latitudes. Thick ice on the ABM and CB during OIS 5b matches the age of an inferred widespread Beringia glaciation [Brigham-Grette et al., 2001]. There has not been conclusive evidence to date that this glacial ice propagated across the whole Arctic basin. Sediment cores recovered from widely separated areas across the Arctic Ocean display a succession of Pleistocene glaciations [e.g., Poore et al., 1993; Phillips and Grantz, 1997; Jakobsson et al., 2000b] as indicated by extremely low or absent biogenic remains and low sedimentation rates or hiatuses [Darby et al., 1997; Nørgaard-Pedersen et al., 1998; Polyak et al., 2004]. Unfortunately, where sediments have not clearly been affected by grounded ice it is not possible to distinguish whether these breaks in sedimentation were caused by a cover of thick sea ice or by a coherent ice shelf. However, it seems unlikely that ABM ice could be confined along the margin by sea ice alone. Whether this confinement resulted from a classical ice shelf, or rather from an amalgam of huge icebergs and thickened sea ice, the implication is that high ice volumes existed at least along the periphery of the Canada basin and almost certainly across the Amerasia basin if the source of the large icebergs was the St. Anna Trough (Figure 3). It is important to clarify whether ice covered the entirety of the basin because of the very different ways that open water and ice reflect and store heat. Continuous sea ice cover would have similar albedo to an ice shelf, but discontinuous ice coverage would allow heat and gas exchange to ventilate surface waters, influencing sediment stratigraphy and thermohaline circulation.

It is critical to our understanding of the relationship between sea level, oxygen isotopes, and global ice volume that we know the source of ice in the Arctic during glacial periods [Broecker, 1975; Williams et al., 1981; Shackleton, 1987]. Biases in $\delta^{18}O$-based sea level reconstructions may be large for pre-LGM glaciations when very little geologic data exist to independently verify paleo-sea level. Expansion of ice sheets into the ocean affects the relationship between marine $\delta^{18}O$ composition and sea level. Floating ice shelves store $\delta^{18}O$ in the same way as land-based ice but without affecting sea level. If deep draft ice in the Arctic Ocean came from surrounding marine or land based ice sheets instead of an ice shelf, it wouldn't represent a separate contribution in the global oxygen isotope record, but rather an increase in sea level. The global $\delta^{18}O$-based sea level curve is used by scientists in a variety of fields including oceanography, geology, biology, and by the oil exploration industry, and as such improvements to this curve for past glacials would impact a broad range of disciplines.

With this project I have contributed to our understanding of the scale and possibly the timing of ice advance across the ABM. Predictably, these findings have generated a whole new series of questions whose answers are needed to improve our grasp of the Arctic-wide cycle of ice growth and retreat. The size and broad distribution of ABM glaciogenic features imply that the whole north coast of Alaska was glaciated (Figures 3 and 7), possibly during OIS 5b. In order to further verify these findings sediment cores need to be collected across the breadth of the margin to better constrain the timing and number of ice advances and their relation to ice streams in the CAA. In addition a series of high-resolution seismic transects should be run perpendicular to the margin in water depths from 200-1000 m to identify possible ice-eroded surfaces as well as lenses of ice-bulldozed debris. Both coring and seismic work are scheduled to take place during a summer cruise to the Chukchi Borderland in 2005. Though some work is underway on similar topics for the Kara/Barents Sea region [Jakobsson et al., in press], it would be useful to numerically model the interactions between land-based and marine ice along the Alaskan coast to see what configurations might replicate the margin-parallel path of the ABM glaciogenic features. The northern Siberian margin remains the least studied or understood component of the Arctic ice picture. Its proximity to the ABM suggests a plausible interaction between the two coastlines and as such any further land or marine data for this area would shed light on Amerasian basin-wide ice dynamics (Figure 3). Any one of the above avenues has the potential to greatly improve our understanding of glacial history in the Arctic, and should therefore be pursued in a multi-disciplinary, multi-country effort to better understand this important region.

FIGURE CAPTIONS

Figure 1 – Earth's radiation balance as measured at the top of the atmosphere. Blue line shows net incoming shortwave (solar) radiation; red line shows outgoing longwave (sensible heat) radiation as measured using Earth Radiation Budget Experiment (ERBE) data [Barkstrom, 1984]. The surplus of heat energy in equatorial latitudes is transferred to the poles via ocean and atmospheric meridional heat transfer. (Figure modified from Pidwerny [2004]).

Figure 2 – One of the primary mechanisms of meridional heat transport in the oceans is the global thermohaline circulation "conveyor belt" illustrated schematically in this image. Saline waters cooled in the North Atlantic sink to form North Atlantic Deep Water which transits south and into the Pacific before upwelling off the coast of Alaska and Canada. THC acts to stabilize climate, and is known to be influenced by fresh water fluxes from the Arctic Ocean.

Figure 3 - Arctic Ocean with IBCAO depths. Orientation of ice shelf and/or iceberg flows across submarine ridges shown by red, pink, and orange arrows/lines [Polyak et al., 2001; Kristoffersen et al., 2004], ABM study area and inferred ice flow direction shown in yellow box, projected ice sources shown by dashed white arrows, clockwise drift path of Beaufort Gyre shown by dashed blue circle, Pacific water ABM boundary current shown by dashed purple arrow [Pickart, 2004]. In the modern system, the Arctic Ocean exchanges water with the Atlantic via the Fram Strait and Barents seaway, and with the Pacific via the Bering Strait. White box indicates approximate location of cores sampling Lomonosov Ridge sediments from Jakobsson et al. [2001]. CB –

Chukchi Borderland, MT – Mackenzie Trough, AG – Amundsen Gulf, MS – M'Clure Strait, YP – Yermak Plateau.

Figure 4 – Figure shows the extent of North American continental ice sheets at the Last Glacial Maximum in white, ocean areas in light gray, landmasses in dark gray, and modern sea level is indicated by the black line, after Clark et al. [2001]. Note that large shelf areas in the Arctic Ocean were exposed above sea level at this time. (Figure modified from Clark et al. [2001]).

Figure 5 - (upper panel) Sidescan sonar image of linear sub-glacial flutes marked with black arrows on the crest of the Lomonosov ridge; (lower panel) Stratigraphic cross-section of the fluted area in the upper panel as indicated with blue arrows; note the planed/eroded surface of the ridge and the accumulation of eroded sediments on the lee side of the ridge crest. West of the 1500 m contour, the ridge is characterized by dendritic gullies similar to those mapped on the ABM. (Figure modified from Polyak et al. [2001]).

Figure 6 – SCICEX-99 sidescan sonar imagery of the crest of the Chukchi Borderland showing a predominance of SE-NW trending sub-glacially generated fluted bedforms. Dragged blocks indicate the paleo-flow direction of the eroding ice from SE to NW. (Figure modified from Polyak et al. [2001]).

Figure 7 - Figure 2: ABM survey area with sidescan sonar image of zig-zagging survey track, IBCAO depth contours, and orientations of mapped seafloor glaciogenic features. Green lines

indicate the presence and preferred orientations of sub-glacially generated flutes, red lines indicate iceberg scours or grooves.

Figure 8 - Photograph of USS Hawkbill surfacing through 1.5 m thick sea ice at the North Pole. Note the vertical orientation of the sails on the upper part of the submarine that allow the vessel to break through the ice. (Photograph courtesy of M. Edwards).

Figure 9 – Healy/Scicex nav comparison

Figure 10 - Upper rose diagram shows the average trends of all bathymetric landmarks for the areas sampled using spectral analysis techniques. Lower rose diagram shows the average trends of ice flow direction inferred from sidescan sonar imagery for each sample area. In each rose diagram the mean trend of flow features is indicated by the green line. The average trend of all bathymetric landmarks is identical to the average trend of all inferred ice flow indicators at 113˚ (significant at the 95% level, Rayleigh's test, Swan and Sandilands [1995]). In some areas where there are no sub-glacial flutes visible in the sidescan sonar data it is not possible to directly assess ice flow directions, so there are fewer samples on the lower rose diagram.

Figure 11 – a) SCICEX single beam bathymetric profile of area H (Figure 7) in meters vs. distance in kilometers. Note the prominent low slope angle bathymetric bench with ~25 m surface relief coincident with the location of seafloor flutes. b) Average Beaufort Sea slope profile showing seafloor surface area in accumulative percent vs. depth in meters (figure

modified from Jakobsson [2002]). Note the 'second shelf break' at ~450 mwd. Box shows location of Figure 11a.

Figure 12 – a) Area C showing chaotically oriented iceberg scours <400 mwd, dendritic gullies >500 mwd, IBCAO contours. b) Area A showing deeply incised grooves <500 mwd, dendritic gullies, IBCAO contours. c) Area G showing parallel flutes <700 mwd, IBCAO contours. d) Area E showing flutes at the shallowest water depths and dendritic gullies that coalesce into broad trunk channels near 1000 mwd, IBCAO contours.

Figure 13 – Map of 'second shelf break' generated using IBCAO data; bench contours highlighted in yellow. SCICEX survey track shown with black line, black boxes indicate regions where glaciogenic bedforms have been mapped in sidescan sonar imagery. Bench is widest at the eastern end of the survey where it underlies fluted terrain in areas G and H, then narrows at the center before widening again near Barrow Canyon.

Figure 14 – Figure shows sample outcrop map for the Flaxman Formation (indicated in red), the deposits that mantle the ABM coastline up to 7 m above modern sea level. The Flaxman Formation shows no isostatic tilt over its length. Red star over inset map shows the inferred central location of isostatic loading over Arctic Canada from old models used to explain the provenance of the Flaxman Formation glaciomarine transgressive deposits [Brigham-Grette, J., pers. comm. 2004]. An ice shelf located on the ABM would depress the coastline locally allowing for tilt-free isostatic rebound along the Alaskan coastline, which is more consistent with

the modern distribution of the Flaxman Formation deposits. (Figure modified from Brigham-Grette, J. [2004]).

Figure 15 - Map view of the ABM showing the location of the gully transect in pink, IBCAO contours, and sidescan sonar data.

Figure 16 - SCICEX narrow-beam bathymetric gully transect of the submarine track shown in Figure 15. Yellow box shows water depths between 200-600 m. Red arrows indicate typical examples of gullies vs. gullied canyons.

Figure 17 - Figure 10: Map showing the largest gullied canyon in the survey area whose location is indicated by the red arrows in Figure 16. SCICEX single beam bathymetric data are overlaid on the sidescan sonar track, and IBCAO bathymetric contours are shown.

Figure 18 – a) EM120 (12 kHz) multibeam swath bathymetry shaded relief image of mega-scale glacial lineations (flutes) in outer Marguerite Trough, west Antarctic Peninsula continental shelf. Maximum height of the flutes is 15 m, and widths range from 130-300 m. Grid cell size = 25 m. The flutes shown in the image occur in water depths of about 520–550 m. b) EM120 shaded relief image of the continental slope offshore of Marguerite Trough. Grid cell size = 50 m. The location of the continental shelf break is marked by the dashed line. Water depths range from 430-3656 m. The Marguerite trough palaeo-ice stream extended to the shelf edge, infilling the area defined by the box on the continental shelf. Note the similarity in style and distribution of

the gullies and flutes mapped in this location to the ABM (Figure 12) and the Lomonosov Ridge (Figure 5). (Figure modified from Ó Cofaigh et al. [2003]).

Figure 19 - Two samples of the graphical user interface used to hand-edit bathymetry with the BTYP program. The upper section of each panel shows the port sidescan sonar swath. The green line represents the bottom detect values used to generate seafloor depths; in the upper panel seafloor depths shallow in the center of the hour and deepen at either end. The lower section of each panel shows full resolution port and starboard bathymetry for the same hour of data. Port is on top, starboard is on bottom, the submarine travels from left to right where the y-axis equals zero (nadir). Color change interval in the bathymetry is 5 m. Red arrows denote glacial features in both the sidescan sonar and in the bathymetry.

Figure 20 - Upper panel shows sidescan sonar imagery for area A; lower panel shows swath bathymetry data with a contour interval of 10 m for area A, grid cell size = 25 m. Note the prominent bathymetric bench denoted by the pink colors in the lower panel. On the surface of the bench are abundant iceberg scours and grooves, as seen in the sidescan sonar imagery, but the glacial features are not resolvable in the bathymetry at this scale.

Figure 21 - Slope profile of raw bathymetry bottom detects. On the x-axis is the acoustic ping number associated with each starboard bathymetry bottom detect. On the y-axis is seafloor depths in meters.

Figure 22 - Upper panel shows from west-east in pink the locations of samples a1, a2, b1, b2a, b2b, c3, and e1; lower panel shows from west-east the locations of samples g1, g2, h1a, h1b, h1c, h2b, h2a, h3a, h3b, h3c, h4b, h4c, and h4a. In each sample location full resolution bathymetry data were used as input to a spectral analysis of seafloor roughness. Sample locations are overlaid on sidescan sonar data and contours indicate depths from IBCAO bathymetry in meters.

Figure 23 - Geometry used to project the azimuth of the ship's track perpendicular to the strike of glaciogenic features. Feature direction indicates the direction of ice flow inferred from sidescan sonar imagery. Apparent wavelength is the apparent spacing between glaciogenic features. True wavelength is the correct spacing between glaciogenic features when measured perpendicular to strike.

Figure 24 - Sample power spectrum generated from the output of spectral analysis of seafloor bathymetry. On the x-axis is harmonic number. On the y-axis is the amplitude of each wavelength squared, i.e. power in m^2. The total number of samples divided by the harmonic number and multiplied by the distance between samples yields wavelength. Inset shows DC shifted raw seafloor bathymetry (blue) that was detrended by subtracting the values shown in pink to produce the detrended line shown in green. The data were next filtered to produce the values shown in orange, then input to the FFT program to produce the power spectrum shown in this figure.

Figure 25 - Figure 15: Upper panel shows the spectral estimates from west-east for the sample areas listed on the x-axis with distance from the eastern end of the survey area. The lower panel

shows the same spectral estimates from west-east but with equal spacing between samples to facilitate analysis. The strength of the signal of seafloor excavation within a sample area and across the margin is shown in dB using rainbow colors. Red values indicate repeating wavelengths of most deeply excavated terrain, and dark blue values indicate wavelengths that are least excavated. The spacing between glaciogenic features is denoted on the y-axis in meters.

Figure 26 – Cartoon showing the Arctic Ocean region during the period of ice shelf advance across the Alaska Margin. Inferred ice flow orientations from seafloor bedforms indicated with yellow, orange, and red arrows. Hypothesized extents of the Eurasian and Greenland/Laurentide ice sheets are shown in white. Schematic icebergs (not to scale) are indicated calving off the margin of the Eurasian ice sheet. Icebergs frozen into thick glacial sea ice transit across the crest of the Lomonosov Ridge and accumulate against the margin of the CAA in response to northward-flowing Atlantic waters (indicated by salmon arrows).

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