Seafloor evidence for ice shelf flow across the Alaska/Beaufort margin of the Arctic Ocean

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ABSTRACT

New imagery of ~14,100 km^2 of seafloor along a 640-km stretch of the Alaska and Beaufort margins (ABM) in water depths from 250-2800 m depicts a repetitive association of glaciogenic bedforms (flutes and iceberg scours), broad erosional bathymetric features, and adjacent down slope turbidite gullies. These bedforms have styles, depths, and orientations similar to features discovered earlier on the Chukchi Borderland, up to 800 km northwest of the ABM. Flutes occur across the surface of a flattened bathymetric bench interpreted to have formed by an ice shelf sliding along the continental slope. The glacial geology of surrounding areas suggests that an ice shelf likely flowed from the mouths of overdeepened glacial troughs in the Canadian Arctic Archipelago westward along the ABM and across the Chukchi Borderland. This curved pathway indicates an obstruction to ice flow in the central Canada Basin, possibly caused by either a basin-wide ice shelf or by a pile-up of mega-bergs originating from the Eurasian side of the Arctic Ocean. The ice shelf that affected the ABM may have formed during an inferred intra-stage-5 widespread Beringian glaciation. Evidence for glaciogenic features on the ABM corroborates suggestions that large ice volumes and extents existed in the Arctic during Pleistocene glacial periods. These findings have far-reaching implications for Arctic climate studies, ocean circulation, sediment stratigraphy, and the stability of circum-arctic continental ice masses.

1. INTRODUCTION

Over the past decade, new datasets have been collected on high-latitude continental margins in both the Arctic and Antarctic, where glaciogenic processes result in distinctive and diverse seafloor morphology [e.g., *Davies et al.*, 1997; *Anderson*, 1999; *Ó Cofaigh et al.*, 2003]. However, the high-latitude margins of the Arctic Ocean have remained only sparsely studied because of the operational constraints imposed by sea ice cover. As a result, the morphology of the Alaska/Beaufort margin (ABM), except for the nearshore zone, has not been mapped adequately despite the significance of this region for important socio-economic and scientific issues such as hydrocarbon deposits, gas-hydrate stability, and the history of Arctic Ocean circulation and ice cover [e.g., *Grantz et al.*, 1979; *Blasco et al.*, 1990; *Dinter et al.*, 1990; *Kayen & Lee*, 1991].

Recent marine geophysical evidence from widely separated regions of the Arctic Ocean (Figure 1) suggests that larger and thicker ice volumes than previously envisioned by most Arctic geologists may have existed in the basin during glacial periods [*Vogt et al.*, 1994; *Jakobsson et al.*, 1999; *Polyak et al.*, 2001; in press; *Kristoffersen et al.*, 2004; *Jakobsson et al.*, 2005; *Darby et al.*, 2005]. Mapped seabed features indicate that both mega icebergs and coherent ice shelves eroded the seafloor across the Arctic Ocean in water depths as great as 1000 meters. These recent data lend credence to the long-standing, controversial hypothesis that vast ice shelves up to >1 km thick once existed in the Arctic Ocean, rivaling or even exceeding in size the largest modern ice shelves of Antarctica [*Mercer*, 1970; *Hughes et al.*, 1977; *Grosswald and Hughes*, 1999]. However, the actual extent, timing, and formation mechanisms of these ice shelves remain largely unknown. A key location for reconstructing the history of Pleistocene Arctic ice shelves is the Chukchi Borderland (CB), a complex of geological structures extending north of the

Chukchi continental shelf, that shows ice shelf related bedforms at depths reaching 1000 m and chaotic iceberg scours at shallower depths (Figure 1) [*Polyak et al.*, 2001; in press; *Jakobsson et al.*, 2005; *Darby et al.*, 2005]. The northwest-trending orientation of glacially-derived flutes on the CB suggests that the northwestern margin of the Laurentide ice sheet was the likely source of eroding ice. This trajectory further implies that the ABM could have been affected by the same large ice masses and it may therefore be an important area to clarify the extent and pathways of Pleistocene ice shelves in the western Arctic Ocean.

1.1 Physiographic and geological background for the Alaska/Beaufort margin

The ABM extends from the base of the Chukchi Borderland eastward to the Mackenzie River delta (Figures 1 and 2). This aseismic, Atlantic-type continental margin has a continental shelf measuring on average 75 km wide that ranges in depth from 0-50 m. The margin is bounded to the north by a steep, heavily canyoned continental slope that drops to 1000 mwd in the Canada Basin over an average distance of just 13 km. To the south, the continental shelf grades into a broad, low relief plain that extends to the foothills of the Brooks Range. The plain is mantled by 10s-100 m of the unconsolidated clastic materials of the Gubik Formation, interpreted to be a series of glaciomarine transgressive deposits dating from 3 Ma through the Last Glacial Maximum [*Dinter et al.*, 1990 and references therein]. Coarse clasts, from gravel to boulders, with Canadian Arctic Archipelago (CAA) provenance are abundant in some sedimentary units on the Alaskan shelf and coast, notably in the Late Pleistocene Flaxman Member of the Gubik Formation that reaches up to 7 m above sea level [*Leffingwell*, 1919; *Rodeick*, 1979]. This deposition has been explained via the basal rainout of numerous icebergs originating from the northern margin of the Laurentide ice sheet during the Early Wisconsinian

81 or an earlier glaciation [*Dinter et al.*, 1990; *Brigham-Grette and Hopkins*, 1995]. However,
82 these authors also pointed out that a large-scale glaciation cannot easily be reconciled with high
83 sea level in a generally stable tectonic environment, and the mapped formation shows no
84 isostatic tilt along its length towards the Laurentide ice sheet, which requires some other
85 mechanism for the delivery of glaciogenic deposits on the Alaskan shelf and coastal plain.
86
87 **2. METHODS**
88
89 Until recently, many portions of the Arctic Ocean had never been or were only sparsely
90 surveyed due to the constraints of operating in a perennially ice-covered and climatically
91 inhospitable ocean. Traditional arctic surveying methods such as the use of floating ice islands,
92 icebreakers, and aircraft suffer from restricted operations because they are subject to seasonal ice
93 and weather conditions. Between 1995-1999 the U.S. Navy invited researchers in the academic
94 community to design and participate in a series of cruises entitled the SCience ICe EXercises
95 (SCICEX) [*Edwards and Coakley*, 2003], conducted aboard Sturgeon-class submarines that
96 travel below and surface through sea ice. During the 1999 SCICEX cruise ~2000 km of 12 kHz
97 sidescan sonar and interferometric bathymetry data were collected by the nuclear submarine *USS*
98 *Hawkbill* along a 680 km portion of the ABM between 140-160° W (Figures 1 and 2). The
99 submarine mapped seafloor from 200-2800 m deep along the ABM in a zigzagging pattern
100 designed to accommodate a water-sampling program (Figure 2). Sidescan swath widths vary
101 from 6 km on the continental shelf to >20 km at the base of the continental slope. Bathymetry
102 swath width is less than the sidescan swath width, averaging ~4 times the water depth. Where
103 seafloor insonification angles exceed 80° in the shallowest survey areas, low-relief morphologic
104 features are emphasized at the outer edges of the sidescan swaths. Unfortunately, subbottom

105 sonar (chirp) data were not recorded in conjunction with the 1999 sidescan survey on the ABM,
106 but several chirp lines were collected west of this area during the 1998 SCICEX cruise. Sidescan
107 and bathymetry data were iteratively processed to remove artifacts that reduce image quality
108 [*Davis et al.*, 2001]. Sidescan data were gridded in 8 m cells and bathymetry data were gridded
109 in 25 m cells to produce swath imagery for ~14,100 km^2 of seafloor. The gridded bathymetry
110 data have a vertical resolution of ~1% of water depth. Higher resolution (0.5% of water depth)
111 soundings collected directly under the submarine every 12 seconds were sub-sampled to create
112 twenty slope profiles used for quantifying seafloor morphology.
113 Geospatial reference for the ABM survey track locations was generated by the
114 submarine's inertial navigation system (INS) and was then re-navigated using surface GPS fixes
115 [*Edwards and Coakley*, 2003]. The INS incurs errors that propagate and exacerbate with time
116 spent under the ice and away from the last GPS reading. The ABM survey took 6 days and the
117 submarine surfaced only at the beginning and end of the survey, resulting in relative positional
118 navigation errors of ~3 km towards the western end of the survey based on data crossover
119 analyses. SCICEX data were co-registered with the International Bathymetric Chart of the
120 Arctic Ocean (IBCAO) [*Jakobsson et al.*, 2000], which is the most comprehensive compilation
121 dataset of Arctic bathymetry currently available and is generated from icebreaker and submarine
122 data soundings contained in archives in the U.S., Canada, Denmark, Russia, Germany, and
123 Iceland. Co-registration of SCICEX and IBCAO data shows good correspondence between the
124 two datasets, and we have used them in concert to characterize bathymetry along the ABM
125 (Figure 2).
126 To document the extent and distribution of seafloor bedforms in the ABM sidescan
127 imagery we analyzed measurements of bedform length, width, elongation ratio, transverse
128 wavelength, and percentage bedform area per unit area using the Environmental Systems

Research Institute's (ESRI) ArcMap geographic information system (GIS) program. Measurements represent minimum lengths in all cases because sidescan sonar data were projected onto a flat surface, when in reality the seafloor surface is sloped. A margin-parallel transect taken by the submarine on its returning east-west survey track followed the 200 fathom (366 m) contour. This transect is used to characterize bathymetric roughness of gullies observed along the continental slope including the frequency, distribution, style, depth and width of gullies, 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, and this example is used to infer gully dynamics and configuration elsewhere.

In order to quantify whether there is a systematic progression of seafloor morphological changes along the ABM, we performed a one-dimensional Fourier analysis on transects of the full resolution swath bathymetry. To minimize outside factors that degrade data quality, we chose twenty profiles showing distinctive bedforms in the sidescan imagery that were characterized by submarine movement in a single direction at a constant speed. We corrected for the oblique submarine orientations (relative to dominant bedform trend) by projecting each vehicle track perpendicular to the bedform trend, and then resampled the data to generate consistent data densities for each sample area. The slope trend and mean value for each sample area were removed and the projected, resampled, and detrended data were filtered to remove long wavelength noise. These data were analyzed using a Fast Fourier Transform (FFT) program and plotted as power spectra to highlight recurring wavelengths (Figure 3). The FFT amplitudes serve as proxies for the variance in height of seafloor morphological features.

3. RESULTS

The SCICEX data display a repetitive association of seafloor bedforms, distinctive bathymetry, and slope mass-wasting features along the length of the ABM. Eight separate regions (Figure 2: A-H) totaling ~550 km^2 show evidence of disturbed seafloor sediments in <250-700 mwd. A range of linear to arcuate bedforms are observed, from randomly oriented, overlapping scours (Figure 4) to fields of parallel, low-relief lineations (flutes) in excess of 10 km long (Figure 5). The basinward terminations of the disturbed zones are marked by deeply incised dendritic gullies that drain the continental slope (Figures 4-6). The disturbed seafloor is always at the shallowest end of the survey track on or immediately above a prominent low-angle bathymetric bench that ranges from 5 to >35 km wide (Figure 7).

3.1 Bedforms

Morphological features in zones of disturbed terrain fall into three distinct types (Figures 4-5). Type 1 bedforms are chaotically oriented, overlapping scours <2 km in length and <25 m wide (Figure 4) concentrated in water depths <500 m, predominantly <400 m (Figure 2). Type 2 bedforms are more deeply incised, arcuate to nearly linear individual grooves up to 5 km long, <50 m wide, and occur throughout the study area in <500 mwd (Figure 4). In all locations where both Type 1 scours and Type 2 grooves are observed, the grooves overprint the chaotically oriented scours. Bathymetry data show that negative-relief scour and groove furrows are commonly bordered by positive-relief levees; total relief measured from the high-resolution bathymetric profiles is up to 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 ~100º and most trend obliquely down slope (Figures 2 and 8). A chirp-sonar subbottom record west of the study area illustrates multiple gouges at the upper slope reaching to ~350-400 mwd (Figure 9). Seafloor sediment in the gouged area is reworked to a depth of several tens of meters; the bottom of this sedimentary unit commonly forms a prominent reflector that truncates underlying strata.

Type 3 bedforms, parallel flutes, are mapped between ~400-700 mwd in three regions on the eastern half of the survey (Figure 2: E, G, and H). They reach >10 km in length and measure ~50 m wide yielding very high length to width ratios of up to 200:1 (Figure 5). The true coverage and length of the flutes are underrepresented as they extend beyond the edges of the sidescan data. Flutes parallel isobaths trending 103-147º from north (average trend 112˚, Figure 8) and occur in fields that are 10->30 km wide to cover a minimum total area of 330 km^2. The total areal coverage of the Type 3 fluted bedforms can be inferred from their distribution along the ABM (Figure 2): if flutes continue beyond mapped areas in similar water depths between areas E, G, and H, their along-margin extent totals at least ~300 km. The flutes documented in areas G and H are overprinted by Type 2 grooves in the shallowest parts of the survey area.

Spectral analysis is used on twenty bathymetric lines along the length of the survey area to compare the distances between and the amplitudes of repeated seafloor bedforms (Figure 10). The amplitude of spectral estimates gives an indication of the relative seafloor roughness at each wavelength. In Figure 10, 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 flute trend, while in sample G2 the highest amplitude seafloor roughness within that area (which is lower amplitude than that observed in area C3) occurs with spacing of ~60 m. In Figure 10 high amplitude roughness centers around the 50 m wavelength,

consistent with the distance between flutes measured independently from the sidescan imagery. On the western end of the survey in areas A, B, and C where scours and grooves prevail, the seafloor is more deeply excavated than in areas E, G, and H where fields of flutes occur. Depths at which the scours and grooves are observed in sample areas A, B, and C are shallower than 370 m with a mean of 290 m; flutes in areas E, G, and H are found in water depths averaging 400 m and as great as 570 m.

3.2 Bathymetric bench

The SCICEX bathymetry documents 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 (Figures 4 and 5). The bench morphology is also visible in IBCAO bathymetry for the Beaufort Sea [*Jakobsson*, 2002] and in a new bathymetry compilation for the Arctic Ocean (Figure 7, [*Edwards et al.,* 2005]). The basinward margin of the bench is associated with the onset of dendritic gullies and a precipitous increase in slope to 13º. Within the study area the bench width varies from 5->35 km, with the broadest portions mapped for ~100 km along the easternmost end of the surveyed area (Figure 7). On a subbottom record west of the study area the continuation of the bench is identified at ~480 m depth with a width of ~3-4 km (Figure 9). The truncation of the slope face strata suggests the erosional character of the bench. The widest portion of the bench underlies areas G and H (Figures 2 and 7) where the most areally extensive examples of fluted seafloor are documented. Bathymetric features on the bench surface are mostly subdued, but in fluted areas the bench exhibits undulations in the slope profiles with relief as great as 25 m.

223 Figure 8 shows the close match of the average trend of flutes measured from sidescan
224 sonar imagery with the average trends of broad-scale morphological landmarks (for example the
225 shelf/slope break on the northern margin of the bench) for the same areas. The average trend of
226 flutes as mapped from sidescan imagery is 112˚, and the average trend of bathymetric landmarks
227 for the same locations is 114˚ from true north (significant at the 95% level using Rayleigh's test
228 for trend [*Swan and Sandilands*, 1995]).
229
230 **3.3 Slope gullies**
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232 Throughout the survey area a well-developed series of gullies cuts the continental slope
233 starting at the northern margin of the bathymetric bench (Figures 4-6) with an average of one
234 gully every 7 km (59 gullies in total over the length of the ~400 km margin-parallel transect
235 taken by the submarine along the 366 m contour). Gully systems are easily recognizable in
236 sidescan sonar imagery due to the marked contrast in acoustic reflectivity between the highly
237 reflective gully channels and the surrounding sedimentary substrate that absorbs sound and
238 yields low acoustic echo returns (Figure 6 a-b). We interpret this contrast in reflectivity to
239 indicate that the gully channels are filled with hard materials such as coarse gravel lag deposits,
240 while finer sediments have been winnowed away and distributed overbank or further
241 downstream. Gully channel depths measured vertically range from 80-770 m (average 307 m),
242 and gully widths range from 0.8-18.1 km (average 5.5 km) (Figure 6c). Nearly 80% percent of
243 the gullies initiate in water depths <700 m. Gullies are deep, with average width to depth ratios
244 of 15:1, and occur in steep terrain, where average continental slope angles are 10˚ from 500-600
245 mwd and 13˚ from 600-700 mwd. The gully system whose length is mapped in Figure 6 (a and
246 b) measures in excess of 40 km long and covers depths ranging from 250->1500 mwd before

continuing out of the sidescan swath image. The thalwegs of the gullies in the dendritic catchment system (ranging in width from 0.7-10 km) coalesce to form broad trunk channels [*Goff and Nordfjord*, 2004] in deeper water depths from ~1000-1500 m (Figure 6 a-b).

To better understand the physical controls on gully formation we analyzed their morphologies as well as their relationship to topographic landmarks using the ~400 km margin-parallel transect taken by the submarine along the 366 m contour (Figure 6c). Gully walls generally are the same height on their east and west sides, with no trend of asymmetry as a function of distance along the ABM. A striking attribute of the gully transect is the flattened character of the shallowest portions of the seafloor compared to the gullied terrain (Figure 6c). 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 previously mentioned bathymetric bench. As seen in Figures 6 and 7, gully density is highest where the bathymetric bench is narrowest at ~148˚ W, and then decreases westward as the bathymetric bench widens again. In addition, the character of the gullies changes from predominantly shallow, narrow gullies where the bathymetric bench is narrowest, to wider and deeper gullies at the westernmost end of the survey. Where multiple individual gullies coalesce, the walls of several successive gullies do not reach shallow enough depths to intersect the flattened bench surface (Figure 6c). Two prominent examples of gullies that coalesced to form trunk channels ~1000m below sea level are indicated in Figure 6 (a and b).

4. DISCUSSION

4.1 Bedform origin

The bedform ensemble mapped on the ABM shows 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; *Dowdeswell et al.*, 2004]. Chaotically oriented to sub-parallel scours and grooves (Figure 4) are iceberg plowmarks documented in regions of modern and past glaciation, where more linear grooves result from the persistent effects of winds or oceanic currents on icebergs [e.g., *Barnes and Lien*, 1988; *Vogt et al.*, 1994]. Parallel lineations or flutes (Figure 5) have been described in numerous glaciated regions and are interpreted to form by an ice sheet or ice shelf sliding over the seafloor [*Josenhans and Zevenhuizen*, 1990; *Polyak et al.*, 1997; *Anderson*, 1999; *Shipp et al.*, 1999]. These bedforms can be associated with larger bathymetric features indicative of extensive erosion of seafloor deposits such as erosional troughs and truncated ridge tops [*Polyak et al.*, 2001; *Canals et al.*, 2002; *Lowe and Anderson*, 2002]. Many studies document a consistent pattern in the bathymetric distribution of glaciogenic bedforms: flutes are typically 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 agreement with the associations described in other studies, the ABM exhibits iceberg scours at the shallowest depths in the study area (mostly <400 m), parallel flutes further down slope (400-700 m), and gullies originating at the lower edge of the glacially-disturbed areas.

290 A peculiarity of the ABM flutes is their orientation parallel to, rather than transverse to,
291 the continental slope. Margin-parallel seafloor lineations on some continental slopes (for
292 example the Greenland margin) are produced by contour currents [*Kuijpers et al.*, 2002].
293 Oceanographic data from the ABM show evidence for a seasonal subsurface boundary current
294 [*Pickart*, 2004], but it occurs at depths of 100-150 m, much shallower than the observed ABM
295 flutes. The existence of deeper margin-parallel currents capable of producing slope furrows
296 would have to be related to the formation and transit of deep waters, such as along the Greenland
297 margin, and is not observed in the Canada Basin today. It is even less likely that deep currents
298 would have formed during glacial periods when the closure of the Bering Strait and CAA
299 channels further restricted circulation in the Arctic Ocean [*Goosse et al.*, 1997a, b].
300 It has been proposed that current-driven, massive tabular icebergs could have carved the
301 flutes observed at ~980 mwd on the Lomonosov Ridge in the central Arctic Ocean (Figure 1)
302 [*Kristoffersen et al.*, 2004], which are morphologically similar to the features we document on
303 the ABM. However, in contrast to the only ~10 km span of fluted terrain observed on the
304 Lomonosov Ridge, the physically separated but consistent, margin-wide distribution of ABM
305 flutes implies the coherent movement of an eroding ice mass over at least 300 km (Figure 2). It
306 is unlikely that even icebergs of the dimensions documented around Antarctica would have the
307 structural integrity or the momentum to ground along this entire stretch of the ABM without
308 changing the direction of drift. We infer that the ABM flutes were formed by a large ice mass
309 (ice shelf) moving parallel to the continental margin and sliding over the seafloor at depths
310 reaching ~700 m. Rose diagrams comparing the trend of bedform directions inferred from
311 sidescan maps with the trend of bathymetric landmarks on this bench are closely aligned (Figure
312 8), suggesting that the eroding ice mass actually carved the bench out of the continental slope.
313 We note that the bench forms a 'second shelf break' along its lower edge, 300+ m deeper than

the actual modern shelf break of the ABM (e.g., Figures 7 and 9). Similar overdeepened 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 maximal water depths of flutes on the ABM and the southern part of the Chukchi Borderland are similar, reaching almost 700 m and ~750 m, respectively. We infer that this depth marks the bottom surface of an ice shelf during its contact with the ABM. The landward limit of an ice shelf buttressed against the ABM cannot be determined from our data since the data do not extend to depths shallower than ~200 m (Figure 2); furthermore, directional bedforms are overprinted by subsequent chaotic iceberg scours in the shallows. A buried hummocky ridge has been noted near the modern ABM shelf break (~50-60 mwd) south of the fluted areas G and H, and was tentatively interpreted as a barrier island chain [*Dinter*, 1985]. Instead, this formation might be a moraine that marks the landward limit of grounded ice. A dedicated study involving the collection of high-resolution seismic reflection records and sediment cores is needed to delineate the boundaries of this grounding.

Spectral analysis of bathymetric data contributes a quantitative and margin-wide characterization of the bedforms (Figure 10). Where we document the best examples of sub-glacial flutes on the eastern half of the survey in areas E, G and H, amplitudes of seafloor roughness are moderate and cover a narrow range of dominant wavelengths from ~45-100 m. In the westernmost, shallower half of the survey, where iceberg scours and grooves erode the substrate, samples from areas A, B, and C show the strongest signals of seafloor excavation and a wider range of wavelengths from ~20-300 m. These results corroborate the change in bedform character along the ABM seen in sidescan and bathymetry images (Figures 4-5).

It is possible that areas A, B, and C were impacted by the same large ice mass that eroded and/or molded the seafloor in areas E, G, and H; however, the chaotically oriented iceberg scours

338 and grooves that primarily characterize the western end of the margin likely overprinted the
339 fluted topography. The two types of morphologically distinct scours and grooves observed
340 suggest the possibility of two generations of iceberg activity, one during the collapse of an ice
341 shelf that eroded the ABM, and the other during a later episode of massive iceberg discharge
342 from the Laurentide Ice Sheet. Both types have a similar azimuth range (95-100˚ from north)
343 (Figures 2 and 8) indicating that icebergs drifted clockwise along the ABM and impacted the
344 slope at an oblique angle from the northeast. This drift pattern is consistent with the distribution
345 of ice-rafted debris in Pleistocene sediments of the Canada Basin assumed to have been
346 dispersed by icebergs carried by the clockwise Beaufort Gyre [*Phillips and Grantz*, 2001]
347 (Figure 1).
348
349 **4.2 Formation of ABM Slope Gullies**
350
351 The co-existence of the slope gullies with the glacially eroded portion of the ABM
352 (Figures 4-7), and their morphologic and contextual similarity to gully systems documented on
353 glaciated continental margins around Antarctica [*Ó Cofaigh et al.*, 2003], East Greenland [*Ó*
354 *Cofaigh et al.*, 2002], and the Gulf of Alaska [*Carlson et al.*, 1990], suggest that the gullies may
355 be related to glaciogenic processes. In each of the above localities, ice is inferred to have
356 extended to the continental shelf edge during glacial periods, releasing sub-glacially entrained
357 sediments and subglacial meltwater where the ice front separated from the seafloor. Sediments
358 were subsequently remobilized by down slope gravity flows and turbidite gullies were eroded
359 into the upper continental slope. *Ó Cofaigh et al.* [2003] argue that where slope gradients are
360 sufficiently steep, any sub-glacial sediments available at the shelf edge may be transported
361 directly to the abyssal plain, bypassing the slope altogether. This results in a sediment-starved

362 upper slope environment conducive to the formation of erosional gullies, regardless of the total
363 amount of sediment available from glacial activity. Thus even though the ABM is known to be
364 mantled with up to 2700 m of unconsolidated sediments [*Grantz et al.*, 1979; *Dinter et al.*, 1990;
365 *Kayen and Lee*, 1991; *Brigham-Grette et al.*, 2001], the continental slope in this area may simply
366 be too steep (up to 13˚ between 600-700 mwd) to retain any glacially deposited sediments on the
367 shelf edge.

368 The geomorphology of the ABM shows evidence for the flow of grounded ice parallel to
369 slope isobaths (Figure 2), instead of transverse to the shelf edge, as in the ice stream localities
370 described above. As a result, it appears that ice probably did not deliver sediments to the ABM
371 slope directly below the area of erosion. Some sediments may have arrived at the shelf edge at
372 an oblique angle where pre-existing topographic variability along the shelf/slope break caused
373 the sliding ice to float over open water, but the majority of the sediments were probably moved
374 parallel to the slope in the direction of dominant ice flow. Thus most sub-glacially entrained
375 sediments may not have been released along the portion of the ABM margin mapped during
376 SCICEX-99, effectively starving the upper continental slope of sediments.

377 Large-scale mass-wasting features on the ABM, involving >200-m of sedimentary cover,
378 have been described in early seismic-reflection studies [e.g., *Grantz et al.*, 1979, 1981], but their
379 exact relationship to slope morphology and the mechanisms of their formation have yet to be
380 identified. Seismic processes on the ABM and the decomposition of gas hydrates during low
381 sea-level stands have been discussed as likely controls on slope instabilities [*Grantz et al.*, 1981;
382 *Kayen and Lee*, 1991]. From the existing data we cannot conclude whether the mapped ABM
383 gullies and other large-scale slumps are related; high-resolution subbottom data from the gullied
384 slope are needed to clarify this relationship.

385

386 **4.3 Ice flow provenance and trajectory**

387

388 The observed flute orientation along the ABM could indicate ice flow in either an
389 easterly or westerly direction. In the absence of unidirectional bedforms such as drumlins,
390 dragged boulders, or larger-scale erosional and depositional features, we have used the broader
391 geological context to determine the likely ice flow direction. Fields of flutes on the Chukchi
392 Borderland (Figure 1) with a similar, somewhat more northward orientation (130-135° at the
393 southern margin) and water depths down to 750+ m, show direct evidence for a southeast to
394 northwest ice flow path [*Polyak et al.*, 2001; *Jakobsson et al.*, 2005]. The broad, over-deepened
395 troughs traversing the western part of the CAA (Amundsen Gulf, M'Clure Strait) and the
396 Mackenzie Trough are the likely outlets for large Laurentide ice streams capable of feeding ice
397 shelves in the Canada Basin (Figure 1). This interpretation is consistent with the glacial geology
398 and geomorphology of these troughs and adjacent areas [*Blasco et al.*, 1990; *Dyke et al.*, 1992;
399 *Stokes et al.*, 2005; in press] and with models of ice discharge [*Bigg and Wadley*, 2001], all
400 indicating the likelihood of large ice streams channeling the northwest margin of the Laurentide
401 Ice Sheet. The CAA provenance of the ice is corroborated by the composition of coarse clastic
402 material in diamict and lag deposits along the ABM [*Rodeick*, 1979; *Naidu and Mowatt*, 1992],
403 although the exact origin of these deposits (subglacial or iceberg-derived) still needs to be
404 identified.

405 The latitudinally-oriented, margin-parallel glaciogenic bedforms on the ABM appear to
406 provide a 'missing link' between the inferred northwest Laurentide sources of ice and the
407 northwestward erosion of the Chukchi Borderland. This interpretation implies that the eroding
408 ice mass flowed along the periphery of the Canada Basin rather than into its center. In order to
409 achieve this pathway, ice flow had to be blocked or diverted by some type of barrier. *Polyak et*

410 *al.*, [2001] suggest that ice was propelled along the coastline and across the CB by the force of
411 CAA/Mackenzie ice streams alone (Figure 1). However, based on ice mass balance
412 considerations [*Mercer*, 1970], this ice would need to be buttressed on both sides of the stream to
413 maintain its full thickness without calving and breaking apart on its northern, oceanward margin.
414 Continuous sea ice cover during glacial periods might briefly retard the calving rate along an ice
415 sheet margin [*Reeh et al.*, 2001] but it is unlikely that sea ice with a maximal thickness of ~7 m
416 when concentrated by wind [*Bourke and Garrett*, 1987], could buttress an advancing ice shelf
417 whose thickness is two orders of magnitude greater.
418 An alternative mechanism to explain the circuitous flow path of ice along the basin
419 margin invokes the existence of a thick ice shelf capping the entire Arctic Ocean [*Mercer*, 1970;
420 *Hughes et al.,* 1977; *Grosswald and Hughes*, 1999]. In these models, the basin-wide ice shelf,
421 fed by multiple sources around its periphery, could have prevented land-sourced ice streams
422 from penetrating directly into the central basins [*Grosswald and Hughes*, 1999]. Evidence for
423 glacial erosion on top of the Lomonosov Ridge (Figure 1) [*Jakobsson*, 1999; *Polyak et al.*, 2001]
424 makes this scenario possible, provided there were synchronously growing ice shelves on both the
425 Eurasian and Laurentide sides of the ocean. The limited extent of the Eurasian ice sheet in the
426 Arctic during the Last Glacial Maximum [*Svendsen et al.*, 2004] refutes the concept of the pan-
427 Arctic ice shelf during that time period, but earlier glacial periods could have been more
428 favorable for its formation. Notably, the age of glacial erosion on the Lomonosov Ridge is
429 estimated at Oxygen Isotope Stage (OIS) 6, >130 ka [*Jakobsson et al.*, 2001].
430 A different interpretation of the Lomonosov Ridge erosion suggests that it was produced
431 by an armada of huge tabular icebergs, rather than an ice shelf [*Kristoffersen et al.*, 2004]. In
432 this scenario, Antarctic-type icebergs purged from the glaciated Barents-Kara continental margin
433 were entrained in thick sea ice and then dragged across the crest of the Lomonosov Ridge by

northward flowing Atlantic water advected through the Fram Strait (Figure 1). We expand upon this model to suggest that these tightly packed mega-bergs could then have proceeded into the Amerasian Basin where they contributed to a pile-up of wind and current-driven ice that deflected the path of ice streams coming out of the mouths of the CAA troughs.

4.4 Age control

The age of maximal glacial erosion on the Chukchi Borderland has been tentatively estimated between early Wisconsinian to Illinoian, or OIS 4 to 6 [*Polyak et al.*, 2003; in press]. Because the orientation of glaciogenic flutes on the CB and ABM (Figure 1) align spatially and have the same depth range, we infer that they were formed by the same glacial event. The placement of this glacial event in the context of the ABM stratigraphy will help to test and refine our age controls. An ice shelf encroachment on the ABM could have affected sedimentation on the margin by directly depositing sub/proglacial till and/or by causing a glacioisostatic depression that was infilled by glaciomarine sediments. *Grosswald and Hughes* [1999] hypothesize that the Flaxman Formation mantling the ABM coastline to an altitude of 7 m and containing a high content of CAA clastic material is a glacial till that was formed by the lateral margin of a transiting ice shelf. However, lithological and paleontological data suggest a glaciomarine rather than sub-glacial origin for this deposit [*Dinter et al.*, 1990; *Brigham-Grette and Hopkins*, 1995]. We infer that the grounding of several hundred meter thick ice on the continental shelf and slope isostatically depressed the ABM, allowing for a glaciomarine transgression of the coastline. This interpretation is consistent with the lack of any mapped isostatic tilt in the Flaxman Formation deposits along the margin, which indicates a local source of isostatic loading at the ABM, rather than a distal Laurentide source as proposed previously

458 [*Dinter et al.*, 1990; *Brigham-Grette and Hopkins*, 1995]. If this scenario is correct, ice shelf
459 encroachment on the ABM would have immediately preceded the deposition of the Flaxman
460 Formation dated to OIS 5a, 70-80 ka [*Dinter et al.*, 1990; *Brigham-Grette and Hopkins*, 1995].
461 Accordingly, the ice-grounding event would likely have occurred during OIS 5b, which fits the
462 age range estimated for the erosion of the Chukchi Borderland [*Polyak et al.*, 2003, in press] and
463 matches the age of an inferred widespread glaciation in the Beringia region [*Brigham-Grette et*
464 *al.*, 2001].

465 The inferred age of ice shelf passage along the ABM is younger than the OIS 6 age
466 estimate for the erosion found on the Lomonosov Ridge [*Jakobsson et al.*, 2001]. However, it is
467 conceivable that glacial periods postdating OIS 6 could have generated ice shelves or iceberg
468 armadas nearly comparable in size to prior surge events, but not quite thick enough to have
469 scoured the deep crest of the Lomonosov Ridge at ~1000 mwd. Furthermore, the mega-berg
470 purge trajectory could vary slightly from one glacial period to another, allowing some very large
471 icebergs to miss the shallowest portions of the Lomonosov Ridge but still collect in the
472 Amerasian Basin.

473

474 **5. CONCLUSIONS**

475

476 Sidescan and bathymetry data obtained during the SCICEX program elucidate the
477 detailed morphology of the Alaska/Beaufort Margin in the depth range of 200-2800 mwd, and
478 help to constrain the sedimentary and glacial history of this part of the Arctic Ocean. The ABM
479 data show a repetitive association of glaciogenic seafloor bedforms and larger-scale erosional
480 bathymetric features in water depths reaching 700 m, with multiple turbidite gullies further down
481 slope. Similar geomorphic associations are known from glaciated continental margins in the

Arctic and Antarctic [e.g., *Anderson*, 1999; *Shipp et al.*, 1999; *Polyak et al.*, 2001; *Ó Cofaigh et al.*, 2003; *Dowdeswell et al.*, 2004]. 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 large ice mass (sliding ice shelf) buttressed against the ABM. Flutes occur on the surface of a bathymetric bench at 400-550 m that may have been formed by ice erosion, an interpretation that is corroborated by the coincident trends of flutes and broader-scale bathymetric landmarks on the bench surface (Figure 8). We note also that this large-scale erosion of the continental slope makes the ABM similar to overdeepened continental margins that are characteristic of formerly glaciated areas. Iceberg scours and grooves observed on the ABM in shallower water depths were probably generated during ice shelf break-up as well as later glacial event(s). The bench is truncated to the north by steep dendritic gullies that initiate in branching catchment areas up to 18 km across and coalesce 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 steep continental slope.

ABM glaciogenic bedforms have depths, styles, and orientations similar to features discovered earlier on the Chukchi Borderland (Figure 1) [*Polyak et al.*, 2001; in press; *Jakobsson et al.*, 2005] and were probably formed during the same glacial episode. ABM bedforms link CB glaciogenic features with the over-deepened troughs of the CAA that were the likely outlets for ice streams from the northwest sector of the Laurentide ice sheet into the Canada Basin. The margin-parallel pathway of this ice flow indicates that incoming ice was diverted from streaming directly into the central part of the basin by a barrier. We infer that such a barrier was formed by either a basin-wide ice shelf fed by multiple sources, or by a pile-up of Antarctic-style icebergs driven by winds and currents from the Eurasian side of the Arctic Ocean. Corroboration of the movement of very thick ice (~1000 m), be it an ice shelf or iceberg armadas, towards the Canada

506 Basin, is seen in the eroded crest of the Lomonosov Ridge at the center of the Arctic Ocean
507 [*Jakobsson*, 1999; *Polyak et al.*, 2001; *Kristoffersen et al.*, 2004].

508 We propose that the sliding of a Canada Basin ice shelf over the ABM is causally linked to
509 the diamict Flaxman Member of the Gubik Formation that mantles the ABM coast up to 7 m
510 above sea level and contains numerous coarse clasts of CAA provenance [*Dinter et al.*, 1990;
511 *Brigham-Grette and Hopkins*, 1995]. This linkage can explain the unusual geology of the
512 Flaxman Member and its lack of glacioisostatic tilt towards the CAA. We propose that the
513 grounding of several hundred meter thick ice on the ABM shoulder isostatically depressed the
514 entire margin and allowed for a glaciomarine transgression of the coast. In this case, the age of
515 the ice grounding on the ABM immediately precedes the age of the Flaxman Member, that is
516 OIS 5b. This timing correlates with an inferred widespread glaciation in the Beringian region
517 [*Brigham-Grette et al.*, 2001] and fits the age range estimated for glacial erosion on the Chukchi
518 Borderland [*Polyak et al.*, 2003; in press].

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520

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679 **FIGURE CAPTIONS**

680

681 Figure 1 - Overview map of the Arctic Ocean with bathymetry from the International
682 Bathymetric Chart of the Arctic Ocean (IBCAO) [*Jakobsson et al.*, 2000]. Fields of flutes on the
683 Chukchi Borderland and Lomonosov Ridge shown by red, pink, and orange arrows/lines [from
684 *Polyak et al.*, 2001]; arrows show direction of eroding ice, where reconstructed. ABM study area
685 and inferred ice flow direction shown in yellow box (this study). Subbottom profile (Figure 9)
686 location shown by yellow star. Maximal limit of Late Pleistocene Eurasian and North American
687 ice sheets shown with double white line (Dyke et al., 2002; Svendsen et al., 2004). Projected
688 deep-draft ice sources shown by dashed white arrows. Beaufort Gyre shown by dashed blue
689 circle, Pacific water ABM boundary current shown by dashed purple arrow [*Pickart*, 2004]. CB
690 – Chukchi Borderland, MT – Mackenzie Trough, AG – Amundsen Gulf, MS – M'Clure Strait,
691 YP – Yermak Plateau.

692

693 Figure 2 - ABM survey area with bathymetry, survey track (in black), and the schematic
694 orientations of mapped seafloor glaciogenic features within each study area. Heavy black lines
695 indicate the presence and orientation of flutes, red lines indicate the mean orientations of iceberg
696 scours or grooves. The bathymetry data are a combination of gridded data from IBCAO
697 [*Jakobsson et al.,* 2000] and swath data collected by the *USCGC HEALY, R/V Nathaniel B.*
698 *Palmer* and the *USS Hawkbill* [*Edwards and Coakley,* 2003*; Edwards et al.,* 2005]. Bathymetry
699 is displayed as color-coded contours overlaid on a sun-shaded digital terrain model illuminated
700 from the east.

701

Figure 3 - Sample power spectrum generated from the output of spectral analysis of seafloor bathymetry. The x-axis represents long to short wavelengths; the y-axis is the amplitude of each wavelength squared, i.e., power (in m^2). Inset shows DC shifted raw seafloor bathymetry that was detrended and filtered, then input to the FFT program to produce the power spectrum shown in this figure.

Figure 4 – Sidescan (upper) and bathymetry (lower) SCICEX swath data for area A (see Figure 2 for location). The bathymetry data are displayed as color-coded contours overlaid on sun-shaded topography illuminated from 005° at an altitude of 10° above the horizon to emphasize small-scale relief. Scours and grooves (indicated) are most apparent in the higher resolution sidescan data with its higher angle of insonification, but all features show some relief in the lower resolution (25 m grid cells) bathymetric data.

Figure 5 –Sidescan (upper) and bathymetry (lower) SCICEX swath data for area H (see Figure 2 for location). The bathymetry data are displayed as color-coded contours overlaid on sun-shaded topography illuminated from 355° at an altitude of 10° above the horizon to emphasize small-scale relief. The presence of flutes in the bathymetry (indicated) is more strongly correlated with their apparent width in the sidescan data rather than the total vertical relief. Deep, narrow flutes in the 8-m resolution gridded sidescan data can be difficult to detect in the 25-m bathymetry gridded data, but wider flutes are apparent across the entire swath.

Figure 6 – (a) sidescan image of an extensive dendritic gully system coalescing into trunk channels (area E on Figure 2). (b) Interpreted outline of the gully system. (c) Highest resolution SCICEX (nadir) bathymetry of ABM topography collected along the returning submarine track,

726 which surveyed approximately parallel to the 366-m isobath. Water depths between 200-600 m
727 are highlighted. Arrows indicate typical examples of gullies along the margin.
728
729 Figure 7 - 3-D view of the ABM that highlights the bathymetric bench most prominent along the
730 margin in Area H (see Figure 2 for location and description of the bathymetric datasets).
731 SCICEX survey track shown in black, the 350 m and 600 m contours that bound the bench are
732 shown in red and blue, respectively. The bathymetry data are displayed as color-coded contours
733 overlaid on a sun-shaded digital terrain model illuminated from the north. The diagonal stripes
734 in the bathymetry data are artifacts that result from resampling the IBCAO data at higher
735 resolution than their 2.5 x 2.5 km grid cell size. CB – Chukchi Borderland. Inset shows SCICEX
736 slope profile of Area H along the transect indicated with heavy black line on the 3-D map.
737
738 Figure 8 - Average trends of bathymetric landmarks for the areas sampled using spectral analysis
739 techniques: a) Bathymetric landmark trend in fluted areas, b) bathymetric landmark trend in
740 scoured and grooved areas, c) bedform orientation from sidescan sonar data in fluted areas, d)
741 bedform orientation from sidescan sonar data in scoured and grooved areas. In each rose
742 diagram the mean trend is indicated by the dashed line.
743
744 Figure 9 - Fragment of a chirp-sonar subbottom record west of the study area (SCICEX-98; see
745 Figure 1 for location) illustrating iceberg scouring to water depths of between 350-400 m and
746 truncation of the inclined bedrock strata on the slope, with an erosional bench between 450-500
747 m.
748

749 Figure 10 – Spectral estimates seafloor roughness for the sites listed on the x-axis (see Figure 2
750 for locations) with distance from the eastern end of the survey area (upper panel), and with equal
751 spacing between bathymetric lines analyzed to facilitate visualization (lower panel). The
752 strength of the signal of seafloor excavation within one site and across the margin is shown in dB
753 using rainbow colors. Color change from red to dark blue values indicates the change in
754 repeating wavelengths from the most deeply excavated to the least excavated terrain. The
755 spacing between glaciogenic features is denoted on the y-axis in meters.