Full title: **Seafloor evidence for ice shelf flow across the Alaska/Beaufort margin of the Arctic Ocean**

Short title: **Ice shelf flow across the Alaska/Beaufort margin**

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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 (lineations 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. Lineations occur across the surface of a flattened bathymetric bench interpreted to have formed by an ice shelf sliding along the continental slope and scraping the seafloor at temporary grounding locations. 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 Oxygen Isotopic Stage 4 or 5b, possibly correlating to 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). The high-latitude margins of the Arctic Ocean have remained only sparsely studied, however, 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 adequately mapped 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 and 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, 1999; Polyak *et al.*, 2001; 2007; 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). Clarifying the full extent, timing, and formation mechanisms of these ice shelves is one of the most intriguing tasks for Pleistocene paleoglaciologists. 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 almost 1000 m and chaotic iceberg scours at shallower depths (Figure 1) (Polyak *et al.*, 2001; 2007; Jakobsson *et al.*, 2005; Darby *et al.*, 2005). A commonly northwest-trending orientation of glacially-derived lineations on the CB, especially those at deeper water depths, suggests that the northwestern margin of the Laurentide ice sheet was a 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.

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 meters water depth (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

84 Pleistocene Flaxman Member of the Gubik Formation that reaches up to 7 m above sea level
85 (Leffingwell, 1919; Rodeick, 1979). This deposition has been explained via the basal rainout of
86 numerous icebergs originating from the northern margin of the Laurentide ice sheet during the
87 Early Wisconsinian or an earlier glaciation (Dinter *et al.*, 1990; Brigham-Grette and Hopkins,
88 1995). These authors also pointed out, however, that a large-scale glaciation cannot easily be
89 reconciled with high sea level in a generally stable tectonic environment, and the mapped
90 formation shows no isostatic tilt along its length towards the Laurentide ice sheet, which requires
91 some other mechanism for the delivery of glaciogenic deposits onto the Alaskan shelf and
92 coastal plain.
93
94 **2. METHODS**
95
96 Between 1995-1999 the U.S. Navy invited researchers in the academic community to
97 design and participate in a series of cruises entitled the SCience ICe EXercises (SCICEX)
98 (Edwards and Coakley, 2003), conducted aboard Sturgeon-class submarines that travel below
99 and surface through sea ice. During the 1999 SCICEX cruise ~2000 km of 12 kHz sidescan
100 sonar and interferometric bathymetry data were collected by the nuclear submarine *USS*
101 *Hawkbill* along a 680 km portion of the ABM between 140-160° W (Figures 1 and 2). The
102 submarine mapped seafloor from 200-2800 m deep along the ABM in a zigzagging pattern
103 designed to accommodate a water-sampling program (Figure 2). Sidescan swath widths vary
104 from 6 km on the continental shelf to >20 km at the base of the continental slope. Bathymetry
105 swath width is less than the sidescan swath width, averaging ~4 times the water depth. Where
106 seafloor ensonification angles exceed 80° in the shallowest survey areas, low-relief morphologic

features are emphasized at the outer edges of the sidescan swaths. Unfortunately, subbottom sonar (chirp) data were not recorded in conjunction with the 1999 sidescan survey on the ABM, but several chirp lines were collected west of this area during the 1998 SCICEX cruise. Sidescan and bathymetry data were iteratively processed to remove artifacts that reduce image quality (Davis *et al.*, 2001). Sidescan data were gridded in 8 m cells and bathymetry data were gridded in 25 m cells to produce swath imagery for ~14,100 km^2 of seafloor. The gridded bathymetry data have a vertical resolution of ~1% of water depth. Higher resolution (0.5% of water depth) soundings collected directly under the submarine every 12 seconds were sub-sampled to create twenty slope profiles used for quantifying seafloor morphology.

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 and Coakley, 2003). 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 western end of the survey based on data crossover analyses. SCICEX data were co-registered with the International Bathymetric Chart of the Arctic Ocean (IBCAO) (Jakobsson *et al.*, 2000), which 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 United States, Canada, Denmark, Russia, Germany, and Iceland. Co-registration of SCICEX and IBCAO data shows good correspondence between the two datasets, and we have used them in concert to characterize bathymetry along the ABM (Figure 2).

129 To document the extent and distribution of seafloor bedforms in the ABM sidescan
130 imagery we analyzed measurements of bedform length, width, elongation ratio, transverse
131 wavelength, and percentage bedform area per unit area using the Environmental Systems
132 Research Institute's (ESRI) ArcMap GIS program. Measurements represent minimum lengths in
133 all cases because sidescan sonar data were projected onto a flat surface, when in reality the
134 seafloor surface is sloped. A margin-parallel transect taken by the submarine on its returning
135 east-west survey track followed the 200 fathom (366 m) contour. This transect is used to
136 characterize the bathymetric roughness of gullies observed along the continental slope including
137 the frequency, distribution, style, depth and width of gullies, and depth of gully onset. In one
138 locality it is possible to observe almost the full length of an individual gully system, and this
139 example is used to infer gully dynamics and configurations elsewhere.
140 In order to quantify whether there is a systematic progression of seafloor morphological
141 changes along the ABM, we performed a one-dimensional Fourier analysis on transects of the
142 full resolution swath bathymetry. To minimize outside factors that degrade data quality, we
143 chose twenty profiles showing distinctive bedforms in the sidescan imagery that were
144 characterized by submarine movement in a single direction at a constant speed. We corrected for
145 the oblique submarine orientations (relative to dominant bedform trend) by projecting each
146 vehicle track perpendicular to the bedform trend, and then resampled the data to generate
147 consistent data densities for each sample area. The slope trend and mean value for each sample
148 area were removed and the projected, resampled, and detrended data were filtered to remove
149 long wavelength noise. These data were analyzed using a Fast Fourier Transform (FFT)
150 program and plotted as power spectra to highlight recurring wavelengths. The FFT amplitudes
151 serve as proxies for the variance in height of seafloor morphological features.

3. RESULTS

The SCICEX data display a recognizable association of seafloor bedforms, distinctive bathymetry, and slope mass-wasting features that are found repeatedly 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 to fields of parallel, low-relief lineations (or flutes) in excess of 10 km long. The basinward terminations of the disturbed zones are marked by deeply incised dendritic gullies that drain the continental slope. 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. Each of these bedforms will now be described in turn.

Bedforms

Morphological features in zones of disturbed terrain fall into three distinct types. Type 1 bedforms are chaotically oriented, overlapping scours <2 km in length and <25 m wide (Figure 3) concentrated in water depths <500 m, predominantly <400 m. 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 3). In all locations where both bedform types are observed, the grooves overprint the chaotically oriented scours. Bathymetric 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

175 60-135º from north, but fields have a dominant overall fabric of ~100º and most trend obliquely
176 down slope (Figure 2).

177 A chirp-sonar subbottom record west of the study area (location shown in Figure 1)
178 illustrates multiple gouges at the upper slope reaching to ~350-400 mwd (Figure 4). Seafloor
179 sediment in the gouged area is reworked to a depth of several tens of meters; the bottom of this
180 sedimentary unit commonly forms a prominent reflector that truncates underlying strata.

181 Type 3 bedforms, parallel lineations, are mapped between ~400-700 mwd in three
182 regions on the eastern half of the survey (Figure 2: E, G, and H). They reach >10 km in length
183 and measure ~50 m wide yielding very high length to width ratios of up to 200:1 (Figure 5). The
184 true coverage and length of the lineations are underrepresented as they extend beyond the edges
185 of the sidescan data. Lineations parallel isobaths trending 103-147º from north (average trend
186 112˚, Figure 6) and occur in fields that are 10->30 km wide to cover a minimum total area of 330
187 km^2. The total areal coverage of the Type 3 bedforms can be inferred from their distribution
188 along the ABM (Figure 2): if lineations continue beyond mapped areas in similar water depths
189 between areas E, G, and H, their along-margin extent totals ~300 km. The lineations
190 documented in areas G and H are replaced and apparently at least partially overprinted by
191 grooves and/or scours in the shallowest parts of the survey area (Figure 7).

192 Spectral analysis is used on twenty bathymetric lines along the length of the survey area
193 to compare the distances between and the amplitudes of the seafloor bedforms described above
194 (Figure 8). The amplitude of spectral estimates gives an indication of the relative seafloor
195 roughness at each wavelength. In Figure 8, colored amplitudes indicate the scale of roughness of
196 the seafloor at the wavelengths indicated in meters on the y-axis. For example, in sample area
197 C3, there is high amplitude roughness of the seafloor with repeated spacing of ~45 m

198 perpendicular to the sample transect and parallel to the lineation trend, while in sample G2 the
199 highest amplitude seafloor roughness within that area (which is lower amplitude than that
200 observed in area C3) occurs with spacing of ~60 m. In Figure 8 high amplitude roughness
201 centers around the 50 m wavelength, consistent with the distance between lineations measured
202 independently from the sidescan imagery. On the western end of the survey in areas A, B, and C
203 where scours and grooves prevail, the seafloor is more deeply excavated than in areas E, G, and
204 H where fields of lineations occur. Depths at which the scours and grooves are observed in
205 sample areas A, B, and C are shallower than 370 m with a mean of 290 m; lineations in areas E,
206 G, and H are found in water depths averaging 400 m and as great as 570 m.
207
208 **Bathymetric bench**
209
210 The SCICEX bathymetry documents the co-occurrence of the seafloor bedforms
211 described above with the location of a flattened bathymetric bench that is 400-550 m deep and
212 has an average slope of just 1.6º (Figures 3, 5, and 9). The bench morphology is also visible in
213 IBCAO bathymetry for the Beaufort Sea (Jakobsson, 2002) and in a new bathymetry compilation
214 for the Arctic Ocean (Figure 9) (Edwards *et al.,* 2005). The basinward margin of the bench is
215 associated with the onset of dendritic gullies and a precipitous increase in slope to 13º. Within
216 the study area the bench width varies from 5->35 km, with the broadest portions mapped for
217 ~100 km along the easternmost end of the surveyed area (Figure 9). On a subbottom record west
218 of the study area the continuation of the bench is identified at ~480 m depth with a width of ~3-4
219 km (Figure 4). The truncation of the slope face strata suggests the erosional character of the
220 bench. The widest portion of the bench underlies areas G and H (Figures 2 and 9) where the

most areally extensive examples of lineated seafloor are documented. The morphology of the bench surface is mostly subdued, but in lineated areas the bench exhibits undulations in the slope profiles with relief as great as 25 m (Figure 9). The average trends of bathymetric contours between ~250-500 mwd, which includes part of the continental slope; the bench surface, and the basinward margin of the bench, show a close match with the average trend of lineations measured from sidescan sonar imagery; 114˚ and 112˚ from true north, respectively (Figure 6).

Slope gullies

Throughout the survey area a well-developed series of gullies erodes the continental slope starting at the northern margin of the bathymetric bench (Figures 2, 3, 5, and 10) with an average of one gully every 7 km (59 gullies in total over the length of the ~400 km margin-parallel transect taken by the submarine along the 366 m contour). Gully systems are easily recognizable in sidescan sonar imagery due to the marked contrast in acoustic reflectivity between the highly reflective gully channels and the surrounding sedimentary substrate that absorbs sound and yields low acoustic echo returns (Figure 10 a and b). We interpret this contrast in reflectivity to indicate that the gully channels are filled with hard materials such as coarse gravel lag deposits, while finer sediments have been winnowed away and distributed overbank or 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 10c). Nearly 80% percent of the gullies initiate in water depths <700 m. Gullies are deep, with average width to depth ratios of 15:1, and occur in steep terrain, where average continental slope angles are 10˚ from 500-600 mwd and 13˚ from 600-700 mwd. The gully system whose length

is mapped in Figure 10 a and 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 10 a and b).

To better understand the physical controls on gully formation we analyzed their morphologies as well as their relationship to large-scale topographic features using the ~400 km margin-parallel transect taken by the submarine along the 366 m contour (Figure 10c). 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 10c). 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 2, 9, and 10, gully density is highest where the bathymetric bench is narrowest at ~148˚ W, and then decreases westward as the bathymetric bench widens. 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 10c). Two prominent examples of gullies that coalesced to form trunk channels ~1000m below sea level are indicated in Figure 10 a and b.

4. DISCUSSION

Bedform origin

The bedform ensemble mapped on the ABM generally resembles glaciogenic features found on continental margins and oceanic plateaus elsewhere in the Arctic and in the 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 3) are iceberg plowmarks similar to those documented elsewhere 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 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). The scale of the ABM lineations (Figure 5) is similar to features described elsewhere as 'mega-scale glacial lineations' (Anderson, 1999; Shipp et al., 1999), but we refrain from using this term until the full extent of these features is clarified. Images of lineations in our data (Figure 5) are not as sharp as in some other glaciated shelf areas, possibly due to thicker post-glacial sedimentary cover on the Alaskan margin as corroborated by a recent survey west of the study area (Darby *et al.*, 2005). Nevertheless, the linear nature of these bedforms is conspicuous (Figure 5). Glaciogenic bedforms such as lineations and drumlins 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). We infer that the association of ABM lineations with the topographic

bench is also not random, but reflects their formation by related erosional processes. Many studies document a consistent pattern in the bathymetric distribution of glaciogenic bedforms: lineations 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 lineations further down slope (400-700 m), and gullies originating at the lower edge of the glacially-disturbed areas.

A peculiarity of the ABM erosional features is their orientation parallel to, rather than transverse to, the continental slope, which excludes the possibility of their formation by an ice sheet centered on the Alaskan coast upslope from the study area. Instead, the eroding force had to be directed along the slope. Margin-parallel seafloor furrows on some continental slopes (for example the Greenland margin) 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 lineations. The existence of deeper margin-parallel currents capable of producing slope furrows 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).

Glaciogenic agents generating bedforms in the open ocean, such as the slope-parallel lineations on the ABM discussed here, can be either free-floating icebergs or larger coherent ice masses (ice shelves). It has been proposed that current-driven, massive tabular icebergs could have carved the lineations observed at almost 1000 mwd on the Lomonosov Ridge in the central Arctic Ocean (Figure 1) (Kristoffersen *et al.*, 2004), which are morphologically similar to the features we document on the ABM. However, in contrast to the only ~10 km span of eroded terrain observed on the Lomonosov Ridge, the physically separated but consistent, margin-wide distribution of ABM lineations implies the coherent movement of an eroding ice mass over ~300 km (Figure 2). It is unlikely that even icebergs of the dimensions documented around Antarctica would have the structural integrity or the momentum to ground along this entire stretch of the ABM without changing their drift direction. We infer therefore that the ABM lineations were formed by a large floating ice mass (ice shelf) moving parallel to the continental margin from a grounding line located further east or west. This ice shelf had to have been buttressed against the ABM during a re-grounding event at depths reaching ~700 m (maximal depth of lineated seafloor), which is >500 m deeper than sea level during any of the Quaternary glaciations. This paleoglaciological scenario is somewhat comparable to the history of the Ross Ice Shelf which expanded during some Pleistocene glaciations and impinged on the adjacent coasts (Denton *et al.*, 1989; Hall *et al.*, 2000).

The maximal water depths of lineations 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 eroding ice mass (ice shelf) during its contact with the ABM and the Borderland. 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

333 (Figure 2); furthermore, directional bedforms are overprinted by subsequent chaotic iceberg
334 scours in the shallows. A buried hummocky ridge has been noted near the modern ABM shelf
335 break (~50-60 mwd) south of the lineated areas G and H, and was tentatively interpreted as a
336 barrier island chain (Dinter, 1985). Instead, this formation might be a moraine that marks the
337 landward limit of grounded ice. A dedicated study involving the collection of high-resolution
338 seismic reflection records and sediment cores is needed to delineate the boundaries of this
339 grounding.

340 Rose diagrams comparing the trend of lineations inferred from sidescan maps with the
341 trend of bathymetric contours in the bench areas are closely aligned (Figure 6), suggesting that
342 the eroding ice mass actually carved the bench out of the continental slope. We note that the
343 bench forms a 'second shelf break' along its lower edge, 300+ m deeper than the actual modern
344 shelf break of the ABM (e.g., Figures 4 and 9). Similar overdeepened continental shelf breaks
345 occur only on glaciated margins of the Arctic Ocean (Barents Sea and the Greenland/Canadian
346 margin) (Jakobsson, 2002) and in Antarctica (Anderson, 1999).

347 Spectral analysis of our bathymetric data contributes a quantitative and margin-wide
348 characterization of the bedforms (Figure 8). Where we document the best examples of sub-
349 glacial lineations on the eastern half of the survey in areas E, G and H, amplitudes of seafloor
350 roughness are moderate and cover a narrow range of dominant wavelengths from ~45-100 m. In
351 the westernmost, shallower half of the survey, where iceberg scours and grooves erode the
352 substrate, samples from areas A, B, and C show the strongest signals of seafloor excavation and
353 a wider range of wavelengths from ~20-300 m. These results corroborate the change in bedform
354 character along the ABM seen in sidescan and bathymetry images (Figures 3 and 5).

355 It is possible that areas A, B, and C were impacted by the same large ice mass that eroded
356 and/or molded the seafloor in areas E, G, and H; however, the chaotically oriented iceberg scours
357 and grooves that primarily characterize the western end of the margin likely overprinted the
358 lineated topography. The two types of morphologically distinct scours and grooves observed
359 suggest the possibility of two generations of iceberg activity, one during the collapse of an ice
360 shelf that eroded the ABM, and the other during a later episode of massive iceberg discharge
361 from the Laurentide Ice Sheet. Both types have a similar azimuth range (95-100˚ from north)
362 (Figure 2) indicating that icebergs drifted clockwise along the ABM and impacted the slope at an
363 oblique angle from the northeast. This drift pattern is consistent with the distribution of ice-
364 rafted debris in Pleistocene sediments of the Canada Basin assumed to have been dispersed by
365 icebergs carried by the clockwise Beaufort Gyre (Phillips and Grantz, 2001) (Figure 1).
366
367 **Formation of ABM Slope Gullies**
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369 The co-existence of the slope gullies with the glacially eroded portion of the ABM
370 (Figures 3, 5; 9, and 10), and their morphologic and contextual similarity to gully systems
371 documented on glaciated continental margins around Antarctica (Ó Cofaigh *et al.*, 2003), East
372 Greenland (Ó Cofaigh *et al.*, 2002), and the Gulf of Alaska (Carlson *et al.*, 1990), suggest that
373 the gullies are related to glaciogenic processes. In each of the above localities, ice is inferred to
374 have extended to the continental shelf edge during glacial periods, releasing sub-glacially
375 entrained sediments and subglacial meltwater where the ice front separated from the seafloor.
376 Sediments were subsequently remobilized by down slope gravity flows and turbidite gullies were
377 eroded into the upper continental slope. Ó Cofaigh *et al.* (2003) argue that where slope gradients

are sufficiently steep, any sub-glacial sediments available at the shelf edge may be transported directly to the abyssal plain, 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 (up to 13˚ between 600-700 mwd) to retain any glacially deposited sediments on the shelf edge.

The geomorphology of the ABM shows evidence for the flow of grounded ice parallel to slope isobaths (Figure 2), instead of transverse to the shelf edge, as in the ice stream localities described above. As a result, it appears that ice probably did not deliver sediments to the ABM slope directly below the area of erosion. 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 sliding ice to float over open water, but the majority of the sediments were probably moved parallel to the slope in the direction of dominant ice flow. Thus most sub-glacially entrained sediments may not have been released along the portion of the ABM margin mapped during SCICEX-99, effectively starving the upper continental slope of sediments.

Large-scale mass-wasting features on the ABM, involving >200 m of sedimentary cover, have been described in early seismic-reflection studies (e.g., Grantz *et al.*, 1979, 1981), but their exact relationship to slope morphology and the mechanisms of their formation have yet to be identified. Seismic processes on the ABM and the decomposition of gas hydrates during low sea-level stands have been discussed as likely controls on slope instabilities (Grantz *et al.*, 1981; Kayen and Lee, 1991). From the existing data we cannot conclude whether the mapped ABM

401 gullies and other large-scale slumps are related; high-resolution subbottom data from the gullied
402 slope are needed to clarify this relationship.
403
404 **Ice flow provenance and trajectory**
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406 The observed lineation orientation along the ABM could indicate ice flow in either an
407 easterly or westerly direction. In the absence of unidirectional bedforms such as drumlins,
408 dragged boulders, or larger-scale erosional and depositional features, we have used the broader
409 geological context to determine the likely ice flow direction. At least two fields of lineations on
410 the Chukchi Borderland (Figure 1) with a similar, somewhat more northward orientation (130-
411 135° at the southern margin) and water depths down to 750+ m, show direct evidence for a
412 southeast to northwest ice flow path in the form of drumlins and dragged boulders (Polyak *et al.*,
413 2001; Jakobsson *et al.*, 2005). The broad, over-deepened troughs traversing the western part of
414 the CAA (Amundsen Gulf, M'Clure Strait) and, possibly, the partially buried Mackenzie Trough
415 are the likely outlets for large Laurentide ice streams capable of feeding ice shelves in the
416 Canada Basin (Figure 1). This interpretation is consistent with the glacial geology and
417 geomorphology of these troughs and adjacent areas (Blasco *et al.*, 1990; Dyke *et al.*, 1992;
418 Stokes *et al.*, 2005; 2006) and with models of ice discharge (Bigg and Wadley, 2001), all
419 indicating the likelihood of large ice streams channeling the northwest margin of the Laurentide
420 Ice Sheet. The CAA provenance of the ice that contacted the ABM is corroborated by the
421 composition of coarse clastic material in diamict and lag deposits along this margin (Rodeick,
422 1979; Naidu and Mowatt, 1992), although the exact origin of these deposits (subglacial or
423 iceberg-derived) still needs to be identified.

424 The latitudinally-oriented, margin-parallel glaciogenic bedforms on the ABM appear to
425 provide a 'missing link' between the inferred northwest Laurentide sources of ice and the
426 northwestward erosion of the Chukchi Borderland. This interpretation implies that the eroding
427 ice mass flowed along the periphery of the Canada Basin rather than into its center. In order to
428 achieve this pathway, ice flow had to be blocked or diverted by some type of barrier. Polyak *et*
429 *al.*, (2001) suggest that ice was propelled along the coastline and across the CB by the force of
430 CAA/Mackenzie ice streams (Figure 1). Based on ice mass balance considerations, however,
431 this ice would need to be buttressed on both sides of the stream to maintain its full thickness
432 without calving and breaking apart on its northern, oceanward margin. Continuous sea ice cover
433 during glacial periods might briefly retard the calving rate along an ice sheet margin (Reeh *et al.*,
434 2001) but it is unlikely that sea ice with a maximal thickness of ~7 m when concentrated by wind
435 (Bourke and Garrett, 1987), could buttress an advancing ice shelf whose thickness is two orders
436 of magnitude greater.
437 An alternative mechanism to explain the circuitous flow path of ice along the basin
438 margin invokes the existence of a thick ice shelf capping the entire Arctic Ocean (Mercer, 1970;
439 Hughes *et al.,* 1977; Grosswald and Hughes, 1999). In these models, the basin-wide ice shelf,
440 fed by multiple sources around its periphery, could have prevented land-sourced ice streams
441 from penetrating directly into the central basins (Grosswald and Hughes, 1999). Evidence for
442 glacial erosion on top of the Lomonosov Ridge (Figure 1) (Jakobsson, 1999; Polyak *et al.*, 2001)
443 makes this scenario possible, provided there were synchronously growing ice shelves on both the
444 Eurasian and Laurentide sides of the ocean. Based on preliminary, simplified numerical
445 modeling (Jakobsson *et al.*, 2006), the growth of an ice shelf several hundred meters thick from
446 the Eurasian margin all the way to the Lomonosov Ridge is possible provided there is a condition

447 that constrains the stretching of ice. Such a constraint might be provided by abnormally thick
448 sea ice or piled-up icebergs during glacial stages, but this hypothesis needs to be tested by more
449 refined modeling based on comprehensive, age-specific paleogeographic reconstructions. For
450 example, the limited extent of the Eurasian ice sheet in the Arctic during the Last Glacial
451 Maximum (Svendsen *et al.*, 2004) refutes the concept of the pan-Arctic ice shelf during that time
452 period, but earlier glacial periods could have been more favorable for its formation. Notably, the
453 age of glacial erosion on the Lomonosov Ridge is estimated at Oxygen Isotope Stage (OIS) 6,
454 >130 ka (Jakobsson *et al.*, 2001).
455 An alternative interpretation of the Lomonosov Ridge erosion suggests that it was
456 produced by an armada of huge tabular icebergs, rather than an ice shelf (Kristoffersen *et al.*,
457 2004). In this scenario, Antarctic-type icebergs purged from the glaciated Barents-Kara
458 continental margin were entrained in thick sea ice and then dragged across the crest of the
459 Lomonosov Ridge by northward flowing Atlantic water advected through the Fram Strait (Figure
460 1). If this were the case, we suggest that these tightly packed mega-bergs could then have
461 proceeded into the Amerasian Basin where they contributed to a pile-up of wind and current-
462 driven ice that deflected the path of ice streams coming out of the mouths of the CAA troughs.
463
464 **Age control**
465
466 The age of maximal glacial erosion on the Chukchi Borderland has been tentatively
467 estimated between OIS 4 to 5d (Polyak *et al.*, 2007). Because the prevalent orientation of
468 glaciogenic lineations representing this erosion on the CB align spatially with those at the ABM
469 (Figure 1) and have a similar 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. The encroachment of grounded ice onto the ABM could have affected sedimentation on the margin by directly depositing sub/proglacial till and/or by causing a glacioisostatic depression that was later 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 hypothesize 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 (Dinter *et al.*, 1990; Brigham-Grette and Hopkins, 1995). If this scenario is correct, ice shelf encroachment on the ABM would have immediately preceded the deposition of the Flaxman Formation dated to ca. 70-80 ka, that is OIS 5a (Dinter *et al.*, 1990; Brigham-Grette and Hopkins, 1995). Accordingly, the ice-grounding event would likely have occurred during OIS 4 or an interstadial within OIS 5, which matches the age range estimated for the erosion of the Chukchi Borderland (Polyak *et al.*, 2007) and for an inferred widespread glaciation in the Beringia region (Brigham-Grette *et al.*, 2001).

The inferred age of ice shelf passage along the ABM is younger than the OIS 6 age estimate for the erosion found on the Lomonosov Ridge (Jakobsson *et al.*, 2001). It is conceivable nevertheless that glacial periods postdating OIS 6 could have generated ice shelves

493 or iceberg armadas nearly comparable in size to prior surge events, but not quite thick enough to
494 have scoured the deep crest of the Lomonosov Ridge at ~1000 mwd. Furthermore, the mega-
495 berg purge trajectory could vary slightly from one glacial period to another, allowing some very
496 large icebergs to miss the shallowest portions of the Lomonosov Ridge but still collect in the
497 Amerasian Basin.
498
499 **5. CONCLUSIONS**
500
501 Sidescan and bathymetry data obtained during the SCICEX program illustrate the
502 morphology of the Alaska/Beaufort margin in the depth range of 200-2800 m, and help to
503 elucidate the sedimentary and glacial history of this part of the Arctic Ocean. Along several
504 transects the ABM data show an orderly association of glaciogenic seafloor bedforms such as
505 scours, grooves, and lineations, and larger-scale erosional bathymetric features in water depths
506 reaching 700 m, with multiple turbidite gullies further down slope. Although our data coverage
507 is more fragmentary than in some other survey areas, these bedforms are recognizable as
508 geomorphic associations known from glaciated continental margins in the Arctic and Antarctic
509 (e.g., Anderson, 1999; Shipp *et al.*, 1999; Polyak *et al.*, 2001; Ó Cofaigh *et al.*, 2003;
510 Dowdeswell *et al.*, 2004). At 400-700 mwd, fields of sub-glacially generated lineations cover an
511 estimated minimum seafloor area of 330 km^2, indicating margin-parallel flow of a large ice mass
512 buttressed against the ABM. Lineations occur on the surface of a bathymetric bench at 400-550
513 m that may have been formed by ice erosion, an interpretation that is corroborated by the
514 coincident trends of lineations and bathymetric contours on the bench surface. We note also that
515 this large-scale erosion of the continental slope makes the ABM similar to overdeepened

516 continental margins that are characteristic of formerly glaciated areas. Iceberg scours and
517 grooves observed on the ABM in shallower water depths were probably generated during ice
518 shelf break-up as well as later glacial event(s). The bench is truncated to the north by steep
519 dendritic gullies that initiate in branching catchment areas up to 18 km across and coalesce to
520 form broad trunk channels in water depths >1500 m. Gullies are interpreted to have formed in
521 response to sediment-laden sub-glacial meltwater cascading down the steep continental slope.
522 We acknowledge that the data presented are fragmentary and therefore are not as conclusive as
523 those documented for some other glaciated margins. Further ABM surveying is needed to
524 delineate the areal distribution of glaciogenic bedforms and to verify their spatial consistency.
525 ABM glaciogenic bedforms have depths, styles, and orientations similar to features
526 discovered earlier on the Chukchi Borderland (Figure 1) (Polyak *et al.*, 2001; 2007; Jakobsson *et*
527 *al.*, 2005) and were probably formed during the same glacial episode. ABM bedforms link CB
528 glaciogenic features with the over-deepened troughs of the CAA that were the likely outlets for
529 ice streams from the northwest sector of the Laurentide ice sheet, and which could have
530 propelled ice shelves into the Canada Basin. The margin-parallel pathway of this ice flow along
531 the ABM indicates that incoming ice was diverted from streaming directly into the central part of
532 the basin by some obstruction. We infer that such an obstruction was formed by either a basin-
533 wide ice shelf fed by multiple sources, or by a pile-up of Antarctic-style icebergs driven by
534 winds and currents from the Eurasian side of the Arctic Ocean. Corroboration of the movement
535 of very thick ice (~1000 m), be it an ice shelf or iceberg armadas, towards the Canada Basin, is
536 seen in the eroded crest of the Lomonosov Ridge at the center of the Arctic Ocean (Jakobsson,
537 1999; Polyak *et al.*, 2001; Kristoffersen *et al.*, 2004; Jakobsson *et al.*, 2006).

538 We propose that the sliding of a Canada Basin ice shelf over the ABM is causally linked to
539 the diamict Flaxman Member of the Gubik Formation that mantles the ABM coast up to 7 m
540 above sea level and contains numerous coarse clasts of CAA provenance (Dinter *et al.*, 1990;
541 Brigham-Grette and Hopkins, 1995). This linkage can explain the unusual geology of the
542 Flaxman Member and its lack of glacioisostatic tilt towards the CAA. We propose that the
543 grounding of several hundred meter thick ice on the ABM shoulder isostatically depressed the
544 entire margin and allowed for a glaciomarine transgression of the coast. In this case, the age of
545 the ice grounding on the ABM immediately precedes the age of the Flaxman Member, that is
546 OIS 4 or 5b. This timing may indicate a correlation with an inferred widespread glaciation in the
547 Beringian region (Brigham-Grette *et al.*, 2001) and fits the age range estimated for glacial
548 erosion on the Chukchi Borderland (Polyak *et al.*, 2007).

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550

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708 **FIGURE CAPTIONS**

709

710 Figure 1 - Overview map of the Arctic Ocean with bathymetry from the International
711 Bathymetric Chart of the Arctic Ocean (IBCAO) (Jakobsson *et al.*, 2000). Fields of lineations on
712 the Chukchi Borderland and Lomonosov Ridge shown by red, pink, and orange arrows/lines
713 (from Polyak *et al.*, 2001); arrows show direction of eroding ice, where reconstructed. ABM
714 study area and inferred ice flow direction shown in yellow box (this study). Subbottom profile
715 (Figure 4) location shown by yellow star. Maximal limit of Late Pleistocene Eurasian and North
716 American ice sheets shown with double white line (Dyke *et al.*, 2002; Svendsen *et al.*, 2004).
717 Projected deep-draft ice sources shown by dashed white arrows. Beaufort Gyre shown by
718 dashed blue circle, Pacific water ABM boundary current shown by dashed purple arrow (Pickart,
719 2004). CB – Chukchi Borderland, MT – Mackenzie Trough, AG – Amundsen Gulf, MS –
720 M'Clure Strait, YP – Yermak Plateau.

721

722 Figure 2 - ABM survey area with bathymetry, survey track (in black), and the schematic
723 orientations of mapped seafloor glaciogenic features within each study area. Heavy black lines
724 indicate the presence and orientation of lineations, red lines indicate the mean orientations of
725 iceberg scours or grooves. The bathymetry data are a combination of gridded data from IBCAO
726 (Jakobsson *et al.,* 2000) and swath data collected by the *USCGC HEALY, R/V Nathaniel B.*
727 *Palmer* and the *USS Hawkbill* (Edwards at Coakley, 2003*;* Edwards *et al.,* 2005). Bathymetry is
728 displayed as color-coded contours overlaid on a sun-shaded digital terrain model illuminated
729 from the east. Note the deeply eroded gullies visible in the bathymetry data north of the
730 continental shelf.

731 Figure 3 – Sidescan (upper) and bathymetry (lower) SCICEX swath data for area A (see Figure 2
732 for location). The bathymetry data are displayed as color-coded contours overlaid on sun-shaded
733 topography illuminated from 005° at an altitude of 10° above the horizon to emphasize small-
734 scale relief. Scours and grooves (indicated) are most apparent in the higher resolution sidescan
735 data (8 m grid cells) with its higher angle of ensonification, but all features show some relief in
736 the lower resolution (25 m grid cells) bathymetric data.
737
738 Figure 4 - Fragment of a chirp-sonar subbottom record west of the study area (SCICEX-98; see
739 Figure 1 for location) illustrating iceberg scouring to water depths of between 350-400 m and
740 truncation of the inclined bedrock strata on the slope, with an erosional bench between 450-500
741 m.
742
743 Figure 5 –Sidescan (upper) and bathymetry (lower) SCICEX swath data for area H (see Figure 2
744 for location). The bathymetry data are displayed as color-coded contours overlaid on sun-shaded
745 topography illuminated from 355° at an altitude of 10° above the horizon to emphasize small-
746 scale relief. The presence of lineations in the bathymetry (indicated) is more strongly correlated
747 with their apparent width in the sidescan data rather than the total vertical relief. Deep, narrow
748 lineations in the 8 m resolution gridded sidescan data can be difficult to detect in the 25 m
749 bathymetry gridded data, but wider lineations are apparent across the entire swath.
750
751 Figure 6 - Average trends of bathymetric features for the lineated areas E, G, and H: a) large-
752 scale bathymetric features such as such as the slope bathymetric contours, the bench alignment;
753 and the shelf/slope break, b) lineations in sidescan sonar data. In each rose diagram the mean

754 trend is indicated by the dashed line and is significant at the 95% level using Rayleigh's test for
755 trend (Swan and Sandilands, 1995).
756
757 Figure 7 – 8 m resolution gridded sidescan data for Area G (see Figure 2 for location). Note the
758 depth-dependence of lineations (~350-600 mwd in this image) versus grooves (shallower than
759 350 mwd).
760
761 Figure 8 – Spectral estimates of seafloor roughness for the sites listed on the x-axis (see Figure 2
762 for locations) with distance from the eastern end of the survey area (upper panel), and with equal
763 spacing between bathymetric lines analyzed to facilitate visualization (lower panel). The
764 strength of the signal of seafloor excavation within one site and across the margin is shown in dB
765 using rainbow colors. Color change from red to dark blue values indicates the change in
766 repeating wavelengths from the most deeply excavated to the least excavated terrain. The
767 spacing between glaciogenic features is denoted on the y-axis in meters.
768
769 Figure 9 - 3-D view of the ABM that highlights the bathymetric bench most prominent along the
770 margin in Area H (see Figure 2 for location and description of the bathymetric datasets).
771 SCICEX survey track shown in black, the 350 m and 600 m contours that bound the bench are
772 shown in red and blue, respectively. The bathymetry data are displayed as color-coded contours
773 overlaid on a sun-shaded digital terrain model illuminated from the north. The diagonal stripes
774 in the bathymetry data are artifacts that result from resampling the IBCAO data at higher
775 resolution than their 2.5 x 2.5 km grid cell size. CB – Chukchi Borderland. Inset shows SCICEX
776 slope profile of Area H along the transect indicated with heavy black line on the 3-D map.

777 Figure 10 – (a) sidescan image of an extensive dendritic gully system coalescing into trunk
778 channels (area E on Figure 2). (b) interpreted outline of the gully system. (c) Highest resolution
779 SCICEX (nadir) bathymetry of ABM topography collected along the returning submarine track,
780 which surveyed approximately parallel to the 366-m isobath. Water depths between 200-600 m
781 are highlighted. Arrows indicate typical examples of gullies along the margin.