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**

Jennifer L. Engels, University of Hawaii, Department of Geology and Geophysics, 1680 East-West Road, Honolulu, HI 96822, phone: (808) 956-8489, fax: (808) 956-5512, engels@hawaii.edu

Margo H. Edwards, University of Hawaii, Hawaii Institute of Geophysics and Planetology, 1680 East-West Road, Honolulu, HI 96822, margo@soest.hawaii.edu

Leonid Polyak, Ohio State University, Byrd Polar Research Center, Columbus, OH, polyak.1@osu.edu

Paul D. Johnson, University of Hawaii, Hawaii Mapping Research Group, 1680 East-West Road, Honolulu, HI 96822, pjohnson@soest.hawaii.edu

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 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; 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

61 reconstructing the history of Pleistocene Arctic ice shelves is the Chukchi Borderland (CB),
62 a complex of geological structures extending north of the Chukchi continental shelf, that
63 shows ice shelf related bedforms at depths reaching 1000 m and chaotic iceberg scours at
64 shallower depths (Figure 1) (Polyak *et al.*, 2001; in press; Jakobsson *et al.*, 2005; Darby *et*
65 *al.*, 2005). The northwest-trending orientation of glacially-derived flutes on the CB
66 suggests that the northwestern margin of the Laurentide ice sheet was the likely source of
67 eroding ice. This trajectory further implies that the ABM could have been affected by the
68 same large ice masses and it may therefore be an important area to clarify the extent and
69 pathways of Pleistocene ice shelves in the western Arctic Ocean.
70
71 **1.1 Physiographic and geological background for the Alaska/Beaufort margin**
72
73 The ABM extends from the base of the Chukchi Borderland eastward to the
74 Mackenzie River delta (Figures 1 and 2). This aseismic, Atlantic-type continental margin
75 has a continental shelf measuring on average 75 km wide that ranges in depth from 0-50 m.
76 The margin is bounded to the north by a steep, heavily canyoned continental slope that
77 drops to 1000 meters water depth (mwd) in the Canada Basin over an average distance of
78 just 13 km. To the south, the continental shelf grades into a broad, low relief plain that
79 extends to the foothills of the Brooks Range. The plain is mantled by 10s-100 m of the
80 unconsolidated clastic materials of the Gubik Formation, interpreted to be a series of
81 glaciomarine transgressive deposits dating from 3 Ma through the Last Glacial Maximum
82 (Dinter *et al.*, 1990 and references therein). Coarse clasts, from gravel to boulders, with
83 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 or an earlier glaciation (Dinter *et al.*, 1990; Brigham-Grette and Hopkins, 1995). However, these authors also pointed out that a large-scale glaciation cannot easily be reconciled with high sea level in a generally stable tectonic environment, and the mapped formation shows no isostatic tilt along its length towards the Laurentide ice sheet, which requires some other mechanism for the delivery of glaciogenic deposits onto the Alaskan shelf and coastal plain.

2. METHODS

Until recently, many portions of the Arctic Ocean had never been or were only sparsely surveyed due to the constraints of operating in a perennially ice-covered and climatically inhospitable ocean. Traditional arctic surveying methods such as the use of floating ice islands, icebreakers, and aircraft suffer from restricted operations because they are subject to seasonal ice and weather conditions. Between 1995-1999 the U.S. Navy invited researchers in the academic community to design and participate in a series of cruises entitled the SCience ICe EXercises (SCICEX) (Edwards and Coakley, 2003), conducted aboard Sturgeon-class submarines that travel below and surface through sea ice. During the 1999 SCICEX cruise ~2000 km of 12 kHz sidescan sonar and interferometric bathymetry data were collected by the nuclear submarine *USS Hawkbill* along a 680 km

107 portion of the ABM between 140-160° W (Figures 1 and 2). The submarine mapped
108 seafloor from 200-2800 m deep along the ABM in a zigzagging pattern designed to
109 accommodate a water-sampling program (Figure 2). Sidescan swath widths vary from 6 km
110 on the continental shelf to >20 km at the base of the continental slope. Bathymetry swath
111 width is less than the sidescan swath width, averaging ~4 times the water depth. Where
112 seafloor ensonification angles exceed 80° in the shallowest survey areas, low-relief
113 morphologic features are emphasized at the outer edges of the sidescan swaths.
114 Unfortunately, subbottom sonar (chirp) data were not recorded in conjunction with the 1999
115 sidescan survey on the ABM, but several chirp lines were collected west of this area during
116 the 1998 SCICEX cruise. Sidescan and bathymetry data were iteratively processed to
117 remove artifacts that reduce image quality (Davis *et al.*, 2001). Sidescan data were gridded
118 in 8 m cells and bathymetry data were gridded in 25 m cells to produce swath imagery for
119 ~14,100 km^2 of seafloor. The gridded bathymetry data have a vertical resolution of ~1% of
120 water depth. Higher resolution (0.5% of water depth) soundings collected directly under
121 the submarine every 12 seconds were sub-sampled to create twenty slope profiles used for
122 quantifying seafloor morphology.
123 Geospatial reference for the ABM survey track locations was generated by the
124 submarine's inertial navigation system (INS) and was then re-navigated using surface GPS
125 fixes (Edwards and Coakley, 2003). The INS incurs errors that propagate and exacerbate
126 with time spent under the ice and away from the last GPS reading. The ABM survey took 6
127 days and the submarine surfaced only at the beginning and end of the survey, resulting in
128 relative positional navigation errors of ~3 km towards the western end of the survey based
129 on data crossover analyses. SCICEX data were co-registered with the International

130 Bathymetric Chart of the Arctic Ocean (IBCAO) (Jakobsson *et al.*, 2000), which is the most
131 comprehensive compilation dataset of Arctic bathymetry currently available and is
132 generated from icebreaker and submarine data soundings contained in archives in the
133 United States, Canada, Denmark, Russia, Germany, and Iceland. Co-registration of
134 SCICEX and IBCAO data shows good correspondence between the two datasets, and we
135 have used them in concert to characterize bathymetry along the ABM (Figure 2).

136 To document the extent and distribution of seafloor bedforms in the ABM sidescan
137 imagery we analyzed measurements of bedform length, width, elongation ratio, transverse
138 wavelength, and percentage bedform area per unit area using the Environmental Systems
139 Research Institute's (ESRI) ArcMap GIS program. Measurements represent minimum
140 lengths in all cases because sidescan sonar data were projected onto a flat surface, when in
141 reality the seafloor surface is sloped. A margin-parallel transect taken by the submarine on
142 its returning east-west survey track followed the 200 fathom (366 m) contour. This transect
143 is used to characterize the bathymetric roughness of gullies observed along the continental
144 slope including the frequency, distribution, style, depth and width of gullies, and depth of
145 gully onset. In one locality it is possible to observe almost the full length of an individual
146 gully system, and this example is used to infer gully dynamics and configurations
147 elsewhere.

148 In order to quantify whether there is a systematic progression of seafloor
149 morphological changes along the ABM, we performed a one-dimensional Fourier analysis
150 on transects of the full resolution swath bathymetry. To minimize outside factors that
151 degrade data quality, we chose twenty profiles showing distinctive bedforms in the sidescan
152 imagery that were characterized by submarine movement in a single direction at a constant

153 speed. We corrected for the oblique submarine orientations (relative to dominant bedform
154 trend) by projecting each vehicle track perpendicular to the bedform trend, and then
155 resampled the data to generate consistent data densities for each sample area. The slope
156 trend and mean value for each sample area were removed and the projected, resampled, and
157 detrended data were filtered to remove long wavelength noise. These data were analyzed
158 using a Fast Fourier Transform (FFT) program and plotted as power spectra to highlight
159 recurring wavelengths (Figure 3). The FFT amplitudes serve as proxies for the variance in
160 height of seafloor morphological features.
161
162 **3. RESULTS**
163
164 The SCICEX data display a recognizable association of seafloor bedforms,
165 distinctive bathymetry, and slope mass-wasting features that are found repeatedly along the
166 length of the ABM. Eight separate regions (Figure 2: A-H) totaling ~550 km^2 show
167 evidence of disturbed seafloor sediments in <250-700 mwd. A range of linear to arcuate
168 bedforms are observed, from randomly oriented, overlapping scours (Figure 4) to fields of
169 parallel, low-relief lineations (flutes) in excess of 10 km long (Figure 5). The basinward
170 terminations of the disturbed zones are marked by deeply incised dendritic gullies that drain
171 the continental slope (Figures 4-6). The disturbed seafloor is always at the shallowest end
172 of the survey track on or immediately above a prominent low-angle bathymetric bench that
173 ranges from 5 to >35 km wide (Figure 7).

174 **3.1 Bedforms**

175

176 Morphological features in zones of disturbed terrain fall into three distinct types
177 (Figures 4-5). Type 1 bedforms are chaotically oriented, overlapping scours <2 km in
178 length and <25 m wide (Figure 4) concentrated in water depths <500 m, predominantly
179 <400 m (Figure 2). Type 2 bedforms are more deeply incised, arcuate to nearly linear
180 individual grooves up to 5 km long, <50 m wide, and occur throughout the study area in
181 <500 mwd (Figure 4). In all locations where both Type 1 scours and Type 2 grooves are
182 observed, the grooves overprint the chaotically oriented scours. Bathymetric data show that
183 negative-relief scour and groove furrows are commonly bordered by positive-relief levees;
184 total relief measured from the high-resolution bathymetric profiles is up to 10 m. Scours
185 and grooves occur in fields of several hundred overlapping individual features with a range
186 of trends 60-135º from north, but fields have a dominant overall fabric of ~100º and most
187 trend obliquely down slope (Figures 2 and 8). A chirp-sonar subbottom record west of the
188 study area (location shown in Figure 1) illustrates multiple gouges at the upper slope
189 reaching to ~350-400 mwd (Figure 9). Seafloor sediment in the gouged area is reworked to
190 a depth of several tens of meters; the bottom of this sedimentary unit commonly forms a
191 prominent reflector that truncates underlying strata.

192 Type 3 bedforms, parallel flutes, are mapped between ~400-700 mwd in three
193 regions on the eastern half of the survey (Figure 2: E, G, and H). They reach >10 km in
194 length and measure ~50 m wide yielding very high length to width ratios of up to 200:1
195 (Figure 5). The true coverage and length of the flutes are underrepresented as they extend
196 beyond the edges of the sidescan data. Flutes parallel isobaths trending 103-147º from

197 north (average trend 112˚, Figure 8) and occur in fields that are 10->30 km wide to cover a
198 minimum total area of 330 km^2. The total areal coverage of the Type 3 fluted bedforms can
199 be inferred from their distribution along the ABM (Figure 2): if flutes continue beyond
200 mapped areas in similar water depths between areas E, G, and H, their along-margin extent
201 totals ~300 km. The flutes documented in areas G and H are overprinted by Type 2 grooves
202 in the shallowest parts of the survey area.

203 Spectral analysis is used on twenty bathymetric lines along the length of the survey
204 area to compare the distances between and the amplitudes of the seafloor bedforms
205 described above (Figure 10). The amplitude of spectral estimates gives an indication of the
206 relative seafloor roughness at each wavelength. In Figure 10, colored amplitudes indicate
207 the scale of roughness of the seafloor at the wavelengths indicated in meters on the y-axis.
208 For example, in sample area C3, there is high amplitude roughness of the seafloor with
209 repeated spacing of ~45 m perpendicular to the sample transect and parallel to the flute
210 trend, while in sample G2 the highest amplitude seafloor roughness within that area (which
211 is lower amplitude than that observed in area C3) occurs with spacing of ~60 m. In Figure
212 10 high amplitude roughness centers around the 50 m wavelength, consistent with the
213 distance between flutes measured independently from the sidescan imagery. On the
214 western end of the survey in areas A, B, and C where scours and grooves prevail, the
215 seafloor is more deeply excavated than in areas E, G, and H where fields of flutes occur.
216 Depths at which the scours and grooves are observed in sample areas A, B, and C are
217 shallower than 370 m with a mean of 290 m; flutes in areas E, G, and H are found in water
218 depths averaging 400 m and as great as 570 m.

219 **3.2 Bathymetric bench**

220

221 The SCICEX bathymetry documents the co-occurrence of the seafloor bedforms
222 described above with the location of a flattened bathymetric bench that is 400-550 m deep
223 and has an average slope of just 1.6º (Figures 4, 5, and 7). The bench morphology is also
224 visible in IBCAO bathymetry for the Beaufort Sea (Jakobsson, 2002) and in a new
225 bathymetry compilation for the Arctic Ocean (Figure 7) (Edwards *et al.,* 2005). The
226 basinward margin of the bench is associated with the onset of dendritic gullies and a
227 precipitous increase in slope to 13º. Within the study area the bench width varies from 5-
228 >35 km, with the broadest portions mapped for ~100 km along the easternmost end of the
229 surveyed area (Figure 7). On a subbottom record west of the study area the continuation of
230 the bench is identified at ~480 m depth with a width of ~3-4 km (Figure 9). The truncation
231 of the slope face strata suggests the erosional character of the bench. The widest portion of
232 the bench underlies areas G and H (Figures 2 and 7) where the most areally extensive
233 examples of fluted seafloor are documented. Bathymetric features on the bench surface are
234 mostly subdued, but in fluted areas the bench exhibits undulations in the slope profiles with
235 relief as great as 25 m (Figure 7).

236 Figure 8 shows the close match of the average trend of flutes measured from
237 sidescan sonar imagery with the average trends of broad-scale morphological landmarks
238 (for example the shelf/slope break on the northern margin of the bench) for the same areas.
239 The average trend of flutes as mapped from sidescan imagery is 112˚, and the average trend
240 of bathymetric landmarks for the same locations is 114˚ from true north (significant at the
241 95% level using Rayleigh's test for trend (Swan and Sandilands, 1995)).

242 **3.3 Slope gullies**

243

244 Throughout the survey area a well-developed series of gullies erodes the continental
245 slope starting at the northern margin of the bathymetric bench (Figures 2 and 4-6) with an
246 average of one gully every 7 km (59 gullies in total over the length of the ~400 km margin-
247 parallel transect taken by the submarine along the 366 m contour). Gully systems are easily
248 recognizable in sidescan sonar imagery due to the marked contrast in acoustic reflectivity
249 between the highly reflective gully channels and the surrounding sedimentary substrate that
250 absorbs sound and yields low acoustic echo returns (Figure 6 a and b). We interpret this
251 contrast in reflectivity to indicate that the gully channels are filled with hard materials such
252 as coarse gravel lag deposits, while finer sediments have been winnowed away and
253 distributed overbank or further downstream. Gully channel depths measured vertically
254 range from 80-770 m (average 307 m), and gully widths range from 0.8-18.1 km (average
255 5.5 km) (Figure 6c). Nearly 80% percent of the gullies initiate in water depths <700 m.
256 Gullies are deep, with average width to depth ratios of 15:1, and occur in steep terrain,
257 where average continental slope angles are 10˚ from 500-600 mwd and 13˚ from 600-700
258 mwd. The gully system whose length is mapped in Figure 6 (a and b) measures in excess of
259 40 km long and covers depths ranging from 250->1500 mwd before continuing out of the
260 sidescan swath image. The thalwegs of the gullies in the dendritic catchment system
261 (ranging in width from 0.7-10 km) coalesce to form broad trunk channels (Goff and
262 Nordfjord, 2004) in deeper water depths from ~1000-1500 m (Figure 6 a and b).
263 To better understand the physical controls on gully formation we analyzed their
264 morphologies as well as their relationship to topographic landmarks using the ~400 km

265 margin-parallel transect taken by the submarine along the 366 m contour (Figure 6c). Gully
266 walls generally are the same height on their east and west sides, with no trend of asymmetry
267 as a function of distance along the ABM. A striking attribute of the gully transect is the
268 flattened character of the shallowest portions of the seafloor compared to the gullied terrain
269 (Figure 6c). At ~400-600 mwd the seafloor appears horizontally planed off with gullies
270 downcutting through and dissecting this surface. This upper, flattened surface corresponds
271 to the previously mentioned bathymetric bench. As seen in Figures 2, 6 and 7, gully density
272 is highest where the bathymetric bench is narrowest at ~148˚ W, and then decreases
273 westward as the bathymetric bench widens. In addition, the character of the gullies changes
274 from predominantly shallow, narrow gullies where the bathymetric bench is narrowest, to
275 wider and deeper gullies at the westernmost end of the survey. Where multiple individual
276 gullies coalesce, the walls of several successive gullies do not reach shallow enough depths
277 to intersect the flattened bench surface (Figure 6c). Two prominent examples of gullies that
278 coalesced to form trunk channels ~1000m below sea level are indicated in Figure 6 (a and
279 b).
280
281 **4. DISCUSSION**
282
283 **4.1 Bedform origin**
284
285 The bedform ensemble mapped on the ABM shows a remarkable resemblance to
286 glaciogenic features found on continental margins and oceanic plateaus elsewhere in the
287 Arctic and in the Antarctic (e.g., Davies *et al.*, 1997; Anderson, 1999; Shipp *et al.*, 1999;

288 Polyak *et al.*, 2001; Dowdeswell *et al.*, 2004). Chaotically oriented to sub-parallel scours
289 and grooves (Figure 4) are iceberg plowmarks similar to those documented elsewhere in
290 regions of modern and past glaciation, where more linear grooves result from the persistent
291 effects of winds or oceanic currents on icebergs (e.g., Barnes and Lien, 1988; Vogt *et al.*,
292 1994). Parallel lineations or flutes have been described in numerous glaciated regions and
293 are interpreted to form by an ice sheet or ice shelf sliding over the seafloor (Josenhans and
294 Zevenhuizen, 1990; Polyak *et al.*, 1997; Anderson, 1999; Shipp *et al.*, 1999). Images of
295 flutes in our data (Figure 5) are not as sharp as in some other glaciated shelf areas, possibly
296 due to thicker post-glacial sedimentary cover on the Alaskan margin as corroborated by a
297 recent survey west of the study area (Darby *et al.*, 2005). Nevertheless, the linear nature of
298 these bedforms is conspicuous (Figure 5). Glaciogenic bedforms such as flutes and
299 drumlins can be associated with larger bathymetric features indicative of extensive erosion
300 of seafloor deposits such as erosional troughs and truncated ridge tops (Polyak *et al.*, 2001;
301 Canals *et al.*, 2002; Lowe and Anderson, 2002). Many studies document a consistent
302 pattern in the bathymetric distribution of glaciogenic bedforms: flutes are typically detected
303 in the deepest parts of glacially eroded areas but are truncated and obliterated by iceberg
304 scours in the shallows, while basinward continental slopes show mass-wasting features,
305 with mudflows and turbidite gullies dominating low-angle and steep slopes, respectively
306 (Shipp *et al.*, 1999; Polyak *et al.*, 2001; Ó Cofaigh *et al.*, 2003; Dowdeswell *et al.*, 2004).
307 In agreement with the associations described in other studies, the ABM exhibits iceberg
308 scours at the shallowest depths in the study area (mostly <400 m), parallel flutes further
309 down slope (400-700 m), and gullies originating at the lower edge of the glacially-disturbed
310 areas.

311 A peculiarity of the ABM flutes is their orientation parallel to, rather than transverse
312 to, the continental slope, which excludes the possibility of their formation by an ice sheet
313 centered on the Alaskan coast upslope from the study area. Instead, the eroding force had
314 to be directed along the slope. Margin-parallel seafloor lineations on some continental
315 slopes (for example the Greenland margin) are produced by contour currents (Kuijpers *et*
316 *al.*, 2002). Oceanographic data from the ABM show evidence for a seasonal subsurface
317 boundary current (Pickart, 2004), but it occurs at depths of 100-150 m, much shallower than
318 the observed ABM flutes. The existence of deeper margin-parallel currents capable of
319 producing slope furrows would have to be related to the formation and transit of deep
320 waters, such as along the Greenland margin, and is not observed in the Canada Basin today.
321 It is even less likely that deep currents would have formed during glacial periods when the
322 closure of the Bering Strait and CAA channels further restricted circulation in the Arctic
323 Ocean (Goosse *et al.*, 1997a, b).

324 Glaciogenic agents generating bedforms in the open ocean, such as the slope-
325 parallel lineations on the ABM discussed here, can be either free-floating icebergs or larger
326 coherent ice masses (ice shelves). It has been proposed that current-driven, massive tabular
327 icebergs could have carved the flutes observed at ~980 mwd on the Lomonosov Ridge in
328 the central Arctic Ocean (Figure 1) (Kristoffersen *et al.*, 2004), which are morphologically
329 similar to the features we document on the ABM. However, in contrast to the only ~10 km
330 span of fluted terrain observed on the Lomonosov Ridge, the physically separated but
331 consistent, margin-wide distribution of ABM flutes implies the coherent movement of an
332 eroding ice mass over ~300 km (Figure 2). It is unlikely that even icebergs of the
333 dimensions documented around Antarctica would have the structural integrity or the

334 momentum to ground along this entire stretch of the ABM without changing their drift
335 direction. We infer therefore that the ABM flutes were formed by a large floating ice mass
336 (ice shelf) moving parallel to the continental margin from a grounding line located further
337 east or west. This ice shelf had to have been buttressed against the ABM during a re-
338 grounding event at depths reaching ~700 m (maximal depth of fluted seafloor), which is
339 >500 m deeper than sea level during any of the Quaternary glaciations.

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

352 Rose diagrams comparing the trend of bedform directions inferred from sidescan
353 maps with the trend of bathymetric landmarks on this bench are closely aligned (Figure 8),
354 suggesting that the eroding ice mass actually carved the bench out of the continental slope.
355 We note that the bench forms a 'second shelf break' along its lower edge, 300+ m deeper
356 than the actual modern shelf break of the ABM (e.g., Figures 7 and 9). Similar

357 overdeepened continental shelf breaks occur only on glaciated margins of the Arctic Ocean
358 (Barents Sea and the Greenland/Canadian margin) (Jakobsson, 2002) and in Antarctica
359 (Anderson, 1999).

360 Spectral analysis of our bathymetric data contributes a quantitative and margin-wide
361 characterization of the bedforms (Figure 10). Where we document the best examples of
362 sub-glacial flutes on the eastern half of the survey in areas E, G and H, amplitudes of
363 seafloor roughness are moderate and cover a narrow range of dominant wavelengths from
364 ~45-100 m. In the westernmost, shallower half of the survey, where iceberg scours and
365 grooves erode the substrate, samples from areas A, B, and C show the strongest signals of
366 seafloor excavation and a wider range of wavelengths from ~20-300 m. These results
367 corroborate the change in bedform character along the ABM seen in sidescan and
368 bathymetry images (Figures 4-5).

369 It is possible that areas A, B, and C were impacted by the same large ice mass that
370 eroded and/or molded the seafloor in areas E, G, and H; however, the chaotically oriented
371 iceberg scours and grooves that primarily characterize the western end of the margin likely
372 overprinted the fluted topography. The two types of morphologically distinct scours and
373 grooves observed suggest the possibility of two generations of iceberg activity, one during
374 the collapse of an ice shelf that eroded the ABM, and the other during a later episode of
375 massive iceberg discharge from the Laurentide Ice Sheet. Both types have a similar
376 azimuth range (95-100˚ from north) (Figures 2 and 8) indicating that icebergs drifted
377 clockwise along the ABM and impacted the slope at an oblique angle from the northeast.
378 This drift pattern is consistent with the distribution of ice-rafted debris in Pleistocene

379 sediments of the Canada Basin assumed to have been dispersed by icebergs carried by the
380 clockwise Beaufort Gyre (Phillips and Grantz, 2001) (Figure 1).
381
382 **4.2 Formation of ABM Slope Gullies**
383
384 The co-existence of the slope gullies with the glacially eroded portion of the ABM
385 (Figures 4-7), and their morphologic and contextual similarity to gully systems documented
386 on glaciated continental margins around Antarctica (Ó Cofaigh *et al.*, 2003), East
387 Greenland (Ó Cofaigh *et al.*, 2002), and the Gulf of Alaska (Carlson *et al.*, 1990), suggest
388 that the gullies are related to glaciogenic processes. In each of the above localities, ice is
389 inferred to have extended to the continental shelf edge during glacial periods, releasing sub-
390 glacially entrained sediments and subglacial meltwater where the ice front separated from
391 the seafloor. Sediments were subsequently remobilized by down slope gravity flows and
392 turbidite gullies were eroded into the upper continental slope. Ó Cofaigh *et al.* (2003) argue
393 that where slope gradients are sufficiently steep, any sub-glacial sediments available at the
394 shelf edge may be transported directly to the abyssal plain, bypassing the slope altogether.
395 This results in a sediment-starved upper slope environment conducive to the formation of
396 erosional gullies, regardless of the total amount of sediment available from glacial activity.
397 Thus even though the ABM is known to be mantled with up to 2700 m of unconsolidated
398 sediments (Grantz *et al.*, 1979; Dinter *et al.*, 1990; Kayen and Lee, 1991; Brigham-Grette *et*
399 *al.*, 2001), the continental slope in this area may simply be too steep (up to 13˚ between
400 600-700 mwd) to retain any glacially deposited sediments on the shelf edge.

401 The geomorphology of the ABM shows evidence for the flow of grounded ice
402 parallel to slope isobaths (Figure 2), instead of transverse to the shelf edge, as in the ice
403 stream localities described above. As a result, it appears that ice probably did not deliver
404 sediments to the ABM slope directly below the area of erosion. Some sediments may have
405 arrived at the shelf edge at an oblique angle where pre-existing topographic variability
406 along the shelf/slope break caused the sliding ice to float over open water, but the majority
407 of the sediments were probably moved parallel to the slope in the direction of dominant ice
408 flow. Thus most sub-glacially entrained sediments may not have been released along the
409 portion of the ABM margin mapped during SCICEX-99, effectively starving the upper
410 continental slope of sediments.
411 Large-scale mass-wasting features on the ABM, involving >200 m of sedimentary
412 cover, have been described in early seismic-reflection studies (e.g., Grantz *et al.*, 1979,
413 1981), but their exact relationship to slope morphology and the mechanisms of their
414 formation have yet to be identified. Seismic processes on the ABM and the decomposition
415 of gas hydrates during low sea-level stands have been discussed as likely controls on slope
416 instabilities (Grantz *et al.*, 1981; Kayen and Lee, 1991). From the existing data we cannot
417 conclude whether the mapped ABM gullies and other large-scale slumps are related; high-
418 resolution subbottom data from the gullied slope are needed to clarify this relationship.
419
420 **4.3 Ice flow provenance and trajectory**
421
422 The observed flute orientation along the ABM could indicate ice flow in either an
423 easterly or westerly direction. In the absence of unidirectional bedforms such as drumlins,

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

The latitudinally-oriented, margin-parallel glaciogenic bedforms on the ABM appear to provide a 'missing link' between the inferred northwest Laurentide sources of ice and the northwestward erosion of the Chukchi Borderland. This interpretation implies that the eroding ice mass flowed along the periphery of the Canada Basin rather than into its center. In order to achieve this pathway, ice flow had to be blocked or diverted by some type of barrier. Polyak *et al.*, (2001) suggest that ice was propelled along the coastline and across the CB by the force of CAA/Mackenzie ice streams (Figure 1). However, based on

447 ice mass balance considerations (Mercer, 1970), this ice would need to be buttressed on
448 both sides of the stream to maintain its full thickness without calving and breaking apart on
449 its northern, oceanward margin. Continuous sea ice cover during glacial periods might
450 briefly retard the calving rate along an ice sheet margin (Reeh *et al.*, 2001) but it is unlikely
451 that sea ice with a maximal thickness of ~7 m when concentrated by wind (Bourke and
452 Garrett, 1987), could buttress an advancing ice shelf whose thickness is two orders of
453 magnitude greater.

454 An alternative mechanism to explain the circuitous flow path of ice along the basin
455 margin invokes the existence of a thick ice shelf capping the entire Arctic Ocean (Mercer,
456 1970; Hughes *et al.,* 1977; Grosswald and Hughes, 1999). In these models, the basin-wide
457 ice shelf, fed by multiple sources around its periphery, could have prevented land-sourced
458 ice streams from penetrating directly into the central basins (Grosswald and Hughes, 1999).
459 Evidence for glacial erosion on top of the Lomonosov Ridge (Figure 1) (Jakobsson, 1999;
460 Polyak *et al.*, 2001) makes this scenario possible, provided there were synchronously
461 growing ice shelves on both the Eurasian and Laurentide sides of the ocean. Based on
462 preliminary, simplified numerical modeling (Jakobsson *et al.*, in press), the growth of an ice
463 shelf several hundred meters thick from the Eurasian margin all the way to the Lomonosov
464 Ridge is possible provided there is a condition that constrains the stretching of ice. Such a
465 constraint might be provided by abnormally thick sea ice or piled-up icebergs during glacial
466 stages, but this hypothesis needs to be tested by more refined modeling based on
467 comprehensive, age-specific paleogeographic reconstructions. For example, the limited
468 extent of the Eurasian ice sheet in the Arctic during the Last Glacial Maximum (Svendsen
469 *et al.*, 2004) refutes the concept of the pan-Arctic ice shelf during that time period, but

earlier glacial periods could have been more favorable for its formation. Notably, the age of glacial erosion on the Lomonosov Ridge is estimated at Oxygen Isotope Stage (OIS) 6, >130 ka (Jakobsson *et al.*, 2001).

An alternative interpretation of the Lomonosov Ridge erosion suggests that it was produced by an armada of huge tabular icebergs, rather than an ice shelf (Kristoffersen *et al.*, 2004). In this scenario, Antarctic-type icebergs purged from the glaciated Barents-Kara continental margin 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). If this were the case, we 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. 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.

493 Grosswald and Hughes (1999) hypothesize that the Flaxman Formation mantling the ABM
494 coastline to an altitude of 7 m and containing a high content of CAA clastic material is a
495 glacial till that was formed by the lateral margin of a transiting ice shelf. However,
496 lithological and paleontological data suggest a glaciomarine rather than sub-glacial origin
497 for this deposit (Dinter *et al.*, 1990; Brigham-Grette and Hopkins, 1995). We hypothesize
498 that the grounding of several hundred meter thick ice on the continental shelf and slope
499 isostatically depressed the ABM, allowing for a glaciomarine transgression of the coastline.
500 This interpretation is consistent with the lack of any mapped isostatic tilt in the Flaxman
501 Formation deposits along the margin, which indicates a local source of isostatic loading at
502 the ABM, rather than a distal Laurentide source as proposed previously (Dinter *et al.*, 1990;
503 Brigham-Grette and Hopkins, 1995). If this scenario is correct, ice shelf encroachment on
504 the ABM would have immediately preceded the deposition of the Flaxman Formation dated
505 to OIS 5a, 70-80 ka (Dinter *et al.*, 1990; Brigham-Grette and Hopkins, 1995). Accordingly,
506 the ice-grounding event would likely have occurred during an interstadial within OIS 5,
507 which matches the age range estimated for the erosion of the Chukchi Borderland (Polyak
508 *et al.*, in press) and for an inferred widespread glaciation in the Beringia region (Brigham-
509 Grette *et al.*, 2001).

510 The inferred age of ice shelf passage along the ABM is younger than the OIS 6 age
511 estimate for the erosion found on the Lomonosov Ridge (Jakobsson *et al.*, 2001). However,
512 it is conceivable that glacial periods postdating OIS 6 could have generated ice shelves or
513 iceberg armadas nearly comparable in size to prior surge events, but not quite thick enough
514 to have scoured the deep crest of the Lomonosov Ridge at ~1000 mwd. Furthermore, the
515 mega-berg purge trajectory could vary slightly from one glacial period to another, allowing

516 some very large icebergs to miss the shallowest portions of the Lomonosov Ridge but still
517 collect in the Amerasian Basin.
518
519 **5. CONCLUSIONS**
520
521 Sidescan and bathymetry data obtained during the SCICEX program elucidate the
522 detailed morphology of the Alaska/Beaufort margin in the depth range of 200-2800 m, and
523 help to constrain the sedimentary and glacial history of this part of the Arctic Ocean. The
524 ABM data show a repetitive association of glaciogenic seafloor bedforms and larger-scale
525 erosional bathymetric features in water depths reaching 700 m, with multiple turbidite
526 gullies further down slope. Similar geomorphic associations are known from glaciated
527 continental margins in the Arctic and Antarctic (e.g., Anderson, 1999; Shipp *et al.*, 1999;
528 Polyak *et al.*, 2001; Ó Cofaigh *et al.*, 2003; Dowdeswell *et al.*, 2004). At 400-700 mwd,
529 broad fields of sub-glacially generated flutes cover a minimum seafloor area of 330 km^2,
530 indicating widespread margin-parallel flow of a large ice mass buttressed against the ABM.
531 Flutes occur on the surface of a bathymetric bench at 400-550 m that may have been formed
532 by ice erosion, an interpretation that is corroborated by the coincident trends of flutes and
533 broader-scale bathymetric landmarks on the bench surface (Figure 8). We note also that
534 this large-scale erosion of the continental slope makes the ABM similar to overdeepened
535 continental margins that are characteristic of formerly glaciated areas. Iceberg scours and
536 grooves observed on the ABM in shallower water depths were probably generated during
537 ice shelf break-up as well as later glacial event(s). The bench is truncated to the north by
538 steep dendritic gullies that initiate in branching catchment areas up to 18 km across and

539 coalesce to form broad trunk channels in water depths >1500 m. Gullies are interpreted to
540 have formed in response to sediment-laden sub-glacial meltwater cascading down the steep
541 continental slope.

542 ABM glaciogenic bedforms have depths, styles, and orientations similar to features
543 discovered earlier on the Chukchi Borderland (Figure 1) (Polyak *et al.*, 2001; in press;
544 Jakobsson *et al.*, 2005) and were probably formed during the same glacial episode. ABM
545 bedforms link CB glaciogenic features with the over-deepened troughs of the CAA that
546 were the likely outlets for ice streams from the northwest sector of the Laurentide ice sheet,
547 and which could have propelled ice shelves into the Canada Basin. The margin-parallel
548 pathway of this ice flow along the ABM indicates that incoming ice was diverted from
549 streaming directly into the central part of the basin by some obstruction. We infer that such
550 an obstruction was formed by either a basin-wide ice shelf fed by multiple sources, or by a
551 pile-up of Antarctic-style icebergs driven by winds and currents from the Eurasian side of
552 the Arctic Ocean. Corroboration of the movement of very thick ice (~1000 m), be it an ice
553 shelf or iceberg armadas, towards the Canada Basin, is seen in the eroded crest of the
554 Lomonosov Ridge at the center of the Arctic Ocean (Jakobsson, 1999; Polyak *et al.*, 2001;
555 Kristoffersen *et al.*, 2004; Jakobsson *et al.*, in press).

556 We propose that the sliding of a Canada Basin ice shelf over the ABM is causally linked
557 to the diamict Flaxman Member of the Gubik Formation that mantles the ABM coast up to
558 7 m above sea level and contains numerous coarse clasts of CAA provenance (Dinter *et al.*,
559 1990; Brigham-Grette and Hopkins, 1995). This linkage can explain the unusual geology
560 of the Flaxman Member and its lack of glacioisostatic tilt towards the CAA. We propose
561 that the grounding of several hundred meter thick ice on the ABM shoulder isostatically

562 depressed the entire margin and allowed for a glaciomarine transgression of the coast. In
563 this case, the age of the ice grounding on the ABM immediately precedes the age of the
564 Flaxman Member, that is OIS 5b. This timing correlates with an inferred widespread
565 glaciation in the Beringian region (Brigham-Grette *et al.*, 2001) and fits the age range
566 estimated for glacial erosion on the Chukchi Borderland (Polyak *et al.*, 2003; in press).

567 **ACKNOWLEDGMENTS**

568

569 National Science Foundation awards to Edwards and Polyak were used to support this

570 project (OPP-9813535, 0136238, 9813532, and 0401692).

571 **REFERENCES**

572

573 Anderson JB. 1999. *Antarctic Marine Geology*. Cambridge University Press: Cambridge.

574 Barnes PW, Lien R. 1988. Icebergs rework shelf sediments to 500 m off Antarctica.
575 *Geology* **16**: 1130-1133.

576 Bigg GR, Wadley MR. 2001. The origin and flux of icebergs released into the Last Glacial
577 Maximum Northern Hemisphere oceans: the impact of ice-sheet topography.
578 *Journal of Quaternary Science* **16**: 565-573.

579 Blasco S, Fortin G, Hill P, O'Connor M, Brigham-Grette J. 1990. The late Neogene and
580 Quaternary stratigraphy of the Canadian Beaufort continental shelf. In *The Geology*
581 *of North America, Volume L, the Arctic Ocean Region*, Grantz A, Johnson L,
582 Sweeney J (eds). Geological Society of America: Denver; 491-501.

583 Bourke R, Garrett R. 1987. Sea ice thickness and distribution in the Arctic Ocean. *Cold*
584 *Regions Science and Technology* **13**: 259-280.

585 Brigham-Grette J, Hopkins D. 1995. Emergent marine record and paleoclimate of the last
586 interglaciation along the northwest Alaskan coast. *Quaternary Research* **4**: 159-173.

587 Brigham-Grette J, Hopkins D, Ivanov V, Basilyan A, Benson S, Heiser P, Pushkar S. 2001.
588 Last Interglacial (isotope stage 5) glacial and sea-level history of coastal Chukotka
589 Peninsula and St. Lawrence Island, Western Beringia. *Quaternary Sci. Rev* **20**: 419-
590 436.

591 Canals M, Casamor J, Urgeles R, Calafat A, Domack E, Baraza J, Farran M, De Batist M.
592 2002. Seafloor evidence of a subglacial sedimentary system off the northern
593 Antarctic Peninsula. *Geology* **30**: 603-606.

Carlson P, Bruns T, Fisher M. 1990. Development of slope valleys in the glacimarine environment of a complex subduction zone, Northern Gulf of Alaska. In *Glacimarine Environments: Processes and Sediments, Geological Society Special Publication No 53*, Dowdeswell J, Scourse J (eds). Geological Society of America: Denver; 139-153.

Darby D, Jakobsson M, Polyak L. 2005. Icebreaker expedition collects key Arctic seafloor and ice data. *Eos* **86** *(52)*: 549, 552.

Davies T, Bell T, Cooper A, Josenhans H, Polyak L, Solheim A, Stoker M, Stravers J (eds). 1997. *Glaciated Continental Margins: An Atlas of Acoustic Images*. Chapman and Hall: London.

Davis R, Edwards M, Rognstad M, Appelgate T, Chayes D. 2001. New tools for processing sidescan and interferometric bathymetry data. *Sea Technology* **42**: 21-27.

Dinter D. 1985. Late Quaternary depositional history of Alaskan Beaufort Shelf. *American Association of Petroleum Geologists Bulletin* **69**: 662-662.

Dinter D, Carter L, Brigham-Grette J. 1990. Late Cenozoic geologic evolution of the Alaskan North Slope and adjacent continental shelves. In *The Geology of North America, Volume L, the Arctic Ocean Region*, Grantz A, Johnson L, Sweeney J (eds). Geological Society of America: Denver; 459-489.

Dowdeswell J, Ó Cofaigh C, Pudsey C. 2004. Continental slope morphology and sedimentary processes at the mouth of an Antarctic paleo-ice stream. *Marine Geology* **204**: 203-214.

Dyke A, Morris T, Green D, England J. 1992. Quaternary geology of Prince of Wales Island, Arctic Canada. *Geological Survey of Canada Memoir* **433**: 77-89.

617 Dyke AS, Andrews JT, Clark PU, England JH, Miller GH, Shaw J, Veillette JJ. 2002. The
618 Laurentide and Innuitian ice sheets during the Last Glacial maximum. *Quatern. Sci.*
619 *Rev*. **21**: 9-31.
620 Edwards M, Coakley B. 2003. SCICEX investigations of the Arctic Ocean System. *Chemie*
621 *der Erde* **63**: 281-328.
622 Edwards M, Johnson P, Wright D, Dailey M. 2005. A Distributed Data Management
623 System to Support Arctic Geology and Geophysics: A Resource of Extended EEZ
624 Claims. *EOS Trans. AGU,* **86** *(52), Fall Meeting Supplement,* Abstract T13D-0497.
625 Goff J, Nordfjord S. 2004. Interpolation of Fluvial Morphology Using Channel-Oriented
626 Coordinate Transformation: A Case Study from the New Jersey Shelf. *Mathematical*
627 *Geology* **6**: 643-658.
628 Goosse H, Campin J, Fichefet T, Deleersnijder E. 1997a. Sensitivity of a global ice-ocean
629 model to the Bering Strait throughflow. *Climate Dynamics* **13** *(5)*: 349-358.
630 Goosse H, Fichefet T, Campin J. 1997b. The effects of the water flow through the Canadian
631 Archipelago in a global ice-ocean model. *Geophysical Research Letters* **24** *(12*):
632 1507-1510.
633 Grantz A, Eittreim S, Dinter D. 1979. Geology and tectonic development of the continental
634 margin north of Alaska. *Tectonophysics* **59**: 263-291.
635 Grantz A, Eittreim S, Whitney O. 1981. Geology and physiography of the continental
636 margin north of Alaska and implications for the origin of the Canada Basin. In *The*
637 *Ocean Basins and Margins, v. 5*, Nairn A, Churkin M, Stehli F (eds). Plenum: New
638 York; 439-492.

639 Grosswald M, Hughes T. 1999. The case for an ice shelf in the Pleistocene Arctic Ocean.
640 *Polar Geography* **23**: 23-54.
641 Hughes T, Denton G, Grosswald M. 1977. Was there a late Wurm ice sheet? *Nature* **266**:
642 596-602.
643 Jakobsson M. 1999. First high-resolution chirp sonar profiles from the central Arctic Ocean
644 reveal erosion of Lomonosov Ridge sediments. *Marine Geology* **158**: 111-123.
645 Jakobsson M, Cherkis N, Woodward J, Coakley B, Macnab R. 2000. A new grid of Arctic
646 bathymetry: a significant resource for scientists and mapmakers. *EOS* **81** *(9)*: 89, 93,
647 96.
648 Jakobsson M, Lovlie R, Arnold E, Backman J, Polyak L, Knutsen J-O, Musatov E. 2001.
649 Pleistocene stratigraphy and paleoenvironmental variation from Lomonosov Ridge
650 sediments, central Arctic Ocean. *Global and Planetary Change* **31**: 1-22.
651 Jakobsson M. 2002. Hypsometry and volume of the Arctic Ocean and its constituent seas.
652 *Geochemistry, Geophysics, Geosystems* **3** *(2)*: 1-18.
653 Jakobsson M, Gardner J, Vogt P, Mayer L, Armstrong A, Backman J, Brennan R, Calder B,
654 Hall J, Kraft B. 2005. Multibeam bathymetric and sediment profiler evidence for ice
655 grounding on the Chukchi Borderland, Arctic Ocean. *Quaternary Research* **63**: 150-
656 160.
657 Jakobsson M, Siegert MJ, Paton M. In press. Is grounding of an ice shelf possible in the
658 central Arctic Ocean? A modeling experiment. *ICAM IV (International Conference*
659 *on Arctic Margins) Proceedings*.
660 Josenhans H, Zevenhuizen J. 1990. Dynamics of the Laurentide Ice Sheet in Hudson Bay,
661 Canada. *Marine Geology* **92**: 1-26.

662 Kayen R, Lee H. 1991. Pleistocene slope instability of gas hydrate-laden sediment on the
663 Beaufort Sea margin. *Marine Geotechnology* **10**: 125-141.
664 Kristoffersen Y, Coakley B, Jokat W, Edwards M, Brekke H, Gjengedal J. 2004. Seabed
665 erosion on the Lomonosov Ridge, central Arctic Ocean: a tale of deep draft icebergs
666 in the Eurasia Basin and the influence of Atlantic water inflow on iceberg motion?
667 *Paleoceanography* **19**: PA3006.
668 Kuijpers A, Hansen B, Huhnerbach V, Larsen B, Nielsen T, Werner F. 2002. Norwegian
669 Sea overflow through the Faroe-Shetland gateway as documented by its bedforms.
670 *Marine Geology* **188**: 147-164.
671 Leffingwell E de K. 1919. The Canning River region, northern Alaska. *U.S. Geological*
672 *Survey Professional Paper 109*. U.S. Geological Survey: Washington.
673 Lowe A, Anderson J. 2002. Evidence for abundant subglacial meltwater beneath the paleo-
674 ice sheet in Pine Island Bay, Antarctica. *Journal of Glaciology* **49**: 125-138.
675 Mercer J. 1970. A former ice sheet in the Arctic Ocean. *Palaeo* **8**: 19-27.
676 Naidu A, Mowatt T. 1992. Origin of gravels from the southern coast and continental shelf
677 of the Beaufort Sea, Arctic Alaska. *1992 ICAM Proceedings Programs with*
678 *Abstracts*: 351-356.
679 Ó Cofaigh C, Taylor J, Dowdeswell J, Rosell-Mele A, Kenyon N, Evans J, Mienert J. 2002.
680 Sediment reworking on high-latitude continental margins and its implications for
681 paleoceanographic studies: insights from the Norwegian-Greenland Sea. In *Glacier-*
682 *Influenced Sedimentation on High-Latitude Continental Margins, Geological*
683 *Society, London, Special Publications 203*, Dowdeswell J, Ó Cofaigh C (eds).
684 Geological Society of London: London; 325-348.

685 Ó Cofaigh C, Taylor J, Dowdeswell J, Pudsey C. 2003. Palaeo-ice streams, trough mouth
686 fans and high-latitude continental slope sedimentation, *Boreas* **32**: 37-55.
687 Phillips R, Grantz A. 2001. Regional variations in provenance and abundance of ice-rafted
688 clasts in Arctic Ocean sediments: implications for the configuration of late
689 Quaternary oceanic and atmospheric circulation in the Arctic. *Marine Geology* **172**:
690 91-115.
691 Pickart R. 2004. Shelfbreak circulation in the Alaskan Beaufort Sea: mean structure and
692 variability. *Journal of Geophysical Research* **109**: C04024.
693 Polyak L, Forman S, Herlihy F, Ivanov G, Krinitsky P. 1997. Late Weichselian deglacial
694 history of the Svyataya (Saint) Anna Trough, northern Kara Sea, Arctic Russia.
695 *Marine Geology* **143**: 169-188.
696 Polyak L, Edwards M, Coakley B, Jakobsson M. 2001. Ice shelves in the Pleistocene Arctic
697 Ocean inferred from glaciogenic deep-sea bedforms. *Nature* **410** *(6827)*: 453-457.
698 Polyak L, Brachfeld S, Darby D, Bischoff J, Phillips R, Febo L, Coakley B, Edwards M,
699 Engels J. 2003. Multiple glacial erosion on the Chukchi Borderland, Arctic Ocean,
700 as indicated by preliminary data on sediment stratigraphy and seafloor morphology.
701 *XVI INQUA Congress, Programs with Abstracts*: 195.
702 Polyak L, Darby D, Bischoff J, Jakobsson M. In press. Evidence for late Pleistocene
703 glaciogenic erosion of the Chukchi Margin, Arctic Ocean. *Quaternary Research*.
704 Reeh N, Thomsen HH, Higgins AK, Weidick A. 2001. Sea ice and the stability of north and
705 northeast Greenland floating glaciers. *Annals of Glaciology* **33**: 474-480.

706 Rodeick C. 1979. The origin, distribution, and depositional history of gravel deposits on the
707 Beaufort Sea continental shelf, Alaska. *U.S. Geological Survey, Open File Report*
708 *79-234*: 87.
709 Shipp S, Anderson J, Domack E. 1999. Late Pleistocene-Holocene retreat of the West
710 Antarctic Ice-Sheet system in the Ross Sea: Part 1 – Geophysical Results. *GSA*
711 *Bulletin* **111** *(10)*: 1486-1516.
712 Stokes C, Clark C, Darby D, Hodgson D. 2005 Late Pleistocene ice export events into the
713 Arctic Ocean from the M'Clure Strait Ice Stream, Canadian Arctic Archipelago.
714 *Global and Planetary Change* **49**: 139–162.
715 Stokes C, Clark C, Winsborrow M. In press. Subglacial bedform evidence for a major
716 palaeoice stream in Amundsen Gulf and its retreat phases, Canadian Arctic
717 Archipelago. *Journal of Quaternary Science*.
718 Svendsen J, Alexanderson H, Astakhov VI, Demidov I, Dowdeswell JA, Funder S,
719 Gataullin V, Henricksen M, Hjort C, Houmark-Nielsen M, Hubberton HW,
720 Ingolfsson O, Jakobsson M, Kjaer KH, Larsen E, Lokrantz H, Lunkka JP, Lysa A,
721 Mangerud J, Matiouchkov A, Murray A, Moller P, Niessen F, Nikolskaya O, Polyak
722 L, Saarnisto M, Siegert C, Siegert MJ, Spielhagen RF, Stein R. 2004. Late
723 Quaternary ice sheet history of Northern Eurasia. *Quaternary Science Review* **23***,*
724 *issues 11-13*: 1229-1271.
725 Swan A, Sandilands M. 1995. *Introduction to Geological Data Analysis*. Blackwell Science
726 Ltd.: Oxford.

727 Vogt P, Crane K, Sundvor E. 1994. Deep Pleistocene iceberg plowmarks on the Yermak
728 Plateau: sidescan and 3.5 kHz evidence for thick calving ice fronts and a possible
729 marine ice sheet in the Arctic Ocean. *Geology* **22**: 403-406.

730 **FIGURE CAPTIONS**

731

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

743

744 Figure 2 - ABM survey area with bathymetry, survey track (in black), and the schematic
745 orientations of mapped seafloor glaciogenic features within each study area. Heavy black
746 lines indicate the presence and orientation of flutes, red lines indicate the mean orientations
747 of iceberg scours or grooves. The bathymetry data are a combination of gridded data from
748 IBCAO (Jakobsson *et al.,* 2000) and swath data collected by the *USCGC HEALY, R/V*
749 *Nathaniel B. Palmer* and the *USS Hawkbill* (Edwards at Coakley*,* 2003*;* Edwards *et al.,*
750 2005). Bathymetry is displayed as color-coded contours overlaid on a sun-shaded digital
751 terrain model illuminated from the east. Note the deeply eroded gullies visible in the
752 bathymetry data north of the continental shelf.

753 Figure 3 - Sample power spectrum generated from the output of spectral analysis of seafloor
754 bathymetry. The x-axis represents long to short wavelengths; the y-axis is the amplitude of
755 each wavelength squared, i.e., power (in m^2). Inset shows DC shifted raw seafloor
756 bathymetry that was detrended and filtered, then input to the FFT program to produce the
757 power spectrum shown in this figure.
758
759 Figure 4 – Sidescan (upper) and bathymetry (lower) SCICEX swath data for area A (see
760 Figure 2 for location). The bathymetry data are displayed as color-coded contours overlaid
761 on sun-shaded topography illuminated from 005° at an altitude of 10° above the horizon to
762 emphasize small-scale relief. Scours and grooves (indicated) are most apparent in the
763 higher resolution sidescan data (8 m grid cells) with its higher angle of ensonification, but
764 all features show some relief in the lower resolution (25 m grid cells) bathymetric data.
765
766 Figure 5 –Sidescan (upper) and bathymetry (lower) SCICEX swath data for area H (see
767 Figure 2 for location). The bathymetry data are displayed as color-coded contours overlaid
768 on sun-shaded topography illuminated from 355° at an altitude of 10° above the horizon to
769 emphasize small-scale relief. The presence of flutes in the bathymetry (indicated) is more
770 strongly correlated with their apparent width in the sidescan data rather than the total
771 vertical relief. Deep, narrow flutes in the 8 m resolution gridded sidescan data can be
772 difficult to detect in the 25 m bathymetry gridded data, but wider flutes are apparent across
773 the entire swath.

774 Figure 6 – (a) sidescan image of an extensive dendritic gully system coalescing into trunk
775 channels (area E on Figure 2). (b) interpreted outline of the gully system. (c) Highest
776 resolution SCICEX (nadir) bathymetry of ABM topography collected along the returning
777 submarine track, which surveyed approximately parallel to the 366-m isobath. Water
778 depths between 200-600 m are highlighted. Arrows indicate typical examples of gullies
779 along the margin.
780
781 Figure 7 - 3-D view of the ABM that highlights the bathymetric bench most prominent
782 along the margin in Area H (see Figure 2 for location and description of the bathymetric
783 datasets). SCICEX survey track shown in black, the 350 m and 600 m contours that bound
784 the bench are shown in red and blue, respectively. The bathymetry data are displayed as
785 color-coded contours overlaid on a sun-shaded digital terrain model illuminated from the
786 north. The diagonal stripes in the bathymetry data are artifacts that result from resampling
787 the IBCAO data at higher resolution than their 2.5 x 2.5 km grid cell size. CB – Chukchi
788 Borderland. Inset shows SCICEX slope profile of Area H along the transect indicated with
789 heavy black line on the 3-D map.
790
791 Figure 8 - Average trends of bathymetric landmarks for the areas sampled using spectral
792 analysis techniques: a) Bathymetric landmark trend in fluted areas, b) bathymetric
793 landmark trend in scoured and grooved areas, c) bedform orientation from sidescan sonar
794 data in fluted areas, d) bedform orientation from sidescan sonar data in scoured and grooved
795 areas. In each rose diagram the mean trend is indicated by the dashed line.

796 Figure 9 - Fragment of a chirp-sonar subbottom record west of the study area (SCICEX-98;
797 see Figure 1 for location) illustrating iceberg scouring to water depths of between 350-400
798 m and truncation of the inclined bedrock strata on the slope, with an erosional bench
799 between 450-500 m.
800
801 Figure 10 – Spectral estimates seafloor roughness for the sites listed on the x-axis (see
802 Figure 2 for locations) with distance from the eastern end of the survey area (upper panel),
803 and with equal spacing between bathymetric lines analyzed to facilitate visualization (lower
804 panel). The strength of the signal of seafloor excavation within one site and across the
805 margin is shown in dB using rainbow colors. Color change from red to dark blue values
806 indicates the change in repeating wavelengths from the most deeply excavated to the least
807 excavated terrain. The spacing between glaciogenic features is denoted on the y-axis in
808 meters.