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Earth Surface Processes and Landforms
Earth Surf. Process. Landforms **32**, 000–000 (2007)
Published online in Wiley InterScience
(www.interscience.wiley.com) **DOI:** 10.1002/esp.1601

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

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Abstract

New imagery of ~14 100 km^2 of seafloor along a 640 km stretch of the Alaska and Beaufort margins (ABM) in water depths from 250 to 2800 m depicts a repetitive association of glaciogenic bedforms (lineations and iceberg scours), broad erosional bathymetric features and adjacent downslope 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 probably 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.
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Keywords:

Received 9 January 2007;
Revised 6 July 2007;
Accepted 12 July 2007

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; Darby *et al.*, 2005; Jakobsson *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 m. These recent data lend credence to the long-standing, controversial hypothesis that vast ice shelves up to >1 km thick once existed in the

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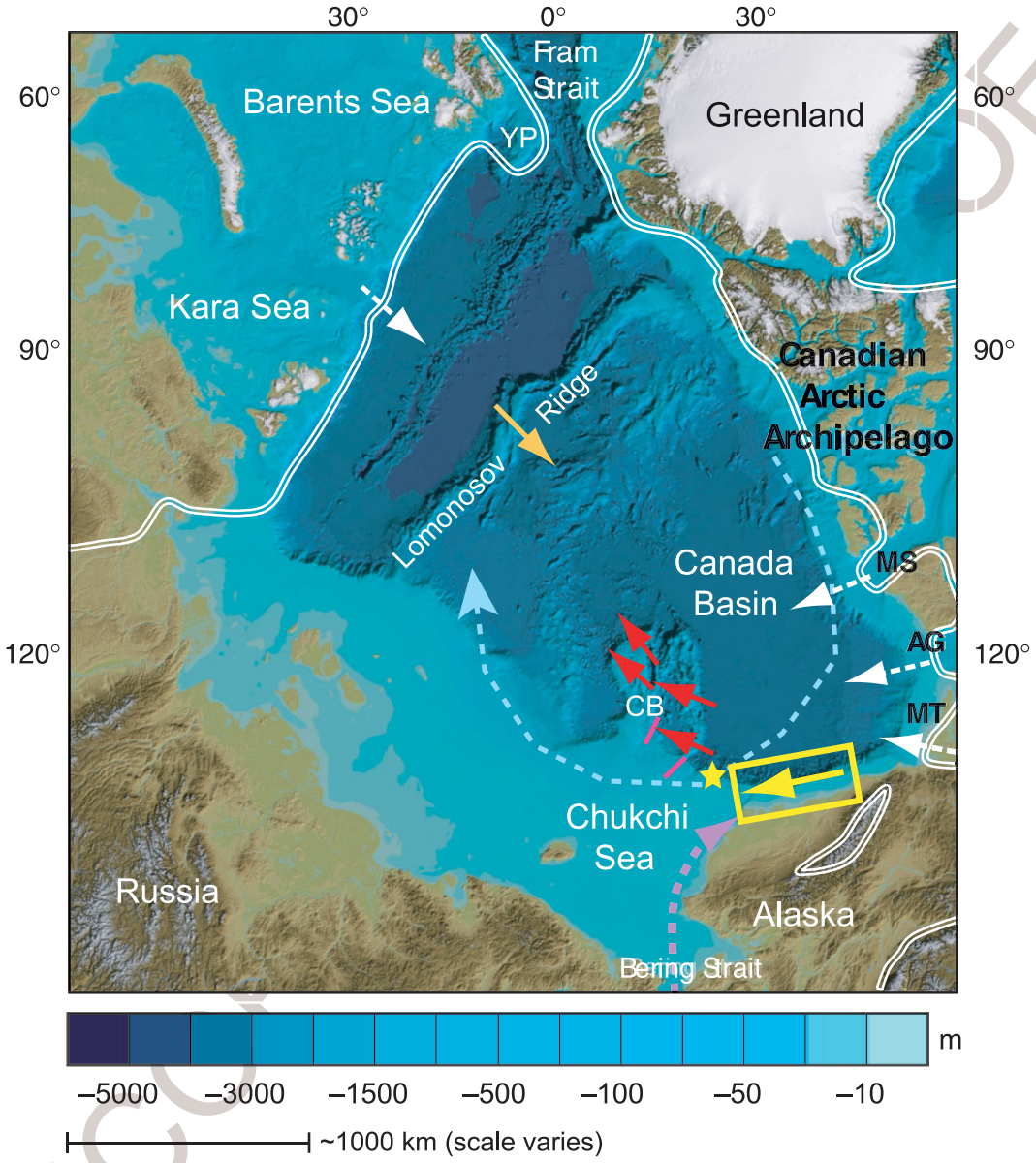


Figure 1. Overview map of the Arctic Ocean with bathymetry from the International Bathymetric Chart of the Arctic Ocean (Jakobsson *et al.*, 2000). Fields of lineations on the Chukchi Borderland and Lomonosov Ridge shown by red, pink, and orange arrows/lines (from Polyak *et al.*, 2001); arrows show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

Arctic Ocean, rivalling 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 palaeoglaciologists. 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

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1000 m and chaotic iceberg scours at shallower depths (Figure 1) (Polyak *et al.*, 2001, 2007; Darby *et al.*, 2005; Jakobsson *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 to 50 m. The margin is bounded to the north by a steep, heavily canyoned continental slope that drops to 1000 m 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 tens to hundreds of metres of the unconsolidated clastic materials of the Gubik Formation, interpreted to be a series of glaciomarine transgressive deposits dating from 3 Ma through to the Last Glacial Maximum (Dinter *et al.*, 1990, and references therein). Coarse clasts, from gravel to boulders, with Canadian Arctic Archipelago (CAA) provenance are

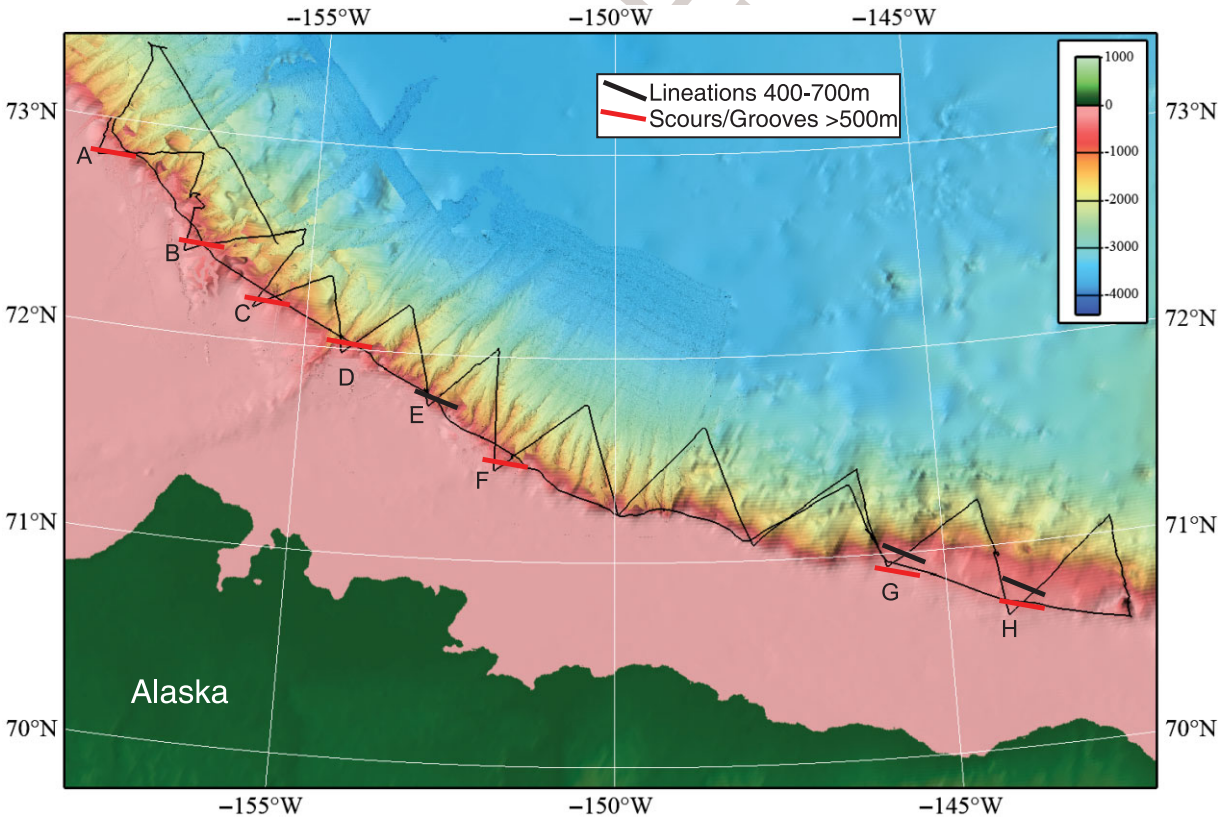


Figure 2. Alaska–Beaufort margin survey area with bathymetry, survey track (in black) and the schematic orientations of mapped seafloor glaciogenic features within each study area. Heavy black lines indicate the presence and orientation of lineations, red lines indicate the mean orientations of iceberg scours or grooves. The bathymetry data are a combination of gridded data from the International Bathymetric Chart of the Arctic Ocean (Jakobsson *et al.*, 2000) and swath data collected by the *USCGC HEALY, R/V Nathaniel B. Palmer* and the *USS Hawkbill* (Edwards and Coakley, 2003; Edwards *et al.*, 2005). Bathymetry is displayed as colour-coded contours overlaid on a sun-shaded digital terrain model illuminated from the east. Note the deeply eroded gullies visible in the bathymetry data north of the continental shelf.

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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). These authors also pointed out, however, 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.

Methods

Between 1995 and 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 portion of the ABM between 140 and 160°W (Figures 1 and 2). The submarine mapped seafloor from 200 to 2800 m deep along the ABM in a zigzagging pattern designed to accommodate a water-sampling programme (Figure 2). Sidescan swath widths vary from 6 km on the continental shelf to >20 km at the base of the continental slope. Bathymetry swath width is less than the sidescan swath width, averaging about four times the water depth. Where seafloor ensonification angles exceed 80° in the shallowest survey areas, low-relief morphological features are emphasized at the outer edges of the sidescan swaths. Unfortunately, sub-bottom 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 artefacts 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 s were subsampled to create 20 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 global positioning system (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. The 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 USA, 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).

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

In order to quantify whether there is a systematic progression of seafloor morphological changes along the ABM, we performed a one-dimensional Fourier analysis on transects of the full resolution swath bathymetry. To minimize outside factors that degrade data quality, we chose 20 profiles showing distinctive bedforms in the sidescan imagery that were characterized by submarine movement in a single direction at a constant speed. We corrected for the oblique submarine orientations (relative to dominant bedform trend) by projecting each vehicle track perpendicular to the bedform trend, and then resampled the data to generate consistent data densities for each sample area. The slope trend and mean value for each sample area were removed and the projected, resampled, and detrended data were filtered to remove long-wavelength noise. These data were analysed using a Fast Fourier Transform (FFT) program and plotted

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as power spectra to highlight recurring wavelengths. The FFT amplitudes serve as proxies for the variance in height of seafloor morphological features.

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) totalling ~550 km^2 show evidence of disturbed seafloor sediments in <250–700 m w.d. 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 depth (w.d.) <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 m w.d. (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 60°–135°N, but fields have a dominant overall fabric of ~100° and most trend obliquely downslope (Figure 2).

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

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

Spectral analysis is used on 20 bathymetric lines along the length of the survey area to compare the distances between and the amplitudes of the seafloor bedforms described above (Figure 8). The amplitude of spectral estimates gives an indication of the relative seafloor roughness at each wavelength. In Figure 8, coloured amplitudes indicate the scale of roughness of the seafloor at the wavelengths indicated in metres on the *y* axis. For example, in sample area C3, there is high amplitude roughness of the seafloor with repeated spacing of ~45 m perpendicular to the sample transect and parallel to the lineation trend, whereas in sample G2 the highest amplitude seafloor roughness within that area (which is lower amplitude than that observed in area C3) occurs with spacing of ~60 m. In Figure 8 high amplitude roughness centres around the 50 m wavelength, consistent with the distance between lineations measured independently from the sidescan imagery. On the western end of the survey in areas A, B and C where scours and grooves prevail, the seafloor is more deeply excavated than in areas E, G and H where fields of lineations occur. Depths at which the scours and grooves are observed in sample areas A, B and C are shallower than 370 m with a mean of 290 m; lineations in areas E, G and H are found in water depths averaging 400 m and as great as 570 m.

Bathymetric bench

The SCICEX bathymetry documents the co-occurrence of the seafloor bedforms described above with the location of a flattened bathymetric bench that is 400–550 m deep and has an average slope of just 1·6° (Figures 3, 5 and 9).

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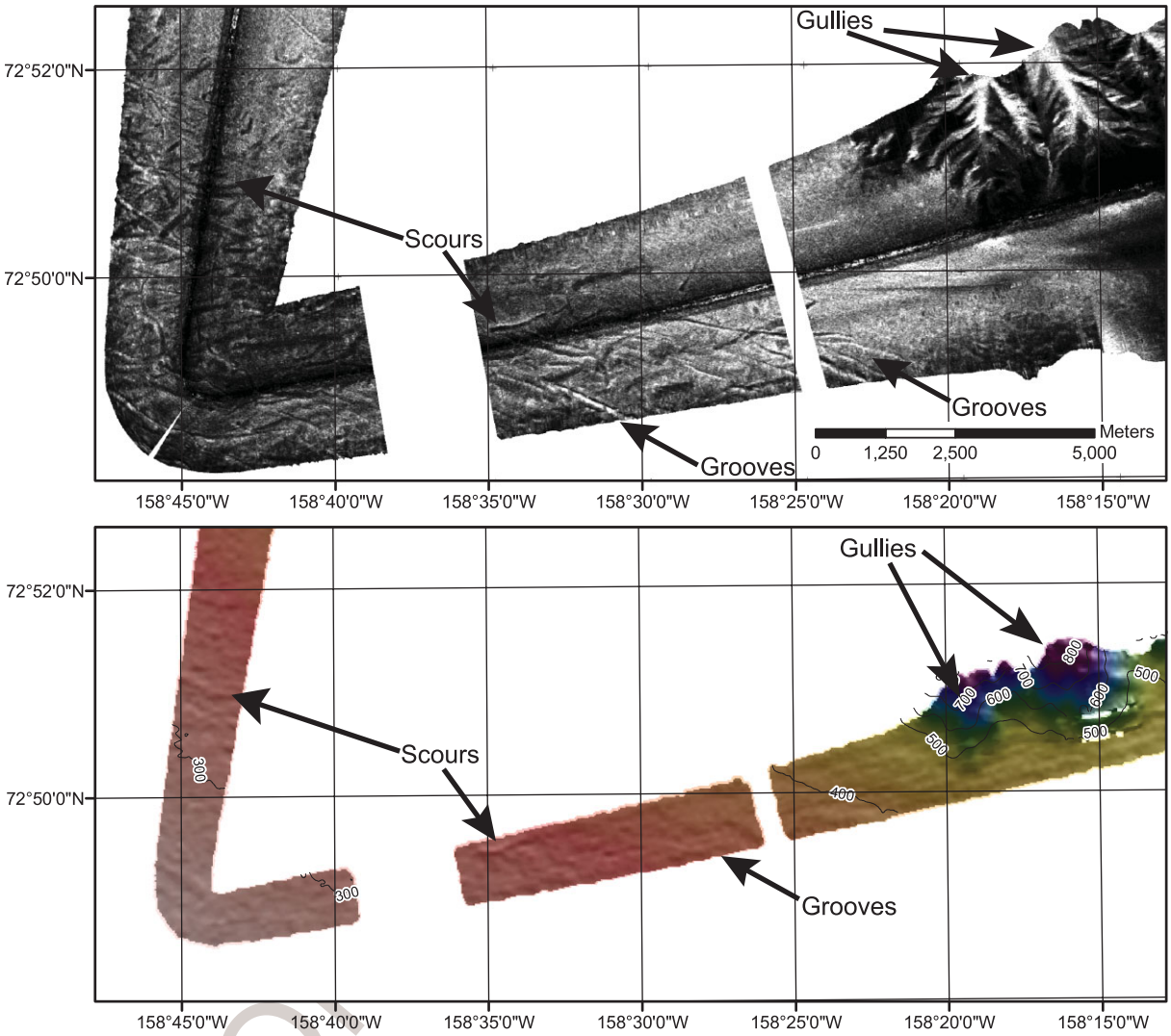


Figure 3. Sidescan (upper) and bathymetry (lower) SCICEX swath data for area A (see Figure 2 for location). The bathymetry data are displayed as colour-coded contours overlaid on sun-shaded topography illuminated from 005° at an altitude of 10° above the horizon to emphasize small-scale relief. Scours and grooves (indicated) are most apparent in the higher resolution sidescan data (8 m grid cells) with its higher angle of ensonification, but all features show some relief in the lower resolution (25 m grid cells) bathymetric data.

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The bench morphology is also visible in IBCAO bathymetry for the Beaufort Sea (Jakobsson, 2002) and in a new bathymetry compilation for the Arctic Ocean (Figure 9) (Edwards *et al.*, 2005). The basinward margin of the bench is associated with the onset of dendritic gullies and a precipitous increase in slope to 13°. Within the study area the bench width varies from 5 to >35 km, with the broadest portions mapped for ~100 km along the easternmost end of the surveyed area (Figure 9). On a sub-bottom record west of the study area the continuation of the bench is identified at ~480 m depth with a width of ~3–4 km (Figure 4). The truncation of the slope face strata suggests the erosional character of the bench. The widest portion of the bench underlies areas G and H (Figures 2 and 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 and 500 m w.d., 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).

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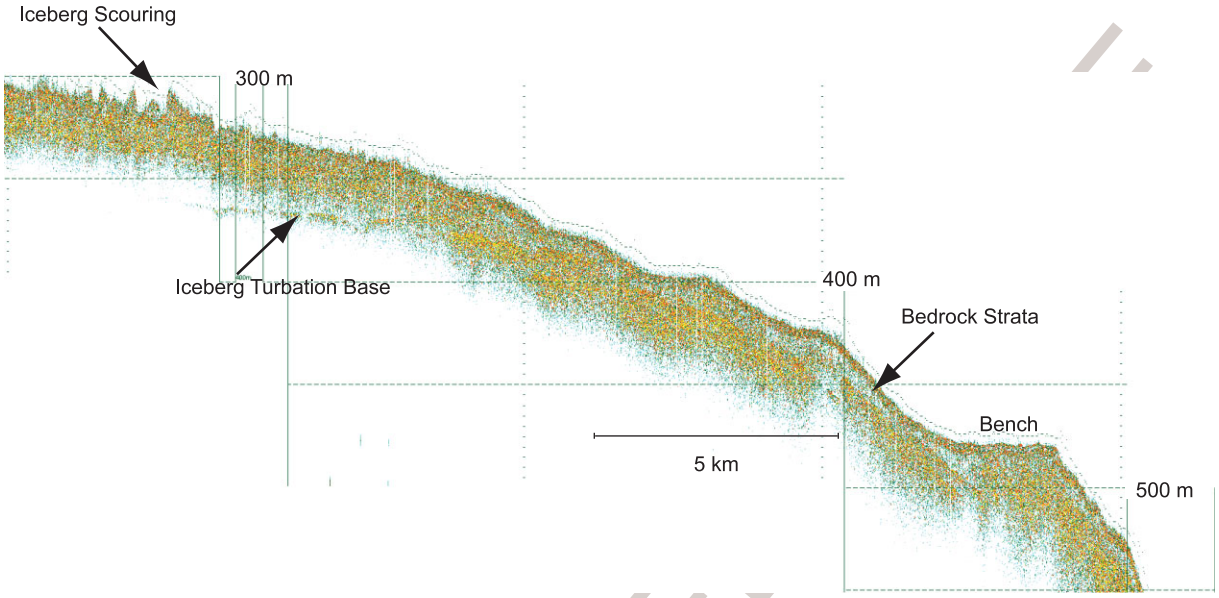


Figure 4. Fragment of a chirp-sonar sub-bottom record west of the study area (SCICEX-98; see Figure 1 for location) illustrating iceberg scouring to water depths of between 350 and 400 m, and truncation of the inclined bedrock strata on the slope, with an erosional bench between 450 and 500 m.

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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 10a 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 to 770 m (average 307 m), and gully widths range from 0·8 to 18·1 km (average 5·5 km) (Figure 10c). Nearly 80% 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 m w.d. and 13° from 600–700 m w.d. The length of the gully system mapped in Figure 10a and b is in excess of 40 km and covers depths ranging from 250 to >1500 m w.d. before continuing out of the sidescan swath image. The thalwegs of the gullies in the dendritic catchment system (ranging in width from 0·7 to 10 km) coalesce to form broad trunk channels (Goff and Nordfjord, 2004) in deeper water depths from ~1000 to 1500 m (Figure 10a and b).

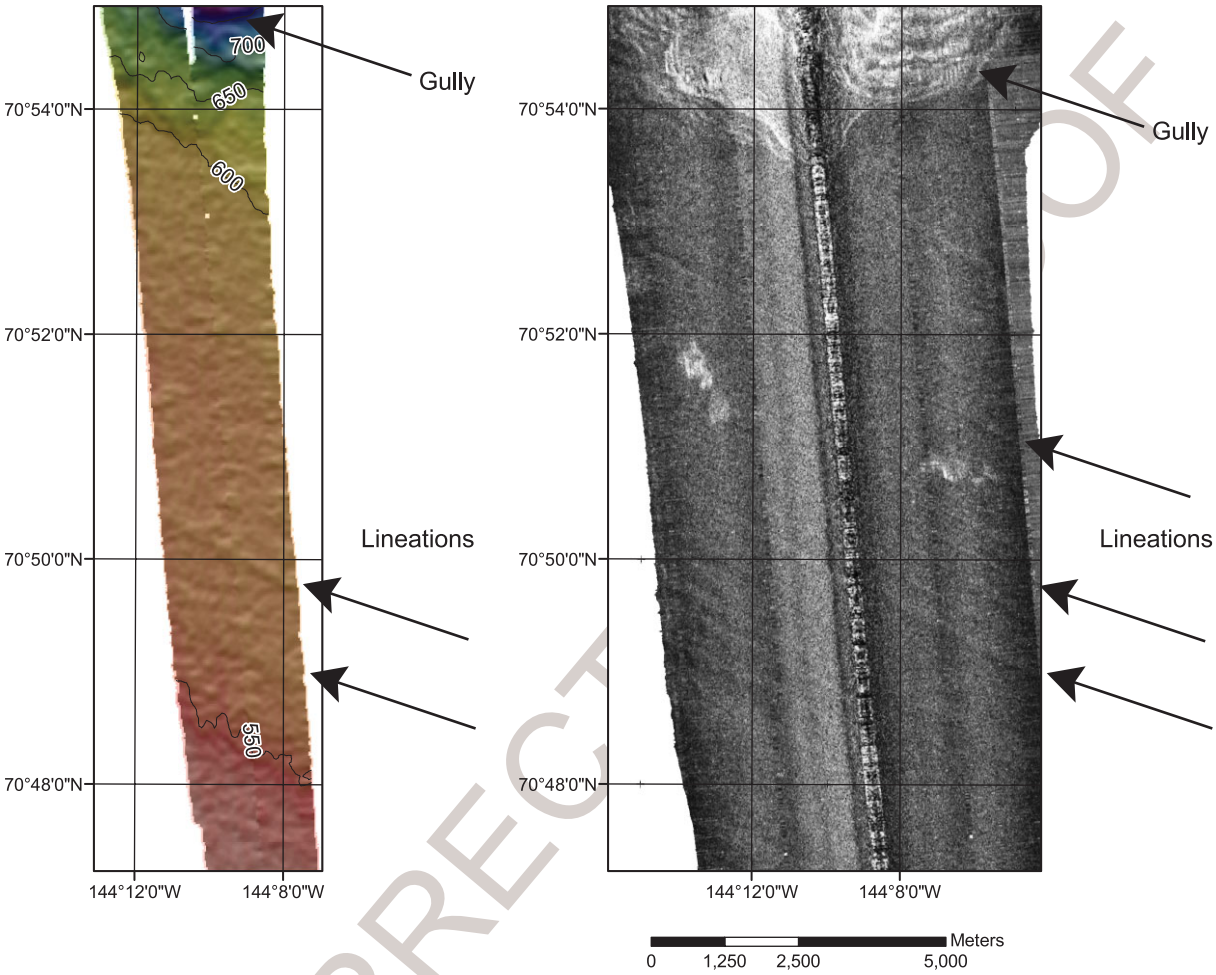
To better understand the physical controls on gully formation we analysed their morphologies as well as their relationship to large-scale topographic features using the ~400 km margin-parallel transect taken by the submarine along the 366 m contour (Figure 10c). Gully walls generally are the same height 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 with the gullied terrain (Figure 10c). At ~400–600 m w.d. 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 ~1000 m below sea level are indicated in Figure 10a and b.

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

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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 subparallel scours and grooves (Figure 3) are iceberg ploughmarks 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

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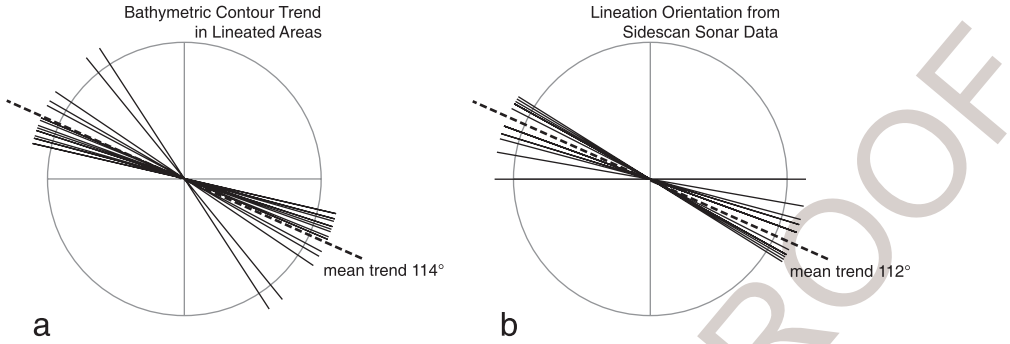


Figure 6. Average trends of bathymetric features for the lineated areas E, G and H: (a) large-scale bathymetric features such as the slope bathymetric contours, the bench alignment and the shelf–slope break; (b) lineations in sidescan sonar data. In each rose diagram the mean trend is indicated by the dashed line and is significant at the 95% level using Rayleigh's test for trend (Swan and Sandilands, 1995).

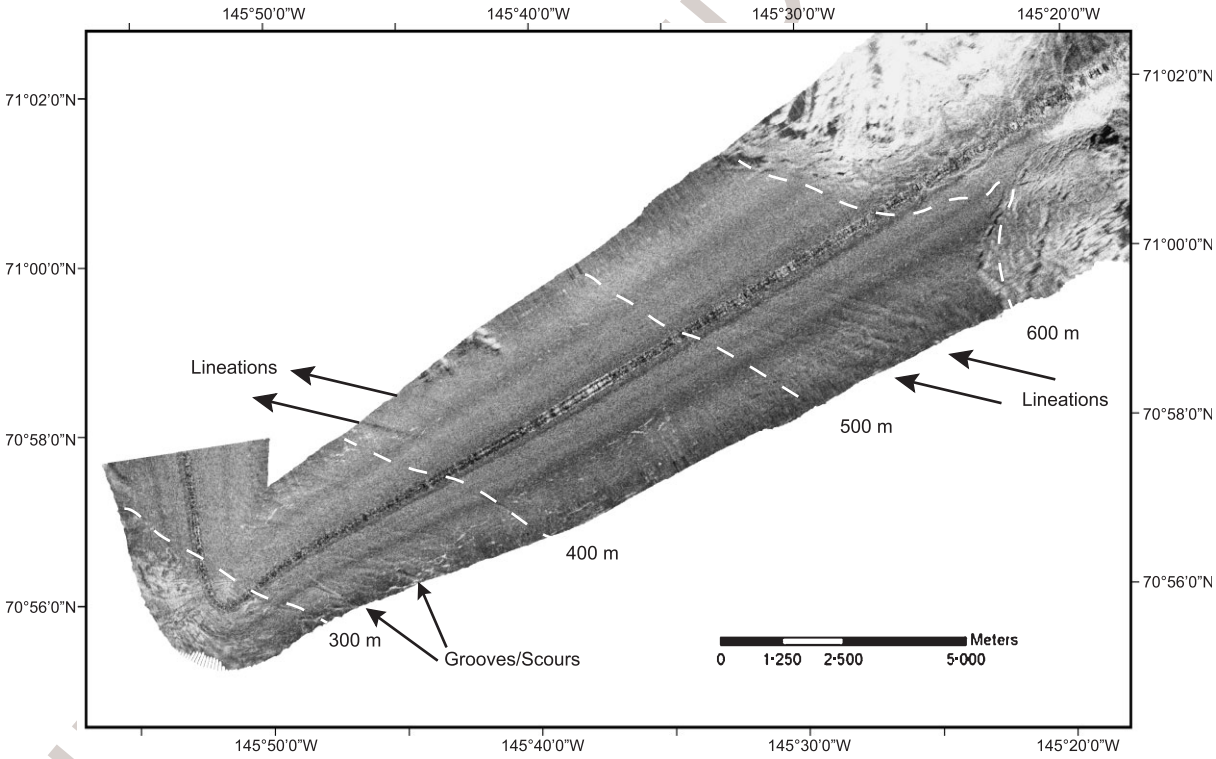


Figure 7. 8-m resolution gridded sidescan data for Area G (see Figure 2 for location). Note the depth-dependence of lineations (~350–600 m w.d. in this image) versus grooves (shallower than 350 m w.d.).

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

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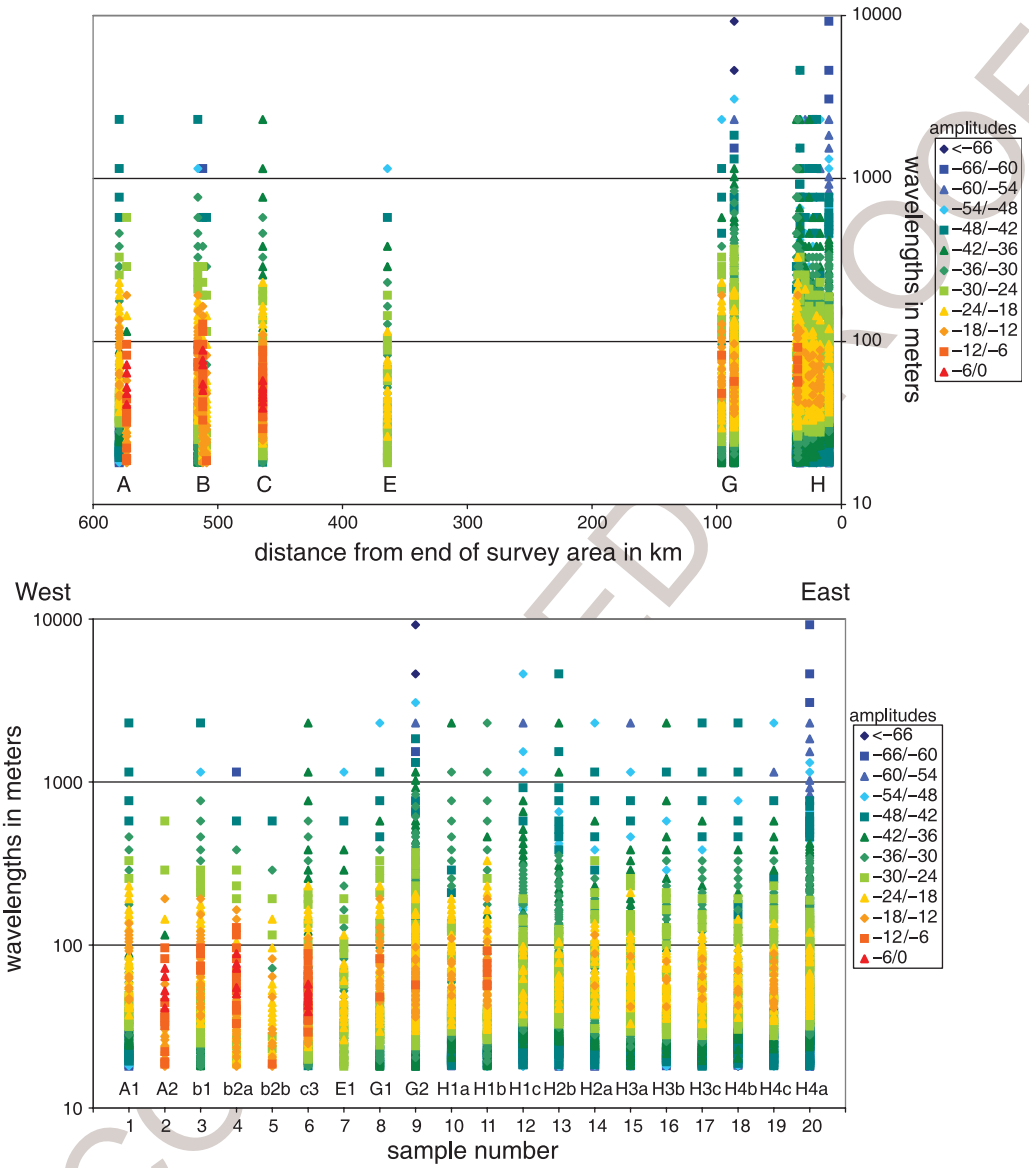


Figure 8. Spectral estimates of seafloor roughness for the sites listed on the *x* axis (see Figure 2 for locations) with distance from the eastern end of the survey area (upper panel), and with equal spacing between bathymetric lines analysed to facilitate visualization (lower panel). The strength of the signal of seafloor excavation within one site and across the margin is shown in dB using rainbow colours. Colour change from red to dark blue values indicates the change in repeating wavelengths from the most deeply excavated to the least excavated terrain. The spacing between glaciogenic features is denoted on the *y* axis in metres.

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 centred 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

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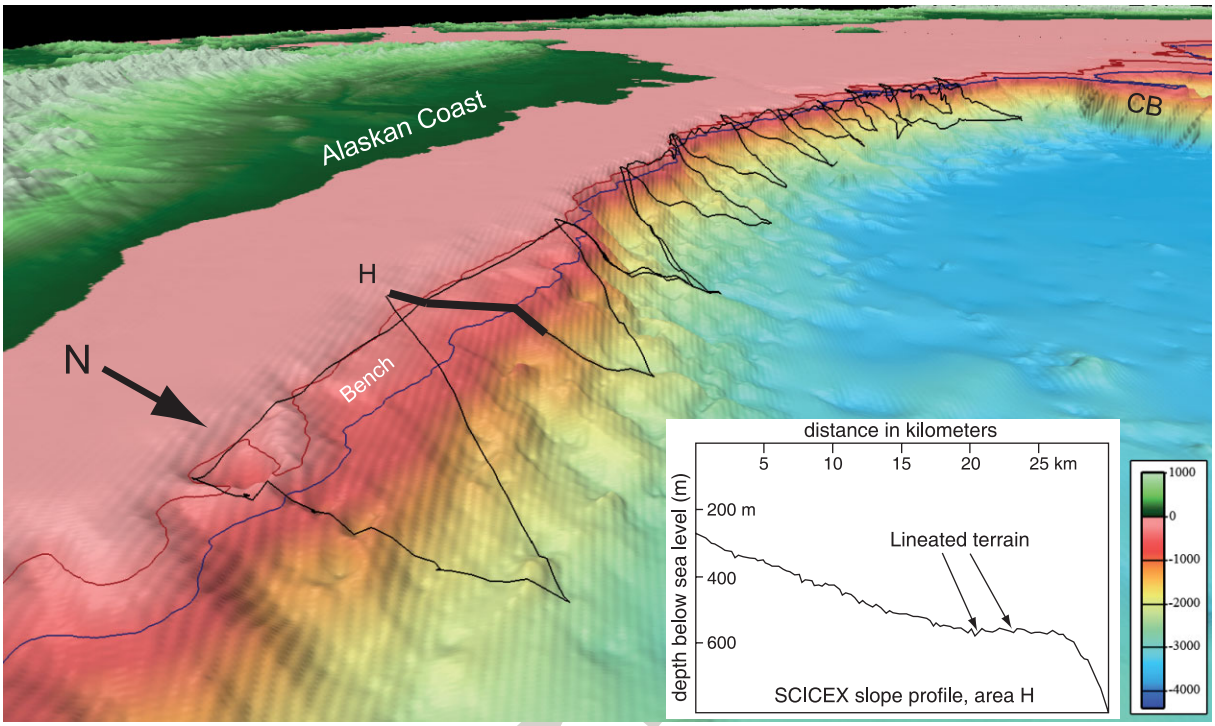


Figure 9. Three-dimensional view of the Alaska–Beaufort margin that highlights the bathymetric bench most prominent along the margin in Area H (see Figure 2 for location and description of the bathymetric datasets). SCICEX survey track shown in black, the 350 m and 600 m contours that bound the bench are shown in red and blue, respectively. The bathymetry data are displayed as colour-coded contours overlaid on a sun-shaded digital terrain model illuminated from the north. The diagonal stripes in the bathymetry data are artefacts that result from resampling the International Bathymetric Chart of the Arctic Ocean data at higher resolution than their 2·5 × 2·5 km grid cell size. CB, Chukchi Borderland. Inset shows SCICEX slope profile of area H along the transect indicated with heavy black line on the three-dimensional map.

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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 m w.d. 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 regrounding 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 palaeoglaciological 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).

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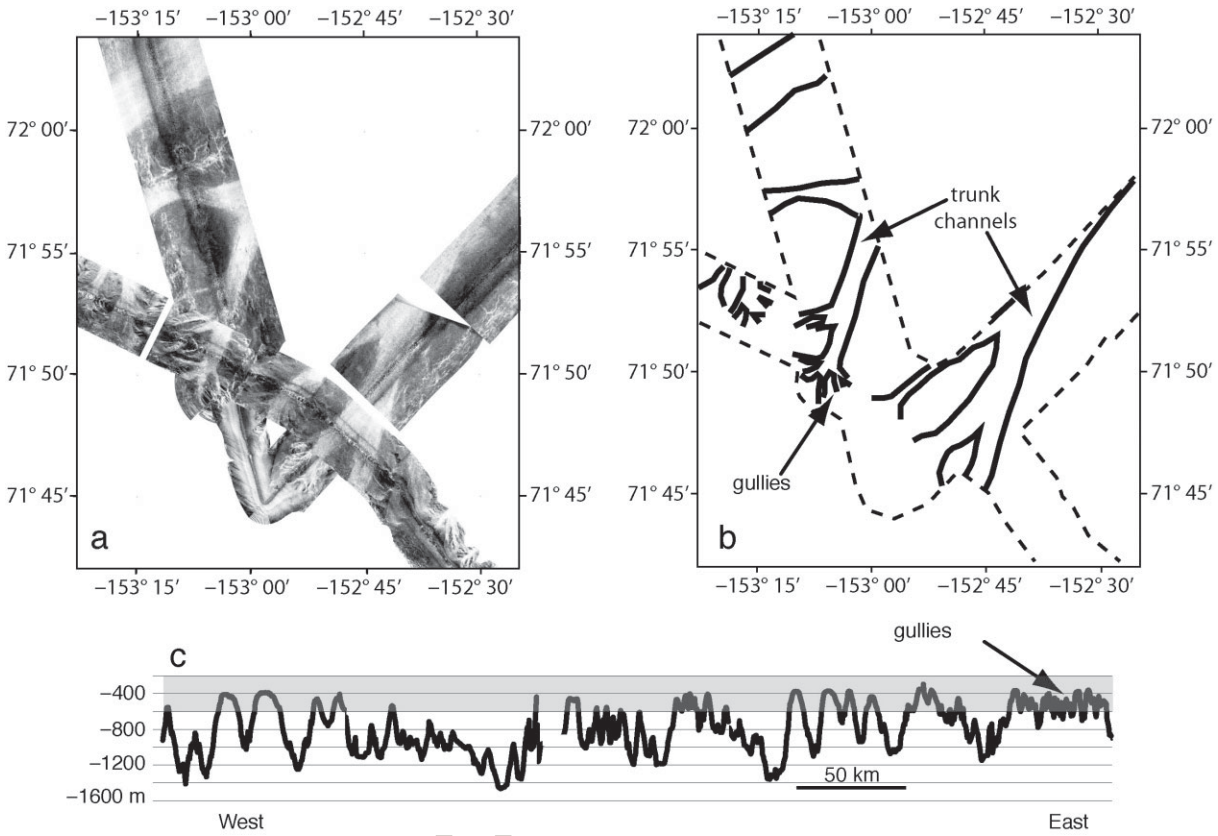


Figure 10. (a) Sidescan image of an extensive dendritic gully system coalescing into trunk channels (area E on Figure 2). (b) Interpreted outline of the gully system. (c) Highest resolution SCICEX (nadir) bathymetry of Alaska–Beaufort margin topography collected along the returning submarine track, which surveyed approximately parallel to the 366-m isobath. Water depths between 200 and 600 m are highlighted. Arrow indicates typical examples of gullies along the margin.

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

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

Spectral analysis of our bathymetric data contributes a quantitative and margin-wide characterization of the bedforms (Figure 8). Where we document the best examples of subglacial lineations on the eastern half of the survey in areas E, G and H, amplitudes of seafloor roughness are moderate and cover a narrow range of dominant wavelengths from ~45 to 100 m. In the westernmost, shallower half of the survey, where iceberg scours and grooves erode the substrate, samples from areas A, B and C show the strongest signals of seafloor excavation and a wider range of wavelengths

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from ~20 to 300 m. These results corroborate the change in bedform character along the ABM seen in sidescan and bathymetry images (Figures 3 and 5).

It is possible that areas A, B and C were impacted by the same large ice mass that eroded and/or moulded the seafloor in areas E, G and H; however, the chaotically oriented iceberg scours and grooves that primarily characterize the western end of the margin probably overprinted the lineated topography. The two types of morphologically distinct scours and grooves observed suggest the possibility of two generations of iceberg activity, one during the collapse of an ice shelf that eroded the ABM, and the other during a later episode of massive iceberg discharge from the Laurentide Ice Sheet. Both types have a similar azimuth range (95–100°N) (Figure 2) indicating that icebergs drifted clockwise along the ABM and impacted the slope at an oblique angle from the northeast. This drift pattern is consistent with the distribution of ice-rafted debris in Pleistocene sediments of the Canada Basin assumed to have been dispersed by icebergs carried by the clockwise Beaufort Gyre (Phillips and Grantz, 2001) (Figure 1).

Formation of Alaska–Beaufort margin slope gullies

The co-existence of the slope gullies with the glacially eroded portion of the ABM (Figures 3, 5; 9 and 10), and their morphological and contextual similarity to gully systems documented on glaciated continental margins around Antarctica (Ó Cofaigh *et al.*, 2003), East Greenland (Ó Cofaigh *et al.*, 2002) and the Gulf of Alaska (Carlson *et al.*, 1990), suggest that the gullies are related to glaciogenic processes. In each of the above localities, ice is inferred to have extended to the continental shelf edge during glacial periods, releasing subglacially entrained sediments and subglacial meltwater where the ice front separated from the seafloor. Sediments were subsequently remobilized by downslope gravity flows and turbidite gullies were eroded into the upper continental slope. Ó Cofaigh *et al.* (2003) argue that where slope gradients are sufficiently steep, any subglacial 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 and 700 m w.d.) 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 subglacially 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 gullies and other large-scale slumps are related; high-resolution sub-bottom data from the gullied slope are needed to clarify this relationship.

Ice flow provenance and trajectory

The observed lineation orientation along the ABM could indicate ice flow in either an 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. At least two fields of lineations 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, overdeepened troughs traversing the western part of the CAA (Amundsen Gulf, M'Clure Strait) and, possibly, the partially buried 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, 2006) and with models of ice discharge

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(Bigg and Wadley, 2001), all indicating the likelihood of large ice streams channelling 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 centre. 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). Based on ice mass balance considerations, however, this ice would need to be buttressed on both sides of the stream to maintain its full thickness without calving and breaking apart on its northern, oceanward margin. Continuous sea-ice cover during glacial periods might briefly retard the calving rate along an ice sheet margin (Reeh *et al.*, 2001) but it is unlikely that sea ice with a maximal thickness of ~7 m when concentrated by wind (Bourke and Garrett, 1987), could buttress an advancing ice shelf with a thickness two orders of magnitude greater.

An alternative mechanism to explain the circuitous flow path of ice along the basin margin invokes the existence of a thick ice shelf capping the entire Arctic Ocean (Mercer, 1970; Hughes *et al.*, 1977; Grosswald and Hughes, 1999). In these models, the basin-wide ice shelf, fed by multiple sources around its periphery, could have prevented land-sourced ice streams from penetrating directly into the central basins (Grosswald and Hughes, 1999). Evidence for glacial erosion on top of the Lomonosov Ridge (Figure 1) (Jakobsson, 1999; Polyak *et al.*, 2001) makes this scenario possible, provided there were synchronously growing ice shelves on both the Eurasian and Laurentide sides of the ocean. Based on preliminary, simplified numerical modeling (Jakobsson *et al.*, 2006), the growth of an ice shelf several hundred metres thick from the Eurasian margin all the way to the Lomonosov Ridge is possible provided there is a condition that constrains the stretching of ice. Such a constraint might be provided by abnormally thick sea ice or piled-up icebergs during glacial stages, but this hypothesis needs to be tested by more refined modelling based on comprehensive, age-specific palaeogeographical reconstructions. For example, the limited extent of the Eurasian ice sheet in the Arctic during the Last Glacial Maximum (Svendsen *et al.*, 2004) refutes the concept of the pan-Arctic ice shelf during that time period, but earlier glacial periods could have been more favourable for its formation. Notably, the age of glacial erosion on the Lomonosov Ridge is estimated at marine 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 megabergs 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.

Age control

The age of maximal glacial erosion on the Chukchi Borderland has been tentatively estimated between OIS 4 to 5d (Polyak *et al.*, 2007). Because the prevalent orientation of glaciogenic lineations representing this erosion on the CB align spatially with those at the ABM (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 palaeontological data suggest a glaciomarine rather than subglacial origin for this deposit (Dinter *et al.*, 1990; Brigham-Grette and Hopkins, 1995). We hypothesize that the grounding of several hundred metre 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 probably have occurred during OIS 4 or an interstadial

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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 post-dating OIS 6 could have generated ice shelves or iceberg armadas nearly comparable in size to prior surge events, but not quite thick enough to have scoured the deep crest of the Lomonosov Ridge at ~1000 m w.d. Furthermore, the megaberg purge trajectory could vary slightly from one glacial period to another, allowing some very large icebergs to miss the shallowest portions of the Lomonosov Ridge but still collect in the Amerasian Basin.

Conclusions

Sidescan and bathymetry data obtained during the SCICEX programme illustrate the morphology of the Alaska–Beaufort margin in the depth range of 200 to 2800 m, and help to elucidate the sedimentary and glacial history of this part of the Arctic Ocean. Along several transects the ABM data show an orderly association of glaciogenic seafloor bedforms such as scours, grooves and lineations, and larger-scale erosional bathymetric features in water depths reaching 700 m, with multiple turbidite gullies further downslope. Although our data coverage is more fragmentary than in some other survey areas, these bedforms are recognizable as geomorphological associations known from glaciated continental margins in the Arctic and Antarctic (e.g., Anderson, 1999; Shipp *et al.*, 1999; Polyak *et al.*, 2001; Ó Cofaigh *et al.*, 2003; Dowdeswell *et al.*, 2004). At 400–700 m w.d., fields of subglacially generated lineations cover an estimated minimum seafloor area of 330 km^2, indicating margin-parallel flow of a large ice mass buttressed against the ABM. Lineations occur on the surface of a bathymetric bench at 400–550 m that may have been formed by ice erosion, an interpretation that is corroborated by the coincident trends of lineations and bathymetric contours on the bench surface. We note also that this large-scale erosion of the continental slope makes the ABM similar to overdeepened continental margins that are characteristic of formerly glaciated areas. Iceberg scours and grooves observed on the ABM in shallower water depths were probably generated during ice-shelf break-up as well as later glacial event(s). The bench is truncated to the north by steep dendritic gullies that initiate in branching catchment areas up to 18 km across and coalesce to form broad trunk channels in water depths >1500 m. Gullies are interpreted to have formed in response to sediment-laden subglacial meltwater cascading down the steep continental slope. We acknowledge that the data presented are fragmentary and therefore are not as conclusive as those documented for some other glaciated margins. Further ABM surveying is needed to delineate the areal distribution of glaciogenic bedforms and to verify their spatial consistency.

The ABM glaciogenic bedforms have depths, styles and orientations similar to features discovered earlier on the Chukchi Borderland (Figure 1) (Polyak *et al.*, 2001, 2007; Jakobsson *et al.*, 2005) and were probably formed during the same glacial episode. The ABM bedforms link CB glaciogenic features with the overdeepened troughs of the CAA that were the likely outlets for ice streams from the northwest sector of the Laurentide ice sheet, and which could have propelled ice shelves into the Canada Basin. The margin-parallel pathway of this ice flow along the ABM indicates that incoming ice was diverted from streaming directly into the central part of the basin by some obstruction. We infer that such an obstruction was formed by either a basin-wide ice shelf fed by multiple sources, or by a pile-up of Antarctic-style icebergs driven by winds and currents from the Eurasian side of the Arctic Ocean. Corroboration of the movement of very thick ice (~1000 m), be it an ice shelf or iceberg armadas, towards the Canada Basin, is seen in the eroded crest of the Lomonosov Ridge at the centre of the Arctic Ocean (Jakobsson, 1999; Polyak *et al.*, 2001; Kristoffersen *et al.*, 2004; Jakobsson *et al.*, 2006).

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

Acknowledgements

National Science Foundation awards to Edwards and Polyak were used to support this project (OPP-9813535, 0136238, 9813532 and 0401692). The manuscript was significantly improved in response to the commentary of two anonymous reviewers.

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