Glacial geomorphology of the Central Arctic Ocean: the Chukchi Borderland and the Lomonosov Ridge

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Abstract
The last decade of geophysical seafloor mapping in the Arctic Ocean from nuclear submarines and icebreakers reveals a wide variety of glaciogenic geomorphic features at water depths reaching 1000 m. These findings provide new and intriguing insights into the Quaternary glacial history of the Northern Hemisphere. Here we integrate multi- and single beam bathymetric data, chirp sonar profiles and sidescan images from the Chukchi Borderland and Lomonosov Ridge to perform a comparative morphological seafloor study. This investigation aims to elucidate the nature and provenance of ice masses that impacted the Arctic Ocean sea floor during the Quaternary. Mapped glaciogenic bedforms include iceberg keel scours, most abundant at water depths shallower than ~350–400 m, flutes and megascale glacial lineations extending as deep as ~1000 m below the present sea level, small drumlin-like features and morainic ridges and grounding-zone wedges. The combination of these features indicates that very large glacial ice masses extended into the central Arctic Ocean from surrounding North American and Eurasian ice sheets several times during the Quaternary. Ice shelves occupied large parts of the Arctic Ocean during glacial maxima and ice rises were formed over the Chukchi Borderland and portions of the Lomonosov Ridge. More geophysical and sediment core data combined with modeling experiments are needed to reconstruct the timing and patterns of these events. Copyright © 2008 John Wiley & Sons, Ltd.

Keywords: Arctic Ocean; seafloor morphology; geophysical mapping; ice shelves; icebergs; glaciogenic bedforms; Pleistocene

Introduction
It is well established that large areas of Polar continental margins as well as some submarine ridges in the adjacent deep oceans have been repeatedly impacted by grounded ice sheets (see, e.g., Davies et al., 1997; Dowdeswell and Ó Cofaigh, 2002). Ice-produced bedforms preserved on the seafloor in such areas provide the keys to our understanding of the marine extension and history of these ice sheets. Over the last two decades, glaciogenic morphology has been investigated at many sites on formerly glaciated seafloors on the continental margins of North America (Josenhans and Zevenhuizen, 1990; Todd et al., 1999), northern Eurasia (Ottesen et al., 2005; Polyak et al., 1997; Solheim et al., 1990; Vogt et al., 1994) and around the Antarctic continent (Anderson, 1999; Canals et al., 2002, 2000; Ó Cofaigh et al., 2002; Shipp et al., 1999; Wellner et al., 2006). New advances in geophysical mapping technologies such as high-resolution multibeam bathymetric echo sounders and sub-bottom profilers can provide an even clearer view of glaciogenic morphology than remote sensing imaging of land areas due to the effects of post-glacial land erosion and obscuring vegetation cover. For example, a variety of glaciogenic bedforms such as de Geer moraines, flutes, drumlins and iceberg scours have been mapped using multibeam sonar with exceptional detail on the Svalbard, Antarctic (Ross Sea) and Nova Scotia continental margins (Ottesen and Dowdeswell, 2006; Shipp et al., 1999; Todd et al., 1999; Wellner et al., 2006).
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Table I. Expeditions when data used in the present study were collected

<table>
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<tr>
<th>Cruise</th>
<th>Vessel</th>
<th>Geophysical mapping/geological sampling</th>
<th>References (e.g.)</th>
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</thead>
<tbody>
<tr>
<td>Arctic Ocean 96</td>
<td>Swedish icebreaker Oden</td>
<td>chirp sonar profiling/sediment coring</td>
<td>Backman et al. (1997)</td>
</tr>
<tr>
<td>HLY0302</td>
<td>USS Pargo and Hawkbill</td>
<td>chirp sonar profiling/multibeam bathymetry</td>
<td>Jakobsson et al. (2005)</td>
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<tr>
<td>HLY0501/0503</td>
<td>USCGC Healy</td>
<td>chirp sonar profiling/multibeam bathymetry/sediment coring</td>
<td>Darby et al. (2005)</td>
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*The bathymetric and sidescan sonar data acquisition was only partly working during the 1998 cruise.

The perennially sea ice covered Arctic Ocean stands out as one critical area for which geomorphic data collection is less extensive, despite its obvious importance for understanding the Quaternary history of both the Arctic Ocean deep central basin and the shallow shelf areas, where large ice sheets extended during glacial maxima. Nevertheless, the use of seafloor mapping equipment from submarines and icebreakers during the last decade has generated a considerable amount of geomorphic data from the Arctic Ocean, only a minor portion of which has been published (Engels et al., 2008; Jakobsson, 1999; Jakobsson et al., 2005; Kristoffersen et al., 2004; Polyak et al., 2007, 2001). Not surprisingly, the interpretation of reported glaciogenic features allows for different viewpoints, and the reconstruction of the history and dynamics of ice masses invading the Arctic Ocean remain poorly understood. The hypothesis of the pan-Arctic ice shelf proposed by Mercer (1970) and further advocated notably by Grosswald and Hughes (1999) postulates the existence of a kilometre-thick floating ice shelf over the entire Arctic Ocean, which probably implies the initial formation of this ice shelf by thickening of sea ice. The opposite point of view is that all deep-draft erosion found in the Arctic Ocean was generated by large icebergs carried in ‘armadas’ by currents (Kristoffersen et al., 2004). There is a need for a synthesis and analysis of the accumulated data in order to further test these hypotheses. Here we report and discuss a compilation of seafloor mapping data acquired during the 1998–1999 SCICEX nuclear-submarine expeditions and several icebreaker cruises including the Oden 1996, the USCGC Healy 2003, and the 2005 Healy–Oden Trans-Arctic Expedition (HOTRAX) (Table I; Figure 1). We focus on the two principal areas that contain the majority of mapped glaciogenic bedforms in the Arctic Ocean – the Chukchi Borderland and the central Lomonosov Ridge. A comparative analysis of glaciogenic morphologies from these areas allows a characterization of the impact of deep-draft ice on the central Arctic Ocean.

**Study Areas**

There are several specifics of the Arctic Ocean that make it different from other oceanic regions that have been affected by large ice masses. The Arctic Ocean basin is nearly landlocked and relatively small, approximately 9.5×10⁶ km² (Jakobsson, 2002), with only one deep-water connection to the world ocean via the Fram Strait (Figure 1). No other ocean experienced such dramatic changes between glacial and interglacial periods; during periods of low sea level during glacial stages the area of the Arctic Ocean shrank by more than 50 per cent (Jakobsson, 2002) and about half of the exposed shallow continental margins were occupied by continental ice sheets (Dyke et al., 2002; Svendsen et al., 2004). The ice and melt-water discharge from these ice sheets had an exceptionally high impact on the Arctic Ocean circulation and sedimentation regimes. Indeed, seafloor mapping data indicate bedforms resulting from glacial erosion and moulding of sea floor at depths reaching 1000 m. These glacially impacted areas are not always immediately adjacent to former ice sheets, but are located inside (Lomonosov Ridge) or protrude into (Chukchi Borderland, Yermak Plateau) the Arctic Ocean basin.

The Chukchi Borderland is comprised of a group of generally less than 1000 m deep, north-trending ridges that surround the extensional Northwind Basin (Figure 2) (Hall, 1990). The easternmost of these ridges is the Northwind Ridge, which is deeper than its western neighbours and is characterized by an exceptionally steep slope towards the Canada Basin and a gently rounded to flat topped ridge crest. The Chukchi Spur–Chukchi Plateau composite ridge lies on the western side of the Northwind Basin and has a wide (>140 km at 78° N) flat topped crest mainly shallower than 600 m (Figure 2(a)). In addition to the large Northwind Ridge and Chukchi composite ridge, several much smaller ridges rise above the floor of the Northwind Basin.

The Lomonosov Ridge is a continental sliver that separated from the Barents–Kara Sea continental margin at about 57 Ma during the opening of the Eurasian Basin through sea floor spreading (Brozena et al., 2003; Jokat et al., 1995;
Figure 1. Bathymetric map of the Arctic Ocean showing track lines of the cruises that collected the geophysical data studied in this work (orange, Healy 2003; yellow, Healy (HOTRAX) 2005; red, SCICEX 1999; purple, SCICEX 1998; light grey, Oden 1996). The types of data collected during each of these cruises are listed in Table I. The black boxes outline the locations of the main study areas of the Chukchi Borderland (Figure 2) and the Lomonosov Ridge (Figure 3). The two main surface circulation patterns that govern the present sea ice drift, the Beaufort Gyre and Transpolar Drift, are marked with grey bold arrows. AG, Amundsen Gulf; AR, Alpha Ridge; CP, Chukchi Plateau; LR, Lomonosov Ridge; MS, M'Clure Strait; MR, Mendeleev Ridge; MV, Mackenzie Valley; NR, Northwind Ridge; StA, Saint Anna Trough; YP, Yermak Plateau; FS, Fram Strait. The 1000 m isobath is shown with a black line.

Vogt et al., 1979). The ridge extends 1650 km across the central Arctic Ocean, from the continental margin of northern Greenland all the way to the shallow shelf of the Laptev Sea. The ridge crest is generally deeper than 1000 m, although there are shallower portions in both the eastern and western ends and in the central part between about 83°50″ and 87°30″ N; 140° and 160° E (Figure 3). The morphology is here characterized by a flat topped ridge crest, with the exception of the smaller areas at about 84°20′ N; 148° E and 86° N; 158° E just extending above the 1000 m isobath.

The first evidence of deep-draft ice impact in the Arctic Ocean was provided by single-beam bathymetric lines showing the extensively scoured surface of the Chukchi Plateau at depths to more than 400 m (Hunkins et al., 1962). Deeper scours, to at least 850 m, were found on the Yermak Plateau, indicating the passage of numerous exceptionally large icebergs exceeding in thickness the biggest icebergs found today off Antarctic coasts (Vogt et al., 1994). The 1996 Oden expedition collected sub-bottom profiler records and sediment cores from the Lomonosov Ridge, at the very centre of the Arctic Ocean (Backman et al., 1997). The surveyed portion of the ridge crest was shown to be truncated to water depths of ~1000 m, with a large stratigraphic hiatus and a layer of over-compacted sediments characterizing the truncated surface (Jakobsson, 1999; Jakobsson et al., 2001).

The SCICEX 1998–1999 submarine expeditions with USS Pargo and Hawkbill included a relatively dense survey (sub-bottom and/or sidescan sonar profiling with approximately 10 km between track lines) of the central Lomonosov Ridge as well as a number of tracks on the Chukchi Borderland (Edwards and Coakley, 2003) (Figure 1). These data provided convincing evidence that the eroded areas on both the Lomonosov Ridge and the Chukchi Borderland
contain widespread, diverse glaciogenic morphology such as sets of parallel lineations (flutes), transverse ridges and drumlins (Polyak et al., 2001). The combination of these features depicts a dramatic impact of large ice masses that once invaded the central Arctic Ocean from both the Laurentide and Eurasian ice sheets. Later surveys performed from the USCGC Healy detailed and expanded the SCICEX results and revealed a complex pattern of ice impact on the eroded areas (Darby et al., 2005; Jakobsson et al., 2005). In particular, combined with stratigraphic investigation of sediment cores (Polyak et al., 2007), these data indicate multiple episodes of glacial erosion on the Chukchi Borderland with different orientations of ice movement.
Methods

USCGC Healy Multibeam bathymetry and chirp sonar sub-bottom profiling

The USCGC Healy is equipped with a 12 kHz Seabeam 2112 multibeam bathymetric sonar. It has 151 beams and produces a swath width of about 2.5–3.5 times the water depth. Multibeam bathymetry was collected from the Alaskan continental shelf and Chukchi Borderland during the HLY0302, HLY0501 and HLY0503 cruises (Table I and Figure 1). The Lomonosov Ridge was mapped during the HLY0503 cruise, but only at water depths deeper than the deepest observed ice erosion, i.e. depths greater than 1000 m. Sound speed profiles of the water column for depth calibration were acquired at least once a day either using the ship’s CTD (SeaBird SBE-911) or expendable probes (XBT, XSV, XCTD). During heavy ice breaking in the central Arctic Ocean (9/10 sea ice cover), the quality of the acquired bathymetric data was substantially reduced. The initial post-processing of the multibeam data was done onboard USCGC Healy directly after data acquisition using the software Caris HIPSTM and FledermausTM. For this present study, further post-processing and data analysis has been carried out at Stockholm University using the same software set-up as onboard USCGC Healy. Grids with optimized resolutions (25–100 m) depending on the general depths of mapped areas were created for the analysis of glaciogenic bedforms.

In addition to a multibeam system, the Healy has two sub-bottom profilers: (1) Knudsen 320B/R dual frequency (centre frequencies 4.5 and 12 kHz) and (2) ODEC Bathy-2000 (centre frequency 3.5 kHz). Both systems are ‘chirp sonar’ sub-bottom profilers operating using frequency modulated (FM) signals and apply match filtering signal processing.
to compress and remove the source signature (Schock et al., 1989). The sub-bottom data in the two systems’ acquisition software can be stored both in native (ODEC and Knudsen respectively) and SEG-Y formats. The SEG-Y files were post-processed in this study using the public domain seismic software package Sioseis (http://sioseis.ucsd.edu/). Envelopes of the sub-bottom traces (the correlates) were computed through standard signal processing procedures, automatic gain control (AGC) was applied when needed, and the sub-bottom profiles were displayed in grey scales.

SCICEX bathymetry and sub-bottom profiles

The Seafloor Characterization and Mapping Pod (SCAMP) was installed on the nuclear submarine USS Hawkbill for the SCICEX 1998–1999 expeditions and contained a sidescan swath bathymetric sonar and a modified ODEC Bathy-2000 sub-bottom profiler (Chayes et al., 1998). The ODEC sub-bottom profiling data were stored in SEG-Y format and post-processed with Sioseis in this present work, applying the same set-up as described above for the USCGC Healy data. The SCICEX sidescan and swath bathymetry was initially processed using the open source software MB-System and Generic Mapping Tools (GMT) (Caress and Chayes, 2001; Wessel and Smith, 1991) by the Hawaii Mapping Research Group (Edwards and Coakley, 2003). Bathymetric grids with resolutions of 250 m \times 250 m on a polar stereographic projection were included in the bathymetric regional compilations described below and used to calibrate the depths of the SCICEX sub-bottom profiles.

Bathymetric regional compilations

Digital bathymetric models (DBMs) with grid resolution of 1250 m \times 1250 m on a polar stereographic projection were compiled for the central Lomonosov Ridge and the Chukchi Borderland with adjacent areas using all available depth information (Figures 2 and 3). In addition to the multibeam bathymetry acquired during the USCGC Healy and SCICEX expeditions, the depth information includes the entire International Bathymetric Chart of the Arctic Ocean (IBCAO) database (Jakobsson et al., 2000) and multibeam bathymetry not yet included in IBCAO from the HLY0405 cruise with USCGC Healy (Mayer, 2004) and RV Nathaniel B Palmer cruises within the Shelf Basin Interactions (SBI) programme (http://www.eol.ucar.edu/projects/sbi/). In order to compile the DBMs at a resolution of 1250 m \times 1250 m, interpolation between the multibeam, single beam and digitized bathymetric contours from bathymetric maps was performed applying the same methodology and algorithms as used to compile the IBCAO DBM (Jakobsson et al., 2000).

Glaciogenic Bedforms

Iceberg scours

The most common type of seafloor feature in polar areas is ice keel scours, also commonly called plowmarks, furrows or gouges (Davies et al., 1997; Duncan and Goff, 2001; Meyer et al., 2000). Sea ice can only scour the seafloor to water depths of less than approximately 50 m, while iceberg erosion can extend to depths of several hundred metres (Dowdeswell and Bamber, 2007; Dowdeswell and Whittington, 1992). Understanding of the formation of iceberg scours has been enhanced by comparison with modern iceberg drift patterns (Tchernia and Jeannin, 1984) and by direct field observations (Woodworth-Lynas et al., 1991). On the Chukchi Borderland and adjacent Alaskan margin scours are very abundant at depths shallower than ~350–400 m (Figure 4) (Engels et al., 2008; Hunkins et al., 1962; Jakobsson et al., 2005; Polyak et al., 2001). These are mostly sinuous, chaotically oriented, overlapping incisions with a v-shaped cross-section, less than 50 m wide, a few metres deep and less than 5 km long – analogous to ‘gouge-type furrows’ at the Antarctic margin (Meyer et al., 2000; Wellner et al., 2006). This is the most common type of modern iceberg scouring around sea-bound glaciers and ice sheets. At greater water depths on the Chukchi Borderland scours are more sparsely distributed, but they are larger and extend linearly for much longer distances. Individual scours reaching water depths of ~500 m are up to 700 m wide, 30 m deep and several kilometres long. They display internal groove-and-ridge topography as characteristic of the ‘plane furrow’ type (Meyer et al., 2000; Wellner et al., 2006). These features must have been produced by very large, possibly tilted tabular icebergs (Dowdeswell and Bamber, 2007). A similar pattern in the distribution of iceberg scours has been reported from the St. Anna Trough at the glaciated Eurasian Arctic margin (Polyak et al., 1997).

Exceptionally deep seafloor areas (to ~850 m below present sea level) displaying multiple, large iceberg scours are found on the Yermak Plateau and the central Lomonosov Ridge (Figures 3 and 5) (Jakobsson, 1999, 2000; Polyak et al., 2001, 1999; Vogt et al., 1994). These scours are closely spaced and have a sub-parallel orientation, most
Figure 4. (a) Chirp profile CS2 from the ramp to the Northwind Ridge (see Figure 2 for location). (b) Multibeam swath from the same profile as shown in (a). Data were collected during the HLY0503 expedition (Table I). (c) Sidescan image (SCICEX 1999 data) from the same location as CS2.
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Figure 5. (a)–(c) Chirp profiles LR1 and LR2 along the iceberg-scoured area on the Lomonosov Ridge and LR3 from a deeper fluted area. The transition from an ice-scoured to lineated seabed at about 850 m water depth is clearly seen in LR1. Data were collected during the SCICEX 1999 expedition (Table I). The redeposited sediment leeward of the eroded ridge top is seen in LR2 and LR3. (d) Sidescan image of iceberg scours (SCICEX 1999 data) from the same location with sub-bottom profile LR1.

probably indicating the passage of (an) armada(s) of large icebergs (Figure 5(d)). The prevalent orientation of closely aligned scours on the Lomonosov Ridge allows for two opposite interpretations of iceberg drift direction – from the Eurasian versus the Amerasian basin of the Arctic Ocean (Jakobsson, 2000; Kristoffersen et al., 2004; Polyak et al., 2001). Accordingly, the mapped scours on the Yermak Plateau may indicate an ice drift direction either from or into the Arctic Ocean.

Flutes and megascale lineations

Another type of bedform common on glacially eroded polar continental margins areas is formed by fields of linear, evenly spaced, ridge-and-groove features – in both the Arctic (Jakobsson et al., 2005; Ö Cofaigh et al., 2002; Ottesen et al., 2005; Polyak et al., 2001, 1997; Solheim et al., 1990) and the Antarctic (Anderson, 1999; Gilbert et al., 2003; Shipp et al., 1999; Wellner et al., 2006). Such features, referred to as flutes or lineations, are characteristic of glaciated terrains with a relatively soft bed (Benn and Evans, 1998; Dowdeswell et al., 2004). Although there is a variety of opinions on the significance of erosional versus depositional mechanisms in the formation of flutes, most authors agree on their relation to relatively fast flowing ice such as at surging ice margins and ice streams, with the length of flutes roughly proportional to ice velocity (Clark et al., 2003; Stokes and Clark, 1999).

In the Arctic Ocean, lineations are found on the Lomonosov Ridge and the Chukchi Borderland with the adjacent Alaskan margin (Figures 4–7) (Engels et al., 2008; Jakobsson, 1999; Jakobsson et al., 2005; Polyak et al., 2001), as well as in the deep troughs traversing the formerly glaciated Arctic margins (Ottesen et al., 2005; Polyak et al., 1997; Stokes et al., 2006). Lineations are typically visible at water depths below the range of intense iceberg scouring and are observed to ~1000 m depth on the Lomonosov Ridge and ~900 m on the Chukchi Borderland (Northwind Ridge) (Figures 2 and 3). Deep-water Arctic Ocean lineations are mostly spaced between less than 50 and 200 m (crest to crest) and have low amplitudes (under 5–10 m) and lengths of more than 15 km. However, it should be noted that these are probably longer as measurements are limited by swath data coverage and/or the width of eroded ridge tops. The largest lineations mapped on the Chukchi Borderland (on the ramp to the Northwind Ridge) are spaced between 200 and 1000 m apart and have a relief (from the ridge top to the groove bottom) as high as 40 m (Figure 6) (Jakobsson et al., 2005). These features are similar to megascale glacial lineations reported from the Antarctic continental margin,
Figure 6. Multibeam lines at the ramp to the Northwind Ridge showing megascalen lineations (see Figure 2 for location). The multibeam data were acquired during the HLY0503 and HLY0302 expeditions (Table I).

where they have been related to palaeo-ice streams (Anderson, 1999; Ó Cofaigh et al., 2002; Shipp et al., 1999; Wellner et al., 2006).

Seafloor lineations comparable to those formed by glacial processes are found in non-glaciated regions as well (Bryant et al., 2007; Flood, 1983; Kuijpers et al., 2002; Poppe et al., 2006). These features typically occur in areas affected by persistent currents, either tidal or topographically steered flows through straits or along slopes. However, despite the general similarity, current-generated lineations have marked differences from their glaciogenic counterparts. While the latter are composed of ridges and troughs with similar cross-sectional shape and dimensions, current lineations are commonly formed by distinct furrows cut into generally flat seafloor and separated with interspaces several times wider than the furrows. Furthermore, these furrows have some peculiar features such as ‘tuning forks’ that are characteristic of the action of flowing water, but not ice (Flood, 1983; Poppe et al., 2006). In addition to morphology, critical information is also provided by the seafloor settings in which the lineations are found. Deep-sea current furrows are typically located along the slopes of continental margins and ridges, whereas glaciogenic flutes can occur in a variety of geomorphological settings, from troughs to ridge crests, and can be oriented at various angles to bathymetric obstacles. Moreover, glacial lineations are always found in an orderly combination with other glaciogenic bedforms including iceberg scours, morainic ridges and drumlins, as well as large-scale erosional features such as truncated strata. The combination of the above allows a confident identification of the glaciogenic origin of the lineations found on the Lomonosov Ridge and Chukchi Borderland.

Unlike chaotic or sub-parallel iceberg scours, subglacial lineations retain their parallel fabric for long distances. This, combined with the width of lineation fields (tens of kilometres or possibly more), indicates that they have been generated by very large, coherently moving ice masses such as ice streams or grounded portions of ice shelves known as ice rises. Lineated areas are typically associated with large-scale seafloor erosion, accentuated by a conspicuous truncation of pre-glacial strata capped with a mostly thin layer of diamict sediment, in which the lineations are formed (Dowdeswell et al., 2004). It must be noted that sub-bottom sparker or sonar records across iceberg-scoured and lineated areas may look similar due to abundant hyperbolic reflections; therefore, swath imagery (sidescan sonar or multibeam bathymetry) is critical for a definitive identification of lineations (Figures 4–7).

Drumlins

Drumlins are not common features of the Arctic Ocean seafloor, unlike on some glaciated continental margins, where drumlins are characteristic of the transitional zone between crystalline bed on the inner shelf and soft sedimentary
strata further offshore (Anderson et al., 2001; Wellner et al., 2006, 2001). Seafloor drumlins can also be formed from ice advances across morainic ridges (Fader et al., 1997). The reason for the rarity of drumlins in the ice eroded areas of the Arctic Ocean may be the soft pre-glacial bed comprised by poorly consolidated Meso-Cenozoic strata (Grantz et al., 1990; Moran et al., 2006). Nevertheless, small drumlin-like forms 300–600 m long are found at some sites, notably on the Northwind Ridge (Figure 7(a)). These bedforms appear to be formed by overriding and overprinting of morainic ridges by a younger ice movement. The presence of drumlins probably indicates a warm based ice sheet, and their extension is particularly important for identification of the direction of ice flow (Hättestrand et al., 2004).

**Morainic ridges and wedges**

In several areas on Chukchi Borderland, the glacially influenced seafloor features nested sets of slightly sinuous or arcuate ridges, transverse or oblique to flutes (Figure 7(a)). Ridges appear to be less than 10 m high (their exact height is difficult to assess due to the insufficient resolution of the SCICEX bathymetry), reach 100+ m in width and are individually spaced between 100 and several hundred metres. Similar features have been mapped on many glaciated shelves and interpreted as transverse morainic ridges (grounding zone wedges) formed by stepwise retreat of an ice-sheet margin or grounding line (Davies et al., 1997; Ottesen and Dowdeswell, 2006; Ottesen et al., 2005; Polyak
et al., 1997; Shipp et al., 2003; Todd et al., 1999; Wellner et al., 2006). Ridges mapped on the southeastern Chukchi Plateau are remarkably parallel to bathymetric contours (Polyak et al., 2001). This may indicate that the formation of these ridges was controlled by sea level and occurred in relation to the grounding line of a retreating ice cap (lift-off moraines) (Ottesen and Dowdeswell, 2006).

At several locations, larger ridges or wedges reach thicknesses of ~30 m. An example is the prominent asymmetric, nested wedges at the northernmost eroded bathymetric high of the Northwind Ridge (Figure 8(c)). These wedges have an acoustically transparent character with virtually no visible internal reflectors, and coring during the HOTRAX expedition revealed a diamict sediment composition. Similar wedges are characteristic of former grounding lines at the maximal ice-sheet extension where significant volumes of sediment have been accumulated (Anderson, 1999; Elverhøi and Maisey, 1983; Shipp et al., 1999). Fairly large ridges are also found at several segments of the Chukchi Plateau perimeter at water depths of 600–700 m (Figures 2 and 8(a)), probably marking the limits of an ice rise that once capped the plateau.

Redeposited sediment accumulations

Several seafloor depressions adjacent to eroded areas on both the Lomonosov Ridge and Chukchi Borderland contain accumulations of acoustically transparent material forming lenses occasionally as much as 75 m thick and several
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Figure 8. Sub-bottom profiles from the Chukchi Borderland. Figure 2 shows the locations of the profiles. (a) Morainic ridges from the Chukchi Plateau. (b), (e) Glacially carved bathymetric benches in the Chukchi continental shelf. (c), (f) Smooth seabed with no signs of glacial erosion on the Northwind Ridge at 620 m water depth and on the Chukchi Shelf at 450 m. (d) Morainic wedge on the Northwind Ridge.

kilometres across, sometimes with a convex relief (Figures 2–5). These accumulations, particularly pronounced on the Lomonosov Ridge, are coupled with an ice eroded seafloor, and we assume that these are bodies of diamict sediment formed by downslope debris flow (creep) of till in grounding-line proximal areas, similar to inferred ‘marine flow-till’ accumulations in the depressions of the Barents–Kara margin that reach the thickness of almost 100 m (Gataullin et al., 1993; Polyak et al., 2002). Coring of these deposits on the Lomonosov Ridge and Chukchi Borderland may provide the necessary material to study their exact formation mechanism.

Large-scale bathymetric features

Seafloor areas that feature small to medium-sized subglacial bedforms such as flutes and morainic ridges also often bear evidence of erosion over large areas. For example, this is expressed in the carving of broad troughs and bathymetric benches on the continental shelves and slopes (Blasco et al., 1990; Canals et al., 2002) and planation of ridge and plateau tops (Jakobsson, 1999; Polyak et al., 2001). Troughs that traverse glaciated continental margins in both the Antarctic and the Arctic typically contain multiple flutes or megascale lineations that indicate that these troughs operated as conduits for fast-flowing ice streams (Canals et al., 2000; Polyak et al., 1997; Stokes et al., 2006). A less
common case of continental slope erosion is a flattened bathymetric bench that is evident on the Alaskan margin at water depths between 400 and 550 m (Figure 8(b), (e)) (Engels et al., 2008). The broadest part of this bench, almost 40 km wide, can be traced for ca. 100 km along the Beaufort/Alaskan slope, and its narrower continuation extends all the way to the Chukchi Borderland, with small additional benches visible at some segments. Lineations found on the Alaskan slope are concentrated within this bench, which, together with its clearly erosional nature (truncation of slope strata), suggest that the bench has been formed by a large ice mass sliding along the margin.

A spectacular large-scale erosional feature that is intrinsic to the interior of the Arctic Ocean is the planation of ridge and plateau tops. This erosion is especially conspicuous atop relative bathymetric highs of the Northwind and Lomonosov Ridges, which feature flat, somewhat tilted surfaces with a pronounced unconformity (Figures 5(c) and 8(d)). The geometry of unconformity allows an estimate of the thickness of eroded strata, which amounts to more than 50 m on the Lomonosov Ridge (Jakobsson, 1999).

**Discussion**

**Regional glacial histories**

*Chukchi Borderland.* Numerous and diverse glaciogenic bedforms indicative of an impact(s) of very large and coherent ice masses are mapped on the Chukchi Borderland, including the Chukchi Plateau, Chukchi Spur and Northwind Ridge, at water depths reaching locally more than 900 m (Figure 2). Sets of bedforms are commonly superimposed upon one another, forming a geomorphic palimpsest indicative of a complex erosion history that probably involves several glacial events. This inference is consistent with the stratigraphic evidence of at least two asynchronous diamict horizons recovered in sediment cores from the eroded area at the ramp to the Northwind Ridge (Polyak et al., 2007). Based on these data, the youngest and least extensive ice-grounding event occurred during the Last Glacial Maximum (LGM), whereas the more widespread glacial impact is dated to Marine Isotope Stage (MIS) 4 or older (Polyak et al., 2007). The preliminary results of the HOTRAX’05 expedition suggest the presence of yet another diamict horizon, tentatively attributed to MIS 6.

Except for iceberg scours that largely obliterate pre-existing bedforms at water depths to 350–400 m, the most common glaciogenic features are lineations or megascule lineations running in a variety of orientations (Figure 2). Lineations are found on all bathymetric highs of the Northwind Ridge that rise above the depth limit for the glacial erosion, henceforth referred to as the ‘trimline’, and on the flanks of the Chukchi Plateau. At least two of the lineated fields have directional markings such as dragged sediment blocks or boulders on the Chukchi Plateau (Polyak et al., 2001) and drumlinized ridges at the northern Northwind Ridge (Figure 7(a)). These markings indicate the SE–NW direction of eroding ice at these sites. An overprinting of one generation of lineations by another is clearly visible in several locations (Figure 7(b)). These features may indicate the impact of separate glacial events or a complex pattern of ice flow during just one event (Dowdeswell et al., 2006; Wellner et al., 2006).

Sets of morainic ridges are also common for the Chukchi Plateau and some sites on the Northwind Ridge (Figures 2, 7(a) and 8(a)). The depth-contour alignment of ridges (lift-off moraines) mapped at the SE edge of the Chukchi Plateau indicates that they probably encircle the plateau (Polyak et al., 2001). This is corroborated by the consistent presence of ridges on sub-bottom sonar lines crossing its edges (Figures 2 and 8(a)). The placement of ridges suggests that they mark the limits of an ice cap that once covered the plateau, whereas the nested, bathymetry-controlled ridge pattern follows the retreat of this ice cap during sea-level rise. On the Northwind Ridge morainic ridges are abundant at the northernmost eroded mount where they are generally transverse to flutes and are partially drumlinized (Figure 7(a)). Furthermore, a stack of thick morainic wedges resides at the northern edge of this shallow area (Figure 8(d)), and large morainic accumulations are found atop eroded highs further south. In addition to lineations and ridges, lenses of displaced diamict material are common for the depressions in the Chukchi Plateau and slopes of Northwind Ridge bathymetric highs (Figures 2 and 4).

The bathymetric distribution of the eroded areas at the Chukchi Borderland is somewhat asymmetric (Figure 2). The eroded surface (trimline) dips to the largest water depths of more than 900 m at the southern part of the Northwind Ridge, from where it shallows to ~620 m along the ridge northwards and to between 760 and 800 m southwards towards the continental margin. Chirp sonar data show a yet shallower trimline at 550–700 m along the western edge of the Chukchi Plateau, whereas less than 200 km SW of the plateau the Chukchi margin appears to be uneroded at depths as shallow as 450 m (Figure 8(f)). This trimline distribution indicates that the impact of eroding ice masses on the sea floor had a complex pattern probably controlled by the interaction of ice geometry and dynamics with the large-scale morphology of the continental slope and the borderland. The upper limit of the distribution of erosional bedforms towards the continental margin appears to be at a depth of ca. 400 m south of the Chukchi Cap; however,
their identification at shallower depths may not be possible because of intense iceberg scouring at deglaciation and wave erosion during sea-level rise.

Overall, the orientation of glaciogenic bedforms suggests the possibility of two mechanisms of glacial impact on the Chukchi Borderland. One orientation of lineations and the spatial pattern of large-scale erosion indicate the impact of ice masses overriding the Chukchi Borderland from the east and southeast (Figure 2). We infer that they were formed by grounding portions of a coherent floating ice masses (ice shelves) propelled by ice streams of North-American Pleistocene ice sheets (Dyke et al., 2002; Kleman and Glasser, 2007). The nearest possible source was the NW sector of the Laurentide Ice Sheet, which contained very large ice streams discharging voluminous amounts of ice into the adjacent Arctic Ocean (Bigg and Wadley, 2001; Stokes et al., 2005, 2006). The east–west oriented lineations suggest the direct flow of ice from these ice streams, whereas the NW-trending flutes and megascale lineations indicate a more curved pattern of ice flow, possibly steered by the continental slope and constrained by very thick ice in the interior of the Arctic Ocean. This pathway for the flow of eroding ice is consistent with the erosional bench at the Alaskan slope between the NW Laurentide margin and the Chukchi Borderland (Engels et al., 2008).

The other pattern that can be inferred from the distribution of glaciogenic bedforms such as the concentric ridges at the edges of the Chukchi Plateau and the orientation of many fluted fields is the impact of an ice cap centred over the plateau or the entire borderland. Such an ice cap may have developed from an ice rise, such as those found around the Antarctic, where they are formed when an ice shelf grounds on a shallow portion of seafloor (Fahnestock et al., 2000). After its inception, an ice rise may start to act as an independent, local ice cap expanding from the centre outwards, with the ice shelf flowing around. From the available data we cannot elucidate conclusively the complex interaction between these potential sources of erosion. More detailed mapping and stratigraphic investigation is needed to understand the spatial and temporal relationship of bedforms formed by each of these ice masses.

**Lomonosov Ridge.** The major highs of the central Lomonosov Ridge, extending above approximately 1000 m present water depth, bear clear evidence of an impact of grounded ice masses, which extensively eroded the ridge crest and left their imprint in the form of glaciogenic bedforms including lineations, iceberg scours and associated lenses of redeposited sediment (Figure 3). The unconformable truncation of the strata revealed by the sub-bottom profiles provides an unambiguous erosional explanation for the overall flattened ridge crest morphology seen in bathymetric compilations (Figures 3 and 5) (Jakobsson et al., 2000; Narayshkin, 1999). The seafloor of the truncated areas contains multiple flutes and iceberg scours. There is a marked morphological transition at approximately 850 m water depth seen in the shallower southern eroded area around 86° N (Figures 3 and 5(a), (b)). Above this depth, the seabed is dominated by irregular scours with widths exceeding 100 m, suggesting rough under-ice topography. Sidescan images of the scoured surface show subparallel iceberg scours with a prevalent orientation across the ridge (Figure 5(d)) (Jakobsson, 2000; Polyak et al., 2001). Below the transition, the seafloor appears much more even and displays a fluted fabric on sidescan images. We suggest that this change in morphology marks the limit for deep draft icebergs, whereas the deeper fluted seabed (down to 1000 m present water depth) requires larger, coherent ice masses such as moving ice shelves that ground over bathymetric highs and form ice rises.

Icebergs that scoured the Lomonosov Ridge at depths shallower than 850 m could have been clustered together in armadas (Kristoffersen et al., 2004) during periods of intense iceberg discharge from disintegrating ice sheets. The same depth of relict iceberg scours with a similar orientation is found on the Yermak Plateau (Vogt et al., 1994). Iceberg scours down to 850 m below present water depth imply iceberg drafts of about 700–730 m during glacial maxima (120–150 m lower sea level). Today, icebergs off Antarctica and Greenland reach depths of 500–550 m (Barnes and Lien, 1988; Dowdeswell and Bamber, 2007; Dowdeswell et al., 1993), but relict scours off Greenland have been found to reach substantially larger water depths (Kuipjers et al., 2007). The iceberg dimensions required to ground on the Lomonosov Ridge are thus not exceptional. However, deep draft icebergs are usually found close to the grounding line of a glacier, whereas the distance between the scoured areas of the Lomonosov Ridge and the nearest continental margin that could have provided hundreds of metres thick icebergs during glacial maxima was more than 700 km away.

The position of redeposited sediments on only one side of the Lomonosov Ridge indicates that the eroding ice moved across the ridge from the Eurasian to the Amerasian side (Figures 3 and 5). The Barents and Kara Sea ice sheet, which reached the shelf edge at least during Marine MIS 2, 4 and 6 (Svendsen et al., 2004), seems the most likely source for this ice. We note also that this continental margin features very large erosional troughs such as the St. Anna Trough that probably contained major ice streams during glacial maxima (Siegert and Dowdeswell, 2004). Sediment core data show that the undisturbed 2–3 m sedimentary drape atop the ice-eroded surface of the Lomonosov Ridge below ca. 800 m water depth has been deposited since Marine Isotope Stage (MIS) 5-5 (Jakobsson et al., 2001). This result indicates that the deepest erosional event in this area occurred during MIS 6, which was the time of the largest known glaciation on the Barents–Kara shelf (Svendsen et al., 2004). One core collected during the Arctic Ocean 96 expedition with Oden at a shallower water depth of approximately 600 m has a shorter post-erosional stratigraphy, thus indicating a younger erosional event caused by a thinner ice.

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Figure 9. Sub-bottom profiles from the Lomonosov Ridge showing uneroded seafloor above 1000 m. Critical seafloor depths used in our interpretations of the glacial morphology are shown in the profiles. The depths have been derived by depth calibration of the sub-bottom profiles using multibeam bathymetry.

We conclude that the least contradictory interpretation of the existing data is that the major erosion that impacted the central Lomonosov Ridge to 1000 m water depth was related to the extension of an ice shelf from the Barents–Kara ice sheet during MIS 6. After grounding on the proximal side of the ridge crest, the ice shelf could have formed an ice rise, so that further ice dynamics were controlled by the interaction of an ice shelf and ice rise. This might explain the lower thickness of the eroding ice on the distal (Amerasian) side of the ridge as indicated by the uneroded surface in this area at water depth as shallow as ca. 900 m (Figures 3 and 9). Subparallel scours above 850 m depth were probably gouged during deglaciation when armadas of icebergs formed upon disintegration of the ice shelf. The direction of iceberg movement is not apparent. Kristoffersen et al. (2004) hypothesized that icebergs were pushed across the ridge from the Eurasian into Amerasian basin by currents due to intermediate Atlantic water influx. However, such a circulation pattern is not observed in the Arctic Ocean at present (Rudels et al., 1994), while glacial-time circulation is not understood and needs to be modelled. We infer that icebergs, as opposed to the ice shelf, may have rather moved towards the Fram Strait during disintegration of circum-Arctic ice sheets. This pattern is consistent with the orientation of scours on the Yermak Plateau (Vogt et al., 1994).

Pan-Arctic perspective

The widespread occurrence of conspicuous glaciogenic bedforms on ridges and plateaus in the interior of the Arctic Ocean indicates that very large masses of thick ice covered considerable portions or the entirety of this basin at several times during the Pleistocene. The exact mechanisms, timing and provenance of these ice masses are not yet well understood, but some inferences can be made. We find both the hypothesis of a pan-Arctic ice shelf as first proposed by (Mercer, 1970) and the opposite position that all deep-draft glacial erosion was generated by large ‘armadas’ of icebergs (Kristoffersen et al., 2004) difficult to reconcile with the observed patterns of seafloor erosion described in this paper (Figures 2 and 3). The diversity of mapped glaciogenic bedforms, their orderly distribution over large areas and association with large-scale seafloor erosion indicate that most of these features were formed by voluminous, cohesive ice masses grounded on the seafloor rather than by disparate icebergs. On the other hand, the distribution pattern of the bedforms is not consistent with an extensive kilometre-thick ice shelf covering the entire basin, but suggests a more complex erosional history involving ice shelves flowing from the ice sheet margins at the Arctic Ocean continental margin and the formation of ice rises on the impacted ridges. This pattern is especially explicit on the Chukchi Borderland, where the uneven depth distribution of erosional trimline and a complex superposition of various bedforms indicate a repetitive impact of ice shelves protruding from the northern margin of the Laurentide ice sheet combined with an ice rise formed over the Chukchi Plateau or the entire borderland (Figure 10). Based on cross-sectional area and the size of the associated trough-mouth fans, McClure Strait and Amundsen Gulf ice streams...
probably delivered most of the ice to ice shelves in the western Arctic. The three ice stream catchments that drained the Inuitian Ice Sheet northwards (Figure 10) had much smaller catchments than the McClure and Amundsen Gulf ice streams. The extent to which Mackenzie ice streams reached the marine environment is poorly understood. The lineation swarms on land cannot be reliably traced onto the shelf. As shown in Figure 10, there is direct landform evidence for large shifts in catchment location and size and the spatial arrangement of ice streams within the McClure Strait, Amundsen Gulf and Mackenzie valley ice stream corridors. The important implication is that ice discharge from the northern Laurentide Ice Sheet to the Arctic Ocean was highly variable in space and time, possibly on much shorter timescales than the major stadial/interstadial pacing of terrestrial ice volume variations. This shifting dynamics, where the entrance point of major ice influx in the western Arctic probably shifted drastically in location, may have had important consequences for ice shelf thickness distribution, and may be an explanation for the shifting movement pattern indicated by the mapped glacial morphology on the Chukchi plateau. The implications from such a scenario immediately raise the question of how an ice shelf extending into the Canada Basin and fed from the major ice streams of the western Canadian Arctic Archipelago could reach the Chukchi Borderland. We hypothesize that a
strong Beaufort Gyre could help by pressing sea ice and the ice shelf towards the Beaufort margin and, thus, the ice shelf would flow along the Alaskan continental margin as suggested by Engels et al. (2008). An ice shelf covering the southern part of the Canada Basin up to ca. 76° N would be of the order of $500 \times 10^3$ km$^2$, which is similar in size to the currently $487 \times 10^3$ km$^2$ Ross Ice Shelf of Antarctica. It should be noted that the initial grounding on the Chukchi Borderland may not necessarily have been from a coherent ice shelf extending all the way across the Canada Basin. Instead, large floating icebergs could have grounded on the Chukchi Plateau, which is partly shallower than 300 m and thus facilitated the growth of an ice cap.

The erosion mapped on the Lomonosov Ridge displays a simpler pattern, which is dictated by the narrowness of the ridge that was oriented nearly transverse to the impacting ice flow. The principal question arising from these results is how to get an Arctic Ocean ice shelf thick and large enough in order to ground on the central Lomonosov Ridge. Floating ice shelves fed by large ice sheets thin out with increasing distance from the grounding line due to the internal ice shelf strain rate (Dowdeswell and Bamber, 2007). The mechanism of calving involves the propagation of a fracture through the ice and, typically, icebergs calve from an ice shelf when the thickness has been reduced to around a couple of hundred metres (Dowdeswell and Bamber, 2007). In order to test the hypothesis that an ice shelf once grounded on the Lomonosov Ridge, a numerical ice sheet model was applied (Jakobsson et al., 2006). The experiments, albeit not including the complex dynamics of floating ice shelves, indicate that free-flowing ice emanating from the Barents–Kara shelf could not have been thick enough to ground on the Lomonosov Ridge if the ice strain rate is not reduced. However, if this ice shelf was supported or ‘buttressed’ by a counter-force such as a substantially thickened sea ice, it could become grounded on the ridge in the time frame of major glacial maxima (Jakobsson et al., 2006). A circum-Arctic modelling experiment shows similar preliminary results (Ritz et al., 2007). It should also be noted that surface mass balance and the basal melt/ freeze-on of Arctic ice shelves during glacial maxima are completely unknown, but would have been strong controls on their ability to reach for example the Lomonosov Ridge.

An accurate age constraint for the erosional events and their correlation between various parts of the Arctic Ocean is critical for incorporating these events into the overall paleoglaciological context for the Northern Hemisphere. The age of the Lomonosov Ridge erosion has been estimated as MIS 6 (Jakobsson et al., 2001) and on the Chukchi Borderland two erosional events have been dated to the LGM and MIS 4 or older (Polyak et al., 2007). Additionally, the new results from HOTRAX’05 core logs (Darby et al., 2005) reveal a third diamicton on the Northwind Ridge, which may be MIS 6 in age. The latter suggests the possibility of a synchronous expansion of ice shelves from the Eurasian (Barents–Kara) and North American ice sheets into the central Arctic Ocean during MIS 6, which was possibly the most extensive Pleistocene glaciation in the Arctic, at least in Eurasia (Svendsen et al., 2004). This synchronicity would make it possible for the ice shelves to coalesce somewhere in the Arctic Ocean and consequently expand over the entire basin. Limited evidence from other areas of the circum-Arctic margins provides some support for this scenario. Notably, recently re-investigated glaciotectonic features in pre-Quaternary sediments and in buried ice at the northern part of new Siberian Islands indicate the advance of an ice sheet from the north-east (Grosswald, 1980; Tumskoy and Basilyan, 2007). This pattern of ice advance might be related to the grounded edge of a thick floating ice shelf that extended to the periphery of the Arctic Ocean due to the interaction of major Eurasian and North-American ice shelves. Another piece of this pan-Arctic mosaic is located at the northern coasts of Greenland and Ellesmere Island. There is evidence of glacier erosion from ice moving eastwards along the Greenland coast (Funder and Hansen, 1996; Funder and Kjær, 2007), but the source of this ice is not understood. If this source turns out to be from the Canadian Archipelago west of the Nares Strait, this would indicate the likelihood of a thick ice shelf over the entire Canada Basin that deflected ice streams of the Inuitian Ice Sheet eastwards. This question should be elucidated by future mapping of the adjacent part of the Lomonosov Ridge, which shows evidence of deep-draft ice erosion on seismic profiles (Kristoffersen and Mikkelsen, 2006). Such mapping is planned for the 2007 LOMROG (Lomonosov Ridge off Greenland) Expedition (Jakobsson, 2007).

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