Marine Geology 258 (2009) 100–114

Contents lists available at ScienceDirect

Marine Geology

journal homepage: www.elsevier.com/locate/margeo

Morphology of the upper continental slope in the Bellingshausen and Amundsen Seas – Implications for sedimentary processes at the shelf edge of West Antarctica

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ARTICLE INFO

Article history:
Received 15 August 2008
Received in revised form 27 November 2008
Accepted 28 November 2008

Keywords:
West Antarctic continental margin
ice streams
cross-shelf troughs
subglacial meltwater
gullies
slides
debris flows
Last Glacial Maximum

ABSTRACT

Swath bathymetric and sub-bottom profiler data reveal a variety of submarine landforms such as gullies, slide scars, subtle shelf edge-parallel ridges and elongated depressions, and small debris flows along the continental shelf break and upper slope of West Antarctica. Gullies cutting through debris flow deposits on the Belgica Trough Mouth Fan (TMF) suggest formation after full-glacial deposition on the continental slope. The gullies were most likely eroded by sediment-laden subglacial meltwater flows released from underneath the ice margin grounded at the shelf edge at the onset of deglaciation. Scarcity of subglacial meltwater flow features on the outer shelf suggests that the meltwater reached the shelf edge mainly either through the topmost layer of soft diamict or in the form of dispersed flow beneath the ice, although locally preserved erosional channels indicate that more focused and higher-energy flows also existed. Concentration of gullies on the upper continental slope in front of the marginal areas of the major cross-shelf troughs, as contrasted to their axial parts, is indicative of higher-energy gully-eroding processes there, possibly due to additional subglacial meltwater flow from beneath the slow moving ice lying over the higher banks of the troughs. The shallow and sinuous gully heads observed on the outermost shelf within the Pine Island West Trough may indicate postglacial modification by near-bed currents resulting either from the subglacial meltwater flow from underneath the ice margin positioned at some distance landward from the shelf edge, or from the currents formed by brine rejection during sea ice formation. On the continental slope outside major troughs, slide scars as well as shelf-edge parallel ridges and elongated depressions indicate an unstable and failure-prone uppermost slope, although failures were probably mainly associated with rapid sediment loading during glacial periods. Complex, cauliflower- and amphitheatre-shaped gully heads biting back into the shelf edge suggest upslope retrogressive, multi-stage small-scale sliding as a contributing factor to the formation of gullies in these areas. Small debris fans immediately downslope of the slide scars suggest that small-scale debris flows have been the main downslope sediment transfer processes in the areas of weak or absent subglacial meltwater flow.

1. Introduction

Repeated advances and retreats of grounded ice sheets across high-latitude continental margins have resulted in the formation of thick, prograding sedimentary sequences (e.g. Larter and Cunningham 1993; Larter et al., 1997; Dowdeswell et al., 1998; Nitsche et al., 2000; Dowdeswell and Elverhøi, 2002). Fast-flowing ice streams within large cross-shelf troughs have been suggested as the main mechanism delivering glacial sediments to the shelf edge during glacial maxima, whereas less sediment is deposited between the ice streams, and during the interglacial periods (Dowdeswell and Siegert, 1999a; Nygård et al., 2002; Ó Cofaigh et al., 2003; Dowdeswell et al., 2004a, 2006). The flow speed of the ice and the rate at which mass is transferred from the ice sheets depends greatly on the hydrological and geological conditions at the ice-bed interface (Kamb, 2001; Anderson and Shipp, 2001; Canals et al., 2002; Fountain et al., 2005; Boulton and Zatsepin, 2006; Alley et al., 2006; Boulton et al., 2007a,b; Fricker et al., 2007; Vaughan and Arthern, 2007).

Geophysical and sedimentological investigations on the continental shelves of West Antarctica have established characteristic associations of elongate submarine landforms suggesting that large ice streams operated in cross-shelf bathymetric troughs during the Last Glacial Maximum, and that abundant meltwater was locally present beneath those ice streams (Wellner et al., 2001, 2006; Lowe and Anderson, 2002; Ó Cofaigh et al., 2002; Lowe and Anderson, 2003; Ó Cofaigh et al., 2003, 2005a,b; Evans et al., 2006). Although the

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doi:10.1016/j.margeo.2008.11.011

presence and role of meltwater beneath ice sheets has been a topic of debate for decades, the basal hydrological conditions and sliding mechanism of ice streams are presently not well understood even for the terrestrial ice sheets, let alone for those grounded well below the sea level (Bentley, 1998; Engelhardt and Kamb, 1998; Tulaczyk et al., 1998; Clark et al., 1999; Domack et al., 1999; Ó Cofaigh et al., 2002; Dowdeswell et al., 2004b; Ó Cofaigh et al., 2005b). However, a number of recent studies suggest that the presence of meltwater and a layer of soft sediment at the ice-bed interface are vital for the fast flow of ice streams (Bell et al., 1998; Anderson and Shipp, 2001; Hulbe, 2001; Conway et al., 2002; Alley et al., 2003; Vaughan et al., 2003; Peters et al., 2007; Bell et al., 2007). For example, subglacial processes inferred from till studies under Ice Stream B, West Antarctica, show very high subglacial meltwater pressures, including pore-water pressures, which are only a few kPa less than the ice overburden pressures. Also, lack of comminution in the soft till layer suggests sliding as the main ice flow mechanism facilitated by lubrication of subglacial meltwater (Tulaczyk et al., 1998).

Extensive subglacial channel systems on the inner shelf of West Antarctica bear traces of significant subglacial meltwater erosion (Anderson and Shipp, 2001; Lowe and Anderson, 2003; Domack et al., 2006). Evidence of channelized meltwater flow from the outer shelf is scarce, although locally sections of tunnel valleys preserved on the seafloor have been reported (King et al., 2004; Wellner et al., 2006).

Considerable variation in the shelf edge and upper slope morphology and sedimentary architecture reflects the complexity of glacial processes on Arctic and Antarctic high-latitude continental margins (Ó Cofaigh et al., 2003). Gully systems incise many formerly glaciated continental margins (Vanneste and Larter, 1995; Wellner et al., 2001; Ó Cofaigh et al., 2003; Elliott et al., 2006; Wellner et al., 2006; Laberg et al., 2007). Along the West Antarctic continental margin, they have been reported in Bellingshausen and Amundsen Seas (Ó Cofaigh et al., 2005b; Dowdeswell et al., 2006; Evans et al., 2006; Dowdeswell et al., 2008), Weddell Sea (Anderson, 1999; Heroy and Anderson, 2005), Ross Sea (Anderson, 1999; Shipp et al., 1999) and the Antarctic Peninsula margin (Ó Cofaigh et al., 2003; Dowdeswell et al., 2004a). In particular, the gullies are typical in front of all cross-shelf troughs except for Vega Trough (Ó Cofaigh et al., 2003; Heroy and Anderson, 2005; Wellner et al., 2006).

Despite scarce evidence, a majority of these studies suggest that the formation of upper slope gullies is related to erosion by sediment-laden subglacial meltwater sourced from beneath the ice margin grounded at the shelf edge (Wellner et al., 2001; Ó Cofaigh et al., 2003; Heroy and Anderson, 2005; Wellner et al., 2006). However, little is known about meltwater discharge at the ice margin around Antarctica (Jennings, 2006). Better understanding of the distribution patterns of various landforms at the shelf edge may shed light on their formation mechanisms and ultimately help understanding the processes associated with meltwater escape from ice sheets.

In this paper, we describe the morphology of the outermost shelf and upper slope in the Amundsen and Bellingshausen Seas, West Antarctica. We present a quantitative analysis of the morphological features observed, analyze their configuration and distribution patterns along the upper slope, and discuss the process implications for their generation.

2. Geological and glaciological setting

The continental margin of the Antarctic Peninsula and Bellingshausen Sea is a relict subduction zone, but the sectors examined in this paper have been tectonically inactive since c. 55 Ma in the vicinity of Belgica Trough and c. 17 Ma in the vicinity of Marguerite Trough (Larter and Barker, 1991; Larter et al., 2002; Eagles et al., 2004, In Press). The Amundsen Sea margin has been passive since its formation through rifting of Chatham Rise away from Antarctic at c. 90 Ma (Larter et al., 2002; Eagles et al., 2004).

Glacial sedimentation on these continental margins, driven by the alternate waxing and waning of the grounded West Antarctic Ice Sheet (WAIS) and Antarctic Peninsula Ice Sheet (APIS), commenced in the Middle Miocene (Barker and Camerlenghi, 2002; Cooper et al., In Press). The subsequent glaciations that continued throughout the Pliocene and Pleistocene deposited a complex sequence of prograding and aggrading sedimentary units on the continental margin of West Antarctica and the Antarctic Peninsula (Larter and Barker, 1989; Rebesco et al., 1997; Larter et al., 1997; Nitsche et al., 2000). Based on seismic reflection and bathymetric data, the continental slope of West Antarctica and the Antarctic Peninsula was formed by sediment release from broad line sources associated with ice grounded at the shelf break (Larter and Cunningham, 1993). The last ice sheets withdrew from the shelf edge along the Antarctic Peninsula c. 18,000–14,000 yr ago (Heroy and Anderson, 2007). The start of the last glacial recession in the Amundsen and Bellingshausen seas is less well constrained (e.g. Lowe and Anderson, 2002).

The continental shelf in the Amundsen and Bellingshausen Seas is c. 150–500 km wide and from c. 1600 to less than 200 m deep. Three major cross-shelf troughs extending to the shelf edge have been described on this part of West Antarctic shelf: Marguerite Trough, Belgica Trough and Pine Island West Trough (Fig. 1). Based on the extensive distribution of streamlined bedforms on the middle and

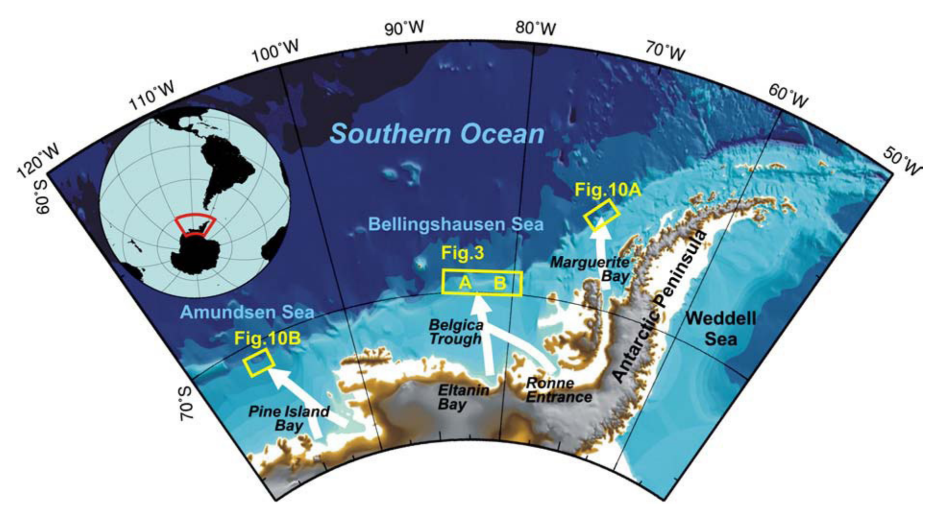

Fig. 1. Location map of the study areas along the continental margin of West Antarctica. Arrows mark the major cross-shelf trough ice streams that are inferred to have flowed across the continental shelf during the last glacial.

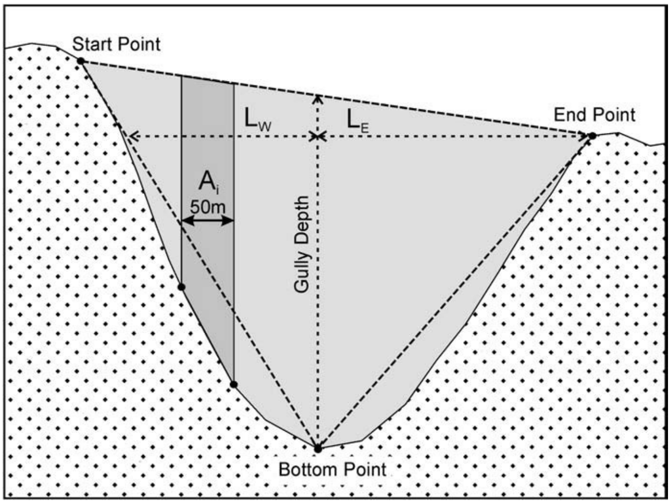

Fig. 2. The morphometric parameters defined in order to characterize the gullies. For the explanation, see text.

outer (offshore) parts of the troughs, all three are suggested to have accommodated large, fast-flowing ice streams, which drained a considerable part of WAIS and the southern APIS during glacial periods (Dowdeswell et al., 2004b; Ó Cofaigh et al., 2005a; Evans et al., 2006). Detailed geophysical investigations in these troughs have revealed a common succession of bedform-sediment associations (from inner to outer shelf) including: a) networks of interconnected channels and cavities eroded into the crystalline bedrock, b) a belt of drumlins at the transition from the crystalline to sedimentary substrate, c) series of elongate glacial lineations on the soft glacial sedimentary substrate that in places extends to the shelf break, d) an outer shelf heavily scoured by iceberg keels, and e) gully/channel systems incised into the glacial deposits of the outermost shelf and upper slope (Wellner et al., 2001; Canals et al., 2002; Lowe and Anderson, 2002, 2003; Ó Cofaigh et al., 2003; Dowdeswell et al., 2004b; Cofaigh et al., 2005b; Wellner et al., 2006; Evans et al., 2006; Dowdeswell et al., 2006). Locally, glacial–marginal grounding-zone wedges on the middle-outer troughs mark stillstands or readvances of the retreating ice-sheet margin during the last deglaciation (Ó Cofaigh et al., 2005a,b; Evans et al., 2006; Dowdeswell et al., 2008).

3. Data and methods

Swath bathymetric and sub-bottom profiler data used in this study were acquired during cruises JR59, JR71, JR84, JR104, JR141 and JR157 of the RRS James Clark Ross from 2001 to 2007. These data were collected using a hull mounted Kongsberg-Simrad EM-120 multibeam echosounder and a TOPAS parametric sub-bottom profiling system. The EM-120 system has working frequencies of 11.75–12.75 kHz and a theoretical maximum swath width of 150° (maximum used in practice was 136°) configured into 191 1°×1° beams. This resulted in horizontal resolution of seabed features in the order of a few meters, degrading somewhat with increasing water depth and towards the edges of swaths. Horizontal and vertical accuracy of the system was in the order of 5 m and 1 m, respectively.

The swath bathymetric data post-processing involved elimination of erroneous soundings, correcting data for vessel motion and sound velocity variations. Occasionally the data were digitally low-pass filtered in order to reduce large scatter due to very soft surficial sediments or rough sea conditions during recording. Post-processed bathymetric data were gridded with variable cell size depending on the level of detail of the seabed features imaged. In all cases, the gridding was done with 50 m by 50 m or finer cell size and, from the resulting isometric grids, cross-sectional profiles were extracted for the submarine landform analysis.

Cross-sectional profiles of the gullies were extracted parallel to the shelf edge where the gullies were best developed: mostly in water depth of 700–800 m, but locally, depending on the gully configuration, also in the 600–700 m depth range. Horizontal distance of 50 m or less between data points along the profiles was maintained in order to assure a close approximation to the actual shape of the gullies. On the cross-sectional profiles, gullies were identified and characterized using morphometric parameters such as width, depth, steepness and cross-sectional area. Width was defined as the distance between the start and end points of a gully (Fig. 2). Gully depth is a vertical distance from the lowest point along the gully profile (bottom point) up to the line defining gully width. L_W and L_E are horizontal distances from the line defining the gully depth to the gully margins to either side of the centre line using an idealized, triangular representation of a gully (Fig. 2). The ratio of gully depth over the sum of L_W+L_E is defined as gully steepness. Gully total cross-sectional area is calculated by integrating the areas of subsequent sections, A_i, defined by the data points maximally 50 m apart along the gully profile. The ratio of L_W/L_E was used to characterize gully asymmetry. The great majority of gullies had the symmetry index of 1 or very close to 1, and the average for all series of gullies analyzed deviated very little from 1. Therefore, we note here that the gullies were essentially symmetric and this parameter is not dealt further in the paper.

4. Results

The coverage of swath bathymetric data from the West Antarctic continental shelf edge is not uniform and the most thoroughly surveyed areas in the Bellingshausen and Amundsen Seas are the major cross-shelf troughs (Fig. 1).

This paper focuses on the morphology of the Bellingshausen Sea shelf break near the mouth of the Belgica Trough. A comparative morphometric analysis of the gullies in front of the Marguerite and the Pine Island West Troughs has also been undertaken in order to generalize the distribution pattern associated with large cross-shelf troughs. A detailed description of the configuration of the gully heads and slide scars incising the shelf edge is also given in order to understand the sediment dynamics at the shelf edge, and shed light on the role of various processes that shape it.

4.1. Shelf break near the Belgica Trough

4.1.1. Background

Belgica Trough is a c. 250 km long, up to 150 km wide and 500–1200 m deep elongate depression on the continental shelf of the Bellingshausen Sea (Fig. 1). Swath bathymetric mapping of streamlined bedforms on the Bellingshausen Sea shelf has identified two main outlet areas, Eltanin Bay and Ronne Entrance, through which West Antarctic and southern Antarctic Peninsula ice sheets, respectively, drained into the fast-flowing Belgica Ice Stream at the LGM. The drainage basin of this ice stream was estimated to exceed 200,000 km^3 (Ó Cofaigh et al., 2005a). The Belgica Trough crosses the shelf edge at c. 70.2°S 86°W. At the shelf edge the trough is c. 140 km wide with the bottom lying at c. 600–680 m water depth (Figs. 3A, 4). Beyond the rather steep banks of the trough the water depth decreases to 400–500 m. On the continental slope in front of the Belgica Trough, the slightly seaward bulging bathymetric contours outline a trough mouth fan (TMF) consisting of prograding deposits delivered to the shelf edge by the fast-flowing ice stream during glacial periods (Ó Cofaigh et al., 2005b; Dowdeswell et al., 2008).

The 415 km long strip of swath bathymetric data collected along the shelf edge and the upper slope at the mouth of the Belgica Trough and to the east of it revealed sets of gullies, slide scars and associated debris flows. In addition, a few subtle ridges sub-parallel to the shelf edge on the upper slope and an elongated depression at the shelf break were also observed (Figs. 1 and 3).

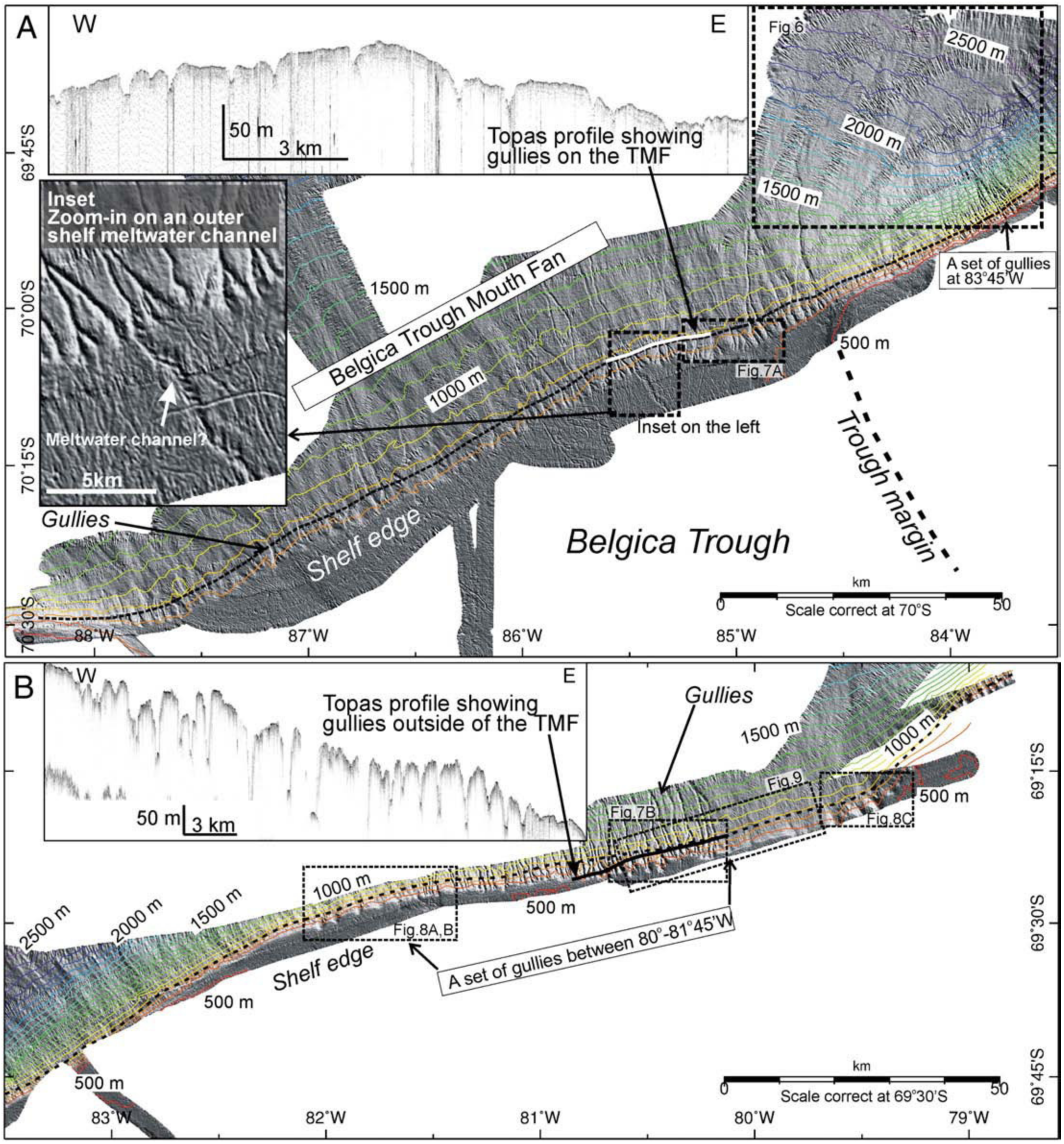

Fig. 3. Morphology of the Bellingshausen Sea shelf edge near the Belgica Trough Mouth (A) and to the east of it (B). TOPAS subbottom profiler records illustrate the dimensions and shape of the gullies in two contrasting upper slope settings: on the low gradient Trough Mouth Fan (A) and on a steeper section of the slope (B). Inset in (A) captures a remnant of a possible channel on the outermost shelf. Black dashed line along the upper slope marks the profile along which the gullies have been analysed. Variability of the morphometric parameters along this line is presented in Fig. 4. The 81 gullies presented in Fig. 5A were identified along this line west of the eastern margin of the Belgica Trough (A). Two other sets of gullies from which statistics are presented in Table 1 are indicated as well (A and B).

4.1.2. Gullies

In total, 232 gullies were identified on a cross-section extracted from the bathymetric grids along the upper slope in water depths between 700 and 800 m (Fig. 3). The distribution of gullies is not uniform and heavily gullied sections alternate with essentially gully-free areas. The gullies cluster mainly in four areas separated by relatively gully-free areas.

The largest set of gullies in the area incise the Belgica TMF, between 84°30′ and 88°W (Dowdeswell et al., 2008) (Fig. 3A). At c. 83°45′W a small cluster of 21, relatively deep and steep gullies cut the upper

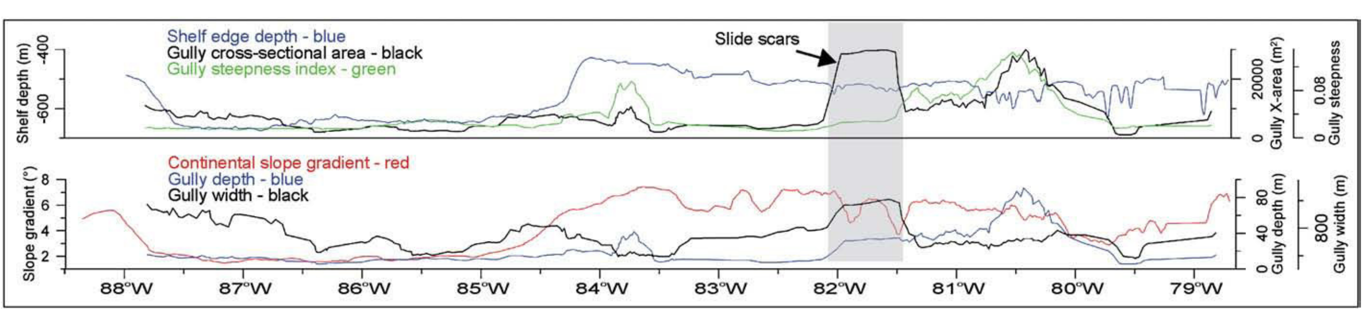

Fig. 4. Variability of morphometric parameters along the uppermost slope of the Bellingshausen Sea. Nine-point running average filter has been applied to the data series in order to reduce scatter. Location of the profile along which the morphometric parameters have been extracted is shown in Fig. 3.

Table 1
Average values of morphometric parameters for all 5 sets of gullies analyzed in this study

Average values	Pine Island West Trough	Belgica TMF	Cluster at 83°45′W	Cluster at 80–81°45′W	Marguerite Trough
Width (m)	631 (223)	744 (413)	419 (110)	678 (192)	891 (331)
Depth (m)	53 (44)	12 (6)	32 (19)	50 (31)	50 (32)
Area (m^2)	20223 (23330)	5666 (5204)	7505 (5698)	14842 (12682)	27854 (25965)
Steepness	0.0860 (0.0479)	0.0195 (0.0080)	0.0796 (0.0363)	0.0869 (0.0469)	0.0615 (0.0300)
Number of gullies	56	81	21	61	63

Standard deviations are given in parenthesis.

slope (Fig. 3A). Further east, after a gully-free interval of relatively steep slope between 81°30′ and 83°30′W, a set of 64 gullies spans from c. 80° to 81°30′W (Fig. 3B). Several rather large gullies appear in the easternmost part of the studied area, at c. 79°W. Unfortunately, the vessel's track bears away from the shelf edge in this area and the coverage of these gullies is incomplete. Therefore, the latter set of gullies will not be dealt further in this paper.

Morphometric parameters (width, depth, cross-sectional area and steepness in Fig. 2) of the gullies, together with the maximum magnitude of the gradient of the upper continental slope and the depth profile along shelf edge are shown in Fig. 4. Belgica Trough can be distinguished clearly on the shelf edge depth profile as well as on the gradient graph of the upper slope, between 84°30′W and 88°W (Figs. 3 and 4).

The lower values of 1–2° of the upper slope gradient between 84°30′W and 88°W result from the large sediment fan deposited in front of the Belgica Trough (Dowdeswell et al., 2008). To either side of the Belgica TMF, the upper continental slope is considerably steeper, reaching on average 5–6° immediately west, and 6–8° east of it (Figs. 3 and 4). The steep upper slope is less than 10–12 km wide and a transition to the significantly lower gradient lower continental slope occurs at about 2000 m water depth (Fig. 3).

The quantitative analysis of gullies along this transect has been carried out in 3 segments: The TMF area, a small cluster of gullies centered at 83°45′W, and the set of gullies between 80° and 81°30′W.

4.1.2.1. Belgica TMF between 84°30′ and 88°W. 81 gullies were identified on a cross-section in front of the Belgica Trough (Figs. 3 and 5A). The gullies commence at the shelf edge in water depth between c. 600 and 680 m (Figs. 3A and 7A). The gullies are from 2–3 to c. 10 km long and on average c. 12 m deep. They are on average 750 m wide and have cross-sectional area of 5700 m^2 (Table 1).

In their upstream parts, the gullies usually have distinct V-shaped cross-sections (Fig. 3A, inset). This shape changes gradually to a U- or box-like shape as the gullies coalesce to form channels further downslope. The number and size of the gullies on the Belgica TMF tend to increase from the trough centerline towards the margins (Fig. 5A). For example, the density of the gullies in the central part of the Belgica TMF is 0.48 gullies/km, whereas in the western and eastern parts of the TMF these values are 0.74 and 0.57 gullies/km, respectively (Table 2).

Although the number of gullies per unit length of the shelf edge increases somewhat from the axial part towards the margins of the

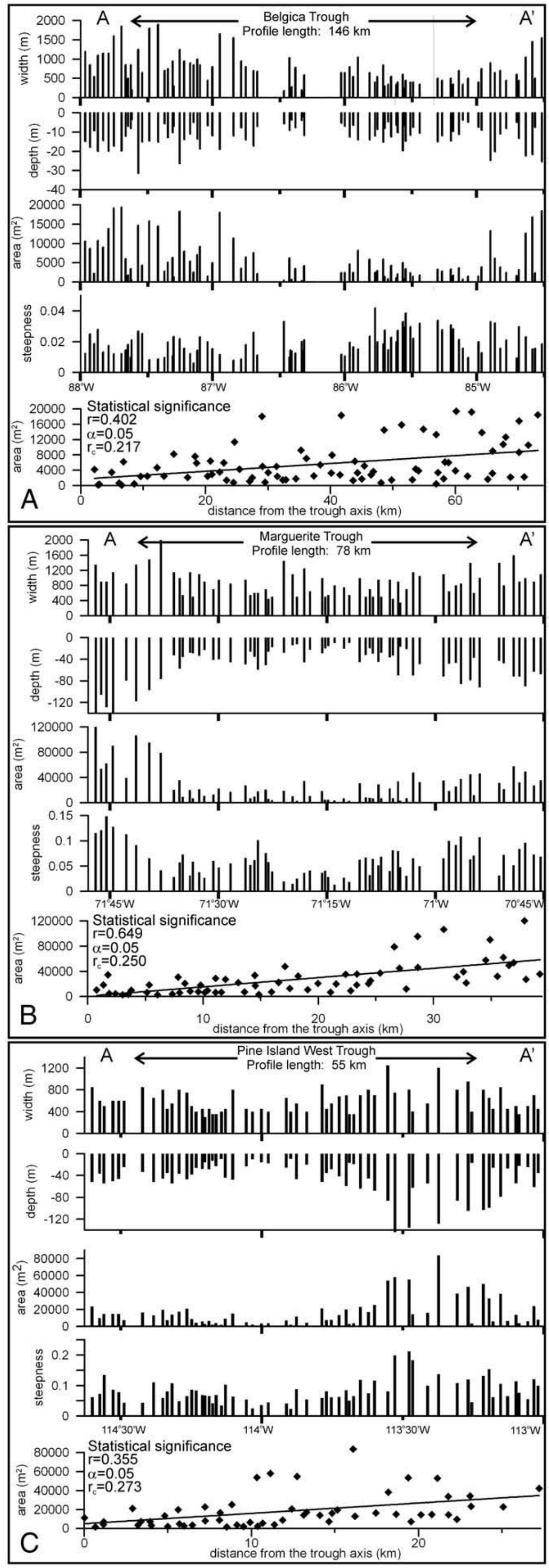

Fig. 5. Distribution of the morphometric parameters of gullies defined in Fig. 2 in front of the three major cross-shelf troughs: Belgica Trough (A), Marguerite Trough (B), and Pine Island West Trough (C). Lowermost graph in each panel presents the statistical significance test (Pearson's r-test) to demonstrate the validity of the increasing trend of gully cross-sectional areas towards the trough margins: α-level of significance, r – Pearson's correlation coefficient, r_c – Pearson's critical correlation coefficient. $r > r_c$ denotes statistically significant trend. Locations of the profiles along which the gullies were identified are shown with dashed lines in Figs. 3A, 10A and B, respectively.

Table 2
Distribution of gullies in front of the three major cross-shelf troughs: the Belgica Trough, the Marguerite Trough and the Pine Island West Trough

Gullies	Belgica Trough			Marguerite Trough			Pine Island West Trough		
	West	Centre	East	West	Centre	East	West	Centre	East
No/km	0.74	0.48	0.57	0.73	0.91	0.76	1.43	1.19	0.81
m^2/km	6289	1742	2563	33412	11720	23188	15550	13278	37770

Upper slope in front of each trough has been divided in three sections of equal length (West, Centre and East) and for each section the number of gullies (No) and the total gully cross-sectional area (m^2) per unit length (km) of the shelf edge is calculated.

Belgica TMF, the pattern is emphasized when considering the dimensions of the gullies. For instance, the average gully cross-sectional areas per unit length of the shelf edge of about 6300, 1750 and 2550 m^2/km for western, central and eastern parts of the TMF, respectively, clearly demonstrate the dominance of the gullied morphology near the margins of the Belgica TMF relative to its central part (Figs. 3A, 5A and Table 2).

Many of the gullies coalesce and form channels that can be traced for tens of km further downslope. The channels often form higher levees and locally the bottoms of these channels are elevated above the surface of the adjacent seabed on the middle and lower continental slope (Dowdeswell et al., 2008). Bottoms of the channels are characterized by higher backscatter values than the levees and adjacent continental slope areas (Fig. 6).

The gully heads within the Belgica Trough are gentle, in plan form nearly isometric incisions, giving the shelf edge a low sinuosity configuration (Fig. 7A). Locally, they cut back up to c. 500 m into the shelf edge. A c. 12 km long channel was recorded on the outermost shelf of the Belgica Trough, feeding directly into one of the gullies. The channel is 10–12 m deep and 200–400 m wide. In several locations the channel is cut by scour marks left by iceberg keels (Fig. 3A, inset).

4.1.2.2. Small cluster of gullies at 83°45′W. A small cluster of gullies, centered at c. 83°45′W, stretches c. 13 km along the shelf break (Fig. 3A). The gullies are on average deeper (c. 32 m), narrower (c. 419 m) and therefore have steeper sides (c. 0.08) here than on the Belgica TMF, although the average cross-sectional area (7505 m^2) only slightly exceeds that of the gullies on the TMF (Table 1). All gullies within the cluster begin at the shelf break and do not bite back into the shelf edge. They are rather straight with low convergence order, up to c. 10 km long and, unlike the gullies on the Belgica TMF, they do not form channels further downslope. They have a distinct V-shape in cross-section throughout their length, from the shelf edge to c. 1800 m water depth where they disappear (Fig. 3A). Average density of the gullies within the cluster is 1.70 gullies/km and the average cross-sectional area of the gullies per unit length of the shelf edge is 12,750 m^2/km.

4.1.2.3. Gullies between 80° and 81°45′W. A series of gullies is located c. 200 km east of the mouth of the Belgica Trough (Fig. 3B). The gullies are distributed along a 60 km stretch of the shelf edge centred at c. 80°30′W. The larger gullies incise the upper slope where the shelf edge is c. 30 m deeper than on the parts of the margin to either side (Fig. 4). Due to the limited swath coverage of the shelf in this area, we were not able to determine whether this depression extends further landward, featuring possibly the mouth of another cross-shelf trough. However, the presence of an outer shelf depression at that location and a small depocentre in front of it has been confirmed by *RV Polarstern* 1994 and 2001 cruises, suggesting that this may well be the case (Scheuer et al., 2006).

The gullies here are significantly deeper and larger as compared with the gullies incised on the Belgica TMF and within the small cluster at c. 83°45′W (Fig. 3B, inset, 4). The average depth and cross-sectional area of the gullies reach 50 m and c. 15,800 m^2, respectively (Table 1). Several gullies cut back to 2.5 km into the shelf edge (Fig. 7B). They also have highest steepness index of c. 0.09 (Table 1). The average density of the gullies is 0.94 gullies/km, which is almost twice as high as the density at the Belgica TMF, and the average cross-sectional area per unit length of the shelf edge is 14,850 m^2/km.

The gully heads biting back into the shelf edge are elongated, amphitheatre-like or complex (cauliflower-shaped) in plan form with diameters of c. 1.0–1.4 km. Complex gully heads feature

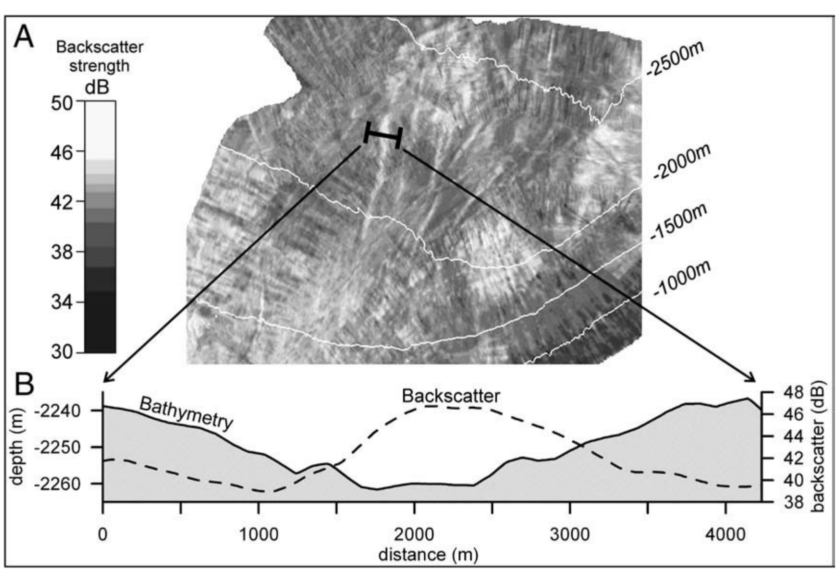

Fig. 6. Backscatter strength extracted from EM120 data (A) and a cross-section across one of the channels on the lower continental slope (B). Bottoms of the channels are characterized by higher backscatter values as compared to the levees. Location of the area is shown in Fig. 3.

secondary incisions within a larger primary head (Fig. 7B). Secondary incisions are usually shallow, up to a few tens of metres deep and bite back a few hundred metres into the side- and headwalls of the main incisions (Fig. 7B).

At 81°30′W a set of a few gullies is linked directly to a small depression in the surface of the outermost shelf (Fig. 8A,B). On the bottom of this depression, bedforms resembling a small network of braided meltwater channels were observed near the shelf edge (Fig. 8B). Finally, at 79°30′W, a single long and narrow gully was observed in the bottom of one of the slide scars for c. 7–8 km (Fig. 8C).

4.1.3. Slide scars and associated debris flows

At two locations, 79°30′W and 82°W, the shelf edge is indented by semicircular to slightly elongated, amphitheatre-like incisions with an approximate diameter of 1.2–2.0 km (Fig. 8). The bottoms of these depressions are smooth and their cross-sections are either box- or U-shaped. The head- and sidewalls of the incisions are c. 35–40 m, locally up to 90 m high, with gradients of up to 10–12°.

The slide scars at 82°W are somewhat smaller and with a simpler planform than those at 79°30′W. In the latter area, several sets of side- or headwalls could locally be distinguished within a single slide

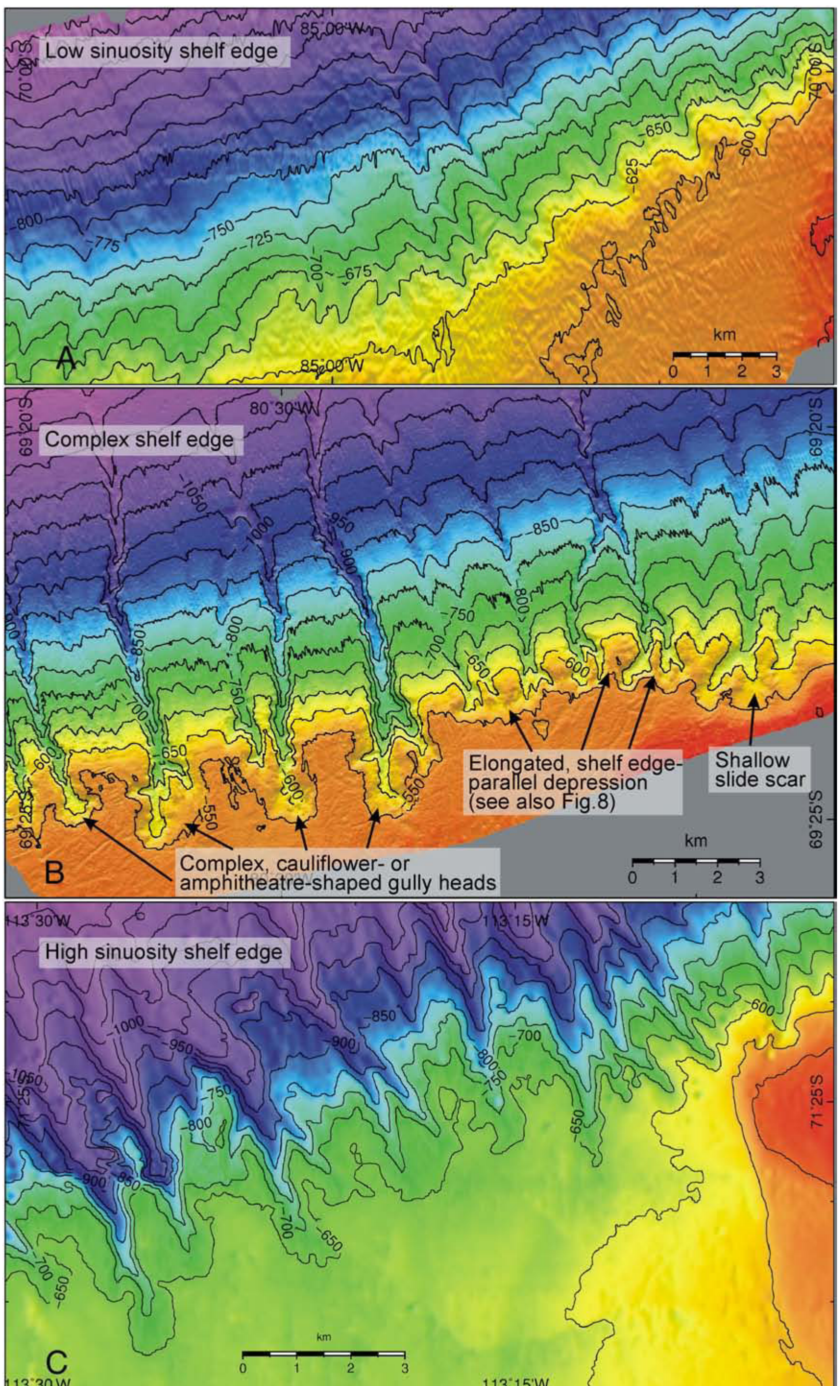

Fig. 7. Morphological features recorded in this study: (A) low sinuosity shelf break featuring gullies commencing at the shelf edge at the mouth of the Belgica Trough, (B) complex gully heads, shelf edge-parallel elongated depression and a shallow slide scar at the shelf edge outside of a major trough, (C) high sinuosity shelf break featuring gullies cutting deep into the shelf edge at the Pine Island West Trough mouth. Locations of the areas are shown in Figs. 3 and 10B.

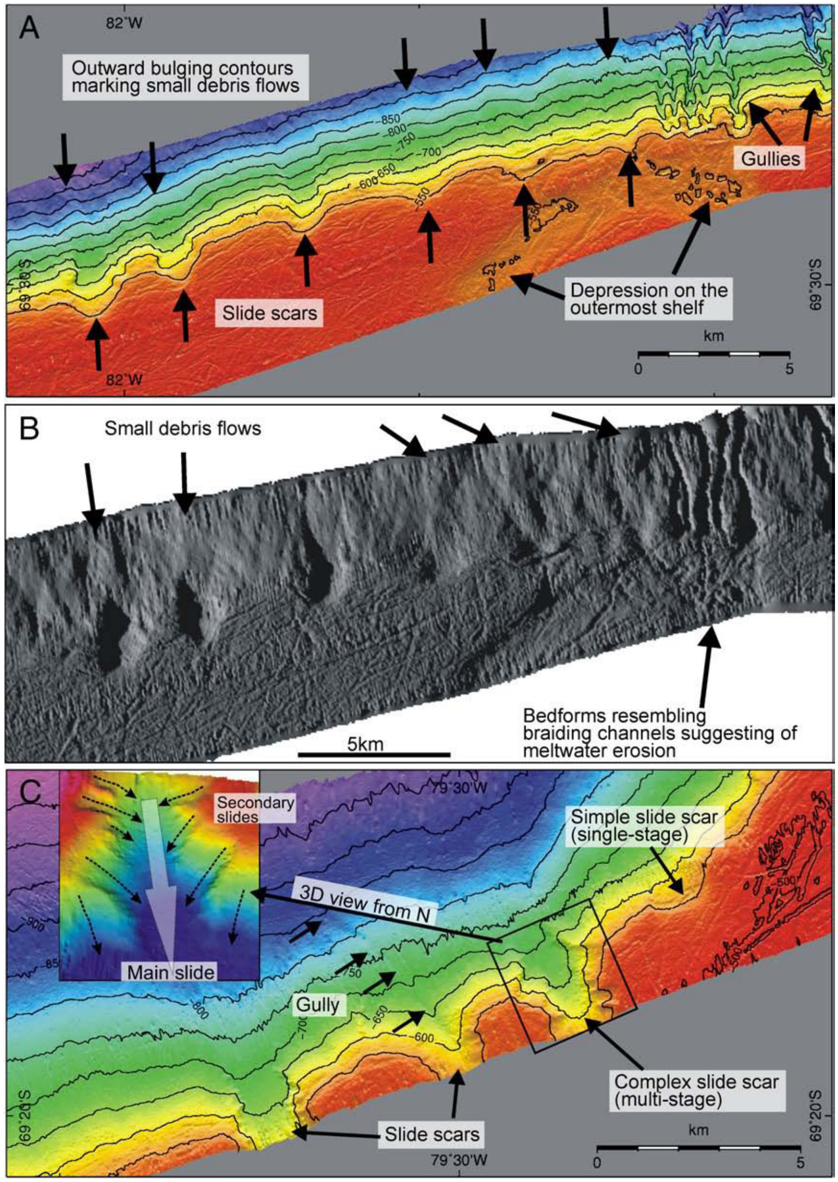

Fig. 8. (A) Series of simple slide scars and associated small debris flows on the upper continental slope of the Bellingshausen Sea. Note the rapid shift from slides to gullies and direct link between the gullies and the small depression on the outermost shelf. (B) Shaded monochrome bathymetry image of the same area as in (A). Note the bedforms resembling braided channels on the outermost shelf linked directly to the gullies, and the small debris flow fans just downslope of the slide scars. (C) A series of larger, complex slide scars. Configuration of the complex scars indicates multiple smaller, secondary slides along the sidewalls of the main slide scar (Inset). Note a simple slide scar in the easternmost part of the area. Locations of the areas in the Bellingshausen Sea are shown in Fig. 3.

scar implying that 2–3 retrogressive slope failure events have taken place there (Fig. 8 C). At 82°W, gentle seaward bulging isobaths immediately downslope of several slide scars mark small debris flow deposits (Fig. 8A, B). These debris flows originate most likely from the collapsed parts of the shelf edge. The measured run-out distance varied from c. 4–6 km. However, it should be noted that these are minimum run-out distances as we were not able to measure longer run-out distances due to the limited swath bathymetric coverage of the continental slope.

4.1.4. Shelf edge-parallel ridges and elongated depression

At c. 80°W, a series of 3–4 gentle, semi-continuous ridges runs sub-parallel to the shelf edge on the uppermost slope. These ridges have a wavelength of c. 200–400 m and height of a few metres, locally up to 5–6 m (Fig. 9). The rather subtle relief of these ridges on the background of the steep upper continental slope makes their imaging difficult. Nevertheless, a 3D perspective view of the shaded bathymetry image illustrates their morphology sufficiently well to visualize their semi-continuous distribution (Fig. 9A). Analysis of the back-scatter strength values extracted from the EM-120 data in that slope section did not reveal any trend or systematic variation associated with the ridge system (Fig. 9A).

At c. 80°20′W, a semi-continuous elongated depression runs along the shelf break, cutting through several inter-gully ridges and forming a joint headwall behind them (Figs. 7B and 9). This depression is c. 20–30 m deep and could be traced for up to 6–7 km. The cross-section of the depression is U-shaped with steep slopes locally up to 17°. The gullies either start in the bottom of the depression or cut a short distance into it. The best developed section of the depression at c. 80°20′W has been visualized in Fig. 9B. The bathymetric cross-sections extracted along three adjacent inter-gully ridges show that the depression is somewhat wider in its central part, as contrasted

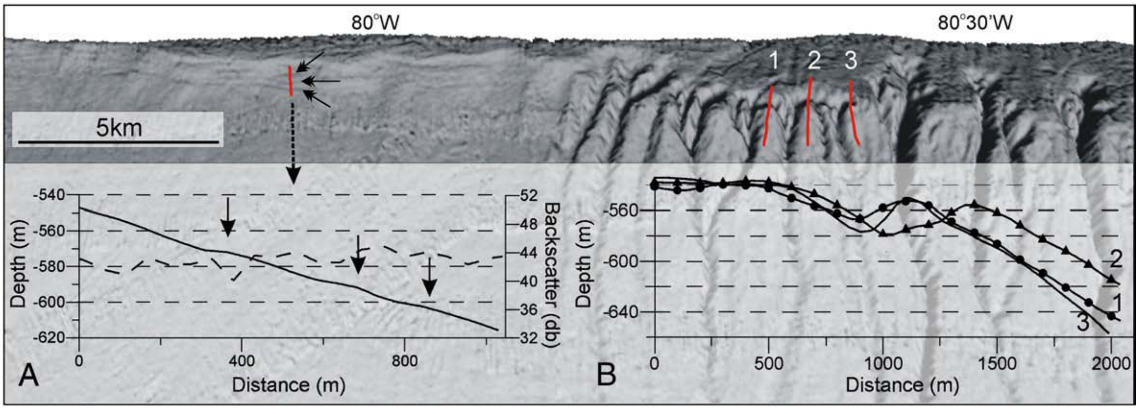

Fig. 9. 3D perspective view at the shelf break in the Bellingshausen Sea showing the gentle, semi-parallel ridges on the upper slope (A) and the elongated depression on the outermost shelf cutting through inter-gully ridges (B). The three arrows in (A) point to the three ridges distinguished along the red line. The dashed and solid lines in (A) signify the backscatter strength and the seabed depth, respectively yielding no correlation between the two. (B) Three bathymetric cross-sections along the crests of inter-gully ridges exemplify the shape and size of the elongated depression.

with its eastern and western parts. Towards the east the elongated depression becomes gradually less well developed until it disappears at c. 80°15′W.

4.2. Gullies at the mouths of the Marguerite and Pine Island West Troughs

4.2.1. The Marguerite Trough

4.2.1.1. Background. The Marguerite Trough is the largest depression on the shelf on the western side of the Antarctic Peninsula shelf, running from George VI Sound to the shelf break. It reaches the shelf edge at c. 66°25′ S 71°25′ W (Fig. 1). At the shelf break, the trough is c. 40 km wide at c. 500–600 m water depth, with relatively well-defined banks to either side (Fig. 10A). The water depth over the banks is 400–500 m. Mega-scale glacial lineations formed in a layer of deformation till several metres thick within the outer Marguerite Trough indicate that a fast-flowing ice stream draining APIS reached the shelf edge during the last glaciation (Dowdeswell et al.,2004b; Ó Cofaigh et al., 2005a). Although the continental slope is on average slightly gentler (9°) in front of the trough mouth as compared to that to either side of it (10–12°), there is no large TMF in this area (Ó Cofaigh et al., 2003; Dowdeswell et al., 2004a). Lack of a sediment fan in front of the mouth of the Marguerite Trough has been explained by the steepness of the continental slope, which probably resulted in rapid down-slope sediment transport and sediment bypassing of the upper slope (Dowdeswell et al., 2004a).

The morphology of the upper continental slope in front of Marguerite Trough differs from that at the Belgica and the Pine Island West Troughs. The steepest interval of the continental slope in front of the Marguerite Trough is located 5–8 km seaward of the shelf break whereas at the Belgica Trough and the Pine Island West Trough the slope is steepest immediately seaward of the shelf break. This morphology is the result of progradation of the upper slope interpreted from the presence of foreset seismic reflectors (Bart and Anderson, 1995; Larter et al., 1997; Nitsche et al., 1997, 2000). This slope configuration, together with the distribution of the gullies, which are better developed on the steeper parts of the slope, has led to suggestions that the formation of gullies may be related to either the slope gradient (Dowdeswell et al., 2004a), or the sediment flux, which

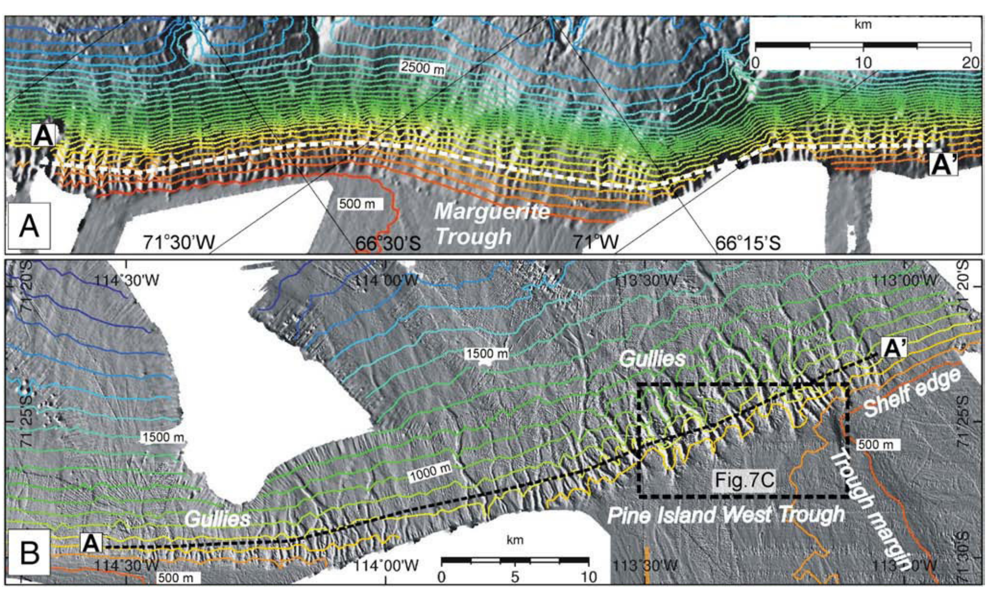

Fig. 10. Gullies in front of the Marguerite Trough (A) and the Pine Island West Trough (B). The dashed lines mark profiles along which the morphometric parameters of the gullies were extracted. The spatial variability of these parameters is presented in Fig. 5.

was higher in the axial part of the trough than near its margins (Ó Cofaigh et al., 2003).

4.2.1.2. Gullies. Seaward of the trough mouth, a set of 63 gullies is incised into the upper slope (Fig. 10A). On average, the gullies are about 900 m wide and 50 m deep, although their dimensions vary considerably (Table 1). Length of the gullies is usually a few km, rarely exceeding 10 km.

The gullies extend to either side of the trough mouth along the shelf edge (Fig. 10A). The gradient of the uppermost continental slope is 5–6° in front of the central part of the trough, and 14°–15° (locally up to 20°) to either side of it. The continental slope has a higher gradient section of up to 20–25° in front of the central part of the trough in water depths of c. 1500–2000 m. The upper slope between the shelf edge and 1500–2000 m water depth in the central part of the trough mouth is typically rather smooth and devoid of gullies (Fig. 10A). Also characteristic of the Marguerite Trough mouth is that, although numerous gullies incise the upper slope in front of the trough, few cut back into the shelf (Fig. 10A). Like the gullies on the upper slope of the Belgica Trough, the gullies have V-shaped cross-sections and tend to cut through the topmost layer of diamict that has been interpreted as debris flow deposits originating from sediments transported to and released at the shelf edge by a major cross-shelf ice stream (Dowdeswell et al., 2004a).

Plotting the measured morphometric parameters of the gullies against their location on the profile along the shelf edge reveals that, similarly to the Belgica TMF, the dimensions of the gullies tend to increase from the centre of the trough towards the margins (Fig. 5B). This trend is most pronounced for the depths and cross-sectional areas of the gullies and somewhat less pronounced for their widths and steepness indices. However, the number of gullies per unit length of the shelf edge does not follow the same sideways increasing trend as was observed in front of the Belgica Trough (Table 2).

4.2.2. The Pine Island West Trough, Amundsen Sea

4.2.2.1. Background. The shelf is c. 500 km wide from the locally more than 1500 m deep and rugged inner Pine Island Bay to the outer shelf with much shallower (400–500 m) and smoother sea floor (Lowe and Anderson, 2002; Evans et al., 2006). The compilation of bathymetric and seismic data from the Amundsen Sea implies that more than one cross-shelf glacial trough may be present there (Nitsche et al., 2000). The one described in this paper is located at the continental margin slightly west of the Pine Island Bay where major glaciers feeding the cross-shelf ice-stream drained (Fig. 1). Therefore, we called this feature the Pine Island West Trough. The Pine Island West Trough is c. 50 km wide at c. 113°45′W 71°25′S, where it reaches the shelf edge (Fig. 1). It accommodated an extensive fast-flowing ice stream during the Last Glacial Maximum (LGM) that drained two major West Antarctic outlet glaciers – the Pine Island Glacier and the Thwaites Glacier (Evans et al., 2006). Presently, these two glaciers together have been estimated to drain over 20% of the area of WAIS (Vaughan et al., 1999) and account for approximately 4% of the total discharge of the ice from the entire Antarctic Ice Sheet (Vaughan et al., 2001).

4.2.2.2. Gullies. Similarly to the Belgica and the Marguerite Troughs, the upper continental slope in front of the Pine Island West Trough is characterised by numerous gullies (Dowdeswell et al., 2006) (Fig.10B). The gullies are on average c. 630 m wide and slightly over 50 m deep with the average cross-sectional area of c. 20,200 m^2 (Table 1). They are typically 5–10 km long and relatively straight (Fig. 10B). A few of them translate directly to channels on the mid-slope (Dowdeswell et al., 2006). The gullies cut into the shelf edge in several places. These incisions are typically longer than anywhere else in the studied areas, exhibit slight sinuosity and reach locally over 3 km in length (Fig. 7C). Due to the deep cutbacks into the shelf, the shelf edge has an extremely jagged configuration in planform.

Distribution of the gullies along the shelf edge is not uniform. As on to the Belgica and Marguerite Troughs, larger gullies tend to be located near the margins of the trough mouth, whereas they are smaller in front of the central part of the trough (Fig. 5C). From Table 2 it appears that the gullies are somewhat larger near the eastern margin of the trough as compared to those at the western margin (Fig. 10B, Table 2). However, the number of gullies per unit length of the shelf edge tends to decrease from west to east (Table 2).

5. Discussion

The morphological features distinguished at the shelf edge and upper slope include:

- Gullies,
- Small-scale slide scars and associated debris flows,
- Sets of gentle ridges running sub-parallel to the shelf edge on a steep section of the uppermost continental slope,
- A local shelf edge-parallel, elongated depression cutting locally through several adjoining inter-gully ridges on the uppermost slope.

In this section these features have been interpreted in terms of the past sedimentary environments and processes at the ice-sheet margin.

5.1. Gullies

Three of the five segments of the shelf edge where gullies were recorded are located in front of the large cross-shelf troughs on the Bellingshausen and Amundsen Sea margins. Only one of the troughs, the Belgica Trough, has an extensive sedimentary trough-mouth fan in front of it (Dowdeswell et al., 2008) (Figs. 3 and 10). This suggests that the formation of gullies is independent of the existence of large TMF. However, all major cross-shelf troughs described in this paper are associated with sets of gullies at the shelf edge, suggesting a possible genetic link between them.

Gullies on the Belgica TMF are distinguished from gullies elsewhere along the margin of West Antarctica by their distinctly lower steepness index and considerably higher degree of down-slope coalescence (Fig. 3, Table 1). Table 1 also shows that the Belgica TMF has the shallowest average gully depth, and a cross-sectional area 3–4 times smaller than the gullies in front of the other two major cross-shelf troughs. This has been attributed to the considerably lower slope gradient of the TMF as compared to the other four areas, where such fans are lacking (Figs. 3 and 10) (Dowdeswell et al., 2004a, 2008).

Exceptions from the positive correlation between the gully depth and slope gradient (e.g. compare the gullies at 80°30′W and 83°45′W) suggest that other factors, besides the slope gradient, play an important role in their development (Figs. 3 and 4). These factors probably include the size of the meltwater drainage system, and the sediment yield as well as the duration of the gully-forming processes, among others. For instance, aside from the gullies on the TMF, the gullies tend to be larger within larger clusters, i.e. with a greater number of gullies (Table 1). This correlation suggests that larger gully systems, in terms of size as well as number of gullies, may reflect larger meltwater sources at the ice sheet margin. For example, the cluster of gullies at 83°45′W is the smallest of all in terms of the number of gullies, as well as their size, and may reflect the lesser amount of subglacial meltwater issuing from underneath the ice margin or merely the more temporary character of the gully formation processes there.

The V-shaped, nearly symmetrical cross-sections and relatively high average steepness indices of gullies, particularly on the steeper slopes outside the TMF, imply high-energy gravity-driven erosion processes. These processes have been envisaged mainly in the form of sediment-laden meltwater flows sourced from beneath the ice margin grounded at the shelf edge (Table 1, Fig. 3) (Anderson et al., 2001; Ó Cofaigh et al., 2005b; Dowdeswell et al., 2006; Wellner et al., 2006; Dowdeswell et al., 2008).

The time of formation of gullies is somewhat uncertain. Based on the evidence from the Belgica TMF, where the gullies cut through debris flows on the continental slope, Dowdeswell et al. (2008) pointed out that if the debris flows originated from downslope transport of diamictic debris during full-glacial conditions, then the gullies must have formed during the deglaciation or in post-glacial time (Fig. 3). A single gully incising the slide scar at 79°30′W also supports this interpretation (Fig. 8C). Hence, it is suggested that the gravity-driven debris-flow deposition that prevailed at full-glacials was replaced by subglacial meltwater-driven turbidity flow sedimentation, which in turn ceased as the ice margin withdrew from the shelf edge in the course of deglaciation. Considering the above, the main erosion of gullies at the shelf edge was probably associated with the meltwater discharge from underneath the ice margin grounded at the shelf edge, although other processes, such as tidally induced flows or bottom currents may have contributed to their evolution as well.

5.1.1. Subglacial meltwater discharge

Many recent studies suggest that water is not only present under ice sheets, but it is also highly mobile (Engelhardt et al., 1990; Hulbe, 2001; Clarke, 2006; Wingham et al., 2006; Fricker et al., 2007). Subglacial lakes relatively close to the present ice margin, and favorable morphological settings on the shelf with linked drainage systems where such water bodies could have existed under the ice during the late glacial or even at LGM, have been described from several locations in Antarctica (Rebesco et al., 1998; Dowdeswell and Siegert, 1999b, 2002; Lowe and Anderson, 2003; Ó Cofaigh et al., 2005a; Siegert et al., 2005; Domack et al., 2006; Wellner et al., 2006; Bell et al., 2007). Although the inner shelf morphology of the Amundsen and Bellingshausen Seas is poorly known, the data collected from the large cross-shelf troughs locally show well-developed relict subglacial meltwater channel systems on the inner- to mid-shelf (Anderson and Shipp, 2001; Anderson et al., 2001; Lowe and Anderson, 2002; Ó Cofaigh et al., 2002; Lowe and Anderson, 2003; Ó Cofaigh et al., 2005a). Similar linked and isolated cavity systems, interpreted as having held subglacial water bodies during the last glacial period, have been described from the Palmer Deep on the Antarctic Peninsula margin (Domack et al., 2006), and in Marguerite Bay (Anderson and Oakes-Fretwell, 2008).

The mechanism of subglacial meltwater flow to the margins of large ice sheets is not fully understood, however (Raymond, 2002). The possible remnant of a meltwater channel at the mouth of the Belgica Trough and the bedforms resembling small braided channels within the small depression on the outermost shelf at c. 81°30′W suggest that subglacial meltwater drainage took place at these locations (Figs. 3A and 8A,B). A similar feature, a tunnel valley eroded into glacial substrate, has been reported from the mid-shelf of the western Ross Sea (Wellner et al., 2006) and small channels have also been reported from beneath the Rutford Ice Stream, western Weddell Sea (King et al., 2004).

However, the scarcity of channels preserved on the outer shelves of Antarctica suggests that subglacial meltwater escapes mainly either through the topmost soft sediment layer (Tulaczyk et al., 1998), or through a distributed (sheet) flow system (Boulton et al., 2007a,b). The subglacial meltwater flow had to adjust to the local bed and pressure conditions and, in situations where meltwater production exceeds the flux that can be discharged through the basal sediment layer, a channelized flow regime may develop temporarily. Although a stable channelized meltwater system is unlikely to form in soft substrate, temporary formation of channels to accommodate excess meltwater may occur (Boulton and Hindmarsh, 1987; Alley et al., 1989; Alley, 1993; Clark and Walder, 1994; Tulaczyk et al., 1998). Seasonal reorganization of flow, from distributed systems in winter to channelized flow in summer, is known from land-based glaciers (Hubbard et al., 1995), and the remnants of channels and tunnel valleys suggest that locally higher-energy meltwater streams capable of eroding soft glacial sediments also existed on the mid- to outer shelf within the large cross-shelf troughs (Fig. 3; Lowe and Anderson, 2003; King et al., 2004; Wellner et al., 2006). Moreover, recent studies on the continental margin of the Orphan Basin have provided evidence to support the link between the glacial debris flow deposition and the glacial meltwater plumes (Tripsanas and Piper, 2008, in press).

The common spatial distribution pattern of the gullies in front of the major cross-shelf troughs may result, at least partly, from the spatial variations in meltwater delivery to the shelf edge and, as such, indicate the subglacial meltwater flow distribution on the outer shelf. As shown in Fig. 5, the depth and total cross-sectional area of the gullies per unit length of the shelf edge increases towards the trough margins in front of all three major cross-shelf troughs (Figs. 3, 5 and 10, Table 2). We chose the total cross-sectional area of gullies per unit length of the shelf edge as the most informative parameter for illustrating the spatial distribution of gullies. It was preferred to other parameters because it combines the number of gullies as well as their size (degree of development), and as such probably best reflects the variability of their generating processes.

A mechanism that could explain the formation of the observed spatial distribution pattern of the gullies may involve the input of additional subglacial meltwater to the marginal areas of the large cross-shelf troughs from the adjacent banks. Fast-flowing ice-streams, irrespective of their mode of formation, draw down water pressure at the ice-bed interface (Röthlisberger, 1972). Due to the difference in subglacial pressure gradient, the subglacial meltwater under the slow-moving ice on the banks tends to flow into the adjacent trough where the pressure gradient is steeper. Therefore, the troughs accommodating ice streams probably act as sinks for subglacial groundwater flow (Boulton et al., 2007b). In the trough, the subglacial meltwater may be incorporated into the till (Wellner et al., 2006), as suggested by the general absence of channels on the floors of the troughs, and then flow through the till. Being driven by the subglacial pressure gradient, it would move towards the ice-sheet margin where it would drain via the gullies. Other factors, such as the additional heat generated due to friction in the marginal shear zones of ice streams (Raymond, 2002), or the subglacial pressure gradient due to the thicker ice in the ice streams that would drive the subglacial meltwater towards the ice stream margins (Dowdeswell and Elverhøi, 2002), may partly be responsible for increased meltwater concentration there as well.

5.1.2. Tidally induced flows

During the late- to postglacial period, the gullies may have been affected by tidal processes. For instance, during deglaciation when ice was grounded on the shallower banks but floating over the trough mouth, strong tidal pumping into and out of the sub-ice cavities could have resuspended sediment and initiated turbidity flows that were strong enough to erode gullies on the continental slope.

5.1.3. Bottom currents

Bottom currents, either sourced from underneath the retreating deglacial ice margin located at some distance landward from the shelf edge or generated by brine rejection during the formation of sea ice could potentially erode gullies. The latter has been suggested as a possible gully generating mechanism during interglacials in the Bellingshausen and Amundsen Seas (Dowdeswell et al., 2006, 2008), in the Bear Island Trough (Vorren et al., 1989; Laberg and Vorren, 1996), and on the western slope of the Crary Fan in the Weddell Sea (Kuvaas and Kristoffersen, 1991). This is also supported by the observation that gullies are typically more abundant in the areas where local depressions have been recorded on the outermost shelf. These bathymetric low points are exactly where such dense water would spill off the shelf if this process occurred. This association is most distinct in front of the major cross-shelf troughs (Figs. 3 and 10), but also occurs on a smaller scale. For instance, at c. 81°30′W gullies are directly linked to a depression few tens of metres deep on the surface of outermost shelf, whereas the slide scars incise the adjacent, somewhat higher shelf edge section within a couple of km of the gullies (Fig. 8A,B). At 113°W the shelf break is deepest on the eastern side of the trough mouth, and again it is interesting to note that this is where the most developed gullies occur (Fig. 10B).

There are no oceanographic observations that point to denser water spilling off the shelf at present in Bellingshausen and Amundsen Seas, although evidence of colder bottom water flowing off the shelf east of the AP, in the Weddell Sea (Patterson and Sievers, 1980; Whitworth et al., 1994) and north of the AP, off the Elephant Island (Meredith et al., 2003) is known. It is worth noting, however that virtually all oceanographic data are from the summer months only, and even if this process is not happening at the present day, it might have happened during deglaciation or earlier in the Holocene.

The higher backscatter values from the bottoms of the channels on the Belgica TMF result most likely from coarser sediment there (Dowdeswell et al., 2008) (Fig. 6). This implies that these gully-channel systems have recently accommodated or are even presently accommodating relatively high-energy currents. Most likely source for such currents would be intermittently downslope sweeping turbidity flows triggered either by small-scale sediment failures on the upper slope or higher salinity bottom water generated by brine rejection during sea ice formation spilling off the shelf. However, it is worth noting that the EM120 uses a 12 kHz signal, and therefore a part of it will penetrate up to a few m into a top layer of very soft sediments. Considering this effect, the backscatter images we see may actually represent the seabed conditions a few thousand years ago rather than those at present.

The effect of reverse (upwelling) currents on the shelf edge morphology cannot be ruled out at this time either. For example, the warmest water in the Amundsen Sea has been shown to be at 400–600 m depth (Bindschadler, 2006; Walker et al., 2007). This warm Circumpolar Deep Water has been shown to intrude onto the shelf along cross-shelf troughs and reach the base of the floating extensions of Pine Island Glacier, Thwaites Glacier and ice shelves in the Amundsen Sea embayment (Jacobs et al., 1996; Anderson, 2007; Walker et al., 2007). However, the effect of these upwelling currents on the shelf edge is poorly known. Considering the envisaged near-bottom outflow of higher salinity water, particularly in the cross-shelf troughs (Dowdeswell et al., 2006; Laberg and Vorren, 1996), the geomorphological impact of the upwelling warmer water on the shelf edge morphology is probably minor, although its effect cannot be discounted at this time. Also, it should be noted that the modern oceanographic regime on the Pacific margin may have only been established during the last few thousand years. If CDW had been coming onto the shelf in the present quantity throughout the Holocene, West Antarctic deglaciation would probably have proceeded much further than it has so far.

5.1.4. Small-scale slope failure induced flows

It is thought that the small-scale sediment failures on the upper slope can transform to (turbid) sediment flows further downslope and could be a possible gully-generation mechanism (Larter and Cunningham, 1993; Vanneste and Larter, 1995). Indeed, the considerable dimensions and elongate, complex planform shape of the gully heads outside the major troughs in the Bellingshausen Sea, (e.g. at 80°30′W), suggests that upslope retrogressive slope failures as well as meltwater erosion have shaped the gullies here (Fig. 7B). Gully heads at 80°30′W are reminiscent of small “embryo” canyons (Figs. 3 and 7B). The locally cauliflower-shaped complex gully heads probably mark sections where small-scale secondary slope failures occurred within the primary slide scar (Mulder and Cochonat, 1996).

5.1.5. Implications of gully head configurations

Another aspect of gully morphology that may shed light on the processes involved in their generation is their head configurations and the variable degree they cut back into the shelf edge. Short cut-back distances most likely infer closeness of the source of the bottom currents generating gullies. For instance, in front of the Belgica Trough, the fact that gullies commence right at the shelf break or bite back into the shelf a very short distance implies they were formed mainly by sediment-laden meltwater (turbidity) flows releasing from beneath the ice margin grounded at the shelf edge (Figs. 3, 7A) (Dowdeswell et al., 2008).

The elongated, slightly sinuous gully heads biting locally back to the shelf for more than 3 km at the mouth of the Pine Island West Trough, on the other hand, may exhibit modification by near-bed currents flowing across the shelf break. Such currents may originate from the glacial meltwater source beneath an ice margin that has withdrawn somewhat from the shelf edge or by brine rejection in the process of sea-ice formation. Although these flows have not been quantified, they have been suggested to be capable of shaping the shelf and slope morphology (Vorren et al., 1989; Kuvaas and Kristoffersen, 1991; Laberg and Vorren, 1996; Dowdeswell et al., 2006, 2008). Even though the bottom flow of higher density water may generate gullies when accelerating down the upper continental slope, the common spatial distribution pattern of increasing size and concentration of the gullies from the center-line toward the margins of the major glacial cross-shelf troughs at the shelf edge is difficult to explain by such near-bed currents alone.

5.2. Slide scars and associated debris flows

Sets of slide scars over two 10–20 km long intervals at c. 82°W and 79°30′W indicate potentially unstable and failure-prone segments of the shelf edge that have collapsed at some point in the past (Figs. 3 and 8). The preconditions for the formation of an unstable shelf edge result from rapid deposition of debris at the shelf edge by the ice sheet during glacial maxima when the ice covered the continental shelves (Norem et al., 1990; Larter and Cunningham, 1993; Laberg and Vorren, 1996; Ó Cofaigh et al., 2005b). High sedimentation rates at the ice margin resulted in thick poorly consolidated sediment successions and depositional oversteepening – a configuration that is inherently unstable. Previous subbottom profiler and multi-channel seismic studies have distinguished gullies and slide scars, as well as debris flow deposits, on the continental slope in the Bellingshausen and Amundsen Seas (Vanneste and Larter, 1995; Nitsche et al., 1997, 2000, 2007). The exact mechanism triggering the debris flows is not known, however, and may vary for different locations. Debris flows are also known from other continental margins with considerably lower slope gradient, but with dimensions usually 1–2 orders of magnitude larger than the debris flows recorded in this study (Aksu, 1984; Aksu and Hiscott, 1989; Laberg and Vorren, 1995, 1996; Vorren et al., 1998; Laberg et al., 2000; Canals et al., 2004; Hüchnerbach and Masson, 2004).

The small debris flows observed immediately downslope of the slide scars at c. 82°W indicate that the run-out distance of the displaced sediment was 4–5 km, although the slope gradient there is relatively high, up to 5–6° (Fig. 8). A statistical study of landslides from the northern Atlantic confirms this, suggesting that longer slides tend to consist of softer and more fluid material as contrasted by shorter ones that are characterized by stiffer sediment (Taylor et al., 2002; Hüchnerbach and Masson, 2004). Hence, the short run-out distance of the debris flows at c. 82°W could well imply a lower water content of the flows as compared to the adjacent shelf edge areas where gullies dominate (Fig. 8A,B). Therefore, we speculate that the main difference between the two contrasting shelf edge configurations, i.e. debris flow vs. gully dominated, may have been the greater input of meltwater from a subglacial source in the latter areas. A high flux of subglacial meltwater escaping from underneath the ice margin may have contributed to the formation of the turbidity currents right at the shelf break which transported glacial debris downslope, eroding gullies in the process. A weaker, less sediment-charged subglacial meltwater source in the adjacent areas resulted in the dumping of the glacial sediments at the shelf break, creating a potentially failure-prone shelf edge configuration.

Although the timing of the slope failures is uncertain, their small size may imply an early stage slope failure in a situation when high LGM sedimentation rates at the shelf edge left little time for sediment compaction and no distinct extensive failure planes had developed within the sedimentary sequence. For instance, based on the inferred high sedimentation rates during glacials, the somewhat larger debris flows recorded on the Belgica TMF have been attributed to the LGM (Dowdeswell et al., 2008). Furthermore, a small gully in the bottom of one of the slide scars at c. 79°30′W indicates that the slope failure took place before the gully incision there, confirming that the small-scale slope failures probably occurred synchronously with the rapid glacial sedimentation at the shelf edge during glacials (Fig. 7C). However, the complex morphology of some of the slide scars, indicative of their multi-stage development, suggests the time-transgressive nature of the slope failure processes that locally may be active to the present day (Fig. 8C).

5.3. Shelf edge-parallel ridges and elongated depression

The gentle ridges in the upper part of the continental slope at c. 80°W could result from a number of slope processes, including successive overlapped slumps, slope failure along the listric faults, or creep (Fig. 9) (Mulder and Cochonat, 1996). All of these processes are likely to have similar morphological expression on the seabed and distinction between them merely on the bases of seafloor morphology is not possible (Mulder and Cochonat, 1996). However, similar structures are known from a variety of slope settings and are indicative of downslope sediment movement (Mulder and Cochonat, 1996; Nitsche et al., 2000; Haflidason et al., 2003; Lastras et al., 2003). Alternatively, such ridges could result from near-bottom current activity on the upper continental slope. However, surficial sediment sorting usually associated with such morphodynamic change would result in variations in the backscatter signal conformal with the ridge morphology (Goff et al., 2004). Based on the fact that the backscatter data from this area exhibited no detectable trend or pattern, we consider this scenario less likely. A high resolution sub-bottom survey combined with sediment sampling would help in determining the origin of these features.

The elongate depression on the outermost shelf may represent an open crown crack, erosionally enhanced normal or listric fault or en echelon (tensile) type of crack (Figs. 8B and 9) (Mulder and Cochonat, 1996; Driscoll et al., 2000). The gully heads cutting locally back into the bottom of the depression suggest that the gullies have been active subsequent to depression formation, although their more or less synchronous genesis cannot be excluded (Figs. 8B and 9).

All these features, the slide scars as well as shelf edge-parallel ridges and the elongate depression, are indicative of potentially unstable and failure-prone slopes where a variety of factors, such as earthquakes, glacioisostatic crustal movements, or ice sheet loading, may potentially trigger large sediment slides (Canals et al., 2004). Interestingly, no large-scale slope failures and canyons have been reported from the modern West Antarctic margin, although some large debris flows of Pliocene age have been imaged in seismic profiles on the Antarctic peninsula margin (Imbo et al., 2003; Diviacco et al., 2006; Hernández-Molina et al., 2006), and are also known from other glacially influenced continental margins (Ó Cofaigh et al., 2003; Dowdeswell et al., 2008).

6. Conclusions

The configuration and distribution of submarine landforms on the outer shelf and upper slope reflects the complexity of sedimentary processes along the Bellingshausen and Amundsen Sea continental margin of West Antarctica since the LGM.

The evidence presented in this study suggests:

1. Slide scars, gullies, shelf-edge parallel deformational sediment ridges and the elongated depression along the shelf edge imply laterally variable sedimentary conditions within distances of a few km. This variability was most likely controlled by the dynamics of subglacial meltwater drainage at the shelf edge.
2. Subglacial meltwater played a key role in the formation of gullies along the upper continental slope. This conclusion is drawn from their distribution, typically linked to the topographically lower sections of the outermost shelf where remnants of submarine landforms locally imply glacial meltwater drainage from beneath the ice sheet. The main areas of subglacial meltwater drainage were the marginal zones of large cross-shelf troughs.
3. Due to a lower slope gradient on the Belgica TMF, the lower energy sediment-laden meltwater flows have formed shallower and gentler gullies that cut back into the shelf edge only a short distance. In contrast, the gullies in front of Marguerite and Pine Island West Troughs are typically deeper and steeper as a result of the steeper continental slope there.
4. Based on the spatial distribution of gullies in front of the large cross-shelf troughs, the meltwater flux was probably higher near the margins of these troughs as compared to their central parts. This is based on the consistent and statistically significant (at 95% confidence level) positive correlation between the gully cross-sectional areas and their respective distances from the centre lines of the troughs. We consider that the most likely mechanism behind this trend is subglacial meltwater drainage from the topographically higher shelf banks that were covered by slower moving ice, into the large cross-shelf troughs, driven by the hydrological pressure gradient at the ice base.
5. Complex, locally amphitheatre- or cauliflower-shaped gully heads outside the major cross-shelf troughs indicate that small-scale retrogressive slope failure processes also contributed to the generation of gullies.
6. Both the gentle ridges in the uppermost part of the continental slope at c. 80°W, and the c. 6–7 km long depression on the outermost shelf at c. 80°20′W, are probably related to deformation and gravitational creep of sediments downslope.
7. Slide scars associated with debris flows, and the shelf-edge parallel ridges and elongated depressions that are indicative of different forms of downslope sediment transport, are typical of relatively elevated sections of the shelf edge. These features probably mark sections of relatively steep and unstable upper slope where the subglacial meltwater flow was weak or absent and had little effect on glacial sedimentation at the shelf edge.
8. Sedimentary evidence suggests that the gully formation post-dates the debris flow and small-scale slope failure events along the West Antarctic continental margin.

No model so far explains unambiguously the formation of gullies on high-latitude continental margins. The morphometric parameters and spatial distribution patterns discussed here place additional constraints on the generation of gullies and slide scars in terms of subglacial hydrology and submarine slope processes.

Acknowledgements

This research was funded by UK Natural Environment Research Council (NERC) grants GR3/JIF/02, NER/G/S/2002/0192 (University of Cambridge), NER/G/S/2002/00009 and the GRADES-QWAD project (British Antarctic Survey). Logistical support was provided by the British Antarctic Survey under the NERC Antarctic Funding Initiative (project AFI4/17). We thank the officers and crew of the RRS James Clark Ross for their expert assistance during cruises JR104 and JR141.

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