THE GEOMORPHOLOGY OF ANTARCTIC SUBMARINE SLOPES

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Abstract

The Antarctic continental margin contains a diverse range of continental slope morphologies, including iceberg keel marks, gullies, channels, mass-wasting features (slides, slumps), ridges, furrows, mounds and trough mouth fans. These features vary significantly in morphology, with bedforms varying in size (width, amplitude and length), shelf incision, sinuosity, branching order, spatial density and cross-sectional shape. The processes which form these features and the environmental controls influencing their morphology are not well documented or well constrained. Understanding the processes operating on the Antarctic continental margin is essential for interpreting seafloor erosion patterns, continental margin evolution, slope instability and sediment core records from the continental slope and rise.

Through quantitative analysis of multibeam bathymetric data along >2670 km of the outer shelf and upper-slope of high latitude continental margins, five distinct Antarctic gully types are identified. Gully morphology was found to vary with local slope character (slope geometry, gradient), regional factors (location of cross-shelf troughs, trough mouth fans and drainage basin size), sediment yield and ice-sheet history. Most gullies are likely formed by: (1) flows generated as a result of the release of subglacial meltwater from beneath an ice-sheet grounded to the shelf edge during glacial maxima; (2) turbidity currents initiated by intense iceberg scouring; or (3) small-scale mass-wasting. Erosion by cascading dense water overflow does not form the deeply incised and V-shaped gullies that occur over much of the Antarctic continental margin. A comparison of some Arctic and Antarctic gully morphologies shows that the Antarctic gullies have much deeper mean incision depths and greater shelf-incisions, suggesting that they either formed over significantly longer periods, or by a greater release of meltwater in the areas with greater gully incision depths.

The first morphological analysis of the southern Weddell Sea outer shelf and upper slope is presented. Two large and relatively recent submarine slides occur on the Crary Fan, the first Quaternary slides to be documented on an Antarctic trough mouth fan. These slides provide evidence for recent large-scale mass-wasting events on the Antarctic continental margin. The interpretation of bedforms on the outer shelf of the southeastern Weddell Sea provide insight into the timing and extent of past ice and points to grounded ice near to the shelf edge during the Late Quaternary.

Declaration

No portion of the work referred to in the thesis has been submitted in support of an application for another degree or qualification of this or any other university or institute of learning.

Copyright Statement

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Finally, I would like to thank my family and partner for their never-ending support and encouragement.

Preface

Jenny Gales holds a BSc (Hons) in Oceanography with Physical Geography (University of Southampton; first-class) and an MRes in Marine Geology and Geophysics (University of Southampton).

This thesis is the result of a Natural Environmental Research Council (NERC) funded PhD project entitled '*The geomorphology of Antarctic submarine slopes'*. The project was joint with the British Antarctic Survey and the University of Manchester, with the majority of the research conducted at the British Antarctic Survey. Between January and March 2011 two months were spent on research cruise JR244 on the RRS *James Clark Ross* to the southern Weddell Sea, Antarctica. Three months of research were conducted at the University of Tromsø Geology Department between September and December 2012. Research was supported by an American Association of Petroleum Geologists (AAPG) Grant In Aid award, the British Antarctic Survey Collaborative Gearing Scheme and the Research Council of Norway under an YGGDRASIL (Young Guest and Doctoral Researchers' Annual Scholarship for Investigation and Learning) Grant. The candidate undertook a two month internship with the oil company Statoil, in Trondheim, Norway in 2010.

Results of this PhD thesis have been presented in nine oral and poster presentations at national and international conferences and seminars. The thesis has resulted in four first author scientific papers and one book chapter that contribute with new scientific knowledge to the fields of continental slope research, slope processes and high latitude continental slope geomorphology.

List of publications

Gales, J. A., Larter, R. D., Mitchell, N. C., Hillenbrand, C.-D., Østerhus, S., Shoosmith, D., 2012. Southern Weddell Sea shelf edge geomorphology: Implications for gully formation by the overflow of high-salinity water. Journal of Geophysical Research – Earth Surface 117, F0421, doi:10.1029/2012JF002357.

Gales, J. A., Larter, R. D., Mitchell, N. C., Dowdeswell, J. A., 2013a. Geomorphic signature of Antarctic submarine gullies: Implications for continental slope processes. Marine Geology 337, 112-124.

Gales, J. A., Forwick, M., Laberg, J. S., Vorren, T. O., Larter, R. D., Graham, A. G. C., Baeten, N. J., Amundsen, H. B., 2013b. Arctic and Antarctic submarine gullies – a comparison of high latitude continental margins. Geomorphology 201, 449-461.

Gales, J. A., Leat, P. T., Larter, R. D., Kuhn, G., Hillenbrand, C.-D., Graham, A. G. C., Mitchell, N. C., Tate, A. J., Buys, G. B., Jokat, W., 2013c. Slides, gullies and channel systems of the southern Weddell Sea, Antarctica. Marine Geology (*Submitted*).

Gales, J. A., Larter, R. D., 2013d. Submarine gullies on the southern Weddell Sea slope, Antarctica. In: Dowdeswell, J. A., Canals, S, M., Jakobbson, M., Todd, B. J., Dowdeswell, E. K., Hogan, K. A., (Eds.), *Atlas of Submarine Glacial Landforms: Modern, Quaternary and Ancient*. Geological Society, London, Memoirs, The Geological Society of London (*Submitted*).

Section 1:

Introduction and background

Chapter 1.

Introduction, motivation and thesis outline.

1.1. Aim of thesis

The work presented in this thesis is based on the analysis and interpretation of over 2670 km of multibeam bathymetric data along the outer shelf and upper slopes of high latitude continental margins. The overall objective of the thesis was to constrain processes operating on Antarctic continental slopes, with the aim of better understanding what processes produced the seafloor morphology present. Processes operating on the Antarctic continental margin and the factors that influence these processes are not well documented or well understood, and the ages of bedforms and the time-scales that they develop over remain largely unknown. Previously glaciated continental margin morphology can provide insight into past ice-sheet dynamics and histories. A better understanding of the processes operating on continental margins and influencing the morphologies present is, therefore, essential not only to better understand past ice-sheet dynamics, but also to predict future ice-sheet change. Below, the background motivation and rationale for the thesis is presented, including a detailed examination of the thesis aims.

1.2. Background motivation for continental slope research

Melting of the Greenland and Antarctic ice sheets has the potential to cause major changes in sea level. If fully melted, these ice sheets would raise global sea levels by approximately 64 m and potentially affect ocean circulation and climate (Bamber et al., 2001; Lythe et al., 2001; Fretwell et al., 2013). Ice sheets in the Arctic and Antarctic have undergone significant phases of growth and decay and are today rapidly changing, having contributed up to 50% of global sea level rise in the last century (Church et al., 2001; Pritchard et al., 2009). Changes in precipitation, atmospheric and oceanographic conditions, including ocean thermal expansion, and melting of the lower latitude ice caps and glaciers, also contribute to this rise (IPCC, 2007). The risks associated with sea level rise to the UK are high, putting billions of pounds worth of infrastructure, resources, and environmental and cultural heritage at risk. Ice sheets are now recognised as the largest source of uncertainty in sea-level rise predictions over the next 50-200 years (IPCC, 2007). Understanding this impact is critical in assessing the future consequences of ice sheet change.

Ice streams are fast-moving corridors of ice which flow into ice shelves, smaller ice tongues or into the ocean (IPCC, 2007). Ice stream dynamics can change rapidly and are poorly understood. Improving our understanding of past extents of ice streams will increase our knowledge of the dynamic behaviour and stability of Antarctic ice sheets. Submarine slopes around Antarctica are strongly influenced by ice sheet dynamics, with sedimentary deposits containing a record of varying ice sheet extent, sub-glacial processes, past erosion and climate change. Slope processes modify this record through erosion, remobilisation and deposition, limiting our ability to interpret past subglacial processes and ice dynamic histories from the depositional record. It is therefore essential to increase our understanding of slope processes in order to improve our knowledge of the sedimentary record and ultimately ice sheet behaviour. This allows past routes of sediment delivery to be mapped and the processes responsible for this delivery to be interpreted, which in turn helps constrain where palaeo-ice streams extended to the continental shelf edge. This will improve our knowledge of past ice-dynamic behaviour and has the potential to be used for testing and refining aspects of numerical models that will predict Antarctic contributions to future sea-level rise. Knowledge of how the Antarctic ice sheets responded to past climate changes is critical in understanding and predicting ice sheet sensitivity to future change.

High latitude continental slopes present significant risk of geo-hazards, including mass-movements (i.e. submarine landslides) and tsunamis (Mitchell, 2012). Numerous submarine slides have been documented on high latitude continental margins e.g. Storegga Slide (Bugge et al., 1987; Bryn et al., 2003), Trænadjuptet Slide (Laberg and Vorren, 2000), Andøya Slide (Kenyon, 1987; Dowdeswell et al., 1996) and Bjørnøyrenna Slide (Laberg and Vorren, 1993), however the mechanisms initiating these slides remain largely unresolved and widely debated. Processes operating on continental slopes, such those leading to weak and unstable sedimentary layers (Long et al., 2003), are able to significantly influence slope instability and may lead to slide initiation. A better knowledge of continental slope processes is therefore essential to increase our understanding of submarine geohazards, the factors influencing slope instability and the risks posed by future instability

1.3. Advances in continental slope research

Advances in multibeam sonar and seabed imaging technology over the past two decades now allow for improved mapping of the seafloor. These improvements include better seafloor resolution (narrower beam widths), with increased accuracy, numbers of beams, swath widths, improved motion sensors and GPS and the potential to measure acoustic backscatter (Caress and Chayes, 1996). Technological advances, such as the use of remotely operated vehicles (ROVs) and autonomous underwater vehicles (AUVs) have allowed logistically difficult areas to be investigated, such as beneath ice-shelf cavities (Jenkins et al., 2010; Ó Cofaigh, 2012).

Global compilation projects such as the General Bathymetric Chart of Oceans (GEBCO, 2003) and regional projects within this such as the International Bathymetric Chart of the Southern Ocean (IBSCO; Arndt et al., 2013) have significantly increased the accuracy of low-resolution global bathymetry using internationally compiled bathymetric data from multibeam and single beam echo sounders, digitized nautical charts and bathymetry predicted from satellite altimetry data (Arndt et al., 2013). The use of fishing vessels to collect bathymetry data, for example the Norwegian company OLEX, and research driven by the oil and gas industry have also enhanced data coverage and resolution. Industry directed research has in particular, led to an increase in high-resolution subbottom data, including 3D seismic data. Programmes such as the Integrated Ocean Drilling Program (IODP) and Ocean Drilling Programme (ODP) and advances in both sediment coring methods and analysis have improved the recovery of longer sediment cores and dating techniques (Ó Cofaigh, 2012).

In Antarctica, although surveying has increased significantly over the last two decades, data extent is still limited by the remoteness of region, with major logistical issues arising from extensive sea-ice cover for much of the year. Many areas in Antarctica remain unsurveyed.

1.4. Rationale

1.4.1. Project originality

The thesis presents a detailed analysis of Antarctic continental slope morphology, with particular emphasis on submarine gullies- the most common morphological features identified on the Antarctic continental margin. Submarine gullies are small-scale, confined channels that form depressions along the continental shelf edge and upper continental slope. Gullies have been identified along much of the Antarctic margin (e.g. Vanneste and Larter, 1995; Shipp et al., 1999; Heroy and Anderson, 2005; Dowdeswell et al., 2004; 2006; 2008; Noormets et al., 2009), however quantitative differences between gully morphologies, the spatial distributions of these features, and the processes that form them, are largely unknown.

The study uses a combination of new and previously collected multibeam bathymetric data from a range of international institutions (Table 3.1), with data spanning > 2670 km along high latitude continental margins. Bedforms, including iceberg keel marks, gullies, channels, mass-wasting features (slides, slumps), ridges, furrows, and mounds, were identified from this data and their morphology quantitatively analysed and spatial distributions mapped.

New data is presented from the southern and south-eastern Weddell Sea, Antarctica collected over two consecutive years (2011-2012). The new data provides insight into palaeo-ice sheet dynamics and histories on the southern and southeastern Weddell Sea margin and improves constraints on past glacial reconstructions within this region. The study presents the first evidence of Quaternary submarine slides observed on an Antarctic trough mouth fan. Recent slides are relatively rare on the Antarctic margin, with few documented examples (e.g. Imbo et al., 2003). This study therefore provides evidence for recent large-scale mass wasting in Antarctica. The thesis provides a step forward in our understanding of Antarctic continental slope morphology, gully forming mechanisms, large-scale slope instability and continental slope processes.

The study provides the first comparison of submarine gullies from Arctic and Antarctic continental margins. The Arctic continental margins used in this study include the northern Norwegian margin, southwest Barents Sea and western Svalbard margin. The areas analysed have similar underlying geologies, with gullies formed in glacigenic and glacially influenced sediments, similar landward sloping shelves and prograded continental slopes. By quantitatively comparing over 1500 gullies from previously glaciated margins and examining similarities and difference between the adjacent slopes, processes operating on continental slopes, factors influencing the different morphologies, and the time-scales that the features formed over, are constrained.

1.4.2. Aims and objectives

The main objective of this thesis is to constrain processes operating on Antarctic continental margins with the aim of better understanding the processes influencing continental slope morphology. This objective can be separated into three separate aims, which together provide a step forward in continental slope research. These are:

1) What are the main morphological features of Antarctic continental slopes and how do these morphologies vary?
2) What factors or processes are responsible for the different morphologies observed on Antarctic continental slopes?
3) What can the slope processes inferred from Antarctic continental margins contribute to the wider field of continental slope research?

Each aim is addressed in the results chapters of this thesis (chapters 4 to 8) and resulting insights are summarised in chapter 9.

Aim 1: What are the main morphological features of Antarctic continental slopes and how do these morphologies vary?

Chapter 4 is a two-page contribution to the Atlas of Submarine Glacial Landforms: Modern, Quaternary and Ancient (Dowdeswell et al., 2014). The contribution provides a detailed description of submarine gullies with particular emphasis on the southern Weddell Sea.

Gales, J. A., Larter, R. D., Leat, P. T., 2013d. Submarine gullies on the southern Weddell Sea slope, Antarctica. In: Dowdeswell, J. A., Canals, S, M., Jakobbson, M., Todd, B. J., Dowdeswell, E. K., Hogan, K. A., (Eds.), *Atlas of Submarine Glacial Landforms: Modern, Quaternary and Ancient*. Geological Society, London, Memoirs, The Geological Society of London (*Submitted*).

Chapter 5 provides a comprehensive analysis of submarine gullies, the most common features on the Antarctic continental margin. Differences are identified in gully morphology by measuring different gully parameters from multibeam bathymetry data along shelf-parallel profiles, including gully length, width, depth, sinuosity, branching order, cross-sectional shape, shelf-incision and density. Using statistical methods, different gully types are recognised from the Antarctic margin and the spatial distribution of these different types mapped by visual pattern recognition. Gully parameters were analysed in conjunction with local slope geology, slope geometry and gradient, large-scale spatial characteristics (i.e. presence of trough mouth fans, cross-shelf troughs and drainage basin size) and oceanographic regimes, to infer processes operating on the Antarctic continental margin. This chapter is published in the journal *Marine Geology*:

Gales, J. A., Larter, R. D., Mitchell, N. C., Dowdeswell, J. A., 2013a. Geomorphic signature of Antarctic submarine gullies: Implications for slope processes. Marine Geology 337, 112-124.

Chapters 7 and 8 use new geophysical data collected in 2011 and 2012 on RRS *James Clark Ross* to identify the main morphological features of the southern Weddell Sea, Antarctica. The small-scale slope morphology of these areas had not been studied previously and thus their slope morphologies were largely unknown. In Chapter 7, small scale and U-shaped gullies that incise the shelf-edge of the southern Weddell Sea are identified. Deeply incised and V-shaped gullies as observed in Chapter 5, are absent from the southern Weddell Sea margin. In chapter 8, a range of bedforms are identified including iceberg scours, gullies, channel systems and submarine slides. Chapter 7 is published in the *Journal of Geophysical Research – Earth Surface* and chapter 8 is submitted for publication in the journal *Marine Geology*:

Gales, J. A., Larter, R. D., Mitchell, N. C., Hillenbrand, C.-D., Østerhus, S., Shoosmith, D., 2012. Southern Weddell Sea shelf edge geomorphology: Implications for gully formation by the overflow of high-salinity water. Journal of Geophysical Research – Earth Surface 117, F0421, doi:10.1029/2012JF002357.

Gales, J. A., Leat, P. T., Larter, R. D., Kuhn, G., Hillenbrand, C.-D., Graham, A. G. C., Mitchell, N. C., Tate, A. J., Buys, G. B., Jokat, W., 2013. Slides, gullies and channel systems of the southern Weddell Sea, Antarctica. Marine Geology (*Submitted*).

Aim 2: What factors or processes are responsible for the different morphologies observed on Antarctic continental slopes?

Chapters 5 and 6 aim to constrain processes operating on high latitude continental margins by examining, firstly, the differences in Antarctic slope morphology (Chapter 5) and, secondly, differences between Arctic and Antarctic slope morphology (Chapter 6). By comparing the spatial distributions of particular gully types (or slope signatures), with the large-scale slope characteristics (i.e. presence of cross-shelf troughs; slope gradient etc.), glaciology and palaeo-glaciology (i.e. drainage basin size; past-glacial history etc.), shallow geology and oceanographic regime, gully-forming mechanisms and factors influencing gully formation are constrained. Chapter 6 is published in the journal *Geomorphology*:

Gales, J. A., Forwick, M., Laberg, J. S., Vorren, T. O., Larter, R. D., Graham, A. G. C., Baeten, N. J., Amundsen, H. B., 2013b. Arctic and Antarctic submarine gullies – a comparison of high latitude continental margins. Geomorphology 201, 449-461.

Chapter 7 tests a popular gully-forming hypothesis in the literature, which suggests that gullies are eroded by cascading flows of cold, dense water, formed during sea-ice formation through brine rejection (Kuvaas and Kristoffersen, 1991; Dowdeswell et al., 2006; 2008; Noormets et al., 2009; Nicholls et al., 2009). New multibeam bathymetric data from the southern Weddell Sea were analysed in a region of highly energetic cascading dense water overflow. Oceanographic measurements from moorings 10 m above the seafloor in this region show that cold, dense water cascades down-slope with velocities up to 1 m s^{-1} (Foldvick et al., 2004). By analysing the surface morphology in this region, chapter 7 tests whether cold, dense water overflow has a distinct seafloor signature and whether deeply incised and V-shaped gullies, which are observed over much of the Antarctic continental margin (shown in chapter 5), occur. If present, this would suggest that cold, dense water overflow is an important gully-forming mechanism which, although not active over much of the continental slope today, may have been active and highly energetic in the past.

Chapter 8 compares the continental slope morphology at the mouths of two cross-shelf troughs in the southern (described in chapter 7), and south-eastern Weddell Sea. Although the troughs are geographically close (separated by 40 km), they display significantly different slope morphologies. By analysing the different slope characteristics (slope gradient; geometry), glacial histories (drainage basin size; ice-sheet size), underlying geologies (source rocks, surface sediments) and oceanographic regimes, chapter 8 identifies differences in slope processes which have led to one slope displaying largely depositional characteristics and the other to display largely erosional characteristics. The morphology of two submarine slides are analysed in chapter 8 using multibeam, subbottom and seismic data. Using these data, the initiation mechanisms and factors influencing slide morphology are discussed, including why slides occur in this region but are largely absent from other Antarctic continental margins (Barker and Austin,1998; Dowdeswell et al., 2002; Nielsen et al., 2005).

Aim 3: ***What can the slope processes inferred from the Antarctic continental margin contribute to the wider field of continental slope research?***

Chapter 9 draws the results of the previous chapters together and discusses insight gained into the processes operating on high latitude continental slopes and the factors influencing these processes. The wider implications of this research are discussed in terms of palaeo-ice sheet history. Finally, recommendations are made for future research directions.

1.5. Thesis structure - thesis by publication

The thesis is structured as a thesis by publication and includes four research papers and one book chapter containing original research in chapters 4, 5, 6, 7 and 8. The research papers are presented in a logical rather than a chronological order and each paper is self-contained, including a separate abstract, introduction, methodology, results, discussion, conclusion and references section. In chapter 4, a detailed description of submarine gullies is provided. Chapter 5 quantitatively identifies different slope morphologies on the Antarctic continental margin and provides insight into the processes that form these slope signatures. Chapter 6 compares Antarctic and Arctic slope morphologies and identifies key differences between high latitude continental margins. Differences in slope processes and gully forming mechanisms between Arctic and Antarctic continental margins are discussed. Chapters 7 and 8 present new multibeam bathymetric data from the southern and southeastern Weddell Sea collected over two consecutive years and constrains slope processes occurring in these regions. Chapter 8 identifies unique bedforms that have not been observed on other Antarctic continental margins. Significant differences in slope morphology are discussed between two Antarctic cross-shelf troughs.

An alternate format thesis was most appropriate as the study naturally separates into multiple research areas (potential gully forming mechanisms), which form the basis of the four research papers and one book chapter. Although separate, these topics interlink to help constrain the mechanisms forming the distinct slope morphologies that we find on the Antarctic continental margin. Together, the following chapters provide a step forward in our knowledge of Antarctic continental slope morphology, slope processes and factors influencing these processes.

Below, I have summarised my contribution, and the contribution of my co-authors for each chapter included.

- **Chapter 4**: As lead author, I analysed and interpreted the data and produced all figures and text. This two page contribution is based on the research presented in Chapters 7 and 8.

- **Chapter 5**: As lead author, I gridded multibeam bathymetric data from a range of data sources (Table 3.1). I analysed and interpreted the data by measuring gully parameters from profiles parallel to the shelf edge, large-scale spatial characteristics and by analysing TOPAS (subbottom) data. Co-authors, Rob Larter, Neil Mitchell and Julian Dowdeswell were involved in some data collection and provided comments on the draft manuscript. This chapter is published in the journal *Marine Geology*.

- **Chapter 6**: As lead author, I processed some of the multibeam bathymetric data. I analysed and interpreted the data by measuring bedform parameters from multibeam data and compared differences between Arctic and Antarctic datasets. Co-authors included Matthias Forwick, Jan Sverre Laberg, Tore Vorren, Rob

Larter, Alastair Graham, Nicole Baeten and Hilde Amundsen. Co-authors were involved in some data collection, data processing and provided comments on the draft manuscript. This chapter is published in the journal *Geomorphology*.

- **Chapter 7**: As lead author, I was involved in data collection during cruise JR244 (2011) to the southern Weddell Sea and all multibeam bathymetric and TOPAS data processing associated with this dataset. I analysed and interpreted the data by measuring gully parameters along profiles extracted from the bathymetry data. Co-authors Rob Larter, Neil Mitchell, Deb Shoosmith, Claus-Dieter Hillenbrand and Svein Østerhus were involved in data collection and provided comments on the draft manuscript. This chapter is published in the *Journal of Geophysical Research – Earth Surface*.

- **Chapter 8**: As lead author, I was involved in some data collection and some data processing. I gridded all multibeam bathymetric data involved and analysed and interpreted the data by measuring a range of bedforms from multibeam, TOPAS and seismic data. Co-authors Rob Larter, Phil Leat, Claus-Dieter Hillenbrand, Neil Mitchell, Gerhard Kuhn, Alastair Graham, Alex Tate and Gwen Buys were involved in some data collection and some data processing, including processing of the seismic data. Co-authors provided comments on the draft manuscript. This chapter is submitted to the journal *Marine Geology.*

- **Chapter 9**: This is a summary of the main conclusions of the research conducted throughout the thesis. The wider implications of this research are discussed and recommendations for future research made.

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Chapter 2.

Scientific context.

2.1. Study Area

The study areas include the western Antarctic Peninsula, Bellingshausen, Amundsen and Weddell Sea continental margin (Fig. 2.1), with data covering > 2126 km of shelf edge and upper continental slope. In Chapter 6, 520 km of Arctic continental margin is also examined, including the continental slope of western Svalbard, southwest Barents Sea and northern Norway.

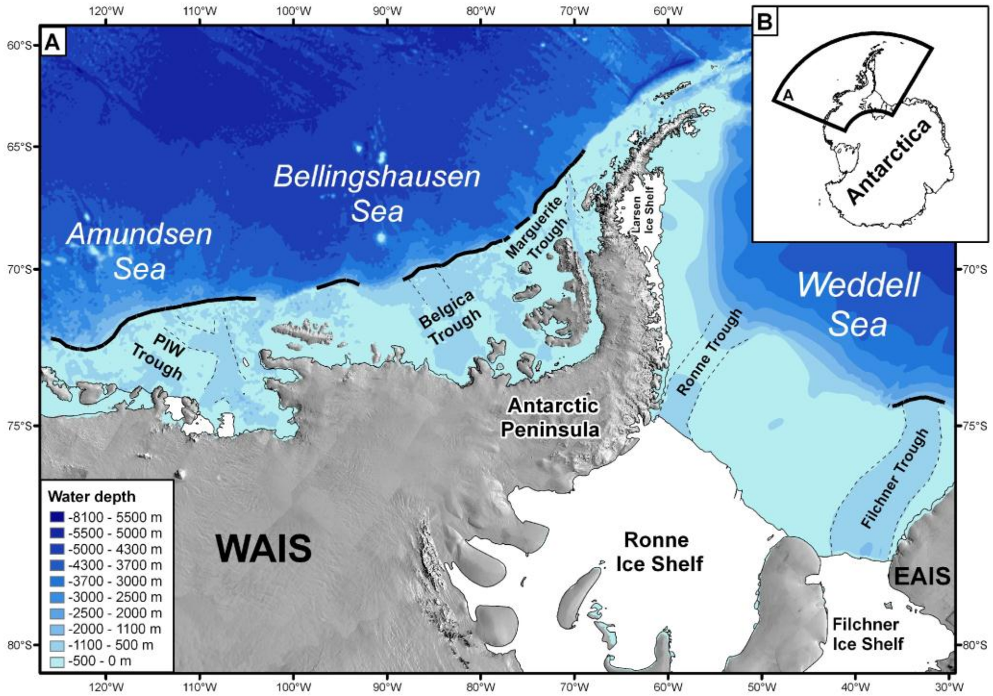


Figure 2.1. A. Antarctic study areas. Black lines indicate areas where data is available. Dotted lines indicate cross-shelf trough boundaries. WAIS is West Antarctic Ice Sheet. EAIS is East Antarctic Ice Sheet. **B**. Inset figure locates (A) on the Antarctic continent. Taken from Gales et al., (2013a).

2.1.1. Antarctic geological setting

Antarctica is formed of two main continental blocks, East and West Antarctica, which are separated by the Transantarctic Mountains. East Antarctica formed during the Archean – Cambrian with West Antarctica forming mainly in the Mesozoic and Cenozoic (Anderson, 1999). East Antarctica is constructed predominantly of the East Antarctic Craton, formed during the Early Archean (Borg et al., 1987). The West Antarctic continent is formed of four discrete/semi-discrete crustal blocks (Antarctic Peninsula, Thurston Island, Ellsworth-Whitmore mountains and Marie Byrd Land). The blocks have separate histories and settled into their present relative positions during the Cretaceous (Dalziel and Elliot, 1982).

The Pacific margin of the Antarctic Peninsula is a relict subduction zone. Subduction occurred throughout the Mesozoic era (Pankhurst, 1990) and ended progressively from the southwest to northeast during the Tertiary period as successive ridge crest segments of the Antarctic-Phoenix spreading centre migrated into the trench (Herron and Tucholke, 1976; Barker, 1982). The Bellingshausen Sea margin is a relict subduction zone, tectonically inactive since 55 Ma (Cunningham et al., 2002). The Amundsen Sea is a divergent plate margin, which has remained passive since New Zealand rifted away from Antarctica at 90 Ma (Larter et al., 2002; Eagles et al., 2004). The Weddell Sea has been a passive margin since the Jurassic period (Bart et al., 1999).

Seismic reflection and coring studies from the Antarctic continental margin show that the Quaternary geology along the slope exhibits a limited range of characteristics, with similar lithology, geotechnical properties and grain size composition (e.g. Vanneste and Larter, 1995; Anderson and Andrews, 1999; Dowdeswell et al., 2004a, 2006, 2008; Hillenbrand et al., 2005, 2009). The continental slopes are constructed largely of thick prograded sedimentary sequences (e.g. Anderson, 1999; Cooper et al., 2008) that reflect cycles of growth and retreat of grounded ice from the West Antarctic Ice Sheet, East Antarctic Ice Sheet and Antarctic Peninsula Ice Sheet.

Terrigenous sediment reaches the shelf via sub-glacial processes, reworking by currents and icebergs and as ice-rafted debris (Vanneste and Larter, 1995). Other mechanisms, such as wind-driven waves and currents have limited influence as extensive sea-ice occurs for much of the year and due to the significant water depths present (Vanneste and Larter, 1995). On the continental slope, less sediment is deposited either side of an ice stream and during interglacial periods (Dowdeswell et al., 2004a). During interglacials, glaciomarine, hemipelagic, and pelagic sediments accumulate on the shelf (Rebesco et al., 1996). The sedimentary regime on the Antarctic continental margin contrasts with that of low latitudes in the large contrast between glacial and interglacial periods in sediment supply to the shelf edge (Tomlinson et al., 1992).

Ice stream flow and location is influenced by substrate type, bed roughness, topography and tectonic controls (Siegert et al., 2005; Smith and Murray, 2009; Bingham and Siegert, 2009; Winsborrow et al., 2010; Livingstone et al., 2012). Fast ice flow is controlled by interactions at the base, with movement facilitated by either subglacial deformation of soft sediments or by basal sliding (Smith and Murray, 2009). Regions of slower moving ice correlate with areas of sedimentary rock or crystalline bedrock, which have a rougher topography (Graham et al., 2009). Studies using seismic reflection data indicate that complex patterns of basal conditions result in strong local variability within an ice stream bed (Smith and Murray, 2009). Geophysical studies show that cross-shelf troughs were repeatedly occupied by ice streams over several glacial cycles, which likely influenced more recent ice stream locations (Larter and Barker, 1989, 1991).

2.1.2. Antarctic glacial setting and history

The Antarctic cryosphere contains 27 million km^3 of ice (Fretwell et al., 2013) and includes the West Antarctic Ice Sheet (WAIS), the East Antarctic Ice Sheet (EAIS) and smaller glacial systems on the Antarctic Peninsula Ice Sheet (APIS) (Lythe et al., 2001). The WAIS is a marine ice sheet, grounded in some locations up to 2 km below sea level (Fretwell et al., 2013). The EAIS is a more stable grounded ice mass with an average thickness of 3 km (Anderson, 1999) and proportionally less flow and discharge than the WAIS (Anderson et al., 2002). The modern APIS is approximately 500 m thick and mostly grounded above sea level (Heroy and Anderson, 2005). This is, however, only a small remnant of the APIS during glacial periods, with much of the area of the glacial APIS below sea level.

Ice has covered parts of Antarctica for the past 34 m.y. (Barrett, 2008) and has since undergone cycles of advance and retreat. The onset of the Antarctic glaciation is suggested to result from: (1) the decline in atmospheric CO_2 levels (DeConto and Pollard, 2003); (2) opening of ocean gateways in the mid-Tertiary enabling ocean circulation to develop around Antarctica (Kennett, 1977); and (3) uplift of the Antarctic mountain ranges, with uplift of the Transantarctic mountains occurring at around 55 Ma (Fitzgerald, 2002).

The extent and volume of past ice sheets remain poorly constrained due to limited onshore and offshore data, difficulties in dating the records and problems associated with inferring ice-sheet volume from the sedimentary record and from glacial isostacy (Nielson et al., 2005). Other measures used to infer changes include the oxygen isotope ($\partial^{18}O$) signal in deep sea sediments, however this is also associated with preservation and diagenetic alteration issues (Pearson et al., 2001). The first continent-wide EAIS developed at around 34 Ma, around the Eocene/Oligocene boundary with ice reaching the coast in Prydz Bay and the Ross Sea (Barron et al., 1988; Barrett, 1989; Barrett et al., 1995; Cooper and O'Brien, 2004). Early history of the WAIS is less well constrained, with no substantial evidence for the presence of the WAIS during the Oligocene (Ingolfsson, 2004). A major step in the development of the WAIS is likely associated with the shift in oxygen isotope ratios in open ocean foraminifera about 14 Ma (Zachos et al., 2001; Shevenell et al., 2004). Compelling evidence from the morphology of volcanoes in Marie Byrd Land, West Antarctica, shows that climate in the interior of this region has not been warm enough to permit significant runoff since the Middle Miocene (Rocchi et al., 2006). Extensive glaciations occurred in both East and West Antarctica during the Miocene, with ice reaching the Antarctic Peninsula continental shelf by the Middle-Late Miocene (Barker and Camerlenghi, 2002). Significant ice volume fluctuations occurred during the Pliocene (Prentice and Fastook, 1990) with seismic records from the Antarctic continental shelf and Ross Sea showing evidence for repeated advance and retreat patterns of grounded ice (Bart and Anderson, 1995). WAIS collapse occurred during some Pleistocene interglacials, possibly including the Eemian interglacial (Mercer, 1968; Scherer et al., 1998).

During the Last Glacial Maximum (LGM), the WAIS extended across the continental shelf of the western Antarctic Peninsula, Bellingshausen and Amundsen seas, reaching the shelf edge in most areas (e.g. Anderson et al., 2002; Ó Cofaigh et al., 2005a, 2005b). Geomorphological evidence and evidence from unconformities indicate that ice streams reached the shelf edge in the Bellingshausen and Amundsen Sea (Ó Cofaigh et al., 2005b; Anderson et al., 2002). Numerical ice sheet models suggest that the grounding line advanced to the Antarctic Peninsula continental shelf edge during the LGM (Stuiver et al., 1981; Pollard and DeConto, 2009). The models correspond with marine seismic and drilling data from the Antarctic continental margin, which display characteristics of an ice sheet grounded at the shelf edge at successive glacial maxima, including a landward sloping shelf, steep continental slope and prograding continental margin (Pudsey et al., 1994).

The EAIS extent during the LGM is widely debated, with estimates ranging from no ice expansion compared to present to ice extent reaching the shelf edge (Hall, 2009). Although also widely debated (e.g. Fütterer and Melles, 1990; Anderson et al., 2002; Bentley et al., 2010), geomorphological and coring studies suggest that ice extended across the Weddell Sea shelf and likely reached the shelf edge during the LGM (Hillenbrand et al., 2012; Larter et al., 2012). Hillenbrand et al., (2012) interpreted some diamictons recovered in cores from the southern Weddell Sea to be of subglacial origin and likely to be of LGM age. The latter conclusion is based on (1) radiocarbon ages of glaciomarine sediments overlying the subglacial sediments giving ages predominantly younger than the LGM; (2) velocity measurements of bottom currents on the shelf, which are deemed unlikely to have eroded a widespread unconformity separating subglacial deposits from the overlying Holocene glaciomarine deposits; and (3) radiocarbon dates of post-LGM age obtained from glaciomarine sediments overlying terrigenous deposits on the continental slope, which indicate that glacigenic detritus was supplied directly to the Weddell Sea shelf edge during the LGM (Hillenbrand et al., 2012).

Antarctic deglaciation is suggested to have commenced around 18 cal. ka BP due to atmospheric warming (Jouzel et al., 2001). Sediment core data show, however, that the timing and rate of deglaciation varied throughout Antarctica, with timing of initial onset as early as 30 cal. ka BP in some locations (Hillenbrand et al., 2010).

2.1.3. Oceanographic setting

The Southern Ocean is a heterogeneous system, formed of physically, biologically and chemically distinct sub-regimes (Treguer and Jacques, 1992) and with boundaries to the South Pacific Ocean, South Atlantic Ocean and the Indian Ocean (Foldvik and Gammelsrod, 1988). The Southern Ocean is strongly affected by the overlying atmosphere, which influences the exchange of heat, momentum and fresh water (Smith and Klinck, 2002) and is the primary forcing factor influencing the salinity-temperature structure west of the Antarctic Peninsula (Klinck et al. 2004).

The main components west of the Antarctic Peninsula include Antarctic Surface Water (upper 100-120 m), Winter Water and Circumpolar Deep Water (CDW) (Smith et al., 1999; Klinck et al., 2004). No Antarctic slope front is present, allowing CDW to penetrate onto the shelf (Whitworth et al., 1998). The Antarctic Circumpolar Current (ACC) is the world’s largest current, usually making up the entire water mass between the surface waters and seafloor, although, on the upper rise of the western Antarctic Peninsula, the ACC does not reach the seafloor as a south-westward current is located beneath it (Hillenbrand et al., 2008). The southern boundary of the ACC west of the Antarctic Peninsula flows along the shelf break (Smith and Klinck, 2002).

In the southern Weddell Sea, High Salinity Shelf Water (HSSW) is produced during sea ice production through brine rejection (Nicholls et al., 2009). HSSW is subsequently supercooled and slightly freshened by circulation beneath the ice shelves, producing cold and dense Ice Shelf Water (ISW) (Nicholls et al., 2009). Mean flow velocities of 0.38 m s^{-1} have been measured on the upper slope at 10 m above the seabed, reaching velocities of 1 m s^{-1} down-slope (Foldvik et al., 2004). This cold, dense water contributes to Weddell Sea Deep Water and Antarctic Bottom Water (AABW), which flow around the northern tip of the Antarctic Peninsula, extending to the South Shetland Trench (Nowlin and Zenk, 1988; Sievers and Nowlin, 1984). AABW forms the southern component of the global thermohaline circulation and is responsible for cooling and ventilating the abyssal world ocean (Foldvik et al., 2004). There is no cold, dense water production west of the Antarctic Peninsula at present due to the inflow of CDW onto the shelf, which prevents Winter Water from extending down to the seafloor, required for the production of cold, dense water (Smith and Klinck, 2002).

Although past volume fluxes of ISW are difficult to estimate, fluxes must have been closely related to circulation changes in the clockwise flowing Weddell Gyre. These, in turn, were closely related to circulation changes within the west-wind driven Antarctic Circumpolar Current. Studies of marine and terrestrial palaeoclimate archives (e.g., Bianchi and Gersonde, 2004; McGlone et al., 2010) did not reveal any significant changes in circulation of the Weddell Gyre or the westerly wind system for the time after 8 cal. ka before present, implying that any major changes in ISW production during the middle and late Holocene are unlikely. During the Last Glacial Maximum (LGM), when the ice sheet is thought to have advanced across the Weddell Sea shelf (Hillenbrand et al., 2012), ISW production likely ceased as ice shelf cavities, needed to supercool HSSW, would not have existed if the ice sheet grounding line had reached the shelf edge. It remains a possibility that HSSW may have been produced at some stages during past glacial cycles in the Amundsen and Bellingshausen seas and west of the Antarctic Peninsula.

South-westward flowing contour-following bottom currents are documented on the slope and rise of the western Antarctic Peninsula margin between 1000-4000 m deep (Camerlenghi et al., 1997; Giorgetti et al., 2003). The currents are inferred from the contourite-like stratigraphic geometry of sedimentary mounds, seafloor photographs and current measurements (Rebesco et al., 1996; Heezen and Hollister, 1971; Tucholke, 1977; Camerlenghi et al., 1997; Giorgetti et al., 2003). The currents likely originate from either Weddell Sea Deep Water (WSDW) or modified WSDW, as a result of mixing with Lower CDW (Giorgetti et al., 2003; Hernández-Molina et al., 2006).

2.2. Continental shelf and slope morphology

2.2.1. Antarctic shelf, slope and rise

The Antarctic continental shelf can be split geologically into three main sectors, the inner, middle and outer shelf (Pudsey et al., 1994). For clarity, the term ‘slope’ will only be used to describe the geomorphological feature, whereas ‘gradient’ will be used to describe the rate of elevation change. The inner shelf is generally characterised by irregular and rugged topography and is formed of crystalline bedrock, largely stripped of sediment by glacial erosion (Heroy and Anderson, 2005). The middle shelf is composed of pre-glacial or thin glacial sediments, forming a zone of older, truncated sedimentary strata (Heroy and Anderson, 2005). The mid-outer shelf is generally landward sloping and shallows toward the shelf break as a result of grounded ice sheet erosion and to a lesser extend due to lithospheric flexture from ice sheet loading (ten Brink and Cooper, 1992; ten Brink and Schneider, 1995; Bart and Iwai, 2012). The outer continental shelf is characterised by prograding sequences, produced by the action of ice sheets grounded at the shelf edge during Glacial Maxima (Larter and Barker, 1989; Heroy and Anderson, 2005).

The continental shelves are dissected by broad cross-shelf troughs, which extend to the shelf edge in places. The troughs were eroded by fast-flowing ice-streams, which transported sediment towards the shelf edge during glacial periods (e.g. Vanneste and Larter, 1995; Dowdeswell and Siegert, 1999; Ó Cofaigh et al., 2003; Dowdeswell et al., 2004a; 2004b). The troughs are characterised by steep sides and flat bottoms eroded into crystalline bedrock and generally become progressively more subdued from the inner to outer shelf, with elevation differences between the trough axes and adjacent banks decreasing from > 1000 m to ~100 m. Substrate changes to a predominantly sedimentary material from inner to outer shelf, providing a more easily erodible surface and resulting in the presence of elongate bedforms such as drumlins and mega scale glacial lineations within some cross shelf troughs (Ó Cofaigh et al., 2005a). Trough mouth fans develop at the mouths of some cross shelf troughs, forming as a result of sediment progradation and composed of mainly glacigenic debris-flow deposits (Ó Cofaigh et al., 2003).

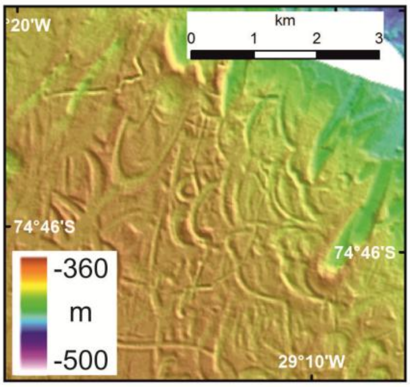
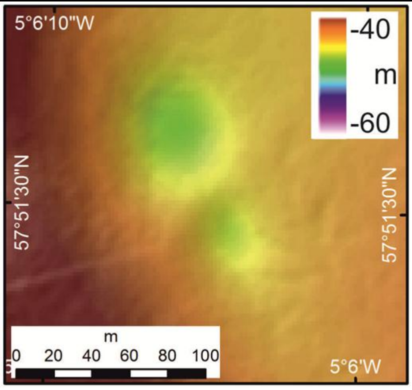
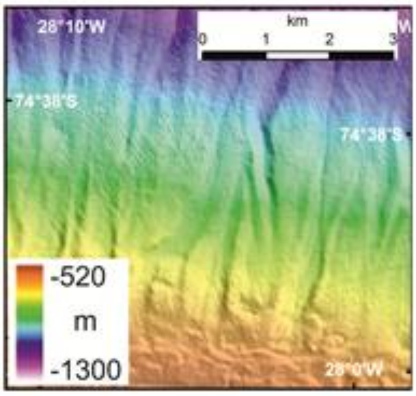
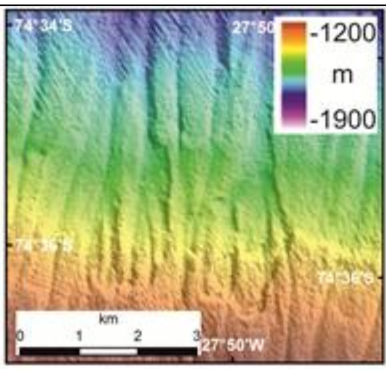
The continental slopes are constructed largely of prograded sedimentary sequences, formed by cycles of growth and retreat of grounded ice (e.g. Anderson, 1999; Cooper et al., 2008). Slope gradient and geometry vary along the continental margin, however gradient is generally steep, reaching 17º (Larter et al., 1997). Seaward of the continental slope, the rise is composed of terrigenous turbidites and contourites interbedded with layers of pelagic sediment containing ice-rafted debris (Larter and Cunningham, 1993; Rebesco et al., 1996; 1997; 2002).

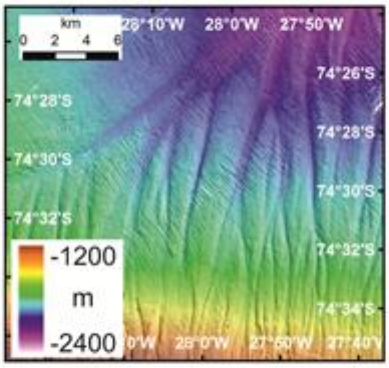
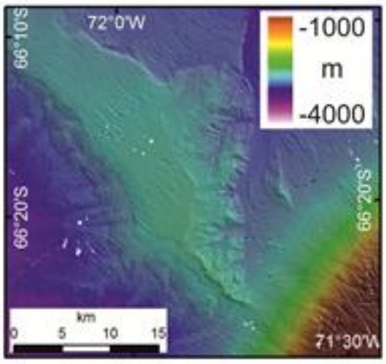
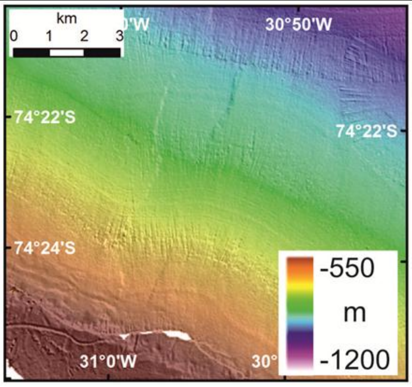
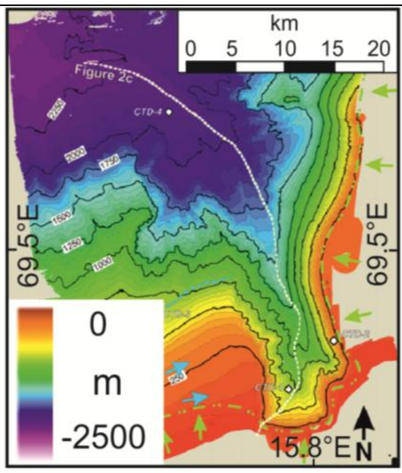
2.2.2. Continental slope bedfoms

The Antarctic continental margin contains a diverse range of slope morphologies. These include iceberg scours or keel marks, gullies, channels, mass-wasting features (slides, slumps), ridges, furrows and mounds (e.g. Ó Cofaigh et al., 2003; Dowdeswell et al., 2004a; 2006; 2008; Noormets et al., 2009; Gales et al., 2013a). Bedforms such as gullies vary in size (width, depth, and length), shelf incision, sinuosity, branching order, density and cross-sectional shape (Noormets et al., 2009; Gales et al., 2012).

Submarine gullies are small-scale, distinct channels in the range of tens of meters deep and are the most common features of Antarctic continental margins (Table 2.1). Submarine channels are defined as larger features, with canyons characterised by steep walls, usually V-shaped cross-sections, sinuous valleys and relief similar to that of their terrestrial counterparts (Shepard, 1963; 1981).

Table 2.1. Morphological features of high latitude continental slopes.

Feature	Example	Description
Iceberg plough mark		Linear to curvilear depressions which occur on the shelf and upper slope which are interpreted to result from scouring by the keels of icebergs. On the Antarctic margin, scours have been measured with mean incision depths of 8 m, widths of 204 m and lengths of 2 km (Gales et al., 2013b).
Pockmark		Rounded, crater-like structures caused by fluid or gas escape. On Arctic margins, pockmarks have been measured with mean diameters of 150 m and depths fo 15 m (Vogt et al., 1994).
Gully		Small-scale, distinct channels which initiate at, or near to, the shelf edge. On Antarctic margins, gullies have lengths ranging from one to tens of km, incision depths of tens of metres and widths of <1 km (Gales et al., 2013a).
Small-scale mass wasting		Small-scale slope failure, with characteristics such as steep headwalls or sidewalls, rounded escarpments, small-scale failing of sediment blocks or slabs and an absence of a well-defined gully thalweg.

Channel		Channels have well-defined thalwegs and do not usually incise the shelf edge. In some cases, levees may be present. Channels may vary in the degree of branching and sinuosity. On Antarctic margins, channels have been measured with widths of >1 km, depths of tens of metres and lengths of tens of km (Gales et al., 2013b).
Sediment mound / drift		Typically asymmetric in shape and with small-scale mass-wasting common on the steeper flank of the mound. On the Antarctic margin, drifts have been measured with lengths of 150 km, widths of >50 km and a relief of 1 km (Rebesco et al., 1996).
Submarine landslide		Slope failure where a coherent mass of sediment moves down slope as a result of planes of failure witihin the underlying sediment (Masson et al., 2006). On the Antarctic margin, landslides measuring 20 km wide, 20 km long and 60 m deep have been measured (Gales et al., 2013b)
Canyon		A steep sided and deep depression which incises the seafloor. Canyons can incise the shelf edge or initiate from further down slope. Canyons have been measured on Arctic margins measuring 1100 m deep, 12 km wide and 40 km long (Laberg et al., 2000; Haflidason et al., 2007). Figure adapted from Laberg et al., (2007).

Different gully morphologies are identified on high latitude continental margins (Noormets et al., 2009; Gales et al., 2012; 2013a), on low latitude margins (i.e. Micallef and Mountjoy, 2011; Vachtman et al., 2012), on hill-slopes in the terrestrial environment (e.g. Hartmann et al., 2003) and on Martian surfaces (Malin and Edgett, 2000). Submarine gullies may be erosional features, formed by incision into the underlying substrate (Micallef and Mountjoy, 2011), or depositional features, developed through aggradation of the gully interfluves (Chiocci and Casalbore, 2011). The processes that form submarine gullies are not well understood or well constrained. Submarine gully formation has been attributed to a variety of processes including mass-wasting (slides, slumps, debris flows, turbidity currents), oceanographic processes such as currents, tides and waves, changes in sea-level, cascading dense water overflows and other hyperpycnal flows, such as sediment-laden subglacial meltwater discharge (Chapter 2.4) (Goodwin, 1988; Kuvaas and Kristoffersen, 1991; Larter and Cunningham, 1993; Shipp et al., 1999; Wellner et al., 2001; Izumi, 2004; Dowdeswell et al., 2006, 2008; Noorments et al., 2009; Piper et al., 2012).

2.3. Knowledge to date: Antarctic continental slope geomorphology

Table 2. 2. Summary of Antarctic cross shelf troughs.

Cross shelf trough	Width[a] (km)	Length[a] (km)	Slope gradient[b] (°)	Slope geometry	TMF	Slope bedforms	Reference (e.g.)
Marguerite Trough	40	370	9.0	Concave; steepest gradient at upper-slope.	No	Gullies; channel system.	Ó Cofaigh et al. 2003; Dowdeswell et al. 2004a; Noormets et al. 2009
Belgica Trough	150	250	1.7	Linear	Yes	Gullies; channels; slide scars.	Dowdeswell et al. 2008; Noormets et al. 2009; Ó Cofaigh et al. 2005b;
Pine Island West Trough	50	~500	4.5	Linear	No	Gullies; Channels	Dowdesewell et al. 2006; Noormets et al. 2009
Filchner Trough	125	450	2.5	Convex-outward	Yes	Gullies	Gales et al. 2012; Larter et al. 2012.

[a]Measured at the shelf edge; [b]mean; TMF is Trough Mouth Fan.

2.3.1. West Antarctic Peninsula slope

The Marguerite Trough extends 370 km from Marguerite Bay to the shelf break, reaching a width of 40 km at the shelf edge (Table 2.1) (Ó Cofaigh et al., 2003; Noormets et al., 2009). The slope geometry at the mouth of the trough is uncommon in that the steepest interval is at ~800 m water depth, compared to that of Belgica Trough (Bellingshausen Sea), Pine Island West Trough (Amundsen Sea) and Filchner Trough (Weddell Sea), where the steepest interval is immediately down-slope of the shelf edge (Noormets et al., 2009). Either side of the trough mouth, the down-slope profile is concave, with upper-slope gradients of 6°, mid-slope gradients of < 14° and gradients of around 3° below this (Dowdeswell et al., 2004a).

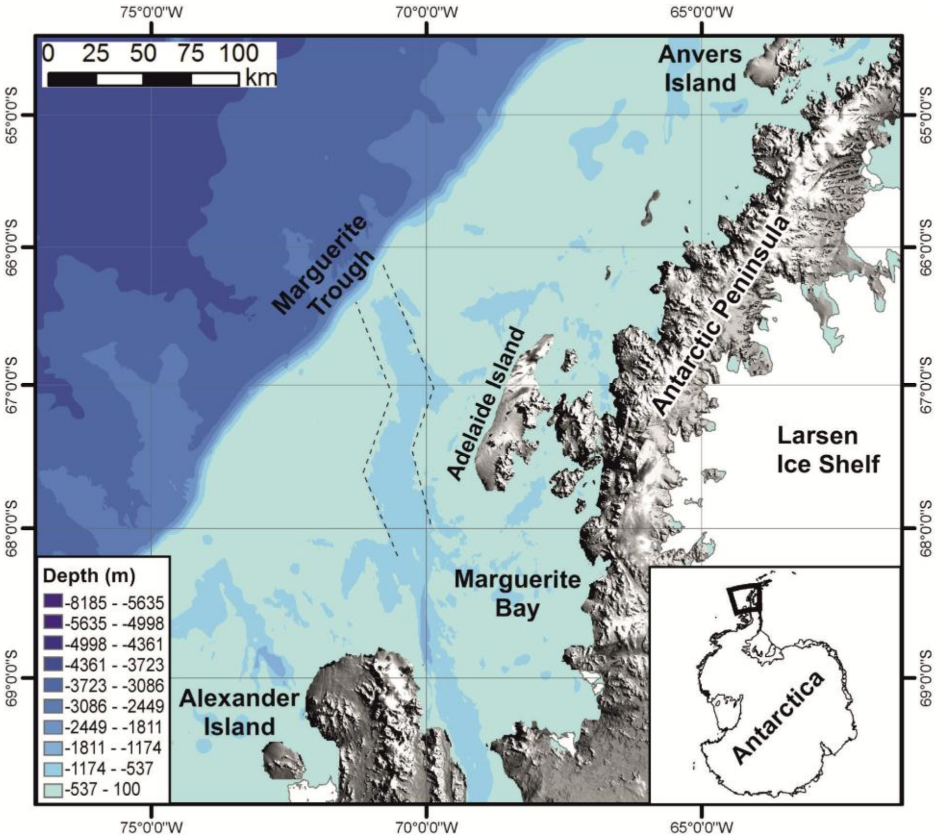


Figure 2. 2. Western Antarctic Peninsula. Dashed black lines indicate Marguerite Trough margins. Inset locates Fig. 2.2 on the Antarctic continent. Regional bathymetry data is from Bedmap2 (Fretwell et al., 2013). Antarctic continent is Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007).

No Trough Mouth Fan (TMF) is present at the Marguerite Trough mouth (Ó Cofaigh et al., 2003). The lack of TMF may result from the steep slope gradient, allowing sediment to bypass the upper-slope due to rapid down-slope sediment transport (Dowdeswell et al., 2004a). West of Anvers Island (Fig. 2.2), seismic reflection data show a change in facies down-slope from weak and chaotic seismic reflections to stronger and continuous reflections, suggesting that processes such as debris flows have acted to transport material down-slope (Larter and Cunningham, 1993).

Gullies are present at the mouth of the Marguerite Trough and on the upper slope (Ó Cofaigh et al., 2003), extending between 3 and 5 km down-slope and reaching amplitudes of > 200 m on either side of the trough and 120 m directly at the trough mouth (Dowdeswell et al., 2004a). The gullies are most developed on either side of the trough, corresponding to increased slope gradients (~11°, upper slope), compared to directly in front of the trough (~9°, upper slope) where the slope is smoother with comparatively fewer and shallower gullies present (Ó Cofaigh et al., 2003). Generally, the gullies do not incise the shelf edge and are characterised by V-shaped cross-sections (Noormets et al., 2009). Dowdeswell et al., (2004a) determined that gullies thalwegs become less well constrained as slope gradients increase toward the mid-slope.

Channel systems are present at the base of the slope along parts of the western Antarctic Peninsula (Dowdeswell et al., 2004a). The channels have amplitudes of 80 to 150 m and widths between 3 and 8 km (Tomlinson et al., 1992; Pudsey et al., 2000). Vanneste and Larter (1995) suggested that the transformation of debris flows to turbidity currents with distance down-slope results in the lack of connection between the upper slope gullies and the channels at the base of the slope, resulting in a continuous apron of debris flow deposits on the lower slope.

2.3.2. Bellingshausen Sea slope

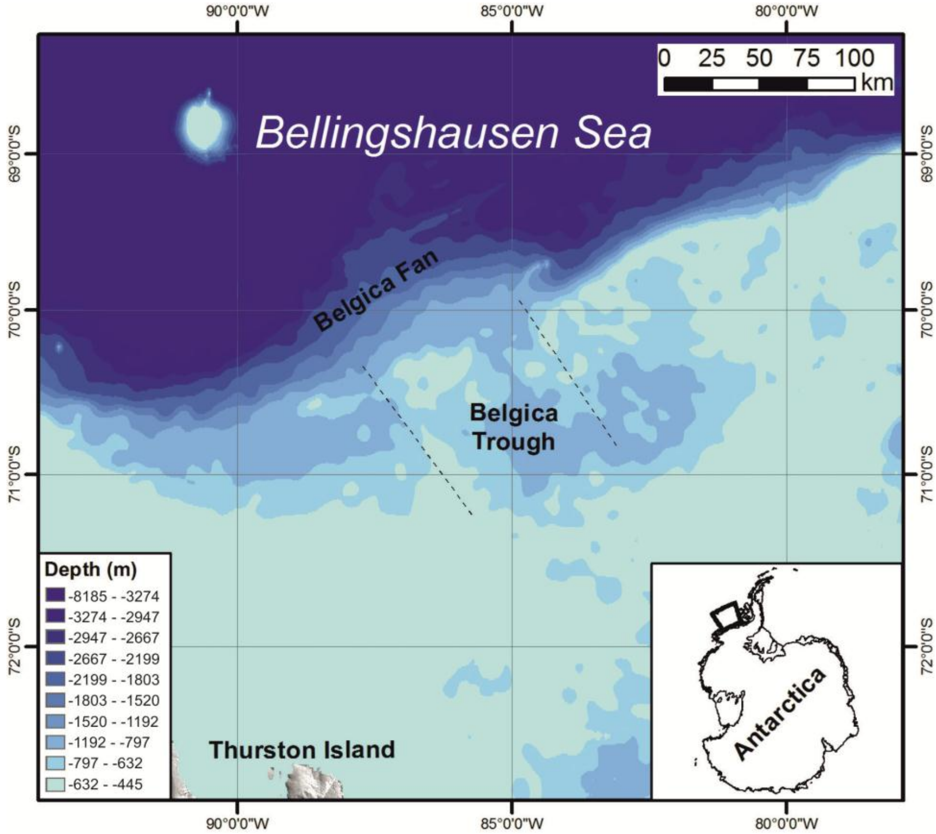


Figure 2. 3. Bellingshausen Sea. Dashed black lines mark Belgica Trough margins. Inset figure shows location on the Antarctic continent. Regional bathymetry data is from Bedmap2 (Fretwell et al., 2013). Antarctic continent is Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007).

The Belgica Trough is 125 km wide and extends 250 km inshore from the shelf edge (Table 2.1) (Ó Cofaigh et al., 2005b). Water depths at the shelf break are between 600 and 680 m inside the trough and between 400 and 500 m on the trough banks (Noormets et al., 2009). The continental shelf of the Bellingshausen Sea reaches a maximum width of 500 km (Dowdeswell et al., 2008). Seismic profiles show progradation of the upper slope and outer shelf due to repeated advances of ice during glacial periods (Nitsche et al., 1997; 2000; Dowdeswell et al., 2008). The presence of mega scale glacial lineations (MSGL) within Belgica Trough indicates that the LGM ice sheet grounded at least near to the shelf break (Ó Cofaigh et al., 2005b).

The gradient of the continental slope ranges from 1-2° at the mouth of the Belgica trough to 5° east of the trough (Dowdeswell et al., 2008). A trough-mouth fan (Belgica Fan) is present seaward of the Belgica Trough, marking a major depocentre in the region

(Dowdeswell et al., 2008). The Belgica Fan was largely formed by glaciogenic debris flows, which transported large amounts of sediment down-slope, leading to slope progradation (Nitsche et al., 1997; Ó Cofaigh et al., 2005b; Dowdeswell et al., 2008).

Gullies are present on the continental slope with most originating from the upper slope, but not significantly incising the shelf edge (Dowdeswell et al., 2008). Gully distribution is irregular, with heavily gullied areas alternating with gully-free regions (Noormets et al., 2009; Nitsche et al., 2000). A major gully system is located seaward of the Belgica Trough, with gullies measuring between 2 and 10 km long, around 750 m wide and 12 m amplitude (Noormets et al., 2009). Noormets et al. (2009) show that the gullies increase in mean gully spacing and size towards the margins of the trough. On the upper slope, the gullies generally form V-shaped cross-sections, transforming to U- or box-shaped down-slope (Noormets et al., 2009). Many of the gully systems show an aborescent pattern, forming stream orders of up to 4th order down-slope (Dowdeswell et al., 2008).

An area of gullies occurs to the east of the trough mouth (83°45'W) (Noormets et al., 2009). Gullies here are significantly straighter, narrower and deeper, and occur in higher densities (approximately 1 gully/km) and with a lower stream order (Dowdeswell et al., 2008) compared to gullies at the trough mouth. The gullies here do not incise the shelf edge and have V-shaped cross-sections which remain V-shaped with distance down-slope (Noormets et al., 2009).

A second outer shelf depocentre is present around 80°30'W (Noormets et al., 2009). Here, gullies incise an area of relatively steep slope, approximately 30 m deeper than the surrounding continental margin (Noormets et al., 2009). The gullies are significantly larger than the gullies on the TMF, with amplitudes reaching > 50 m and with gullies incising the shelf edge by 2.5 km (Noormets et al., 2009). The gullies have complex head shapes with diameters < 1.4 km and with secondary incisions within some gully heads (Noormets et al., 2009).

Gullies are observed initiating from small slide scars (Dowdeswell et al., 2008) and from small shelf edge depressions (e.g. at 81°30'W) (Noormets et al., 2009). Convex-outward isobaths seaward of the slide scars indicate debris flow deposits. Other bedforms such as potential braided meltwater channels, shelf-parallel ridges (at 80°W) and iceberg keel marks are also identified at the shelf edge (Dowdeswell et al., 2008; Noormets et al., 2009). Low sinuosity channels occur on the lower slope of the TMF and can be traced for over 60 km (Dowdeswell et al., 2008). Acoustic backscatter imagery indicates that the channels are floored by sediment that is coarser than the surrounding channel levees and continental slope (Noormets et al., 2009).

2.3.3. Amundsen Sea slope

The Amundsen Sea continental shelf is approximately 500 km wide with depths ranging from 1500 m in Pine Island Bay to 500 m at the shelf edge (Nitsche et al., 2007; Noormets et al., 2009). Bathymetric data show that the shelf is dissected by multiple cross shelf troughs, including the Pine Island West (PIW) Trough at 114°W (Table 2.1) (Nitsche et al., 2007; Noormets et al., 2009). The presence of streamlined bedforms within the troughs indicate that the ice sheet reached the outer shelf during glacial maxima (Lowe and Anderson, 2002; Evans et al., 2006; Graham et al., 2010). Average slope gradient is 4° at the mouth of PIW trough, increasing to 7° on either side of the trough (Dowdeswell et al., 2006).

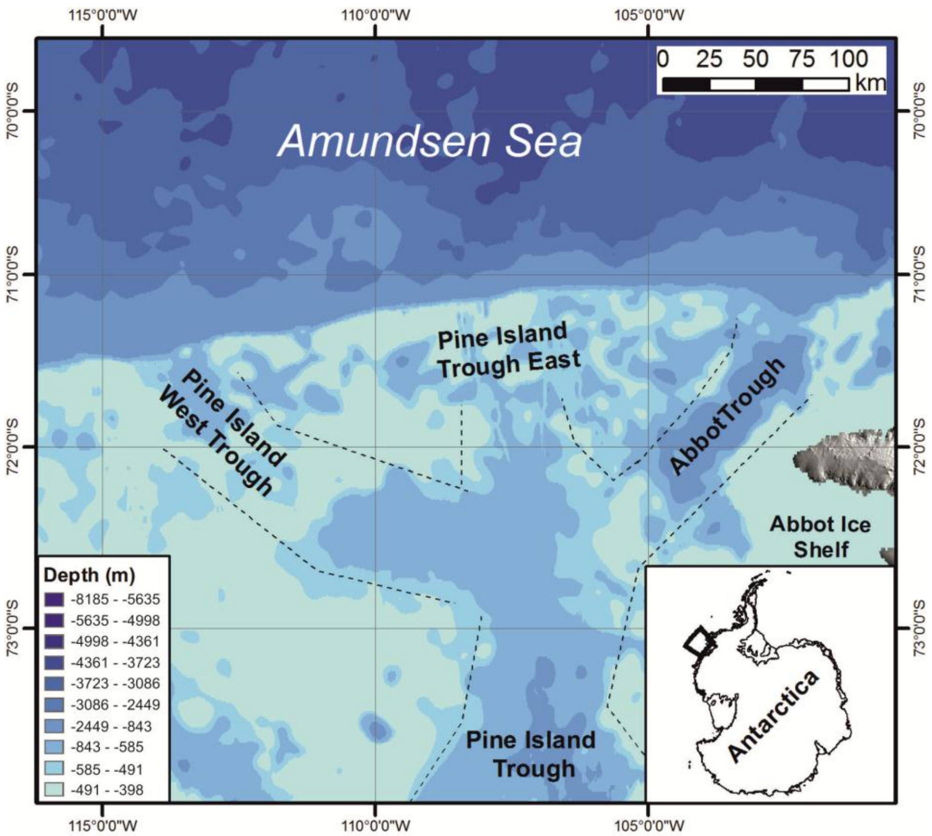


Figure 2. 4. Amundsen Sea. Dashed lines show margins of Pine Island Trough, Pine Island West Trough and Pine Island Trough East. Inset shows location on the Antarctic continent. Regional bathymetry data is from Bedmap2 (Fretwell et al., 2013). Antarctic continent is Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007).

Gullies are irregularly distributed along the Amundsen Sea shelf break, with gully size increasing toward the margins of the PIW Trough (Noormets et al., 2009). At the mouth of PIW Trough, gullies incise the upper slope with average widths of 630 m, lengths of < 10 km and incision depths of over 50 m (Noormets et al., 2009). The gullies are relatively straight, with some forming branching systems (up to 4th order) and transforming into channels down-slope (Dowdesewell et al., 2006). The gullies originate on the outer shelf and incise the shelf edge in some places by up to 3 km, leading to a highly sinuous shelf edge (Noormets et al., 2009). Gullies further east, outside of the trough margins, are straighter with no connections to channels down-slope (Dowdeswell et al., 2006).

Channels occur on the lower slope leading to the continental rise (Dowdeswell et al., 2006). The channels are relatively sinuous, < 1 km wide and around 40 m deep (Dowdeswell et al., 2006). The channel morphologies are not uniform along the slope, with channels becoming wider, straighter and with larger banks eastward of the trough mouth (between 108°W-111°W) (Dowdeswell et al., 2006). Sediment core data indicate that zones of turbidity current deposition are present at the base of the continental slope and rise and on the abyssal plain (Dowsdeswell et al., 2006).

2.3.4. Weddell Sea slope

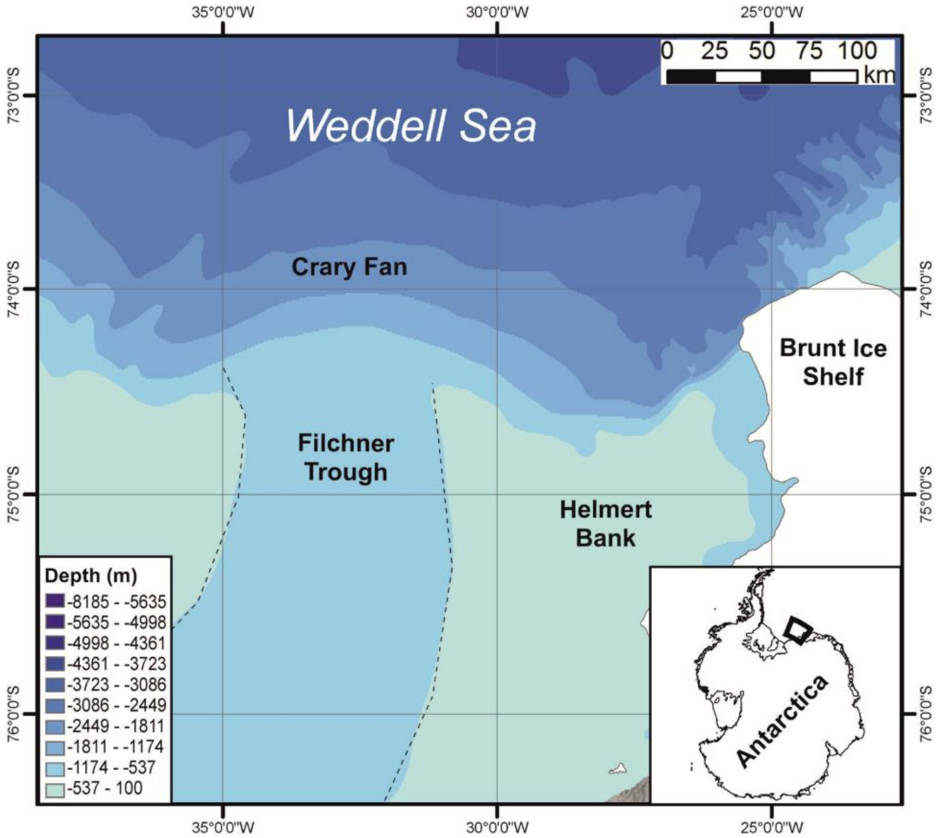


Figure 2. 5. Weddell Sea. Dashed lines show margins of Filchner Trough. Inset shows location on the Antarctic continent. Regional bathymetry data is from Bedmap2 (Fretwell et al., 2013). Antarctic continent is Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007).

The southern Weddell Sea shelf is incised by the Filchner and Ronne cross shelf troughs, which are separated by Berkner Bank. The troughs have landward dipping bathymetric profiles with a trough mouth fan present at Filchner Trough mouth (Crary Fan; Table 2.1) (Bart, 1999). Filchner Trough has a width of 125 km at the shelf edge and length of > 450 km (Larter et al., 2012). The shelf edge occurs at approximately 630 m water depth, with Filchner Trough reaching maximum water depths of 1200 m (Hillenbrand et al., 2012). The southern Weddell Sea upper slope has a gradient of 1.6º decreasing to 0.7º down-slope (Melles et al., 1994). The southeastern Weddell Sea has generally steeper slope gradients (15º average) (Bart et al., 1999).

Large channel-ridge systems are observed on the lower slopes of the southern and southeastern Weddell Sea (Weber et al., 1994). The channels have amplitudes of ~300 m and occur in water depths between 2000 and 3300 m (Melles et al., 1994; Weber et al., 1994). Surface seafloor sediments on the outer shelf consist mainly of sand and gravelly sand due to winnowing of fine grained particles by cold, dense water overflow (Melles et al., 1994). Cores from the outer shelf recovered glaciomarine and subglacial sediments, including over-consolidated deposits and diamictons (e.g. Elverhøi, 1984; Melles and Kuhn, 1993; Hillenbrand et al., 2012). Prior to this study, the morphology of the continental shelf edge and upper slope in the southern and southeastern Weddell Sea were poorly known.

2.4. Slope Processes

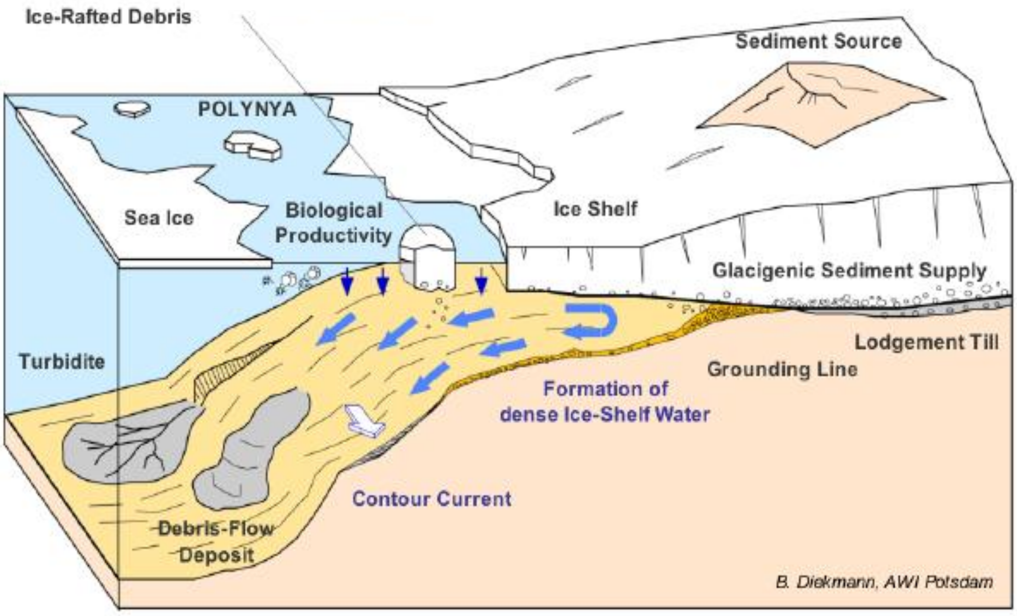


Figure 2. 6. Processes operating on Antarctic submarine slopes. Taken from http://www.awi.de/de/forschung/fachbereiche/geowissenschaften/periglazialforschung/research_themes/limnogeology_palaeoclimate/east_antarctica_prydz_bay/

2.4.1. Submarine slopes

The processes that form complex continental slope morphologies and the factors that influence these processes are not well understood or well constrained. Processes influencing slope morphology include: (1) sedimentary processes, such as mass flows (slides, slumps, debris flows, turbidity currents); (2) oceanographic processes such as geostrophic currents, tides and cascading dense water, formed during sea-ice formation through brine rejection (Kuvaas and Kristoffersen, 1991; Dowdeswell et al., 2006; Noormets et al., 2009; and (3) glacigenic processes, such as subglacial meltwater discharged from beneath an ice sheet or released by sub-glacial lake discharge, iceberg grounding and rapid transport of sediment by ice streams to the shelf edge (Larter and Cunningham, 1993; Vanneste and Larter, 1995; Dowdeswell and Bamber, 2007; Goodwin, 1988; Wellner et al., 2001; Dowdeswell et al., 2006, 2008; Fricker et al., 2007; Piper et al., 2012). These processes may be influenced by factors such as local slope character (slope geometry, gradient), large-scale spatial characteristics (i.e. drainage basin size, location of cross-shelf troughs), ice-sheet history and sediment yield (chapter 2.4.4) (Noormets et al., 2009; Peakall et al., 2012; Gales et al., 2013a). Deciphering the extent that processes and factors influence slope morphology and the time-scales they occur over remains a major challenge.

2.4.2. Mass flows

Mass wasting by the failure of sediment at the shelf break or continental slope results in the transfer of sediment down-slope due to gravity. Mass flows, such as slides, slumps, debris flows and turbidity currents are able to transfer huge volumes of sediment over hundreds of kilometres.

2.4.2.1. Submarine slides

Bates and Jackson (1987) define a slide as: "a mass-movement or descent resulting from failure of earth, snow or rock under shear stress along one or several surfaces that are either visible or may reasonably be inferred". Slides are known to have occurred in the geological past (i.e. Miocene, early Pliocene) on the Antarctic margin; however, there are few modern examples (Barker and Austin, 1998; Imbo et al., 2003; Diviacco et al., 2006). In contrast, slides are abundant on other high latitude slopes, such as on the northern Norwegian and southwest Barents Sea margin (Damuth, 1978; Bugge et al., 1988; Laberg and Vorren, 1993; 2000; Laberg et al., 2000; Dowdeswell et al., 1996; Vorren et al., 1998; Evans et al., 2005; Nielsen et al., 2005). The difference between Arctic and Antarctic slope instability has been attributed to variations in: (1) pore pressure, with Antarctic slopes displaying a greater stability due to ice sheet compaction (Prior and Coleman, 1984; Larter and Barker, 1991); (2) episodic and high quantities of sediment delivered to the shelf edge which may lead to undercompaction and over-steepening (Dowdeswell et al., 2002); (3) the timings of particular stages of glaciation (Nielsen et al., 2005); and (4) the presence of weak contouritic/ hemipelagic or other layers present on many Arctic continental margins (Long et al., 2003; Nielsen et al., 2005; Laberg and Camerlenghi, 2008).

2.4.2.2. Debris flows

Debris flows are dense flows where particles are held in suspension by a viscous matrix, formed by slope failure (Middleton and Hampton, 1973). The formation of debris flows is dependent upon the type of material, the external trigger and the amount of energy involved (Locat and Lee, 2005). Debris flow deposits can extend > 100 km, or may be rapidly diluted until they transform into turbidity currents (Laberg and Vorren, 1995; Dowdeswell et al., 1996). Transformation from a debris flow to a turbidity current may result from mixing of water at the sediment interface or through the initiation of flow turbulence (Hampton, 1972; Allen, 1971) and may be enhanced by hydroplaning, where sediment suspension rises by increasing velocity of the debris flow head (Mohrig et al., 1998). Van Andel and Komar (1969) suggest that a hydraulic jump may form at the base of the continental slope, causing energy to be released from the debris flow as turbulent suspension, generating a turbidity current.

2.4.2.3. Turbidity currents

Turbidity currents are defined as sediment-gravity flows where sediment is suspended by fluid turbulence (Sanders, 1965). Turbidity currents require an initiation mechanism and a suitable gradient to maintain the flow (Normark and Piper, 1991); although studies by Wynn et al. (2012) show that on gradients as low as < 0.1°, flows can travel over > 1000 km. Initiation processes include any mechanism which suspends sediment, including subglacial meltwater released from beneath an ice sheet, storm resuspension, debris flows mixing with ambient water, tectonic influences, iceberg scouring and small-scale slope failure (Piper, 2005). Turbidity currents may cause large-scale slope erosion, resulting in sediment bypassing the slope and delivering sediment to the deep ocean.

2.4.2.4. Triggering mechanisms

2.4.2.4.1. Rapid transfer of sediment to the slope

A major influencing factor on the Antarctic continental margin is the rapid transport of unsorted sediment to the shelf edge by ice streams. This results in a region prone to failure, that a trigger, or combination of triggering mechanisms, may destabilise. Rapid sediment transport to the shelf edge may lead to slope over-steepening and/or excess pore pressure, whereby the effective normal stress and shear strength of the sediment is reduced (Terzaghi, 1962). As the sediment builds up, shear stress and slope gradient increases, which may lead to slope failure (Locat and Lee, 2005). Studies have shown an inverse relationship between slope failure size and slope gradient, where low slope gradients occur due to instability (McAdoo et al., 2000). Winnowing of fine-surface sediments on some high latitude margins have also been shown to stabilise slopes by leaving behind mainly coarse, but poorly sorted material, allowing high-slope gradients to be maintained (Larter and Barker, 1989).

2.4.2.4.2. Tectonic influences

Tectonic influences, including earthquakes and tsunamis, may trigger slope failure (e.g. Piper et al., 1999). Earthquakes cause an increase in shear stress and increase in pore pressure in the underlying sediment due to ground tremors, resulting in sediment transforming to a more fluid-like texture and increasing the susceptibility for failure (Locat and Lee, 2005). Large waves, such as tsunamis, can increase the pore pressure in sediment, causing liquefaction if high enough pressures are reached (Hampton et al., 1996).

2.4.2.4.3. Tidal pumping

Tidal pumping may suspend sediment and initiate mass flows such as turbidity currents through pumping in and out of sub-glacial cavities beneath ice shelves. As ice shelves are floating extensions of ice sheets, the shelves oscillate in response to the tidal regime (Holland, 2008). Turbulence may be generated by the velocity shear, caused by the tidal oscillations (Holland, 2008). As the sub-glacial cavity beneath the ice sheet shallows toward the grounding line, tidal velocity and mixing increases as a result of the continuity effect (MacAyeal, 1984). Similar processes can be expected to occur beneath large, grounded icebergs.

2.4.2.4.4. Ice scour

Ice is able to scour channels or furrows into the seafloor, influencing sedimentation rates, sediment transport and potentially initiating mass flows of resuspended sediment (Piper et al., 2005). Ice scour can result from grounded ice-sheet erosion or from calved icebergs. Grounded ice sheet erosion can produce mega-scale glacial lineations which consist of multiple parallel and elongate ridges and grooves, with average dimensions of ~40 m depth and wavelengths transverse to flow of 300-1000 m (Canals et al., 2000; King et al., 2009). Scour by icebergs is more random, average incision depths of 8 m and widths of < 750 m (Pudsey et al., 1994; Kuijpers et al., 2007).

2.4.2.4.5. Resuspension by currents

Resuspension by shelf and slope currents may influence sedimentation and has the potential to trigger mass flows. South-westward flowing bottom currents are documented on the Pacific margin of the Antarctic Peninsula upper continental rise (McGinnis and Hayes, 1995; Rebesco et al., 1996; Giorgetti et al., 2003), with current meter measurements at 8 m above the seabed measuring mean current velocities of 6 cm s^{-1} (Camerlenghi et al., 1997; Giorgetti et al., 2003). As currents with velocities > 6 cm s^{-1} are able to transport silt and clay sized particles (Young and Southard, 1978; Singer and Anderson, 1984), bottom currents from the Antarctic margin may influence surface sediments by transporting fine grained sediment (Camerlenghi et al., 1997). Episodic high energy currents, for example produced by the sporadic presence of barotrophic eddies, will increase shear stress and may cause short-term increases in sediment resuspension (Giorgetti et al., 2003).

2.4.2.4.6. Gas hydrate dissociation

Gas hydrates are formed of hydrogen-bonded water molecules which form ice-like structures (Sloan, 1998). Under intermediate pressure and low temperatures, these substances are solid and stable, however an increase in water temperature and/or decrease in pressure can cause the gas hydrates to destabilise and dissociate into fluid and gas, leading to slope instability (Sultan et al., 2003). The presence of gas hydrates may also reduce the permeability of sediment present, leading to accumulations of gas beneath the sediment which can lead to under-consolidated sediment prone to failure (Sultan et al., 2003). Although some studies have identified potential gas hydrate reservoirs on parts of the Antarctic margin (e.g. Lodolo and Camerlenghi, 2000; Lodolo et al., 2002; Jin et al., 2003; Tinivella et al., 2008; Geletti and Busetti, 2011), the distribution and incidence of gas hydrates is still poorly known around much of the Antarctic margin (Maslin et al., 2010).

2.4.2.4.7. Isostatic rebound

Isostatic rebound occurs during deglaciation and interglacial periods due to glacial unloading, where a decrease in ice-volume may result in a decrease in overburden pressure on sediments present (Bart et al., 1999). This has the potential to induce extension on the shelf and slope, which may influence slope instability (Bart et al., 1999). However, it is unlikely that ice on the outer shelf was ever thick enough for isostatic rebound to be significant in these areas. The rise in sea-level from glacial to interglacial conditions of ~120 m (IPCC, 2007), may have resulted in a small increase in pressure over many outer shelf and slope areas.

2.4.2.4.8. Weak geological layers

Weak layers within the underlying geology are common on high latitude continental margins and are suggested to be a critical factor in the initiation of the Storegga, Trænadjupet and Afen slides (Bugge et al., 1983; Wilson et al., 2003; Canals et al., 2004). On the Antarctic margin, weak layers are generally thought to be very thin or absent. The distribution of weak layers is controlled by: (1) the oceanographic circulation, where contour currents deposit fine sediments, which form weak contouritic/ hemipelagic layers (Long et al., 2003; Nielsen et al., 2005; Laberg and Camerlenghi, 2008); and (2) changes in climate from a glacial to inter-glacial setting which influences the supply of fine grained sediment, which comes mainly from fluvial discharge and microbiological productivity (Bryn et al., 2005). Weak layers are usually water-saturated, fine grained sediments, which are interbedded with coarser grained, poorly sorted and glacially derived sediment (Masson, 2006). Sedimentation on top of low permability (i.e. clay-rich) weak layers, such as by rapid sediment transfer to the shelf edge during glacial periods, may cause increased pore pressures within the weak layers, leading to increased slope instability (Masson, 2006).

2.4.3. Subglacial meltwater discharge

Recent studies have shown that subglacial meltwater is present and may be highly mobile beneath Antarctic ice sheets (Siegert, 2000; Engelhardt et al., 1990; Fricker et al., 2007; Wingham et al., 2006; Bell et al., 2008; Piper et al., 2012). Subglacial meltwater may be discharged from ice sheet grounding lines during glacial maxima through either of two processes. (1) A continuous supply, where basal water is generated predominantly through geothermal and strain heating, as there was little or no ice-surface melting during full-glacial times, or indeed, on the modern Antarctic ice sheet, to supply the basal drainage system. Typical yields from geothermal heating are, however, in the range of mm/yr of basal melt (e.g. Beem et al., 2010; Pattyn, 2010), making it difficult to sustain continuous discharges over long periods. (2) Episodic water release such as subglacial lake discharges (‘glacial outburst floods’ or ‘jökulhlaups’) (Goodwin, 1988; Wellner et al., 2001; Dowdeswell et al., 2006, 2008; Fricker et al., 2007; Bell, 2008; Noormets et al., 2009; Piper et al., 2012). Mechanisms of meltwater transport are poorly understood and widely debated (Raymond, 2002), with studies suggesting transport through either the top layer of soft sediment as 'Darcian flow' (Tulaczyk et al., 1998), as surface sheet flow (Boulton et al., 2007), via conduits incised into the ice above, known as Rothlisberger 'R' channels or via conduits incised into the bed, known as Nye 'N' channels or 'tunnel valleys' (Ó Cofaigh et al., 1996; Carter et al., 2009).

Subglacial meltwater is able to entrain sediment at the base of an ice sheet and may produce hyperpycnal flows when discharged at the grounding line (Russell and Knudsen, 1999a, b, 2002). Meltwater is buoyant, consisting of cool and fresh water, therefore, enough sediment must be entrained to increase the density of the water, allowing the flow to remain at the seafloor. The critical sediment concentration needed for meltwater to initiate a hyperpycnal flow in seawater is 1–5 kg m^{-3} (Parsons et al., 2001; Mulder et al., 2003). This value is considerably lower than previous estimates of sediment concentrations (e.g. 33 kg m^{-3}; Syvitski, 1989) which are based on buoyancy considerations and do not take into account the effects of fine-scale convective instability (Parsons et al., 2001; Mulder et al., 2003).

Large palaeo-subglacial drainage systems have been documented on the inner Antarctic continental shelf and in onshore areas that are presently ice free including the Palmer Deep (Antarctic Peninsula Margin), Marguerite Bay, Amundsen Sea, on the mid-shelf of the western Ross Sea and beneath the Rutford Ice stream, western Weddell Sea (Sugden et al., 1991; Anderson and Shipp, 2001; Lowe and Anderson, 2002; 2003; Ó Cofaigh et al., 2002; King et al., 2004; Denton and Sugden, 2005; Domack et al., 2006; Wellner et al., 2006; Lewis et al., 2006; Anderson and Oakes-Fretwell, 2008; Larter et al., 2009; Nitsche et al., 2013). The size of meltwater channels documented on the Antarctic continental shelf are in the range of < 2 km wide, up to 300 m deep and < 15 km long (Domack et al., 2006; Ó Cofaigh et al., 2002; Lowe and Anderson, 2002). Very few channel systems have been observed on well-preserved former sub-glacial beds in the outer shelf part of cross-shelf troughs.

2.4.4. Cascading flows of cold, dense water

Dense bottom water formed from sea-ice freezing and brine rejection, and the cascading of this bottom water down-slope may influence slope morphology (Kuvaas and Kristoffersen, 1991; Dowdeswell et al., 2006, 2008; Noormets et al., 2009). Polynyas are areas of sea-ice production that form due to katabatic winds generated by air sinking from ice sheets inshore (Bromwich and Kurtz, 1984). Winds cause sea-ice to be blown offshore, resulting in the continuous production of sea-ice, and therefore formation of cold, dense water.

The Weddell Sea is the largest source of Antarctic Bottom Water (AABW), which is exported from the Southern Ocean and forms a major component of the global thermohaline circulation (Nicholls et al., 2009). High Salinity Shelf Water (HSSW), produced in the Weddell Sea, is supercooled and freshened by circulation beneath the ice shelves, producing cold and dense Ice Shelf Water (ISW). Temperature profiles of the water column west of the Antarctic Peninsula and in the Bellingshausen and Amundsen seas preclude bottom water production at the present day (Hoffman and Klinck, 1998; Smith et al., 1999; Smith and Klinck, 2002; Dinniman and Klinck, 2004).

The production of dense water is more effective in shallower waters allowing HSSW to build up and eventually spill over the shelf, forming down-slope flows of cold and dense plumes. The HSSW forms a gravity flow (cascade), driven by the salinity and temperature contrast and may entrain sediment down-slope (Muench et al., 2009; Caburlotto et al., 2010). Mean flow velocities of 0.38 m s^{-1} have been measured on the upper slope at the mouth of Filchner Trough, southern Weddell Sea (10 m above the seabed), increasing to 1 m s^{-1} down-slope (Foldvik et al., 2004).

2.4.5. Factors influencing slope processes

Environmental controls and local slope characteristics may influence the effect processes have on the surface morphology and include the slope geometry and gradient, underlying geology, drainage basin size and glacial history.

2.4.5.1. Slope geometry and gradient

Wynn et al., (2012) showed that slope gradient can affect the ability of gravity flows to erode, transport and deposit sediment. Increased slope gradients may lead to slope incision, flow confinement and channel narrowing, whereas decreased slope gradients may lead to a decrease in incision and flow constrainment, widening of channels and deposition of load (Friedmann et al., 2000). Subtle changes in slope gradient (~0.05°) have been shown to cause significant changes in flow character (Wynn et al., 2012). In areas with increased slope gradients, sediment may by-pass the slope by transforming from unconfined to confined flows within channels (Ó Cofaigh et al., 2003; Wynn et al., 2012). In regions of lower slope gradient, trough mouth fans are more likely to develop through sediment deposition and debris flow processes (Ó Cofaigh et al., 2003). Alternatively, the low angle slopes may be the result of growth of the trough mouth fans, which in turn result from the nature and rate of delivery of sediments to those parts of the margin.

2.4.5.2. Slope geology

Variation in the near-surface sediments has the potential to affect surface morphology by influencing the threshold for erosion and erosion rate of the surface. The Antarctic continental margins considered in this paper are all underlain by thick Quaternary sediments, most of which are glacially derived (e.g. glacigenic debris flow deposits) or glacially influenced (e.g. glacimarine muds with ice rafted debris) (Wright and Anderson, 1982; Melles and Kuhn, 1993; Bonn et al., 1994; Dowdeswell et al., 2004a, 2004b, 2006; Hillenbrand et al., 2005). Seismic reflection and coring studies from the Antarctic continental slope show that sediments present along the slope exhibit a limited range of characteristics, with similar lithology, physical properties and grain-size composition (e.g. Vanneste and Larter, 1995; Anderson and Andrews, 1999; Michels et al., 2002; Dowdeswell et al., 2004a, 2006, 2008; Hillenbrand et al., 2005; 2009). As the outer-shelf and upper-slope sediments are relatively homogenous in this study, variation in substrate is therefore not likely to be a major factor influencing differences in surface morphology.

2.4.5.3. Drainage basin size

The drainage basin size may influence flow characteristics by largely controlling the volume of the flows, such as the volume of subglacial meltwater released from beneath an ice sheet and the frequency that the flows are released. A larger drainage basin area to grounding line length ratio can therefore be expected to result in a greater rate of delivery of sediment to the grounding line. The volume and frequency of the flow, mean excess density and vertical relief are also factors which influence flow behaviour and erosive ability, such as the bed shear stress (Wynn et al., 2002).

2.4.5.4. Glacial history

The location of cross-shelf troughs, and thus palaeo-ice streams, influences sediment delivery to the shelf edge during glacial periods. The location of cross-shelf troughs may also influence the location where subglacial meltwater is released from beneath an ice sheet due to subglacial pressure gradients created by the ice sheet. Subglacial water pressure gradients beneath ice and additional heat generated by friction at the trough margins focus subglacial meltwater toward the trough margins (Raymond, 2002; Noormets et al., 2009). Under ice streams, pressure at the ice-bed interface is lower than under inter-stream ridges due to the lesser ice thickness (Röthlisberger, 1972) causing meltwater under the slower moving ice at the banks of an ice-stream to be drawn down to the trough margins (Boulton et al., 2007; Vaughan et al., 2008). Thicker ice also commonly occurs in the central part of ice streams, resulting in subglacial pressure gradients that cause subglacial water to flow toward the ice stream margin (Dowdeswell and Elverhøi, 2002). Flow discharge may, therefore, have been higher within the troughs and particularly near the troughs margins.

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Chapter 3.

Materials and Methodology

3.1. Introduction

The interpretation of Antarctic submarine geomorphology depends upon data availability and the type and resolution of the data. Data is restricted in Antarctica due to surveying constrains such as the presence of ice, the remoteness and associated costs, which limit our ability to analyse and interpret the seafloor. The resolution of the data limits the scale of bedforms which can be identified, making the interpretation of small-scale bedforms (< 20 m) difficult. In the literature, it is the lack of high-resolution data, or deep-towed surveys, which limit the extent of geomorphological interpretation, rather than the lack of features present.

3.2. Data and coverage

Multibeam bathymetric data and TOPAS (Topographic Parametric Sounder) subbottom profiler data used in this study were acquired over several cruises (Table 3.1). The data cover > 2126 km along the shelf edge and upper slope of the Antarctic continental margin. For Chapter 6, multibeam bathymetric data from the Arctic were analysed, covering 520 km of continental margin from the northern Norwegian Sea, southwest Barents Sea and western Svalbard margin. Details on the data used are provided in Table 3.1.

Multibeam bathymetric data were collected using a hull-mounted Kongsberg-Simrad EM120 multibeam echosounder with a frequency range of 11.75-12.75 kHz, swath width of up to 150° and a 191 beam array with real-time beam steering and active pitch and roll compensation (RRS *James Clark Ross;* RVIB *Nathaniel B. Palmer),* an Atlas Hydrosweep DS-2 System with a frequency of 15 kHz (RV *Polarstern*), a Kongsberg EM300 echosounder with a frequency of 30 kHz (R/V *Jan Mayen*), a SeaBeam 2112 system with a frequency of 12 kHz (RVIB *Nathanial B Palmer* until 2002), and a Kongsberg EM300 multibeam echosounder with a frequency of 30 kHz (R/V *Jan Mayen*). Navigation data were acquired using Global Positional System receivers. The simrad data were processed by removing erroneous pings and applying a correction for sound-velocity profiles. Sounds velocity profiles are used to model beam ray paths and to determine the path lengths in order to calculate the locations of reflection points. These calculations are carried out in near real time by modern multibeam systems. Vertical measurement accuracy for the multibeam data is 0.2% of the root mean square depth and is in the range of < 1 m at 500 m water depth (for soundings between nadir and $\pm$60°), increasing to > 10 m for soundings at $\pm$70° (de Moustier, 2001). Horizontal resolution varies with ship speed, water depth, beam angle, track spacing and seabed topography, with typical values of 10 to 20 m at 1000 m water depth.

TOPAS data are used to image the sub-surface and were collected using a Kongsberg TOPAS PSO18 system. The TOPAS system works simultaneously alongside the multibeam bathymetric system and the transmission can be configured to any individual frequency, such as ‘burst’ (e.g. 2.8 kHz), or range of frequencies, such as ‘chirp’ (e.g. 2.5-5.5 kHz), within a wide range. The system can resolve the sedimentary layer thicknesses to < 1 m and penetrate to depths $\geq$ 50 m below the seafloor in fine grained sediments. TOPAS data were converted to standard seismic format (SEG-Y) whilst on board.

Table 3. 1. Data used within thesis.

Data set		Reference
Cruise / ID	**Year**	
ANT23-4	2006	Gohl (2006); Nitsche et al. (2007)
JR59	2001	Ó Cofaigh et al. (2002); Dowdeswell et al. (2004)
JR71	2002	Ó Cofaigh et al. (2005a); Dowdeswell et al. (2004)
JR84	2003	Evans et al. (2006); Dowdeswell et al. (2006)
JR97	2005	Gales et al. (2012); Larter et al. (2012)
JR104	2004	Ó Cofaigh et al. (2005b); Dowdeswell et al. (2008)
JR141	2006	Noormets et al. (2009); Graham et al. (2010)
JR157	2007	Noormets et al. (2009)
JR179	2008	Graham et al. (2010)
JR244	2011	Gales et al. (2012); Larter et al. (2012)
NBP9902	1999	Wellner et al. (2001, 2006); Lowe and Anderson (2002)
NBP0001	2000	Nitsche et al. (2007)
NBP0103	2001	Bolmer (2008)
NBP0104	2001	Bolmer (2008)
NBP0201	2002	Wellner et al. (2006)
NBP0202	2002	Bolmer (2008)
NBP0702	2007	Nitsche et al. (2007); Graham et al. (2010)
R/V Jan Mayen	2004/5	Laberg et al. (2007)
R/V Jan Mayen	2010	Baeten et al. (2013)
R/V Jan Mayen	2006/7	Hustoft et al. (2009)
R/V Jan Mayen	2008/9	Forwick et al. (2009a, b)
MAREANO		www.mareano.no

3.3. Methodology

3.3.1. Data processing and interpretation

Multibeam data were processed and gridded using public-access MB-system software (Caress and Chayes, 2003) with cell sizes of 20 m x 20 m and 50 m x 50 m. The grids were extracted and imported into the software ArcGIS, where they were converted to raster grids. TOPAS Kongsberg software was used to visualise the TOPAS data. Limited seismic data were analysed using Promax software and Seismic Unix. Backscatter data were produced using FM Geocoder (Fonseca and Calder, 2005) where the strength of backscattering is dependent upon sediment type, grain size, survey conditions, bed roughness, compaction and slope (Blondel and Murton, 1997).

3.3.2 Quantitative analysis of gully parameters

The slope geomorphology was quantitatively analysed using standard geographic information system (GIS) tools by: (1) extracting profiles parallel to the shelf edge at a range of depths down-slope, along which gully parameters were measured; (2) extracting profiles across gullies to measure cross-sectional shape; and (3) analysing gully density and spatial patterns. Visualisation tools within ArcGIS were used to analyse the general slope character, including slope gradient, aspect (direction of slope facies) and roughness. Cross-shelf profiles were taken at 50 m below the shelf edge and down-slope along the gully length. The gully parameters measured are summarised in Table 3.2.

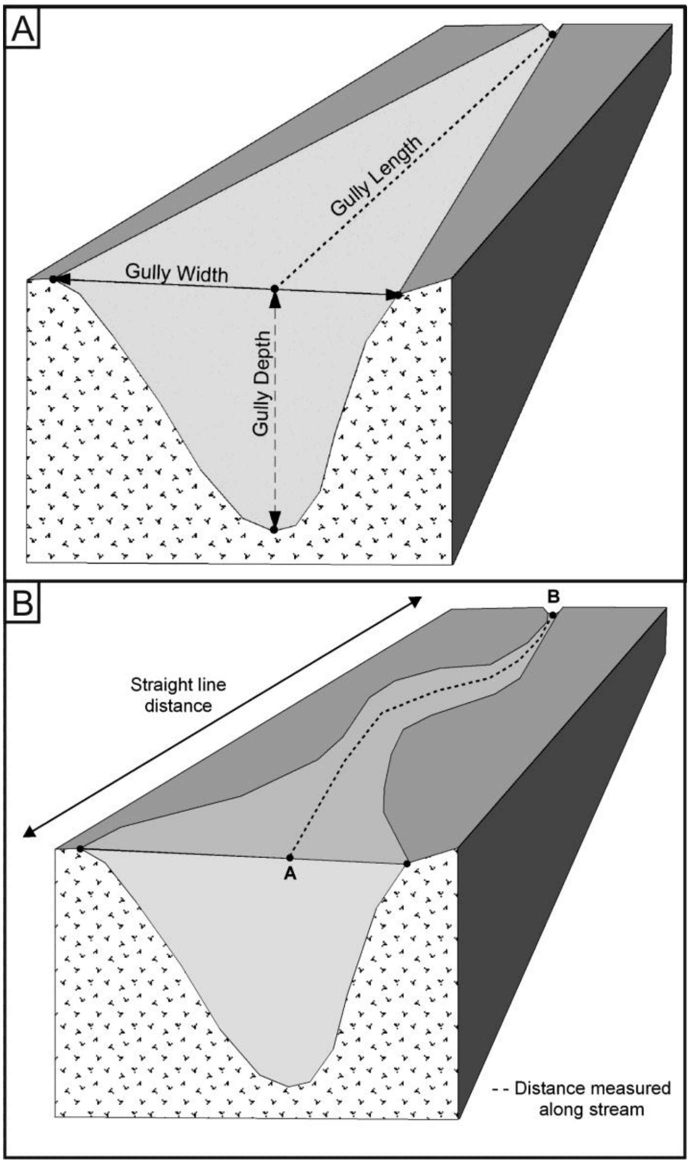

Figure 3. 1. Gully morphometric parameters. **A.** Gully width, length and incision depth. **B.** Gully sinuosity. Adapted from Gales et al., (2013).

Table 3. 2. Description of measured gully parameters from multibeam bathymetric data.

Parameter	Description
Length	Distance a gully can be traced down-slope from gully head. For some gullies (approximately 10% of the gullies measured), this is limited by down-slope data extent (Fig. 3.1).
Width	Distance between points of maximum curvature of gully flanks (Fig. 3.1).
Incision depth	Vertical distance from gully base to line defining gully width (Fig. 3.1).
Width/depth ratio (W/D)	Dimensionless ratio giving an indication of the gully shape.
Branching order	Classification of reaches of a stream network in terms of the number of up-stream tributaries, with order-1 streams being those with no tributaries. Strahler's (1957) method is used to calculate gully stream-order, by assigning each reach of a gully an order of 1. Where two 1st order reaches meet, the down-stream reach is assigned 2nd order. Where two 2nd order reaches meet, the down-stream reach is assigned 3rd order, and so on. Where two reaches with different orders meet, the down-stream reach is assigned the higher order of the two contributing reaches. The gullies were assigned orders in the same way using the automated Stream Order application in ArcGIS 9.3 (hydrology tools), in practice similar to that used by Pratson and Ryan (1996).
Cut-back	Indentation of the shelf edge by gullies. Also referred to as 'shelf incision'.
Slope gradient	General gradient angle measured from the continental slope.
Density	Number of gullies per km along slope (gully/km).
Sinuosity (S)	Dimensionless ratio of length along gully thalweg divided by the linear distance where: Sinuosity index (S) = distance measured along gully / straight line distance. (3.1) We define a sinuosity index greater than 1.04 as sinuous and, by definition, a straight gully would have S = 1 (Schumm, 1963). These values are based on the data available to this study and, therefore, they may not be appropriate for classifying gullies on all Antarctic continental slope (Fig. 3.1)
Cross-sectional shape (U/V)	The shape of a profile taken parallel to the shelf edge, which is characterised using the General Power Law ($^{G}P_{L}$) programme (Pattyn and Van Huele, 1998). $^{G}P_{L}$ approximates the cross-sectional shape of a gully according to: $y - y_0 = a \mid x - x_0 \mid^{b}$ (3.2) The programme calculates a measure of cross-sectional shape (b value) by finding the RMS misfits between the observed cross-section and a large set of symmetrical shapes defined by the equation. In equation (3.2), a and b are constants and x and y are the horizontal and vertical coordinates taken from a cross-sectional profile of a gully. The programme automatically determines x_o and y_o as the coordinates of the minimum elevation of the gully profile. The b value gives a measure of the cross-sectional shape of the gully and ranges from 1 (V-shape) to 2 (parabolic, commonly referred to as U-shape) on the U/V index (Fig. 3.2).

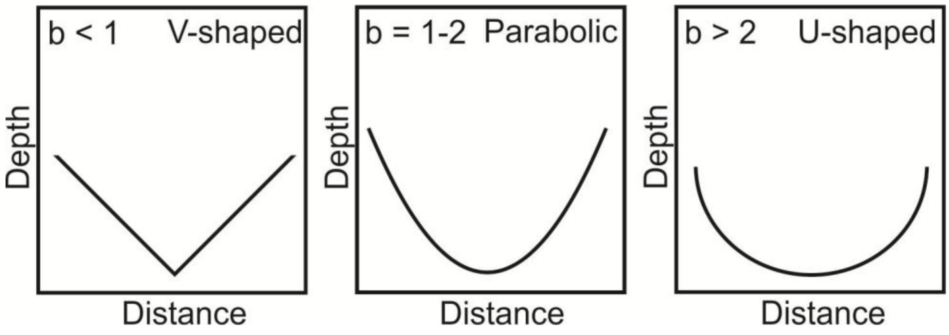

Figure 3. 2. Schematic of cross-sectional shape (U/V) and associated 'b' values derived from the General Power Law programme (see Table 3. 2 for details).

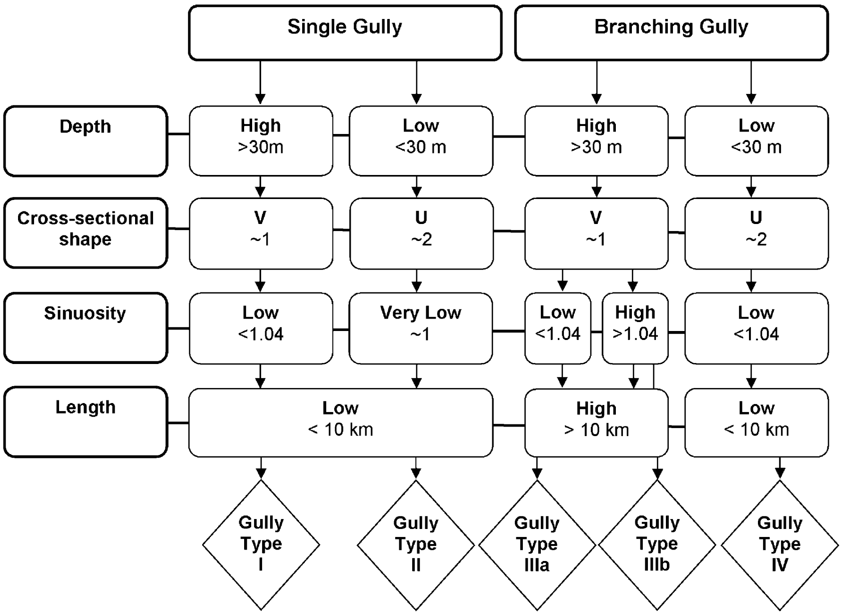

Figure 3. 3. Diagnostic diagram to identify Antarctic gullies based on the quantitative parameters taken at 50 m below the shelf edge including: gully branching order, cross-sectional shape, sinuosity and length. Taken from Gales et al., (2013).

In this study, a gully is identified as a depression of > 5 m and is classified according to its quantitative geomorphic signature. The spatial distribution of individual slope types was mapped by visual pattern recognition. The large-scale continental slope character was assessed by extracting bathymetric slope profiles perpendicular to the continental shelf edge from multibeam bathymetric data. Profiles were used to calculate slope gradient and to analyse slope geometry.

3.3.3. Statistical analysis

Principal Component Analysis (PCA) was used to analyse which gully variables were most important in distinguishing between groups of gullies. PCA is a multivariate analysis tool used to investigate associations between multiple variables within a dataset and aims to explain the maximum amount of variance within a dataset with the fewest Principal Components (PC) (Duntemann, 1989). In PCA, the variables in a dataset are reduced into PCs that represent the highest variance in the data. PCs represent linear combinations of variables that cause maximum variance within the dataset and are calculated until 100% of the variance is explained (Duntemann, 1989). PCA was used to identify which geomorphic gully variables were responsible for the most variance within the dataset, thus identifying which variables are important for classifying different gully types. PCA was carried out in Minitab v15, where the data were normalised into dimensionless units and a correlation matrix calculated. PCA-calculated principal components and eigenvalues of < 1 were excluded from the analysis. Component scores for each gully variable were plotted to show which gully variables cause the most variance within the data.

Statistical significance of the results was tested using analysis of similarities (ANOSIM) in PRIMER 5 (Clarke and Warwick, 2001), by assessing the differences between the groups of gullies, and whether the groupings were significant. Bray–Curtis measure was performed on normalised data (fourth root) to produce a similarity matrix, which was then tested for ANOSIM.

3.4. Considerations and limitations

A major limitation of this study is data availability and data resolution. In many cases, down-slope data is limited to a few km making analysis and interpretation difficult. The highest data resolution used in this study is 20 m, which limits the interpretation of small-scale bedforms.

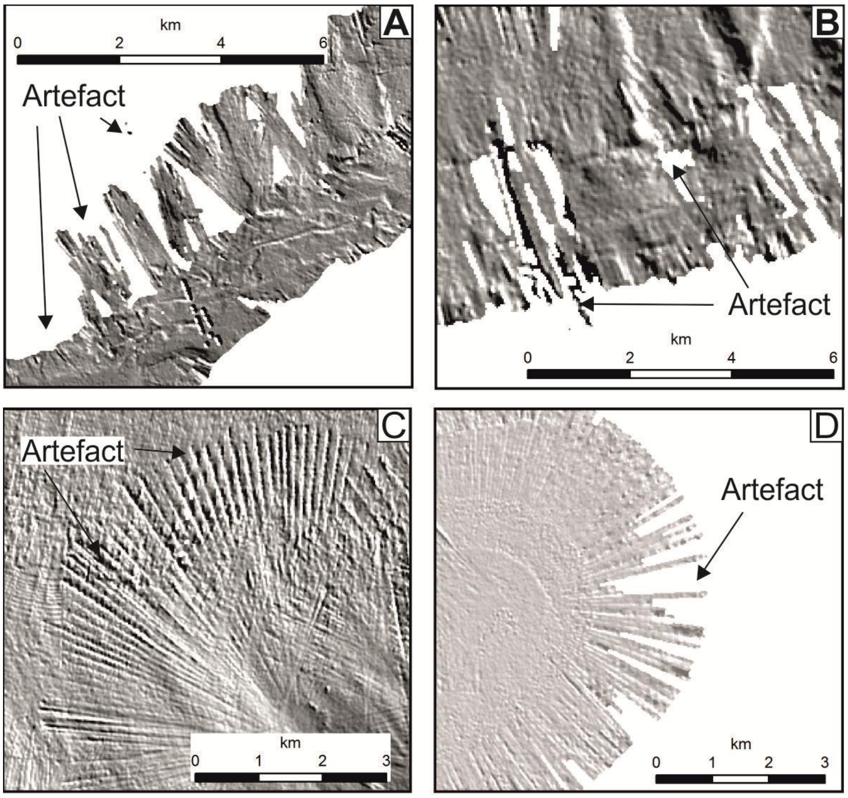

Figure 3. 4. Examples of artefacts within multibeam bathymetric data. **A.** Hillshaded multibeam bathymetric data showing line-drop outs. **B.** Hillshaded multibeam bathymetric data showing line-drops outs and false depth soundings. **C.** Hillshaded multibeam bathymetric data showing artefacts caused by changes in vessel speed. **D.** Hillshaded multibeam bathymetric data showing artefacts caused by changes in vessel speed.

Artefacts and anomalies in the data can limit or hinder interpretation by masking data beneath. Artefacts may be caused by factors influencing the propagation of acoustic waves in the water column, for example due to changes in water temperature, salinity and density (Blondel and Murton, 2009). If the sound-velocity profile is not correct for the area of survey, this may lead to depth errors (observed as ‘smiling’ or ‘frowning’ lines when processing the data), where a deeper or shallower depth is recorded than the true depth. Other variations in the water column, such as due to rough sea conditions, wake, or surveying in sea-ice, can result in bubbles in the water column, leading to holes in the data due to sound attenuation, or shallower depths recorded than the true depth (Fig. 3.4A, 3.4B). Changes in the ships motion (heave, pitch, roll and yaw) may also lead to outliers or spikes in the data. A survey speed which is too high may cause gaps in the data in the along-track direction (Fig. 3.4D).

Another main limitation is the sparseness of seismic data, which would significantly enhance the ability to interpret how bedforms develop and evolve over time. Finally, there is a lack of ground truth data available from the seabed, including photographic stills, video footage and sediment cores.

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Section 2:

Results

Chapter 4.

Submarine gullies on the southern Weddell Sea slope, Antarctica.

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A contribution to the *Atlas of Submarine Glacial Landforms: Modern, Quaternary and Ancient.*

Gales, J. A., Larter, R. D., Leat, P. T., 2013d. Submarine gullies on the southern Weddell Sea slope, Antarctica. In: Dowdeswell, J. A., Canals, S, M., Jakobbson, M., Todd, B. J., Dowdeswell, E. K., Hogan, K. A., (Eds.), *Atlas of Submarine Glacial Landforms: Modern, Quaternary and Ancient*. Geological Society, London, Memoirs, The Geological Society of London (*Submitted*).

4.1. Introduction

Submarine gullies are small-scale, confined channels in the order of tens of metres deep, and form one of the most common morphological features of high latitude continental slopes. Gullies initiate on the upper-slope, erode back into the continental shelf in places and are suggested to be the first features to form on steeply dipping slopes during margin evolution (Laberg et al., 2007). Gully morphology varies in width, incision depth, length, sinuosity, branching order, shelf-incision, cross-sectional shape and spacing, with six distinct gully signatures recognised on high latitude continental slopes (Gales et al., 2013a, b).

4.2. Description

Seventy six gullies incise the upper-slope at the mouth of the Filchner Trough on the Crary Trough-Mouth Fan (Fig. 4.1E). Two types of gully are observed here. Seventy two gullies have distinctive U-shaped cross-sections with a cross-sectional shape that remains U-shaped down-slope (Fig. 4.1E). These gullies have a mean incision depth of 12.5 m, width of 630 m and length of 2.7 km. The gullies initiate at the shelf edge and cut back into the shelf on average by 220 m. The gullies are non-branching with low sinuosities and have a mean gully spacing of 0.6 gully/km. Four gullies display V-shaped cross-sections with longer lengths and lower widths (mean of 283 m) and incision depths (mean of 9.5 m).

At the mouth of the Halley trough, 48 gullies occur with a mean gully spacing of 0.8 gully/km and with a similar morphology to the four V-shaped gullies (Fig. 4.1A). These gullies do not incise the shelf and initiate ~150 m below the shelf edge. The gullies have a low mean incision depth of 11.8 m, a mean width of 560 m, low sinuosity and are generally non-branching. The gullies have a V-shaped cross-section (Fig. 4.1C) and longer mean lengths compared to the U-shaped gullies (Fig. 4.1E).

At the mouth of the Filchner Trough, gullies occur on mean slope gradients of 2.5°, with gradient increasing to 3° at the mouth of the Halley Trough. Deeply incised gullies (mean incision depths $\geq$30 m) that occur on other Antarctic margins are absent from the southern Weddell Sea continental margin (Gales et al., 2012). Sub-bottom data show that gullies at the mouth of the Filchner Trough form in sediment with low acoustic penetration (Fig. 4.1B and 4.1F) and are draped by a layer of semi-transparent sediment of variable thickness. At the mouth of Halley Trough, the gullies initiate below a region of intense iceberg scouring (Fig. 4.1A), whereas the outer shelf of the Filchner Trough is less iceberg-scoured (Fig. 4.1E).

4.3. Interpretation

The predominant morphological signature at the mouth of the Filchner Trough is small-scale and U-shaped gullies that have been categorized as 'type *II*' according to their geomorphic signature. Gales et al. (2013a) describe Type *II* gullies as: non-branching, low incision depth (<30 m), U-shaped, low sinuosity and low length (<10 km). The four small-scale and V-shaped gullies that occur at the mouth of the Filchner Trough and in higher quantities at the mouth of the Halley Trough have been categorized as 'type *I*' gullies, which are characterised by a non-branching, V-shaped, low sinuosity and low length signature (Gales et al., 2013c). Sub-bottom data show that both type *I* and *II* gullies form in poorly stratified or acoustically impenetrable layers, suggesting that the gullies incise the seafloor. This contrasts with good sub-bottom penetration that would be expected if gullies were formed by aggradation of the gully interfluves (Fig. 4.1B and 4.1F). Variation in gully morphology is likely to reflect differences in slope processes, slope character (i.e. substrate, gradient and geometry), ice-sheet history and environmental controls including large-scale spatial characteristics (e.g. drainage basin size, location of cross-shelf troughs).

The processes which form submarine gullies are, however, not well constrained (e.g. Noormets et al., 2009). Suggested mechanisms include: (1) erosion by mass flows (e.g. turbidity currents, debris flows, slides and slumps) initiated by factors such as gas-hydrate dissociation, tectonic influences, presence of weak sedimentary layers within the seabed, re-suspension by iceberg scouring and tidal activity; (2) oceanographic processes such as geostrophic currents, tides and cascading dense water, formed during sea ice formation through brine rejection; and (3) glacigenic processes, such as sediment-laden subglacial meltwater discharged from beneath an ice sheet and the rapid delivery of glacigenic sediment by ice-streams to the shelf edge.

At the mouth of the Filchner Trough, the morphology of type *II* gullies resembles small-scale slide scars, displaying flat-floored cross-sections and steep gully walls. As 60% of the gullies incise the seabed between 5 and 15 m, the gullies are likely to be the result of small-scale slope failure (Gales et al., 2012). Although the gullies form in a region of energetic and cascading cold, dense water overflow, with oceanographic moorings measuring maximum flow velocities of 1 m s^{-1} (Foldvik et al., 2004), their U-shaped and shallow morphology is indicative of small-scale sliding, probably resulting from rapid accumulation and subsequent failure of proglacial sediment during glacial maxima. Deeply incised and V-shaped gullies that are observed along other Antarctic margins, are absent from the southern Weddell Sea margin, suggesting that cold, dense-water overflow is not a likely mechanism forming deeply incised and V-shaped gullies.

The V-shaped type *I* gullies at the mouth of the Halley Trough were probably formed by suspended sediment flows, such as turbidity currents or sediment-laden subglacial meltwater. One mechanism that can initiate turbidity currents is iceberg scouring, which may resuspend sediment deposited at the mouth of the trough. As the upper-slope of the Halley Trough is significantly more iceberg scoured, this may explain why type *I* gullies occur here but are absent from the mouth of the Filchner Trough, where the upper slope is less affected by iceberg scouring.

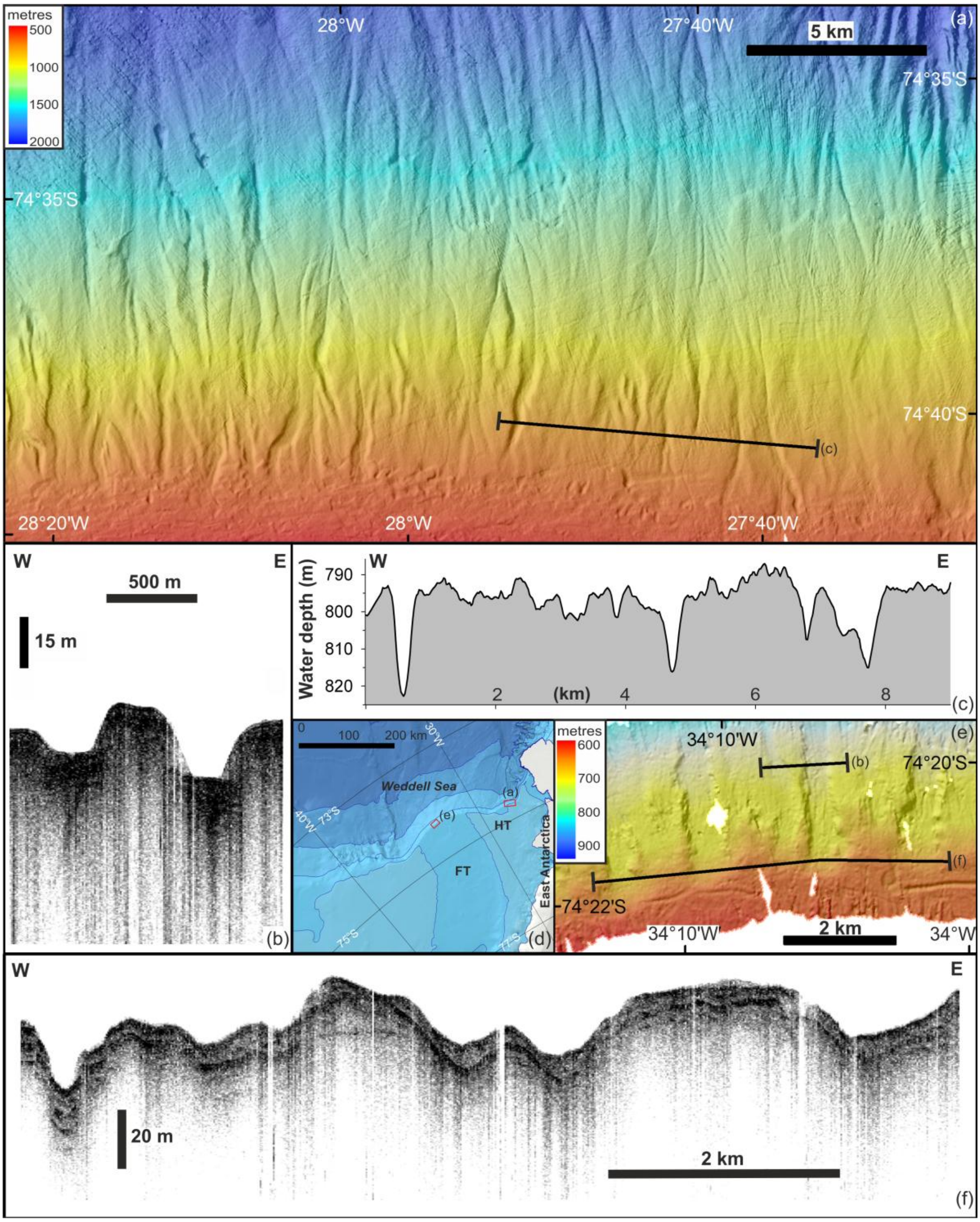

Figure 4. 1. Multibeam bathymetry and along-slope profiles of gullies on the southern Weddell Sea continental margin. **A.** Sun-illuminated multibeam bathymetry showing V-shaped gullies at the mouth of Halley Trough. Acquisition system Kongsberg EM122. Frequency 12 kHz. Grid-cell size 30 m; **B.** Sub-bottom profile of U-shaped gullies. Location is highlighted in part (e). Acquisition system Kongsberg TOPAS PS 018 parametric sub-bottom profiler. Frequency 18 kHz. Secondary beam frequency 1300-5000 Hz; **C.** Along-slope profile through V-shaped gullies. Location is highlighted in part (a). **D.** Map from IBCAO Version 3.0. FT is Filchner Trough. HT is Halley Trough; **E.** Sun-illuminated multibeam bathymetry showing U-shaped gullies at the mouth of Filchner Trough. Acquisition system Kongsberg EM120. Frequency 12 kHz; Grid-cell size 30 m; **F.** Sub-bottom profile of U-shaped gullies. Location is highlighted in part (e). Acquisition system is the same as for part (b).

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Chapter 5.

Geomorphic signature of Antarctic submarine gullies: Implications for continental slope processes

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5. Abstract

Five quantitatively distinct gully types are identified on the Antarctic continental margin from swath bathymetric data of over 1100 individual features. The gullies differ in terms of length, width, depth, width/depth ratio, cross-sectional shape, branching order, sinuosity and spatial density. Quantitative analysis suggests that Antarctic gully morphology varies with local slope character (i.e. slope geometry, gradient), regional factors (i.e. location of cross-shelf troughs, trough mouth fans, subglacial meltwater production rates, drainage basin size), sediment yield and ice-sheet history. In keeping with interpretations of previous researchers, most gullies are probably formed by hyperpycnal flows of sediment-laden subglacial meltwater released from beneath ice sheets grounded at the continental shelf edge during glacial maxima. The limited down-slope extent of gullies on the western Antarctic Peninsula is explained by the steep gradient and slope geometry at the mouth of Marguerite Trough, which cause flows to accelerate and entrain seawater more quickly, resulting in a reduction of the negative buoyancy effect of the sediment load. Due to pressure gradients at the ice-sheet bed caused by variations in ice thickness inside and outside palaeo-ice stream troughs, subglacial meltwater flow was generally focussed towards trough margins. This has resulted in gullies with larger cross-sectional areas and higher sinuosities at the trough margins. A unique style of gullying is observed off one part of the western Antarctic Peninsula, corresponding to an area in which the ice-sheet grounding line is not thought to have reached the shelf edge during the Last Glacial Maximum. We interpret the features in this area as the cumulative result of slope processes that operated over a long period of time in the absence of hyperpycnal meltwater flows.

Keywords: submarine gully, Antarctica, subglacial meltwater, continental margin, geomorphology

1.1. Introduction

Understanding the processes operating on continental margins is essential for interpreting seafloor erosion patterns, continental margin evolution, canyon development and sediment core records from the continental slope and rise. Submarine gullies are the most common morphological features observed on the Antarctic continental margin, and the processes which form these gullies remain a key research question, with important implications for continental slope research.

Submarine gullies are becoming increasingly important in the field of geomorphology, with some studies documenting different gully morphologies on Antarctic continental margins (i.e. Noormets et al., 2009; Gales et al., 2012) and on mid-latitude margins (i.e. Micallef and Mountjoy, 2011; Vachtman et al., 2012). In Antarctica, gullies incise the upper slope over much of the continental margin, in places eroding back into the continental shelf (Noormets et al., 2009). Gullies are present along the western Antarctic Peninsula continental margin and along the shelf edge and upper slopes of the Weddell, Bellingshausen, Amundsen and Ross seas (Vanneste and Larter, 1995; Shipp et al., 1999;

Lowe and Anderson, 2002; Dowdeswell et al., 2004a, 2006, 2008; Heroy and Anderson, 2005; Noormets et al., 2009; Gales et al., 2012).

Variation in gully morphology is likely to reflect differences in the underlying substrate, the spatial characteristics of the slope and slope processes. The continental slopes considered in this paper are all underlain by thick Quaternary sediments, most of which are glacially derived (e.g. glacigenic debris flows) or glacially influenced (e.g. glacimarine muds with ice rafted debris) (Bonn et al., 1994; Hillenbrand et al., 2005; Dowdeswell et al., 2004a; 2006; Wright and Anderson, 1982; Melles and Kuhn, 1993). Seismic reflection and coring studies from the Antarctic continental slope show that sediments present exhibit a limited range of characteristics, with similar lithology, physical properties and grain-size composition (e.g. Vanneste and Larter, 1995; Anderson and Andrews, 1999; Michels et al., 2002; Dowdeswell et al., 2004a, 2006, 2008; Hillenbrand et al., 2005; 2009). As the outer-shelf and upper-slope sediments are relatively homogenous, variation in substrate is therefore not likely to be a factor influencing differences in gully morphology. Although variation in geotechnical properties cannot be dismissed unequivocally, and are in some cases influenced by processes themselves, correlations suggest a strong effect of process rather than substrate in influencing gully morphology.

In this study we quantitatively analyse Antarctic gully morphologies and show that several different gully types are present. Based on characteristics of the different gully types, we infer slope processes that are operating, or have operated in the past, on Antarctic continental margins. The study aims to reduce large uncertainties regarding Antarctic slope processes and to provide a diagnostic tool in which gully types may be identified and used, alongside other indicators, in assessing the processes operating in these high latitude environments.

5.1.1. Slope processes

There are large uncertainties regarding processes operating on different parts of the Antarctic continental margin, with the influence of slope processes and slope character difficult to separate. Suggested processes include mass flows, subglacial meltwater discharge and dense bottom water overflow.

Mass flows, including sediment slides, slumps, debris flows and turbidity currents may be initiated by a range of mechanisms such as gas hydrate dissociation, tidal pumping beneath large icebergs and near ice shelf grounding lines, resuspension by shelf and contour currents, tectonic disturbances, iceberg scouring and rapid accumulations of glacigenic debris at the shelf edge during glacial maxima (Larter and Cunningham, 1993; Vanneste and Larter, 1995; Shipp et al., 1999; Michels et al., 2002; Dowdeswell et al., 2006, 2008; Dowdeswell and Bamber, 2007).

Subglacial meltwater is thought to have been discharged from ice-sheet grounding lines during glacial maxima, through either of two processes. (1) A continuous supply, where basal water is generated predominantly through geothermal and strain heating, as there was little or no ice-surface melting during full-glacial times to supply the basal drainage system. Typical yields for geothermal heating are, however, in the range of mm/yr of basal melt (e.g. Pattyn, 2010; Beem et al., 2010), making it difficult to sustain continuous discharges over long periods. (2) Episodic water release such as subglacial lake discharges (‘glacial outburst floods’ or ‘jökulhlaups’) (Goodwin, 1988; Wellner et al., 2001; Dowdeswell et al., 2006; 2008; Fricker et al., 2007; Bell, 2008; Noormets et al., 2009, Piper et al., 2012). Subglacial meltwater is able to entrain sediment at the base of glaciers and may produce hyperpycnal flows when discharged at the grounding line (Russell and Knudsen, 1999a, b; 2002). The critical sediment concentration needed for meltwater to initiate hyperpycnal flows in seawater is 1-5 kg m^{-3} (Parsons et al., 2001; Mulder et al., 2003). This value is considerably lower than previous estimates of sediment concentrations (e.g. 33 kg m^{-3}; Syvitski, 1989) which are based on buoyancy considerations and do not take into account the effects of fine-scale convective instability (Parson et al., 2001; Mulder et al., 2003). Large palaeo-subglacial drainage systems have been documented on the inner Antarctic continental shelf and in onshore areas that are presently ice free (Sugden et al., 1991; Lowe and Anderson, 2002; Ó Cofaigh et al., 2002; Denton and Sugden, 2005; Domack et al., 2006; Anderson and Oakes-Fretwell, 2008; Larter et al., 2009). However, very few channel systems have been observed on well-preserved outer shelf sediments in cross-shelf troughs.

Dense bottom water formed from sea-ice freezing and brine rejection, and the cascading of this bottom water down-slope may also influence slope morphology (Kuvaas and Kristoffersen, 1991; Dowdeswell et al., 2006, 2008; Noormets et al., 2009). The Weddell Sea is the largest source of Antarctic Bottom Water (AABW), which is exported from the Southern Ocean and forms a major component of the global thermohaline circulation (Nicholls et al., 2009). High Salinity Shelf Water (HSSW), produced in the Weddell Sea through brine rejection during sea-ice production, is supercooled and freshened by circulation beneath the ice shelves, producing cold and dense Ice Shelf Water (ISW). HSSW and ISW contribute to the production of Weddell Sea Bottom Water (WSBW) and Weddell Sea Deep Water (WSDW), which are components of AABW (Nicholls et al., 2009). Temperature profiles of the water column west of the Antarctic Peninsula and in the Bellingshausen and Amundsen seas preclude bottom water production at the present day in those regions (Hoffman and Klinck, 1998; Smith et al., 1999; Smith and Klinck, 2002; Dinniman and Klinck, 2004). Gales et al., (2012) showed that deeply incised and V-shaped gullies, as observed on other parts of the Antarctic continental margins, were absent from the mouth of the Filchner Trough, in a region of active and highly energetic cascading dense water overflow. This absence makes it unlikely that V-shaped and deeply eroded gullies found elsewhere along the Antarctic continental margin were formed by erosion caused by cascading dense water (Gales et al., 2012).

5.2. Regional setting

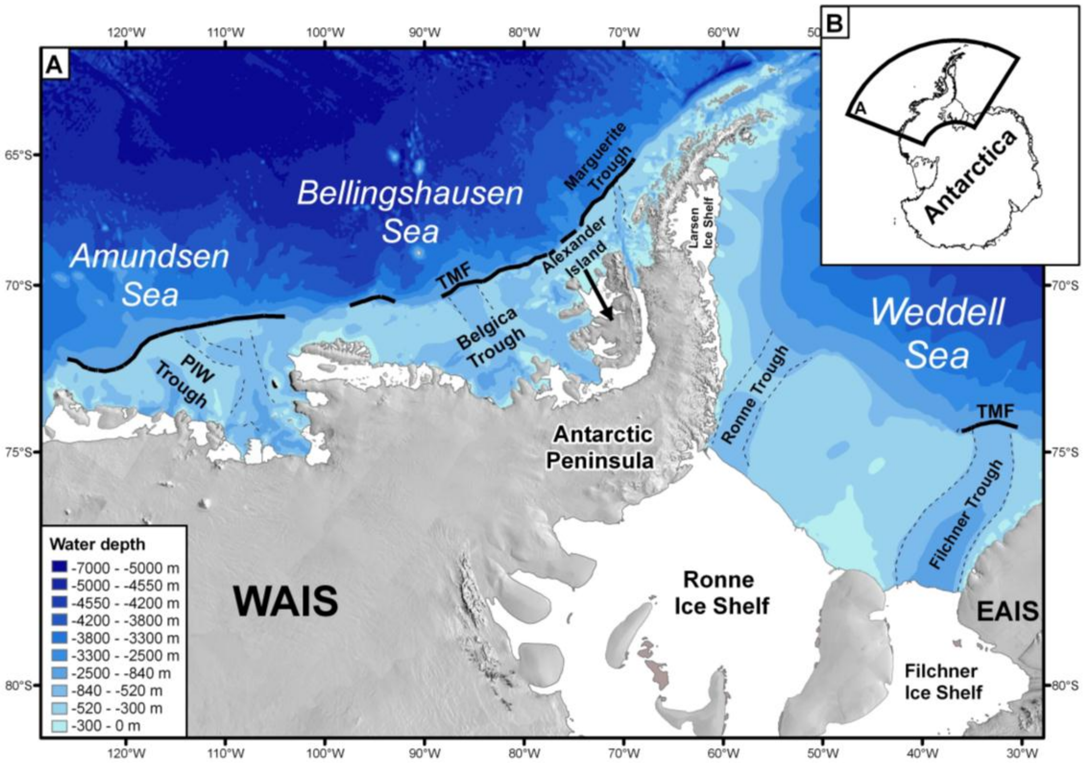

Figure 5. 1. A. Location map of study areas on the Weddell Sea, western Antarctic Peninsula, Bellingshausen and Amundsen Sea continental shelf edge. Thick black lines indicate areas of continental shelf edge analysed. Dashed black lines indicate boundaries of cross-shelf troughs. PIW is Pine Island West. WAIS is West Antarctic Ice Sheet. EAIS is East Antarctic Ice Sheet. TMF is trough mouth fan. **B**. Inset showing location of Fig. 5.1A in relation to the Antarctic continent. Regional bathymetry from Bedmap2 (Fretwell et al., 2013). Antarctic continent is shaded Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007).

The study areas on the Antarctic margin encompass 1919 km of continental shelf edge and upper slope to the west of the Antarctic Peninsula (WAP), in the Bellingshausen and Amundsen seas, and at the mouth of the Filchner Trough, Weddell Sea (Fig. 5.1). The continental shelves of the WAP, Bellingshausen, Amundsen and Weddell seas show characteristic landward-sloping gradients, formed as a result of erosional overdeepening of the inner shelves, and to a lesser extent lithospheric flexture due to ice sheet loading (ten Brink and Cooper, 1992; ten Brink and Schneider, 1995; Bart and Iwai, 2012). The continental shelves are dissected by broad cross-shelf troughs which extend to the shelf edge in places. The troughs were eroded by fast-flowing ice streams which transported sediment towards the shelf edge during glacial maxima (e.g. Vanneste and Larter, 1995; Dowdeswell and Siegert, 1999; Ó Cofaigh et al., 2003; Dowdeswell et al., 2004a, 2004b) and deposited prograding sedimentary sequences along the continental margin and upper slope (e.g. Kuvaas and Kristoffersen, 1991; Larter and Cunningham, 1993; Larter et al., 1997; Cooper et al., 2008). Modal depths of the Antarctic continental shelves are in the range of 400-600 m.

Although the study areas have varying tectonic histories (e.g. Hübscher et al., 1996; Livermore and Hunter, 1996; Eagles et al., 2004, 2009), the underlying Quaternary geology along these Antarctic continental margins is fundamentally similar. The continental slopes are constructed largely of prograded sequences (e.g. Anderson, 1999; Cooper et al., 2008) that reflect cycles of growth and retreat of grounded ice from the West Antarctic Ice Sheet (WAIS), East Antarctic Ice Sheet (EAIS) and Antarctic Peninsula Ice Sheet (APIS). At the mouth of the Belgica and Filchner Trough, sediment progradation has resulted in the formation of trough mouth fans which comprise mainly of glacigenic debris-flow deposits (Ó Cofaigh et al., 2003). The EAIS grew rapidly at the Eocene-Oligocene boundary (34 Ma), reaching the coast in Prydz Bay and the Ross Sea (Barron et al., 1988; Cooper and O'Brien, 2004; Barrett, 1989; Barrett et al., 1995). Early history of the WAIS is less clear; however, it is likely that a major step in its development was associated with the shift in oxygen isotope ratios in open ocean foraminifera about 14 Ma (Zachos et al., 2001). Compelling evidence from the morphology of volcanoes in Marie Byrd Land, West Antarctica, shows that the climate in the interior of the region has not been warm enough to permit significant runoff since the Middle Miocene (Rocchi et al., 2006). During the Last Glacial Maximum (LGM), ice extended across the continental shelf of the WAP and Bellingshausen, Amundsen and Weddell seas, reaching the shelf edge in most areas (e.g. Anderson et al., 2002; Ó Cofaigh et al., 2005a, 2005b; Hillenbrand et al., 2012).

5.3. Data and methods

Multibeam swath bathymetry and TOPAS acoustic subbottom profile data used in this study are summarised in Table 5.1. These data were collected using a range of systems including a Kongsberg EM120 multibeam swath bathymetry system with a frequency range of 11.25-12.75 kHz and swath width of up to 150° (on RRS *James Clark Ross* and, since 2002, on RVIB *Nathaniel B. Palmer*); a Krupp-Atlas Hydrosweep DS-2 which transmits at around 15 kHz (on RV *Polarstern*) and a Seabeam 2112 system which transmits at 12 kHz (on RVIB *Nathaniel B. Palmer* until 2002). TOPAS data were collected using a TOPAS PS 018 acoustic subbottom profiling system, with two primary transmitting frequencies near 18 kHz forming secondary chirp pulses with a frequency range of 1500 to 5000 Hz. TOPAS data have a resolution of better than 1 m and can show signal penetration to depths of > 50 m below the seafloor in fine-grained sediments.

Swath bathymetry data were processed and gridded to a cell size of 50 m or finer using public-access MB-System software (Caress and Chayes, 1996). The slope geomorphology was analysed quantitatively using standard geographic information system (GIS) tools by (1) extracting profiles parallel to the shelf edge at a range of depths down-slope along which gully parameters were measured; (2) extracting profiles across gullies to measure cross-sectional shape; and (3) analysing gully density and spatial patterns.

Table 5. 1. Data sets used in analysis.

Data set		Reference	Principal Investigator
Cruise / ID	**Year**		
ANT23-4	2006	Gohl (2006); Nitsche et al. (2007)	K. Gohl
JR59	2001	Ó Cofaigh et al. (2002); Dowdeswell et al. (2004a, 2004b)	C. Pudsey
JR71	2002	Ó Cofaigh et al. (2005a); Dowdeswell et al. (2004a, 2004b)	C. Pudsey and J. Dowdeswell
JR84	2003	Evans et al., (2006)	A. Jenkins and J. Dowdeswell
JR97	2005	Gales et al. (2012); Larter et al. (2012)	K. Nicholls
JR104	2004	Ó Cofaigh et al. (2005b); Dowdeswell et al. (2008)	R. Larter
JR141	2006	Noormets et al. (2009); Graham et al. (2010)	R. Larter
JR157	2007	Noormets et al. (2009)	J. Dowdeswell
JR179	2008	Graham et al. (2010); Graham and Smith (2012)	R. Larter and P. Enderlein
JR244	2011	Gales et al. (2012); Larter et al. (2012)	R. Larter
NBP9902	1999	Wellner et al. (2001, 2006); Lowe and Anderson (2002)	J. Anderson
NBP0001	2000	Nitsche et al. (2007)	S. Jacobs
NBP0103	2001	Bolmer (2008)	P. Wiebe
NBP0104	2001	Bolmer (2008)	P. Wiebe
NBP0201	2002	Wellner et al. (2006)	J. Anderson
NBP0202	2002	Bolmer (2008)	P. Wiebe
NBP0702	2007	Nitsche et al. (2007)	S. Jacobs

5.3.1. Analysis of gully parameters

Gully parameters, including gully width, depth and length were calculated by extracting profiles from the swath bathymetry data parallel to the shelf edge. Profiles were taken at 50 m depth below the shelf edge and down-slope along the gully length. In this study, we identify a gully as a depression of > 5 m which can be distinguished from a channel by the spatial distribution, where the gully head incises the shelf edge or upper continental slope. Submarine channels are usually larger features, with canyons characterised by steep-walls, sinuous valleys, V-shaped cross-sections, outward sloping axes and relief comparable to that of their terrestrial counterparts (Shepard, 1963; Shepard, 1981).

Length is defined as the distance a gully can be traced down-slope; however, for some gullies (approximately 10% of the gullies measured), this is limited by data extent. Width is defined as the distance between the points of maximum curvature of the gully flanks and depth is the vertical distance from gully base to the line defining gully width (Fig. 5.2) (Noormets et al., 2009). The width/depth ratio (W/D) is a dimensionless ratio which gives an indication of the gully shape. Sinuosity is a dimensionless measure of the ratio of channel length and straight-line distance measured at the top and bottom of the gully, giving a sinuosity index (S) where:

$$\text{Sinuosity index (S)} = \text{distance measured along gully} / \text{straight line distance} \quad (5.1)$$

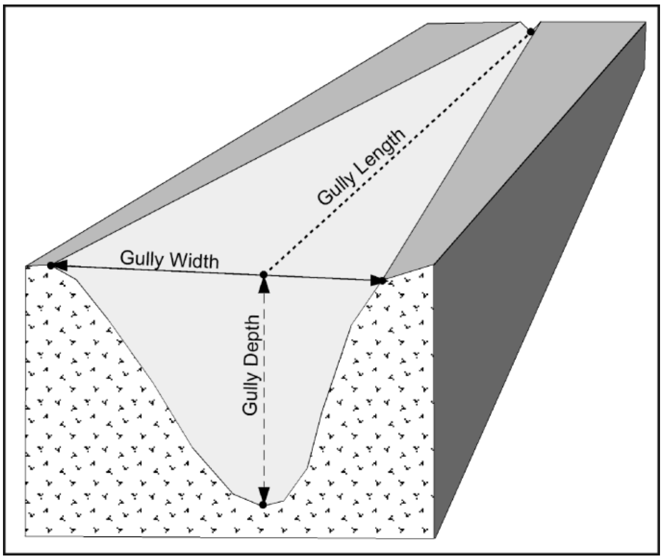

Figure 5. 2. Gully morphometric parameters. For explanation of definitions, see text. Adapted from Noormets et al. (2009).

We define a sinuosity index greater than 1.04 as sinuous and, by definition, a straight gully would have S = 1. These values are based on the data available to this study and, therefore, they may not be appropriate for classifying gullies on all Antarctic continental slopes. The cross-sectional shape of a gully is measured using the General Power Law ($^{G}P_{L}$) programme (Pattyn and Van Huele, 1998), which approximates gully cross-sectional shape according to the general power law equation:

$$y - y_0 = a \mid x - x_0 \mid^{b} \qquad (5.2)$$

where *a* and *b* are constants, *x* and *y* are the horizontal and vertical coordinates taken from a cross-sectional profile of a gully and x_o and y_o are the coordinates of the point of inflection of the gully profile, automatically determined by the programme. The *b* value gives a measure of the shape of the gully, ranging from 1 (V-shape) to 2 (parabolic, commonly referred to as U-shape) on the U/V index. In subsequent sections, we refer to *b* as the 'U/V index'. *b* values of < 1 indicate convex-upward gully flanks whereas values of > 2 indicate a more box-shaped morphology. The *b* value is determined by calculating the minimum RMS misfit between the initial and idealised cross-section from equation 5.2.

Stream-order calculations are used to classify reaches of a stream network in terms of the number of upstream tributaries, with order-1 streams being those with no tributaries. Strahler's (1957) method is used to calculate gully order, by assigning each reach of a gully an order of 1. Where two 1^{st} order reaches meet, the down-stream reach is assigned 2^{nd} order. Where two 2^{nd} order reaches meet, the down-stream reach is assigned 3^{rd} order, and so on. Where two reaches with different orders meet, the down-stream reach is assigned the higher order of the two contributing reaches. The gullies were assigned orders in the same way using the automated Stream Order application in ArcGIS 9.3 (hydrology tools).

The large-scale continental slope character was assessed by extracting bathymetric slope profiles perpendicular to the continental shelf edge from multibeam swath bathymetry data. These were used to calculate slope gradient and to analyse slope geometry.

5.3.2. Statistical analysis of gully variables

Principal Component Analysis (PCA) was used to analyse which gully variables were most important in distinguishing between groups of gullies. PCA is a multivariate analysis tool used to investigate associations between multiple variables within a dataset and aims to explain the maximum amount of variance within a dataset with the fewest Principal Components (PC) (Dunteman, 1989). In PCA, the variables in a dataset are reduced into PCs which represent the highest variance in the data. PCs represent linear combinations of variables that cause maximum variance within the dataset and are calculated until 100% of the variance is explained (Dunteman, 1989). PCA was used to identify which geomorphic gully variables were responsible for the most variance within the dataset (thus identifying which variables are important for classifying different gully types). PCA was carried out in Minitab v15, where the data were normalised into dimensionless units and a correlation matrix calculated. PCA calculated principal components, and eigenvalues < 1 were excluded from the analysis. Component scores for each gully variable were plotted to show which gully variables cause the most variance within the data.

Statistical significance of the results was tested using analysis of similarities (ANOSIM) in PRIMER 5 (Clarke and Warwick, 2001), by assessing the differences between the groups of gullies, and whether the groupings were significant. Bray-Curtis measure was performed on normalised data (fourth root) to produce a similarity matrix which was then tested for ANOSIM.

5.4. Results

We identify six quantitatively different slope types along 1919 km of the Antarctic continental shelf edge from swath bathymetry data (Fig. 5.3 and 5.4). The slope types are classified on the basis of quantitative analysis of the swath bathymetry data (Fig. 5.5 and 5.6), continental slope character and gully spatial distribution (Fig. 5.7). This quantitative analysis supports differences between slope morphological types that we originally identified by qualitative visual pattern recognition. Gully geomorphic parameters were analysed statistically to identify the most important variables in distinguishing different gully types through PCA (Fig. 5.6). The spatial distributions of individual slope types were mapped by visual pattern recognition based on the diagnostic scheme set out in Fig. 5.8.

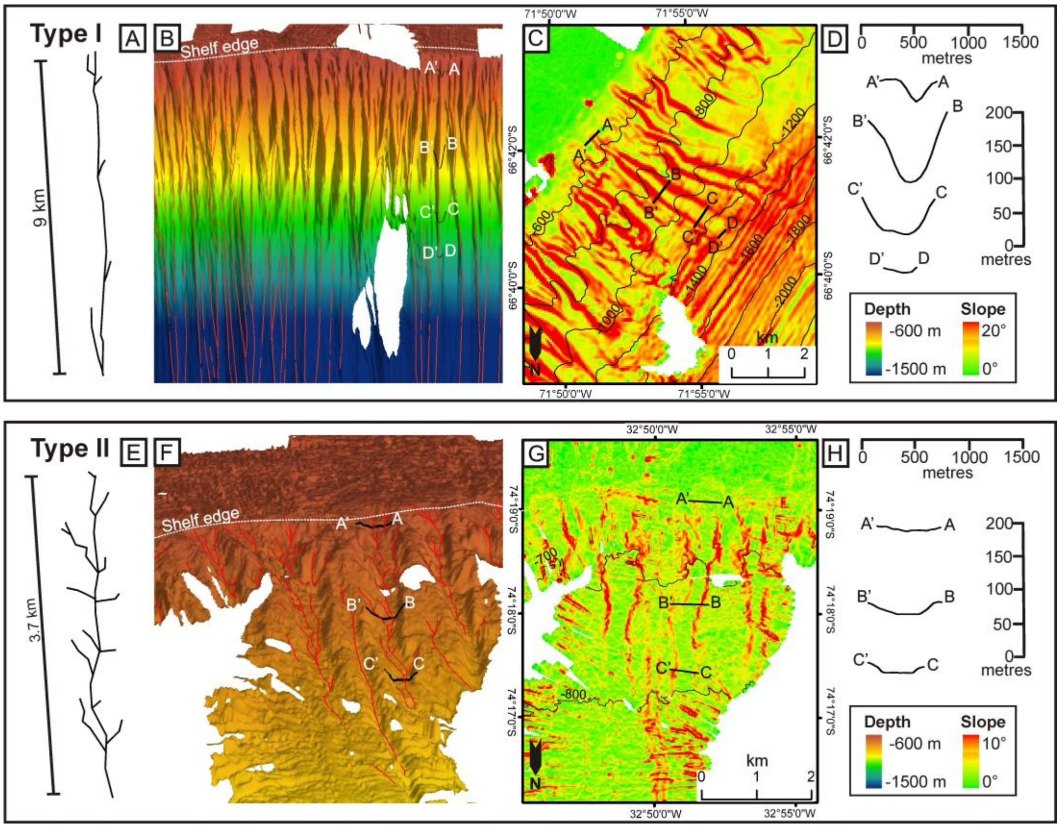

Figure 5. 3. Non-branching gully morphology: Types *I* and *II*. **A**. Schematic of Type *I* gully morphology; **B**. 3D bathymetry overlaid with gully drainage network. Along-slope profiles (labelled A-D') correspond to profiles in C and D. **C**. Local slope angle of the seafloor overlaid with depth contours at 200 m intervals. Along-slope profiles correspond to profiles in B and D. **D**. Along-slope profiles. Locations of profiles are shown in B and C. **E**. Schematic of type *II* gully morphology; **F**. 3D bathymetry overlaid with gully drainage network. Along-slope profiles (labelled A-C') correspond to profiles in G and H. **G**. Local slope angle of the seafloor overlaid with depth contours at 100 m intervals. Along-slope profiles correspond to profiles in F and G. **H**. Along-slope profiles. Locations of profiles are shown F and G.

5.4.1. Morphological identification of gully types

The six slope morphological types, derived from observations of over 1100 individual gullies, include five geomorphically distinct gully types identified as *I*, *II*, *IIIa*, *IIIb* and *IV* and smooth seafloor, defined here as seafloor with topographic depressions of < 5 m and/or with a gully density of < 0.1 gully/km (Fig. 5.3 and 5.4). A matrix plot of gully parameters including gully length, width, depth, width/depth (W/D) ratio, U/V index and sinuosity, shows that from visual inspection, the data form clusters representing the five gully types (Fig. 5.5). Cross-plots of gully length against W/D ratio and depth form well-defined clusters, corresponding with low normalised standard deviations for gully length (1.42), U/V index (2.90) and width/depth ratio (3.04) (Table 5.2). Plots of sinuosity against depth and width, and U/V index against width show a greater spread in the data, corresponding to higher normalised standard deviations (Table 5.2). Gully type *IIIb* forms two clusters along the length axis due to some measured gully lengths being limited by down-slope swath bathymetric data coverage.

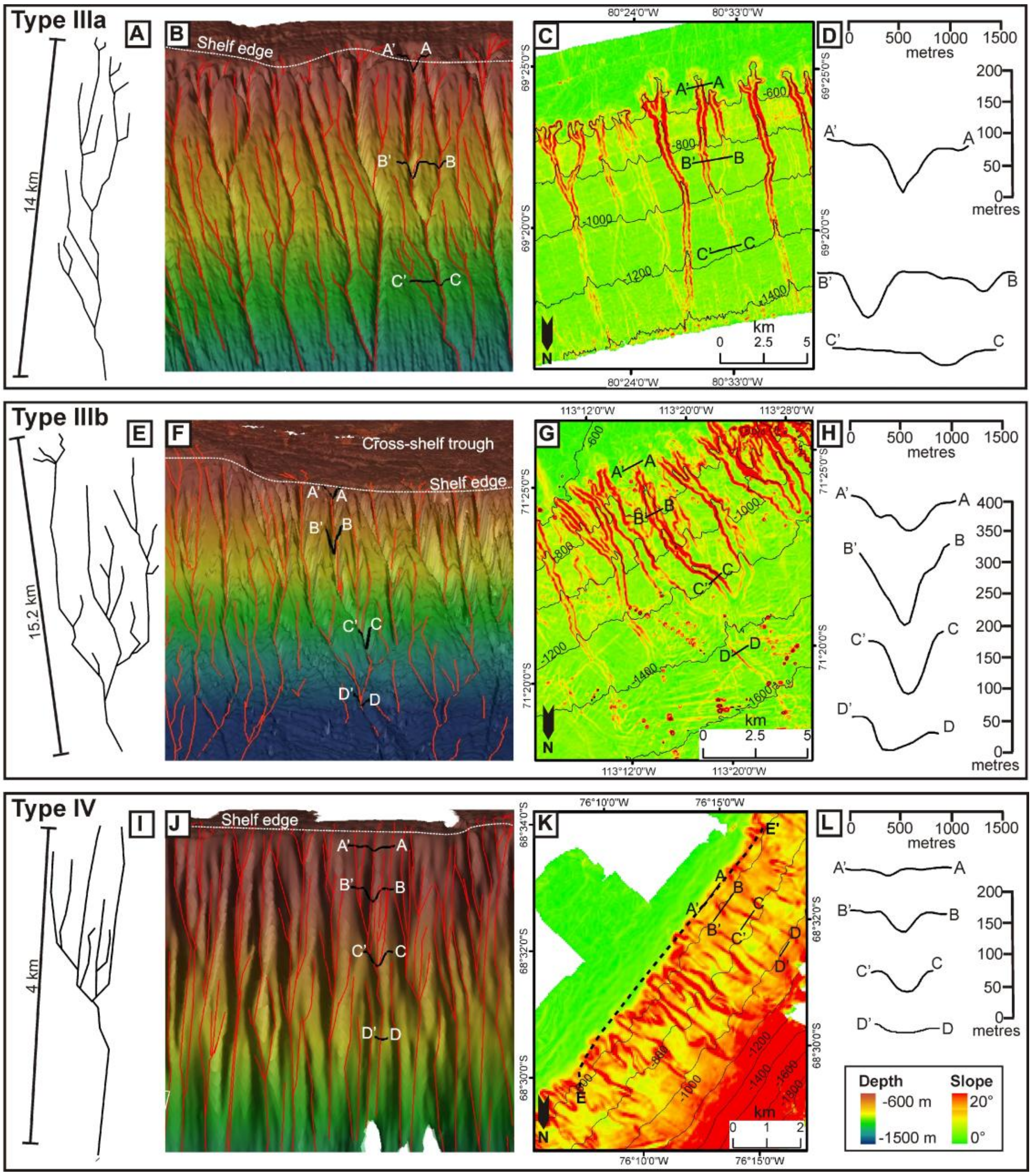

Figure 5. 4. Branching gully morphology: Types *IIIa*, *IIIb* and *IV*. As Fig. 5.3, for type *IIIa* gullies (A-D), type *IIIb* gullies (E-H) and type *IV* gullies (I-L). **K**. Profile E'-E marks location of TOPAS profile in Fig. 5.9.

5.4.2. PCA and statistical analysis

Table 5. 2. Standard deviation of normalised gully variables for each gully type (*I-IV*).

Gully type	N*	U/V	Sinuosity	Length	Width	Depth	W/D	Branching
I	64	2.04	0.61	0.07	0.87	0.59	0.55	0.90
II	61	0.24	0.45	0.07	1.06	0.24	1.24	0.56
IIIa	18	0.16	0.63	0.11	0.91	1.08	0.10	0.00
IIIb	154	0.15	1.19	1.15	1.03	1.15	0.91	0.77
IV	19	0.31	0.70	0.03	0.52	0.38	0.23	0.76
Total	316	2.90	3.62	1.42	4.34	3.45	3.04	2.98

*N is number of observations

PCA conducted on six geomorphic gully parameters (U/V index, sinuosity, gully length, width, depth and degree of branching) calculated a total of six Principal Components (PCs) until 100% of the variance was explained. The first two PCs, with eigenvalues of > 1, were extracted, as these are assumed to be the variables which best characterise gully type (Table 5.3). PCs with eigenvalues of < 1 were excluded from the interpretation, as suggested by Mandal et al. (2008). PC1 and PC2 represent 65.5% of variance within the dataset, with most variability within PC1 (Fig. 5.6, axis 1) due to gully length (score 0.536) and within PC2 (Fig. 5.6, axis 2) due to width (score 0.690), where a higher value indicates a stronger association to the component. Most variability within PC3 is explained by U/V index (score -0.731).

The PCA score plot (Fig. 5.6) shows clusters of samples with similar character, with a tighter cluster representing closer similarity. Five distinct clusters are clearly separated into different sections of the graph and represent the same five gully types shown on the matrix plot (Fig. 5.5, gully types *I, II, IIIa, IIIb* and *IV*).

The results of the matrix and PCA analysis indicate that gullied slopes can be classified into five separate morphological types, which are distinguished by their quantitative parameters. The statistical significance of these groups was tested using analysis of similarities (ANOSIM). The results show that most groups are statistically strong, with *P*-values of 0.1% between all groups apart from between types *I* and *II* (0.2%) and groups *IIIa* and *IIIb* (44.6%). All values, apart from between groups *IIIa* and *IIIb*, lie within the 95% significance level. The high *P*-value between types *IIIa* and *IIIb* suggests that the groupings are not statistically strong and is further support that these are sub-groups of the broader gully type *III*.

Table 5. 3. Eigenvectors for 1st, 2nd and 3rd Principal Component (PC) from PCA after standardisation of variables

Eigenvectors	PC1	PC2	PC3
U/V	-0.252	0.503	0.731
Sinuosity	0.373	0.382	-0.344
Length	0.536	0.084	-0.114
Width	0.117	-0.690	-0.319
Depth	0.500	-0.220	0.383
Branching	0.496	0.263	0.293

5.4.3. Gully signature, abundance and spatial distribution

The most common upper-slope type found on the areas of the Antarctic continental margin studied is topographically smooth seafloor, covering 39% of the margin. Of the five gully types, type *I* is the most abundant, occurring on 21% of the margin and characterised by non-branching and low sinuosity channels which have limited expression down-slope (Fig. 5.3A). These gullies have low W/D ratios and are found in the highest density (1.54 gully/km), predominantly on high-gradient slopes (~9º - 13º). Type *I* is absent from trough mouth fans but is found at trough mouths where no fans have developed (e.g. Marguerite trough-mouth; Dowdeswell et al., 2004a) and elsewhere along the continental margins. Type *I* gullies have V-shaped cross-sections at the shelf edge, becoming progressively deeper and more U-shaped down-slope (Fig. 5.3D).

Type *II* gullies cover 12% of the shelf edge and are characterised by non-branching channels with U-shaped cross-sections (Fig. 5.3H). They are restricted in length to the upper-slope and have high W/D ratios. Type *II* gullies exhibit the lowest gully densities (0.74 gully/km) within the area studied, with generally very low sinuosities and a cross-sectional shape which remains U-shaped with distance down-slope. Type *II* gullies are found in highest concentration at the mouth of the Filchner Trough and are absent from all other trough mouths, apart from in very low densities (along a 3.4 km length of the shelf edge) at Pine Island West (PIW) trough mouth. They are found in considerably lower densities along the shelf edge of the Bellingshausen and Amundsen seas where gully density does not exceed 0.1 gully/km. These gullies often cut back into the shelf edge and have concave shaped gully heads (Noormets et al., 2009). They exist mainly on relatively low slope gradients (~2.5º).

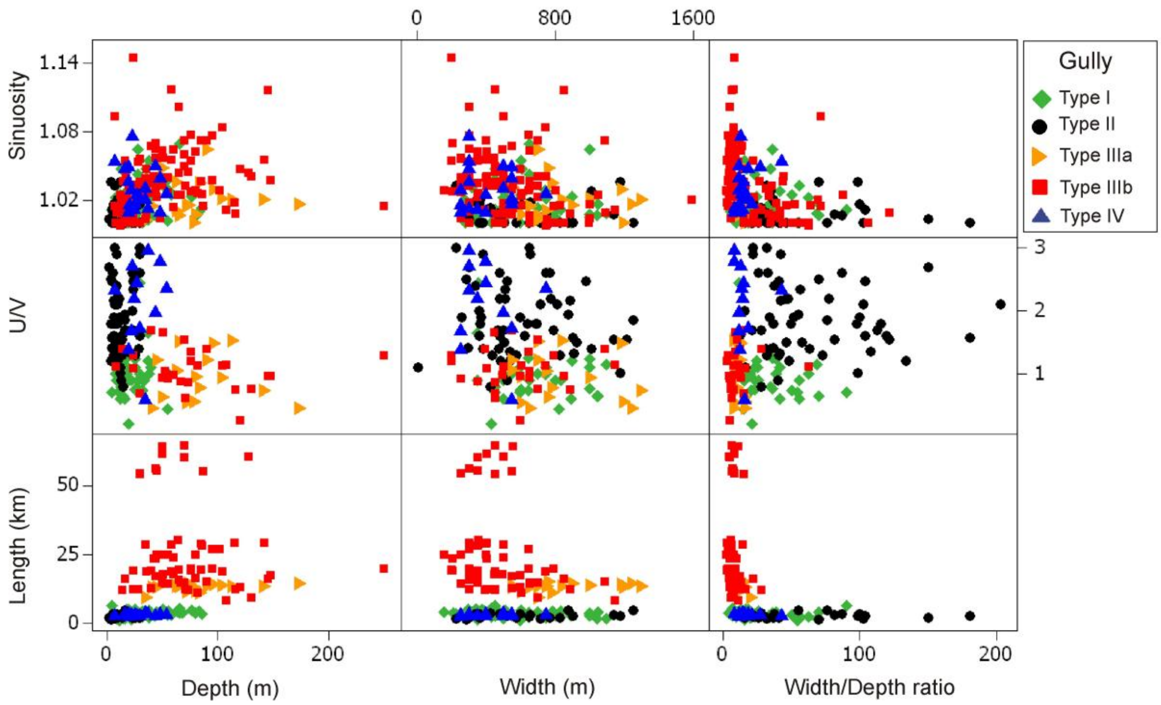

Figure 5. 5. Matrix plot of Antarctic gully parameters, including: gully depth, width/depth ratio, width, length, U/V shape (*b* value) and sinuosity. Gully types *I – IV* are indicated by different colours and symbols.

We classify types *IIIa* and *IIIb* as variants of the same basic gully type because they are morphologically similar in most respects, with V-shaped cross-sections at the shelf edge maintaining shape down-slope, and with long gully lengths (Fig. 5.4). Both types are characterised by low W/D ratios and a branching morphology. The main differences between types *IIIa* and *IIIb* are in sinuosity and spatial location. Type *IIIb* displays high sinuosity (~1.04 at trough margins) and develops both in the presence and absence of trough mouth fans (Belgica Fan). This type of gully is, however, unique to the Belgica and

PIW trough mouths, present only within the confines of the trough margins. Type *IIIa* displays low sinuosity (1.02) and covers 15% of the shelf edge, however is not found in any cross-shelf trough mouths. Type *IIIa* is common in relatively high densities along the shelf edge of the Amundsen and Bellingshausen Seas on gradients of 4-5º. Both types are absent from the WAP continental margin and Filchner Trough mouth.

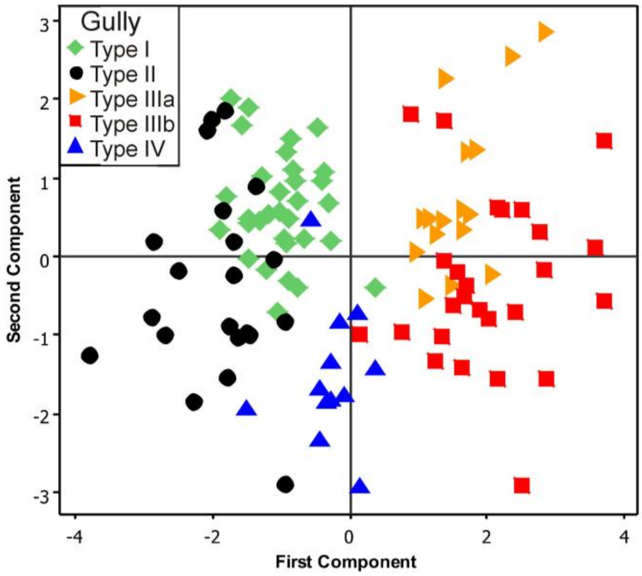

Figure 5. 6. Principal Component Analysis (PCA) of gully U/V index, sinuosity, length, width, depth and presence of branching, with different gully types indicated by different colours and symbols.

Type *IV* is characterised by a unique branching signature (Fig. 5.4K), with tributary branches merging down to approximately 1000 m water depth and forming a 'handprint-like' signature on the seafloor. Type *IV* displays low gully lengths, U-shaped cross-sections, low channel sinuosity and high W/D ratios. This gully type is unique to the WAP shelf edge, where it covers a small section (31 km; 2% of the margin studied) of shelf edge to the west of the Marguerite Trough in moderate density. TOPAS data from this region (Fig. 5.4K and 5.9) show that gullies form in sediment in which there is very low acoustic penetration, with thin lenses of acoustically transparent sediment observed in some areas.

Table 5. 4. Average gully parameters (U/V index, sinuosity, length, width and depth at 50 m below the shelf break)

Gully Type	Stream order	U/V index	Sinuosity	Length (km)	Width (m)	Depth (m)	W/D
I	1	0.9	1.02	3.4	534	33	14
II	1	1.88	1.01	2.7*	640	12	53
IIIa	2	1.00	1.02	14*	831	81	10
IIIb	3	1.07	1.02[a] / 1.04[b]	25.5*	450	64	7
IV	2	2.02	1.03	3	412	29	15

*Limited by data extent; [a]sinuosity at trough axis; [b]sinuosity at trough margins.

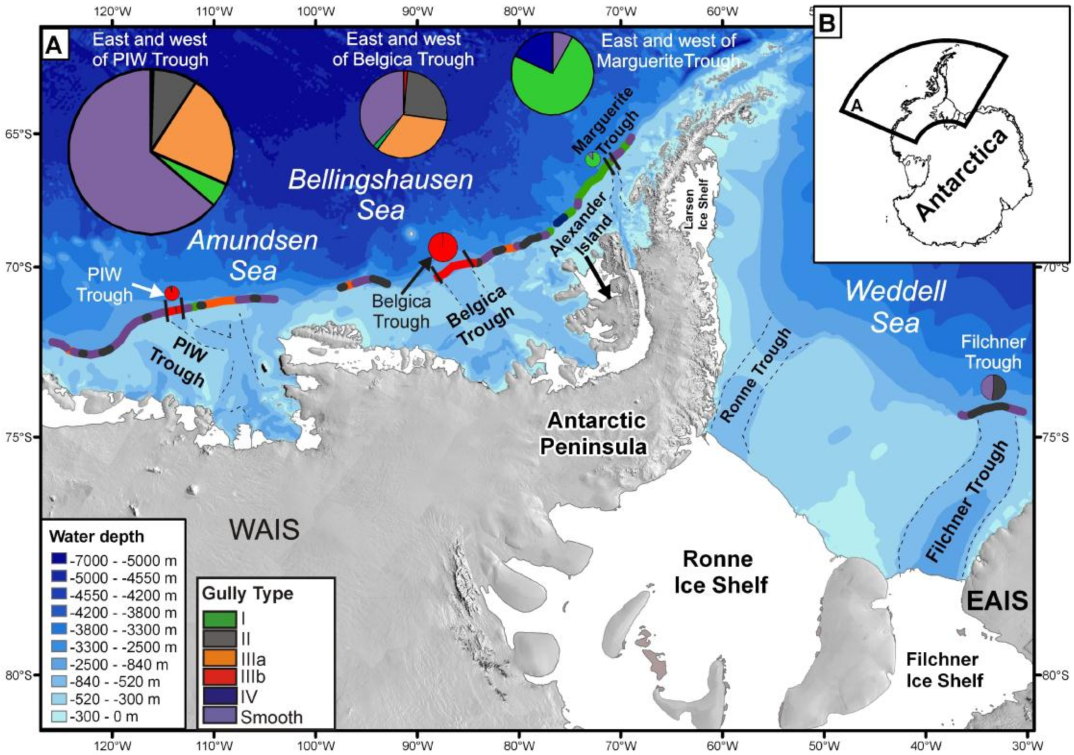

Figure 5. 7. Spatial distribution of gully types *I – IV* and topographically smooth seafloor on the Antarctic continental shelf edge. **A**. Pie charts represent % of seabed covered by each gully type over the Filchner Trough, Marguerite Trough, west Antarctic Peninsula, Belgica Trough, Bellingshausen Sea, Pine Island (West) Trough and Amundsen Sea continental shelf edge. Diameter of pie chart is proportional to km of continental slope analysed (for scale, see Table 5.6 for km of total margin). Colour of line along shelf edge represents predominant gully type present. Dotted black lines mark cross-shelf trough margins. **B.** Location of A on the Antarctic continent. PIW is Pine Island (West). EAIS is East Antarctic Ice Sheet. WAIS is West Antarctic Ice Sheet. Regional bathymetry is from Bedmap2 dataset (Fretwell et al., 2013). Antarctic continent is shaded Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007).

5.5. Discussion

The general form of the shelf edge and upper continental slope of the Antarctic continental margin has been well documented (e.g. Vanneste and Larter, 1995; Lowe and Anderson, 2002; Dowdeswell et al., 2004a; 2006; 2008; Gales et al., 2012), with Noormets et al. (2009) providing the first quantitative analysis of Antarctic gullies. Much uncertainty remains, however, about the variation in gully morphologies which occur on these margins and the processes responsible for shaping them.

Our quantitative analysis of gully parameters shows that gullies can be distinguished based on their geomorphic signature, with five separate gully types identified on parts of the Antarctic continental margin studied. Based on quantitative and statistical analysis, a set of distinguishing parameters is developed to identify gully type (Fig. 5.8). In the following sections, we attempt to interpret these features in relation to the slope characteristics and gully spatial distribution with the aim of constraining the processes that have operated in these environments (Table 5.7).

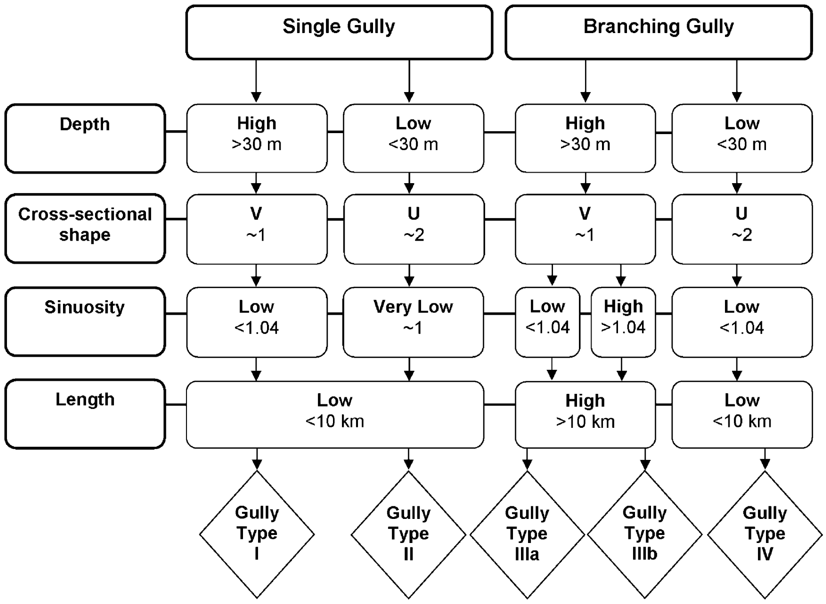

Figure 5. 8. Diagnostic diagram to identify Antarctic gullies based on the quantitative parameters taken at 50 m below the shelf edge including: gully branching order, depth, cross-sectional shape, sinuosity and length.

5.5.1. Influences on gully morphology

5.5.1.1. Subglacial meltwater discharge

Sediment-laden subglacial meltwater has been proposed as an Antarctic gully-forming mechanism (Wellner et al., 2001; 2006; Lowe and Anderson, 2002; Ó Cofaigh et al., 2003; Dowdeswell et al., 2006, 2008; Heroy and Anderson, 2005; Noormets et al., 2009), although there remains a lack of in-situ evidence to confirm this. At the mouths of cross-shelf troughs on the Antarctic continental margin, Noormets et al. (2009) observed a trend of increasing gully cross-sectional area from the centre to the margins of the troughs. Quantitative analysis of Antarctic gully parameters (Table 5.4) shows that not only does cross-sectional area increase toward trough margins, but also, sinuosity is generally higher at trough margins, with sinuosity of type *IIIb* gullies higher towards the eastern margin and decreasing toward the centre of both the PIW and Belgica Trough. This trend is likely to be due to additional focussing of subglacial meltwater flow to trough margins from the adjacent banks, and from within ice streams, due to subglacial water pressure gradients and additional heat generated by friction at the trough margins (Raymond, 2002; Noormets et al., 2009). Under ice streams, pressure at the ice-bed interface is lower than under inter-stream ridges due to the lesser ice thickness (Röthlisberger, 1972), causing meltwater under the slower moving ice at the banks of an ice-stream trough to be drawn down to the trough margins (Boulton et al., 2007; Vaughan et al., 2008). Thicker ice also commonly occurs in the central part of ice streams, resulting in subglacial pressure gradients that cause subglacial water to flow toward the ice stream margin (Dowdeswell and Elverhøi, 2002). Flow discharge may, therefore, have been higher within the troughs and particularly near the troughs margins.

Table 5. 5. Spatial distribution of slope types on Antarctic continental margins.

Gully type	Total gully no.	Total Abundance (km)	Total Abundance (%)	Average density (gully/km)	Within Trough	Inter Trough	TMF present	Average Slope (º)
I	483	398.5	21	1.54	Yes	Yes	No	13
II	138	233.0	12	0.74	Yes	Yes	Yes	2.5
IIIa	292	289.5	15	1.37	No	Yes	No	4-5.3
IIIb	170	213.6	11	1.16	Yes	No	Yes	1.7-4.5
IV	41	31.1	2	1.32	No	Yes	No	9
Smooth margin*	-	753.0	39	-	No	Yes	Yes	-

TMF = Trough mouth fan. *Smooth margin is defined here as topography with < 5 m undulations and/or gully density of < 0.01 gully/km.

Our interpretation of flow processes is, however, based purely on multibeam bathymetry and subbottom profiler data, and targeted high-resolution seismic profiles and sediment cores are needed to test this hypothesis.

5.5.1.2. Slope geometry, gradient and flow entrainment

Local continental slope characteristics may affect gully morphology by influencing processes operating in these environments. At Marguerite Trough mouth, slope geometry and gradient affect gully morphology. Gully type *I* is found in greatest abundance here (87% cover) and along the WAP continental margin. In front of Marguerite Trough, the length of type *I* gullies is limited to the upper slope, dying out at around 1300 m water depth where slope angle increases downslope. The restriction in gully length is likely controlled by the geometry of the slope, which is unique within the data examined. The upper-slope gradient here is relatively low compared with the rest of the WAP margin, increasing in gradient toward the mid-slope (Dowdeswell et al., 2004a), in contrast to slope profiles from Filchner, Belgica and PIW trough mouth which have almost uniform low gradients down-slope (e.g. Dowdeswell et al., 2006). The mid-slope gradient along the WAP margin is also considerably higher than on other continental margins in this study (~13º; Larter and Cunningham, 1993).

The observation that gullies die out downslope here is the opposite of what would be expected if turbidity current 'ignition', (sudden initiation and increase in energy; Parker, 1982; Fukushima et al., 1985), were the process responsible for gully erosion. A critical velocity and critical sediment concentration exists for turbidity current sustainability, and, when the values of these parameters fall below or rise above the critical values, the turbidity current dies or grows (Parker, 1982). In the case of type *I* gullies, the flow velocity must decrease with increasing slope angle for the submarine flow to die out down-slope and not to entrench the gully bed. A simple layer-averaged analysis of submarine flows can be made using the Chezy formula (e.g. Komar, 1969):

$$U = (RgChS / (C_d + e_w))^2 \qquad (5.3)$$

where C is sediment fractional concentration; h is flow thickness; S is bed gradient; e_w is ambient fluid entrainment coefficient; C_d is bed friction factor; U is flow velocity; g is acceleration due to gravity (m s^{-2}) and R is the submerged particle specific gravity ((ρ_s – ρ_w) / ρ_w, the ratio of sediment particle buoyancy density to water density ρ_w) (Mitchell, 2006). This implies that the product *ChS* must decrease downslope. For type *I* gullies, *S* increases down-slope and *h* is constrained by gully depth as flow thickness is limited by over-spilling. Assuming that other factors do not change considerably, a down-slope decrease in the product *ChS* requires the sediment concentration, *C*, to decrease along the flow path. For sediment concentration to decrease, the flow must entrain ambient fluid.

One hypothesis is that turbidity currents are formed by sediment-laden subglacial meltwater, which may erode the seabed and become self-maintained for a discrete period of time (Mulder and Syvitski, 1995). As mean entrainment velocity is proportional to the mean centre-line velocity of a plume (Morton et al., 1956), more seawater is entrained into the plume as flow velocity increases down-slope with increasing slope gradient. This dilutes the sediment concentration and the meltwater temperature anomaly. The reduction of *C* reduces the negative buoyancy effect of the sediment load, causing the flow to diminish and disperse. This reduces the erosive ability of the flow, and may produce the length-limited gullies observed along the WAP margin. This analysis of submarine flows may also be applied to cascading dense water overflow (with modified parameters in the Chezy formula (5.3) to accommodate density differences).

At Marguerite Trough mouth, the depth of the gullies also decreases before the observed increase in slope gradient at around 1300 m water depth. This differs from terrestrial observations of erosion of river knickpoints (i.e. rapids, waterfalls), in which regions of intense erosion are observed in the 'drawdown' reach leading up to the change in gradient, where flow starts to accelerate and shear stress is highest (Bishop and Goldrick, 1992). Although a decrease in gully depth is also observed down-slope for other gully types (Figs. 5.3D, H and 5.4D, H, L), these decreases in depth are accompanied with a small, gradual decrease in slope gradient, as opposed to an increase in slope gradient with type *I* gullies.

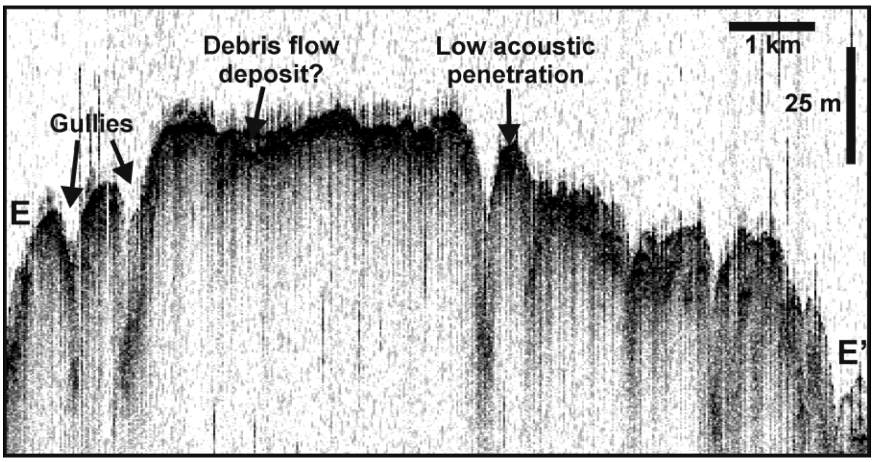

Figure 5. 9. TOPAS profile through shelf edge and upper continental slope offshore from Alexander Island, Antarctica. Location of profile (E-E') is marked by dashed black line in Fig. 5.4K.

Table 5. 6. Spatial distribution of slope types, expressed as km along the margin that they occur.

Gully Type	km cover						
	Filchner Trough	Marguerite Trough	WAP	Belgica Trough	Bellings-hausen	PIW Trough	Amundsen
Location	36º26'W-30 º30'W	71º51'W-70º45'W	77º30'W-71º51'W; 70º45'W-68º40'W	88º23'W-84º25'W	97º30'W-88º23'W; 84º25'W-77º35'W	113º00'W-114º3W	113º00'W-103º42'W; 125º5'W-114º3W
Average upper Slope º	2.5	9	13	1.7	4	4.5	5.3
I	0	74.5	281	0	7.8	0	35.2
II	68.2	0	0	0	97.7	3.4	63.7
IIIa	0	0	0	0	127	0	162.5
IIIb	0	0	0	152	6	51.6	4
IV	0	0	31.1	0	0	0	0
Smooth margin*	63	11.1	67.5	0	144.7	0	466.7
Total margin	131.2	85.6	379.6	152	383.2	55	732.1

WAP is western Antarctic Peninsula. PIW is Pine Island West. *Smooth margin is defined here as topography with < 5m undulations and with gully density of < 0.1.

5.5.1.3. Sediment yield and flow type

Quantitative analysis of Antarctic gully parameters shows a distinct difference in gully spatial distribution (Fig. 5.7), where type *IIIb* gullies occupy lower slope gradients and are restricted to upper slopes offshore of cross-shelf troughs (Belgica, Pine Island West Trough) and type *IIIa* gullies are found on higher slope gradients outside trough margins and display significantly lower average sinuosities (Fig. 5.5 and Table 5.4). Submarine channel sinuosity has been shown to vary on a global scale as a result of latitudinal differences (Peakall et al., 2012). Locally, however, factors including slope gradient, sediment type and flow type may all contribute to small-scale variations in sinuosity and W/D ratio. The difference in morphology between types *IIIa* and *IIIb* gullies may be influenced by increased sediment fluxes and/or subglacial meltwater discharges within cross-shelf troughs, as glacial and deglacial sediment yield varies along the Antarctic continental margin. Ice streams within cross-shelf troughs are major agents of sediment transport, draining huge volumes of ice and delivering sediment to the ice-stream terminus during glacial periods (Larter and Cunningham, 1993; Dowdeswell and Siegert, 1999; Ó Cofaigh et al., 2003; Dowdeswell et al., 2008). Less sediment is deposited outside trough margins due to much slower ice flow velocities in inter-ice-stream areas, and thus lower sediment delivery rates (Bindschandler and Scambos, 1991; Whillans and Van der Veen, 1997; Ottesen and Dowdeswell, 2009).

At Filchner Trough mouth, types *I* and *III* gullies are absent. This suggests that whereas local changes in gully morphology (i.e. between types *IIIa* and *IIIb*) may result from local differences in meltwater discharge or sediment fluxes, other factors such as bedload/ suspended load ratio may also be influencing gully morphology. Gullies with higher W/D ratios and U-shaped cross-sections (types *II* and *IV*) are more likely formed by greater bedload, whereas deeply incised and V-shaped gullies with lower W/D ratios (types *I*, *IIIa* and *IIIb*) are more likely formed by suspended sediment load (Schumm et al., 1963). By analogy to similar features in subaerial environments, we interpret type *II* gullies as slide scars (Gales et al., 2012). Type *II* gullies display flat-floored cross-sections and steep channel walls (Kenyon, 1987) and are probably formed by small-scale sediment failure on the upper slope, which is consistent with high bedload. It is important to consider that the present-day gully morphology may have been influenced by successive infilling which may have altered their original morphology. TOPAS data from the Filchner Trough mouth area, however, show that the gullies form in acoustically impenetrable sediment and are draped by a layer of acoustically transparent (likely post-glacial) sediment and for the most part maintain the original gully morphology (Gales et al., 2012).

5.5.1.4. Ancient gullies

The extent and dynamics of former ice sheets have important implications for shelf edge and upper slope geomorphology, affecting the location and size of ice-streams, sediment transport to the shelf edge, sub-glacial meltwater discharge and production of cold, dense water. Although the glacial histories of the Antarctic ice sheets are difficult to reconstruct and are poorly known in places, previous workers have placed maximum extents of ice during the LGM at, or near, the shelf edge in areas off the western Antarctic Peninsula (Pudsey et al., 1994; Vanneste and Larter, 1995; Ó Cofaigh et al., 2005a; Heroy and Anderson, 2005; Graham and Smith, 2012), Bellingshausen Sea (Ó Cofaigh et al., 2005b; Hillenbrand et al., 2010) and Amundsen Sea (Lowe and Anderson, 2002; Evans et al., 2006; Graham et al., 2010; Kirshner et al., 2012). The extent of grounded ice in the Weddell Sea during the LGM is much debated in published literature, but in some recent studies an extensive ice advance has been interpreted within the Filchner Trough as well as over shallower parts of the shelf (Hillenbrand et al., 2012; Larter et al., 2012).

Graham and Smith (2012) identified a small area near the shelf edge off Alexander Island, western Antarctic Peninsula, which displays no evidence for ice-grounding during the LGM. This area corresponds to the region of type *IV* gullies which are unique to this area and only span ~31 km of the shelf edge. Type *IV* gullies may reflect an ‘ancient’ slope which could be related to earlier glacial advances and which has undergone subsequent gradual modification through mass-wasting processes. The gullies may, therefore, be relict older features compared to other gullies along the Antarctic continental margin which likely date back to the LGM, where it has been established that ice advanced to the shelf edge in many places. If this is the case, this would suggest that different processes were responsible for forming the distinctly different type *I* gullies which dominate the WAP continental margin. If subglacial meltwater was discharged from the ice sheet grounding line, it may not have reached the shelf edge in this area and may have become trapped or diverted by the topography immediately landward of the shelf edge. In this case, the observed slope morphology may have been shaped by the cumulative effects of other processes, such as small-scale mass wasting, over a period of time extending back to

Marine Isotope Stage 6 or earlier. TOPAS data (Fig. 5.9) show that gullies form in sediment in which there is very low acoustic penetration, rather than being depositional features and formed by the aggradation of channel levees. The presence of potential debris flow deposits, indicated by semi-transparent and lens-shaped features on the TOPAS profile, shows that this region has been influenced by mass-wasting processes.

If our interpretation of type *IV* gullies is correct, representing gullies which were formed by earlier glacial advances and which were subsequently modified, this could provide an additional tool for interpreting where ice reached the shelf edge along other Antarctic continental margins during the LGM; the usual indicator is the presence of streamlined subglacial landforms on the outermost shelf (e.g. Ó Cofaigh et al., 2005a, 2005b; Evans et al., 2006; Graham et al., 2010). This has repercussions for reconstructing ice stream history, which may be used for testing and refining ice-sheet models and hence predicting future ice-sheet change.

5.5.1.5. Dense water overflow

Dense bottom water overflow is an active process in the Weddell Sea; however, it is not operative today off the western Antarctic Peninsula or in the Bellingshausen or Amundsen seas. Gales et al. (2012) showed that deeply incised and V-shaped gullies, as observed on other parts of the Antarctic continental margin (e.g. western Antarctic Peninsula, Bellingshausen and Amundsen sea margins), were absent from the mouth of the Filchner Trough, in an area of active and highly energetic cascading dense water overflow. On the upper slope at the mouth of the Filchner Trough, mean flow velocities of 0.38 m s^{-1} have been measured at 10 m above the seabed (Foldvik et al., 2004). Small-scale and U-shaped (type *II*) gullies are observed in this area which have been interpreted as having been formed by small-scale slides (Gales et al., 2012). The absence of deeply incised and V-shaped gullies indicates that there are no fluid erosional processes occurring, assuming that fluid flow produces V-shaped incisions as observed in the terrestrial environment (Simons and Sentürk, 1992). The V-shaped and highly erosive gullies found elsewhere along the Antarctic continental margins are therefore unlikely to have been formed by erosion caused by cold, high-salinity water overflows (Gales et al., 2012).

5.5.1.6. Gully evolution

A further consideration is the timescales over which gullies form, as different morphologies may reflect different stages of gully evolution. Closer examination of gully morphology shows that no significant steps exist between tributary channels and where these join at confluences, implying that either the gullies were active simultaneously, or the channels eroded down to a common resistant layer. If sediment-laden subglacial meltwater were the predominant process forming the gullies, this suggests that water was released over similar time periods and frequencies. Studies of subbottom acoustic profiles and sediment cores show that most gullies are overlain by post-glacial sediment (Melles and Kuhn, 1993), suggesting that they likely formed under full/deglacial conditions.

Table 5. 7. Summary of gully types and associated formation mechanisms.

Gully Type	Gully characteristics*	Key processes and formation mechanisms
I	Non-branching; V-shaped cross section; low sinuosity; low length; high depth.	Sediment-laden subglacial meltwater which is influenced by slope geometry and slope gradient.
II	Non-branching; U-shaped cross section; low sinuosity; low length; low depth.	Small-scale slides. Potential mechanisms for sliding include sediment failure along planes of weakness in the sedimentary structure due to changes in ocean currents or changes associated with glacial-interglacial sedimentation.
IIIa	Branching; V-shaped cross section; low sinuosity; high length; high depth.	Sediment-laden subglacial meltwater, which is associated with turbidity current generation. Morphology may be locally influenced by sediment yield, gradient and flow type.
IIIb	Branching; V-shaped cross section; high sinuosity; high length; high depth.	Sediment-laden subglacial meltwater, which is associated with turbidity current generation. Morphology may be locally influenced by sediment yield, gradient and flow type.
IV	Branching; U-shaped cross section; low sinuosity; low length; low depth.	Formed by previous glacial advance with subsequent modification (small-scale mass-wasting features).

*For explanation of what constitutes as high/low for gully parameters, see text and Figure 5.8.

5.6. Conclusions

Shelf edge and upper slope morphology along the Antarctic continental margin west of the Antarctic Peninsula, in the Bellingshausen and Amundsen seas and at the Filchner Trough mouth, is complex and reflects a balance between local slope character (slope geometry; gradient), large-scale spatial characteristics (i.e. subglacial meltwater production rates, drainage basin size, location of cross-shelf troughs, regional heat flow and strain heating distribution), ice-sheet history and sediment yield. From this study, we conclude that:

1. Five quantitatively distinct gully types are observed along the Antarctic continental margin (identified as *I, II, IIIa, IIIb* and *IV*, ranging from simple channels to complex branching networks), varying in gully length, width, depth, width/depth ratio, sinuosity, channel order, cross-sectional shape and spatial density. An identification scheme is presented that quantitatively and statistically distinguishes the different gully types on the parts of the Antarctic continental margin studied (Fig. 5.8).

2. Erosion by turbidity current flows initiated as a result of sediment-laden subglacial meltwater discharge is an important gully-forming mechanism along the Antarctic continental margin. Its effects are influenced by local slope character (i.e. slope

gradient and geometry) and regional factors (i.e. drainage basin size, location of cross-shelf troughs, ice-sheet extent), causing variation in gully morphology.

3. There is a distinct pattern in spatial distribution of gully types along the Bellingshausen Sea and Amundsen Sea margins. Gullies within trough margins (type *IIIb*) exhibit higher average sinuosity than outside trough margins (type *IIIa*) and sinuosity generally increases from the centre of the trough toward their margins. This pattern is interpreted as reflecting greater sediment delivery to trough mouths due to the presence of ice-streams within cross-shelf troughs at the LGM, focussing of subglacial meltwater towards trough margins due to pressure gradients at the ice bed caused by variations of ice thickness inside and outside the trough margins, and differences in slope gradients offshore from the cross-shelf troughs and on other parts of the margins.

4. Gully type *IV* may be a more ancient continental margin morphology which was not directly affected by glacial sediment deposition or subglacial meltwater discharge during the LGM. If so, this has implications for constraining the drainage pathways of past ice sheets and their full extent at glacial maximum.

Geomorphic parameters and spatial distributions of Antarctic gullies provide insight into the processes operating on Antarctic continental margins, and the factors which influence these processes (Table 5.7). Erosion by turbidity currents triggered as a result of sediment-laden subglacial meltwater provides one coherent hypothesis for gully erosion around much of the Antarctic continental margin. However, the mechanisms behind this, including meltwater availability under cold full-glacial conditions, remain poorly understood. This study presents observations that provide a basis for evaluating hypotheses about gully-forming mechanisms and offers insight into the processes operating on Antarctic continental margins.

5.7. Acknowledgements

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Chapter 6.

Arctic and Antarctic submarine gullies – a comparison of high latitude continental margins.

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6. Abstract

Submarine gullies are common features of high latitude continental slopes and, over the last decade, have been shown to play a key role in continental margin evolution, submarine erosion, down-slope sediment transport, slope deposits, and the architecture of petroleum reservoirs. However, the processes that form these gullies, the timescales over which they develop, and the environmental controls influencing their morphology remain poorly constrained. We present the first systematic and comparative analysis between Arctic and Antarctic gullies with the aim of identifying differences in slope character, from which we infer differences in processes operating in these environments.

Quantitative analysis of multibeam echosounder data along 2441 km of the continental shelf and upper slope and morphometric signatures of over 1450 gullies show that six geomorphically distinct gully types exist on high latitude continental margins. We identify distinct differences between Arctic and Antarctic gully morphologies. In the Arctic data sets, deep relief (> 30 m gully incision depth at 50 m below the shelf edge) and shelf-incising gullies are lacking. These differences have implications for the timescales over which the gullies were formed and for the magnitude of the flows that formed them. We consider two hypotheses for these differences: (1) some Antarctic gullies developed through several glacial cycles; and (2) larger Antarctic gullies were formed since the Last Glacial Maximum as a result of erosive flows (i.e., sediment-laden subglacial meltwater) being more abundant on parts of the Antarctic margin over longer timescales.

A second difference is that unique gully signatures are observed on Arctic and on Antarctic margins. Environmental controls, such as the oceanographic regime and geotechnical differences, may lead to particular styles of gully erosion observed on Arctic and Antarctic margins.

Keywords: geomorphology; submarine gully; continental margin; sedimentary processes; Antarctic; Arctic.

6.1. Introduction

Submarine gullies are small-scale, confined channels in the order of tens of meters deep and form one of the most common morphological features of high latitude continental slopes. Gullies also occur on mid- and low latitude margins and on hillslopes in the terrestrial environment (e.g. Hartmann et al., 2003; Micallef and Mountjoy, 2011; Vachtman et al., 2012; Lonergan et al., 2013). Different gully types are recognised on Arctic and Antarctic continental margins (Fig. 6.1) (Vorren et al., 1989; Laberg and Vorren, 1995; Laberg et al., 2007; Noormets et al., 2009; Pedrosa et al., 2011; Gales et al., 2013), but a distinct lack of knowledge about their formation processes and differences that exist between these regions remains. In Antarctica, gullies vary in length, width, incision depth, branching order, sinuosity, shelf-incision, cross-sectional shape, and mean gully spacing, with gully morphology influenced by environmental controls such as local slope character (i.e. slope gradient, geometry), large-scale spatial characteristics (i.e. drainage basin size, location of cross-shelf troughs, regional heat flow), ice sheet history and sediment yield

(Gales et al., 2013). For Arctic gullies, no systematic studies have to our knowledge been undertaken. By quantitatively examining Arctic and Antarctic gully morphology, we aim to identify variations in slope character that suggest differences in processes operating in these environments, factors influencing slope instability, and the timescales over which these features evolved.

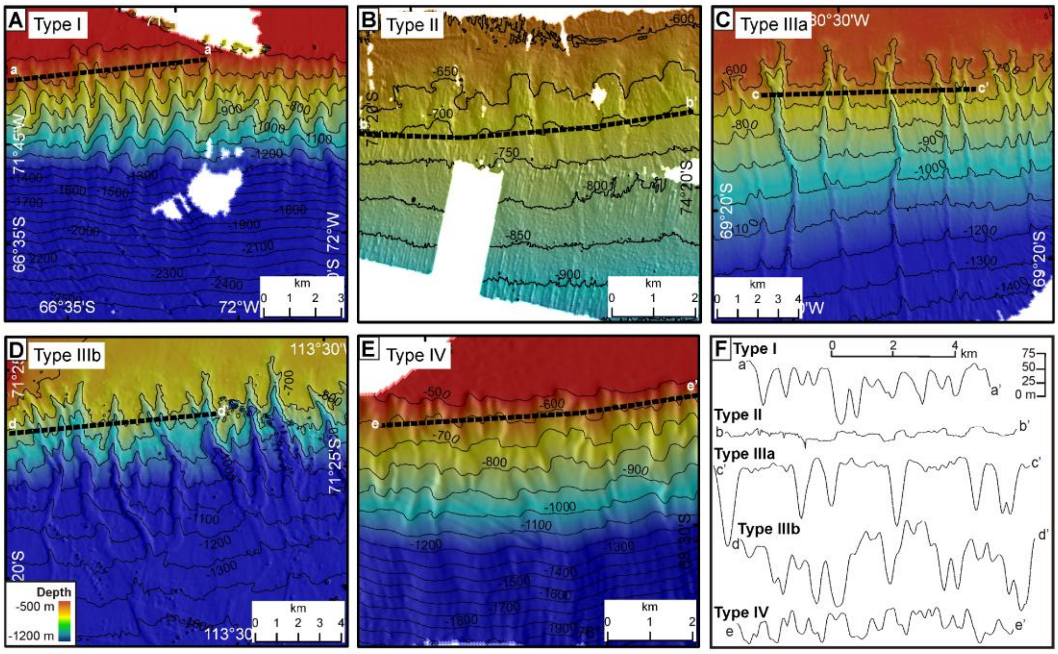

Figure 6. 1. Antarctic slope geomorphology. Different morphological gully styles observed across the Antarctic continental margin. **A.** Gully type *I*; **B.** gully type *II*; **C.** gully type *IIIa*; **D.** gully type *IIIb*; **E.** gully type *IV*; **F.** cross-shelf profiles at 50 m below the shelf edge for gullies A-E. Profile locations are marked by black dashed lines and lower case letters on each figure above. For location of A-E, see Fig. 6.2. Colour scale is the same for A-E. For colour bar, see D. Adapted from Gales et al. (2012).

Although small-scale variations in high latitude slope morphology remain largely unknown, studies examining the large-scale morphology show important differences (Dowdeswell et al., 1998; Ó Cofaigh et al., 2003; Nielsen et al., 2005). One striking difference between some Arctic and Antarctic continental margins is a distinct lack of modern large-scale features of slope instability (i.e. submarine slides) on the latter (Barker et al., 1998; Dowdeswell and Ó Cofaigh., 2002; Nielsen et al., 2005). Large-scale submarine slides are known to have occurred in the geological past (i.e. Miocene, early Pliocene) on the Antarctic margin, but there are few modern examples (Barker and Austin., 1998; Imbo et al., 2003; Diviacco et al., 2006). In contrast, slides are abundant on northern high latitude slopes (Damuth, 1978; Bugge et al., 1988; Laberg and Vorren, 1993; 2000; Dowdeswell et al., 1996; Vorren et al., 1998; Laberg et al., 2000; Evans et al., 2005; Nielsen et al., 2005). These differences have been attributed to variations in (i) the underlying geology, such as the presence of weak contouritic/ hemipelagic layers present on many Arctic continental margins that are mainly controlled by the ocean circulation (Long et al., 2003; Nielsen et al., 2005; Laberg and Camerlenghi, 2008); (ii) pore pressure, with Antarctic slopes displaying a greater stability owing to ice sheet compaction (Prior and Coleman, 1984; Larter and Barker, 1991); (iii) episodic and high quantities of sediment delivered to the shelf edge (Dowdeswell and Ó Cofaigh., 2002); and (iv) the timing of particular stages of glaciation (Nielsen et al., 2005).

Currently, there is a distinct lack of knowledge of the causes behind different gully morphologies observed on high latitude continental slopes. One potential gully-forming mechanism is erosion by hyperpycnal flows initiated as a result of discharges of sediment-laden subglacial meltwater (Wellner et al., 2001; 2006; Lowe and Anderson, 2002; Ó Cofaigh et al., 2003; Dowdeswell et al., 2004; 2006; Heroy and Anderson, 2005; Noormets et al., 2009). Studies of freshwater discharge (such as rivers, jökulhlaups and lahars) and laboratory experiments show that a critical sediment concentration of 1-5 kg m^{-3} is needed to initiate a hyperpycnal flow (taking into account fine-scale convective instability) (Parsons et al., 2001; Mulder et al., 2003). If enough sediment is entrained into subglacial meltwater released from beneath an ice sheet, and a critical sediment concentration is reached, then sediment-laden subglacial meltwater may initiate a hyperpycnal flow with the potential to erode the seafloor. Extensive relict subglacial meltwater channels are present on the Antarctic inner shelf, including the Marguerite and Pine Island troughs (Lowe and Anderson, 2002; Domack et al., 2006; Anderson and Oakes-Fretwell, 2008; Graham et al., 2009; Nitsche et al., 2013), potentially the outer shelf at the Belgica Trough (Noormets et al., 2009), and the inner shelf of the NW Barents Sea (Hogan et al., 2010). Studies of modern ice sheets, however, show that basal meltwater generated is in the range of millimetres per year (e.g. Beem et al., 2010; Pattyn, 2010;), with larger discharges suggested to result from processes such as subglacial lake outbursts or subglacial volcanic activity (Goodwin, 1988; Wellner et al., 2001; Dowdeswell et al., 2006; Fricker et al., 2007; Bell, 2008; Nitsche et al., 2013).

Other potential gully-forming mechanisms include erosion from mass flows initiated by factors such as gas hydrate dissociation, tidal pumping beneath ice shelves, shelf and contour currents, iceberg scouring, and large accumulations of glacigenic debris deposited at, or near to, the shelf edge during glacial maxima (Larter and Cunningham, 1993; Vanneste and Larter, 1995; Shipp et al., 1999; Dowdeswell et al., 2006; Dowdeswell and Bamber, 2007). Gullies may develop by aggradation between gully thalwegs (Field et al., 1999; Chiocci and Casalbore, 2011), with deposits forming from avulsing turbidity currents or levees of debris flows arising from their non linear rheology. Fine sediment, such as silt and clays may be deposited during low shear stress conditions, but may become eroded (resuspended) along gully thalwegs during the passage of the flows. Evidence from subbottom acoustic data from the Antarctic margin, however, shows that in most cases, gullies form in poorly stratified or acoustically impenetrable layers, contrasting good subbottom penetration that is expected on the gully interfluves formed by the aggradation (Gales et al., 2012). Gullies are suggested to be the first features to form on steeply dipping slopes in relation to mass wasting (Laberg et al., 2007), similar to terrestrial settings where gullies are known to represent the first step in the fluvial dissection of landscapes (Bloom, 1991).

The cascading of dense bottom water produced through sea-ice freezing and brine rejection may influence seafloor morphology (Vorren et al., 1988). However, recent studies find that in an area of active dense water overflow in the southern Weddell Sea, Antarctica, deeply incised gullies are absent. This suggests that this mechanism does not form the deeply incised and V-shaped gullies observed over much of the Antarctic continental margin (Gales et al., 2012).

By examining similarities and differences between high latitude gully morphologies, we aim to better understand what predisposes some slopes to particular styles of erosion, while other slopes appear relatively stable. We find that the variation in high latitude gully morphology has important implications for slope processes operating in these environments; the timing and rate of deglaciation and factors influencing slope stability.

6.2. Study areas

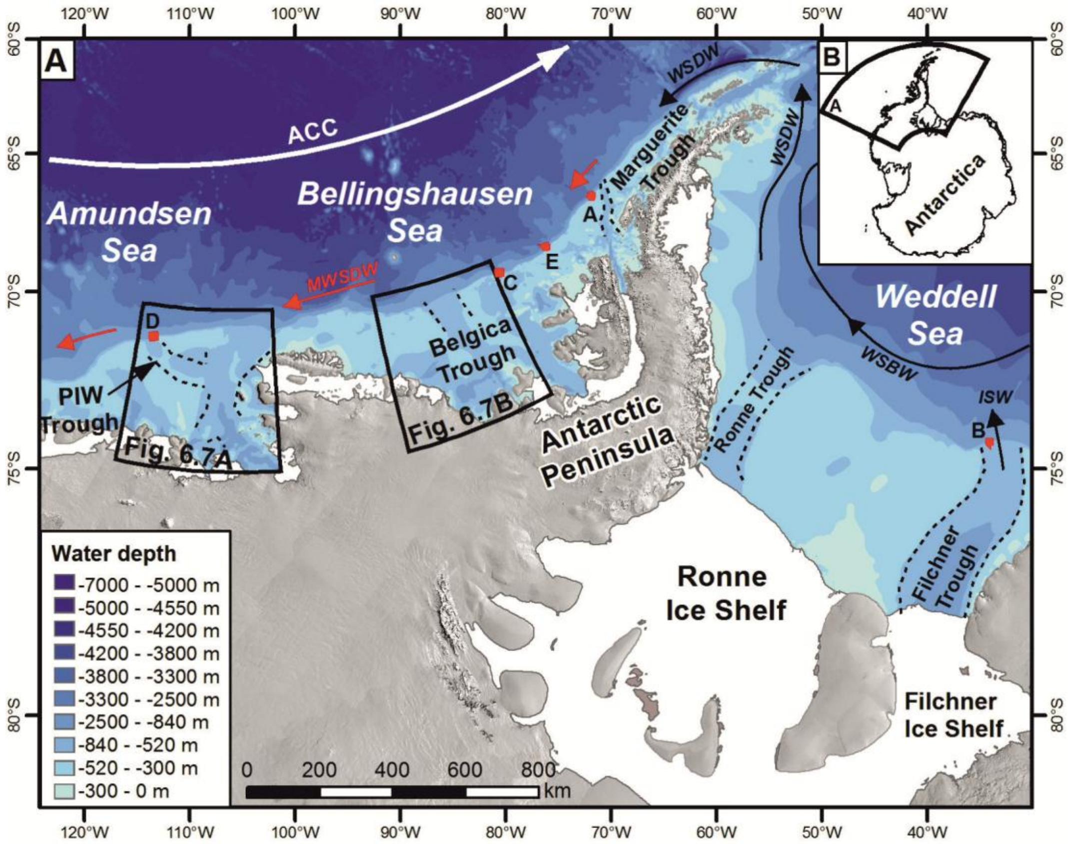

Figure 6. 2. **A**. Study areas from the western Antarctic Peninsula, Bellingshausen, Amundsen and Weddell seas. **B**. Inset shows location of A in relation to the Antarctic continent. Regional bathymetry is from Bedmap2 (Fretwell et al., 2013). The Antarctic continent is Landsat Image Mosaic of Antarctica (LIMA) (U.S Geological Survey, 2007). Black squares mark location of Figs. 6.7A and 6.7B. Red squares mark location of Figs. 6.1A-E. Arrows mark general direction of bottom currents including: Weddell Sea Bottom Water (WSBW), Weddell Sea Deep Water (WSDW) and Ice Shelf Water (ISW; black) and Modified Weddell Sea Deep Water (MWSDW; red). The Antarctic Circumpolar Current (ACC) is marked by a white arrow.

The study includes data along 1921 km of the Antarctic continental shelf and upper slope (Fig. 6.2), including data from the western Antarctic Peninsula, Bellingshausen, Amundsen and Weddell seas and 520 km of Arctic data from western Svalbard, the southwest Barents

Sea, and the northern Norwegian continental margins (Fig. 6.3). The tectonic history and geology of both regions have been extensively studied (e.g. Kenyon, 1987; Hübscher et al., 1996; Livermore and Hunter, 1996; Eagles et al., 2004; 2009; Evans et al., 2005; Stoker et al., 2005; Faleide et al., 2008). The uppermost beds comprise predominantly Quaternary sediment that is glacially derived or glacially influenced (i.e. glacigenic debris flows, glacigenic muds, and ice rafted debris) (e.g. Bonn et al., 1994; Vorren and Laberg, 1997; Vorren et al., 1998; Dahlgren et al., 2002, 2005; Dowdeswell et al., 2004; 2006; Hillenbrand et al., 2005; Cooper et al., 2008; Laberg et al., 2010). One key difference is the greater prevalence of thin hemipelagic/ contouritic sediments on parts of the Arctic continental slope, which form weak interglacial layers interbedded between the glacigenic sediments (Laberg and Vorren, 1995; 2000; Laberg and Camerlenghi, 2008). Although large mounds, interpreted as sediment drifts produced by bottom currents, occur on some areas of the Antarctic continental rise (Rebesco et al., 1996; 2007; Weber et al., 2011), there is little evidence for weak geological layers on the Antarctic upper slope (Nielsen et al., 2005).

Ice has covered much of Antarctica for the last 34 M.y (Barrett, 2008), advancing across the continental shelf from the early Oligocene in east Antarctica and from the Late Miocene on the west Antarctic margin (Nitsche et al., 1997). During the Last Glacial Maximum (LGM), the maximum extents of ice have been placed at, or near to, the shelf edge in areas off the western Antarctic Peninsula (Pudsey et al., 1994; Vanneste and Larter, 1995; Heroy and Anderson, 2005), Bellingshausen Sea (Ó Cofaigh et al., 2005; Hillenbrand et al., 2010b), Amundsen Sea (Lowe and Anderson, 2002; Evans et al., 2006; Graham et al., 2010), and the Weddell Sea, although this remains disputed (Hillenbrand et al., 2012; Larter et al., 2012). In the Arctic, the onset of glaciation occurred during the middle to late Miocene, with glacial-interglacial cycles along the Norwegian margin commencing from around 2.6 Ma (Thiede et al., 1998). The most extensive northern hemisphere ice sheets are suggested to have developed around 600-700 ka (Berger and Jansen, 1994; Wright and Flower, 2002), and have retreated during interglacial periods and since the LGM.

The Arctic and Antarctic have distinctive oceanographic regimes. The northern Norwegian and western Svalbard margins are influenced by the Norwegian and Barents seas. The Barents Sea is dominated by three major currents: the Arctic Current, the Norwegian Atlantic Current (NAC), and the Norwegian Coastal Current. The NAC affects the upper 800 m of the water column and transfers warm and saline water northeastward along the continental slope of the Norwegian Sea, with a branch entering the Barents Sea along with warm coastal waters (Orvik et al., 1995; Blindheim, 2004). The main branch becomes the West Spitsbergen Current to the west of Svalbard (Blindheim, 2004).

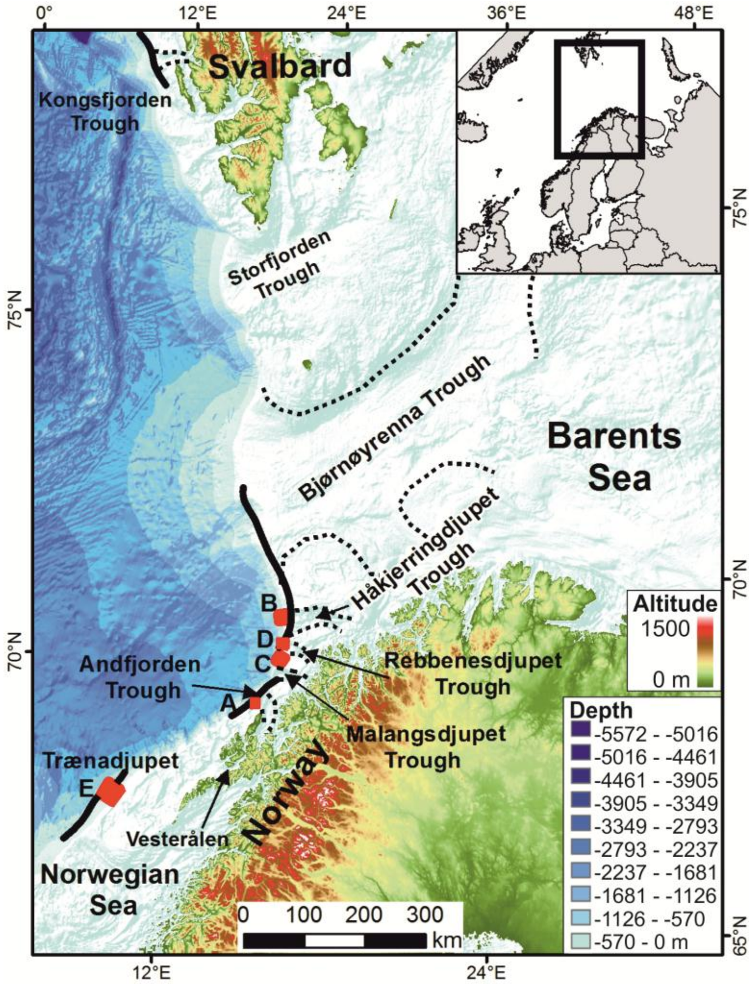

Figure 6. 3. Study areas from western Svalbard, southwest Barents Sea and northern Norwegian continental margins. Inset (upper right) locates the map within the European Arctic. Solid black lines mark extent of multibeam bathymetric data along continental shelf edge. Regional bathymetry is from IBCAO (Jakobsson et al., 2012). Dashed black lines mark boundaries of cross-shelf troughs analysed. Red squares mark locations of Figs. 6.4A-E.

In winter, cold, dense water forms through brine rejection during sea-ice production (Ivanov and Shapiro, 2005). Bottom current velocities of 0.2 m s^{-1} have been measured by current meters along the northern Norwegian slope (Heathershaw et al., 1998), with velocities of 0.2-0.8 m s^{-1} inferred from modelling studies, reaching maximum velocities of 1 m s^{-1} on the outer continental shelf of northern Norway (Bøe et al., 2009).

Cold, dense water overflow is also produced in the southern Weddell Sea, Antarctica (Nicholls et al., 2009). The cold dense water contributes to Weddell Sea Deep Water, which flows around the northern tip of the Antarctic Peninsula (Fig. 6.2; Nowlin and Zenk, 1988). Maximum flow velocities of 1 m s^{-1} were recorded at 2075 m water depth in the southern Weddell Sea (Foldvik et al., 2004). Southwestward flowing bottom currents are also documented on the slope and rise off the western Antarctic Peninsula margin (between 1000 and 4000 m deep) (Camerlenghi et al., 1997; Giorgetti et al., 2003), with maximum current velocities of 0.2 m s^{-1}. As currents with velocities > 0.06 m s^{-1} are able to transport silt- and clay-sized particles (Young and Southard, 1978; Singer and Anderson, 1984), bottom currents may influence the surface morphology by resuspension and transport of fine-grained sediment (Camerlenghi et al., 1997). Episodic and high energy currents, for example, produced by the sporadic presence of barotropic eddies, will increase shear stress and may cause short-term increases in sediment resuspension (Giorgetti et al., 2003).

6.3. Data and methods

Multibeam echosounder data used in this study are summarised in Table 6.1. The data were gridded to cell sizes of 25 to 50 m using either Kongsberg Simrad NEPTUNE software or public-access MB-System software (Caress and Chayes, 1996). In this study, a gully is considered to have an incision depth > 5 m and a gullied slope is one where gullies occur with a mean gully spacing of > 0.1 gully/km. Gully parameters were measured along transects parallel to the shelf edge at 50 m below the shelf edge. Analysis of the Trænadjupet region (Fig. 6.3) required measurements to be taken at 250 m below the shelf edge because of limited data availability. Measured parameters include gully width (distance between points of maximum curvature of gully flanks), incision depth (vertical distance from gully base to line defining gully width), length, branching order, sinuosity (distance measured along gully/straight line distance), mean gully spacing (gully/km), shelf-indentation (cutback), slope angle, and cross-sectional shape (U/V index). Cross-sectional shape was measured using the General Power Law (${}^{G}P_{L}$) programme (Pattyn and Van Huele*,* 1998). ${}^{G}P_{L}$ approximates the cross-sectional shape of a gully according to:

$$y - y_0 = a \mid x - x_0 \mid^{b} \qquad (6.1)$$

where *a* and *b* are constants and *x* and *y* are the horizontal and vertical coordinates taken from a cross-sectional profile of a gully. The programme automatically determines x_o and y_o as the coordinates of the point of inflection of the gully profile. The *b* value gives a measure of the cross-sectional shape of the gully and ranges from 1 (V-shape) to 2 (parabolic, commonly referred to as U-shape) on the U/V index. Branching order was calculated using the ArcGIS automated Stream Order application based on Strahler's (1957) method. Gullies were categorized according to their quantitative geomorphic signature based on the classification proposed by Gales et al. (2013). According to this classification, the following gully parameters are considered as 'high': incision depth > 30 m, sinuosity > 1.04, and length >10 km.

The statistical significances of the results were calculated using the T-test that tests whether differences between Arctic and Antarctic gullies were significant. Standard deviations were calculated to test whether the variances between gully parameters were greater between Arctic and Antarctic gullies than within the data sets.

Table 6. 1. Multibeam data analysed.

Data set		Reference	System
Cruise / ID	**Year**		
ANT23-4	2006	Gohl (2006); Nitsche et al. (2007)	Atlas Hydrosweep DS-2[1]
JR59	2001	Dowdeswell et al. (2004)	Kongsberg EM120[2]
JR71	2002	Ó Cofaigh et al. (2005); Dowdeswell et al. (2004)	Kongsberg EM120[2]
JR84	2003	Evans et al. (2006)	Kongsberg EM120[2]
JR97	2005	Gales et al. (2012); Larter et al. (2012)	Kongsberg EM120[2]
JR104	2004	Ó Cofaigh et al. (2005); Dowdeswell et al. (2008)	Kongsberg EM120[2]
JR141	2006	Noormets et al. (2009); Graham et al. (2010)	Kongsberg EM120[2]
JR157	2007	Noormets et al. (2009)	Kongsberg EM120[2]
JR179	2008	Graham et al. (2010)	Kongsberg EM120[2]
JR244	2011	Gales et al. (2012); Larter et al. (2012)	Kongsberg EM120[2]
NBP9902	1999	Wellner et al. (2001, 2006); Lowe and Anderson (2002)	Seabeam 2112[3]
NBP0001	2000	Nitsche et al. (2007)	Seabeam 2112[3]
NBP0103	2001	Bolmer (2008)	Seabeam 2112[3]
NBP0104	2001	Bolmer (2008)	Seabeam 2112[3]
NBP0201	2002	Wellner et al. (2006)	Kongsberg EM120[2]
NBP0202	2002	Bolmer (2008)	Kongsberg EM120[2]
NBP0702	2007	Graham et al. (2010)	Kongsberg EM120[2]
MAREANO		www.mareano.no	
R/V Jan Mayen	2004/5	Laberg et al. (2007)	Kongsberg EM300[4]
R/V Jan Mayen	2010	Baeten et al. (2013a)	Kongsberg EM300[4]
R/V Jan Mayen	2006/7	Hustoft et al. (2009)	Kongsberg EM300[4]
	2008/9	Forwick (2009a, b)	Kongsberg EM300[4]

[1]Frequency of 15.5 KHz and swath width of up to 120°; [2]Frequency range of 11.75-12.75 kHz and swath width of up to 150°; [3]Frequency of 12 kHz and swath width of up to 120°; [4]Frequency of 30 kHz and swath width of up to 150°.

6.4. Results

Six geomorphically distinct gully types exist along the 2441 km of high latitude continental shelf and upper slope data analysed, with unique gully types recognised on both Arctic and Antarctic slopes. Five geomorphically distinct gully types occur on the Antarctic continental margin (Fig. 6.1). These are described by Gales et al. (2013) as Type *I*: nonbranching with high incision depth (> 30 m), sinuosity (< 1.04) and length (< 10 km) and a V-shaped cross section; Type *II*: nonbranching with low incision depth, sinuosity, and length and a U-shaped cross-section; Type *IIIa*: branching with high incision depth and length, low sinuosity, and a V-shaped cross-section; Type *IIIb*: branching with a high incision depth, sinuosity, and length and a V-shaped cross-section; and Type *IV*: branching with a low incision depth, sinuosity, and length and a U-shaped cross-section.

On the Arctic margins analysed, all gullies have average incision depths of < 30 m, measured at 50 m below the shelf edge. Five geomorphically distinct gully types are identified when gully incision depth is excluded from the identification criteria (Fig. 6.4), including type *I*, *IIIa*, *IIIb*, *IV,* and a previously unclassified gully type '*V*', which is not observed in the Antarctic margin data analysed. Type *V* is characterised by a nonbranching, U-shaped, high length (> 10 km), low incision depth (< 30 m) and low sinuosity (< 1.04) signature. Gully type *I* covers 23% of the Arctic and 21% of the Antarctic margin data analysed; type *II* is not observed in the Arctic but covers 12% of the Antarctic margin data analysed; type *IIIa* covers 9% of the Arctic and 15% of the Antarctic margin data analysed; type *IIIb* covers 31% of the Arctic and 11% of the Antarctic margin data analysed; type *IV* covers 6% of the Arctic and 2% of the Antarctic margin data analysed; and type *V* covers 12% of the Arctic, but is not observed on Antarctic margins analysed (Antarctic values from Gales et al., 2013). Smooth continental slope, defined in this study as having a topography with gully depth incisions < 5 m, or a mean gully spacing of < 0.1 gully/km, covers 19% of the Arctic continental margin analysed and 39% of the Antarctic continental margin analysed. Mean gully spacing is similar between some gully types observed on the Arctic and Antarctic margins, including types *I* and *IIIb*, although spacing varies for types *IIIa* and *IV*, with Arctic gullies generally occurring in lower densities (Table 6.2).

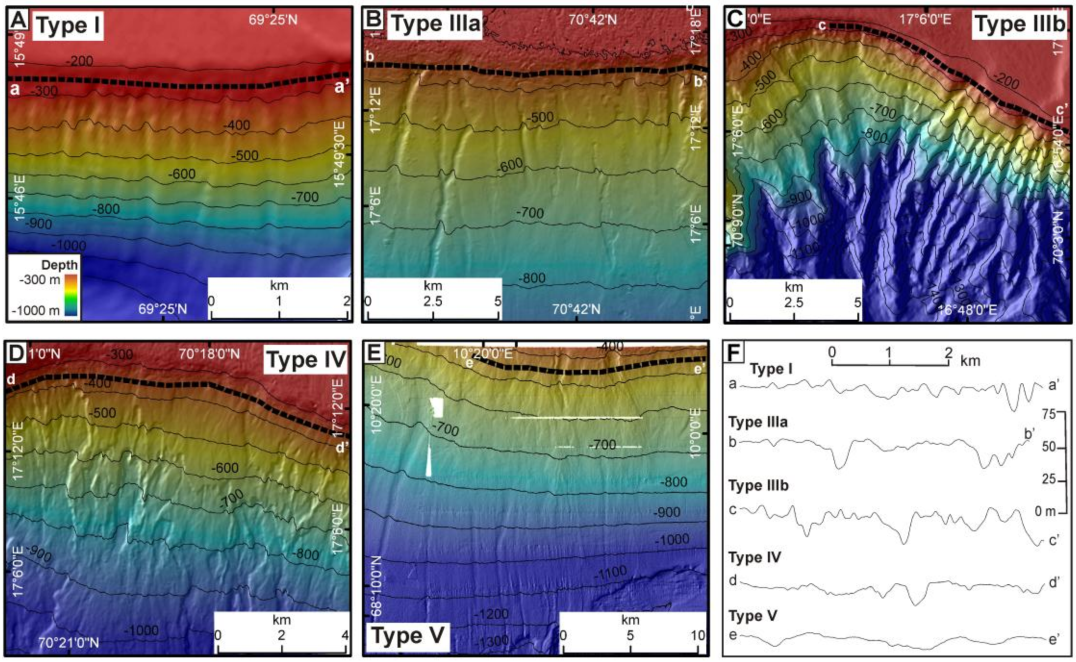

Figure 6. 4. A-E. Arctic gully morphologies from the northern Norwegian, Barents Sea, and western Svalbard margin. **F**. Cross-shelf profiles taken at 50 m* below the shelf edge (*e-e' is taken at 250 m below shelf edge because of data limitations). Profile locations are marked by black dashed lines and lower case letters on each figure above. For location of A-E, see Fig. 6.3. Colour scale is the same for A-E. For colour bar, see A.

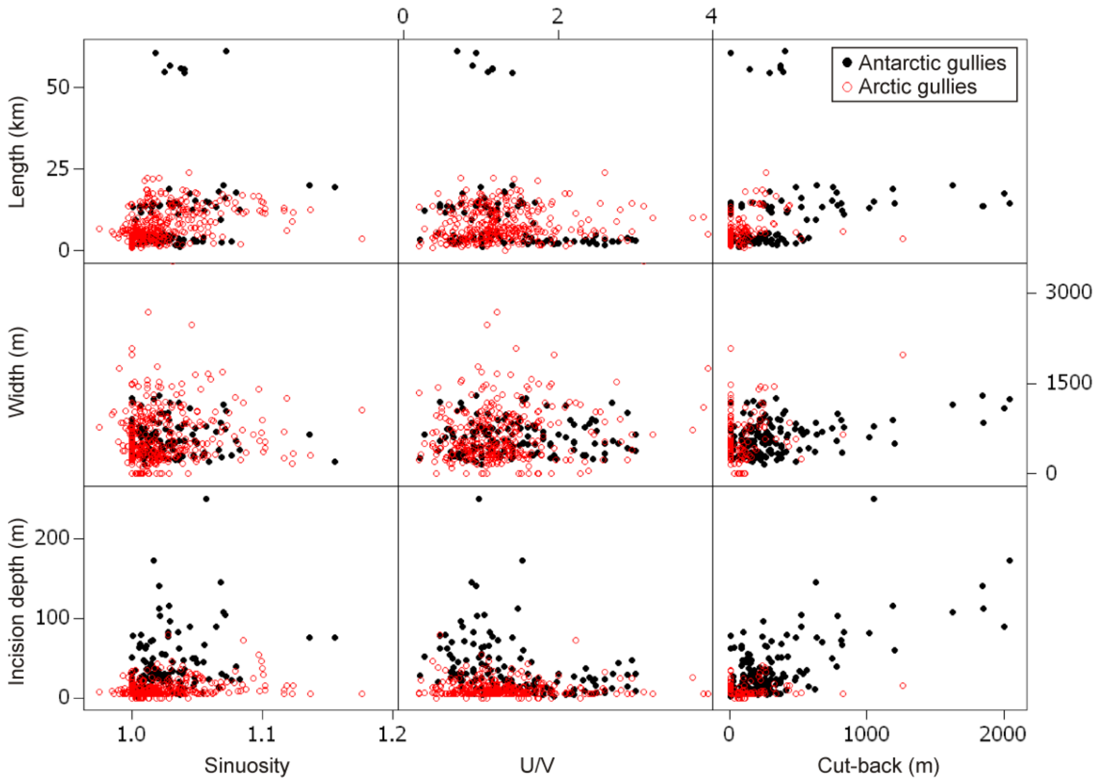

Figure 6. 5. Arctic and Antarctic gully parameters: length (km), width (m), incision depth (m), sinuosity, U/V, and cutback (m). Red circles = Arctic gullies; black circles = Antarctic gullies.

Two distinct differences in gully morphology are recognised between the Arctic and Antarctic datasets. Firstly, there is a distinct lack of deeply incised gullies within the Arctic data, with mean gully depths in each area studied not exceeding 30 m, measured at 50 m below the shelf edge (Fig. 6.5). Mean gully incision depth from the Antarctic margin is 48 m compared to 12 m for Arctic gullies. Secondly, there is a distinct lack of gullies incising the shelf edge, with a mean shelf-indentation of 35 m on the Arctic margins, compared to 324 m on Antarctic margins. The statistical significances of these differences were tested by calculating the standard deviations of the means and by using the T-test. The standard deviations of the mean gully incision depths and shelf-incisions for Arctic and Antarctic gullies were less than the difference between the means. This shows that the differences between Arctic and Antarctic gullies are greater than variances within each dataset. The results of the T-test shows that the P value is < 0.05, and so lies within the 95% significance level, suggesting that there is a significant difference between Arctic and Antarctic gully incision depth and shelf-incision.

Differences also exist in gully spatial distribution. On Antarctic margins, sinuous but long gullies (type *IIIb*) are located at the mouths of cross-shelf troughs, with less sinuous gullies (type *IIIa*) occurring in inter-trough areas (Table 6.3). However, on the Arctic margins analysed, sinuous gullies (type *IIIb*) are observed both in inter-trough regions and at the mouths of cross-shelf troughs (Fig. 6.6). Additionally, type *IV* gullies which are only observed on an isolated inter-trough region offshore from Alexander Island, western Antarctic Peninsula (Fig. 6.1E), are observed more widely on Arctic margins and occur at the mouths of cross-shelf troughs and in inter-trough areas (Fig. 6.6).

Table 6. 2. Mean gully spacing (gully/km) from parts of the Arctic and Antarctic continental shelf and upper slopes.

Gully type	Mean gully spacing (gully/km)	
	Arctic	Antarctic
I	1.56	1.54
II	-	0.74
IIIa	0.74	1.37
IIIb	1.07	1.16
IV	0.7	1.32
V	0.5	-

6.5. Discussion

Quantitative analysis of over 1450 high latitude continental slope gullies show that gullies share similar morphologies and, excluding gully depth, fit a common classification scheme. However, we identify two distinct differences between Arctic and Antarctic gullies. Firstly, there is a significant lack of deeply incised gullies within the Arctic data analysed. Secondly, there is a distinct difference in shelf-indentation with Antarctic gullies displaying greater cutback into the shelf edge. Like submarine canyons, gullies likely evolve through either down slope erosion driven by flows of turbidity currents, which may be initiated by discharges of sediment-laden subglacial meltwater, or through headward erosion by retrogressive mass failures (i.e. slides, slumps), or a combination of both (Harris and Whiteway, 2011). The differences in gully incision depth and shelf-indentation have implications for the timescales over that these features were formed and for the magnitude of the flows that potentially formed them. Questions still remain regarding the processes which form submarine gullies and whether they are formed over multiple glacial cycles by steady-state flows or develop over shorter timescales by larger episodic releases (Ó Cofaigh et al., 2012).

We suggest two hypotheses for the observed differences in gully incision depth and shelf-indentation: (i) some Antarctic gullies were formed over multiple glacial cycles, dating back to before the LGM; and (ii) the stages in which they were formed (i.e. glacial maximum and early deglaciation) prevailed over longer time periods in Antarctica than on Arctic margins during the last glacial cycle. Environmental controls, such as the oceanographic regime and geotechnical differences may also lead to particular styles of gully erosion on Arctic and Antarctic margins.

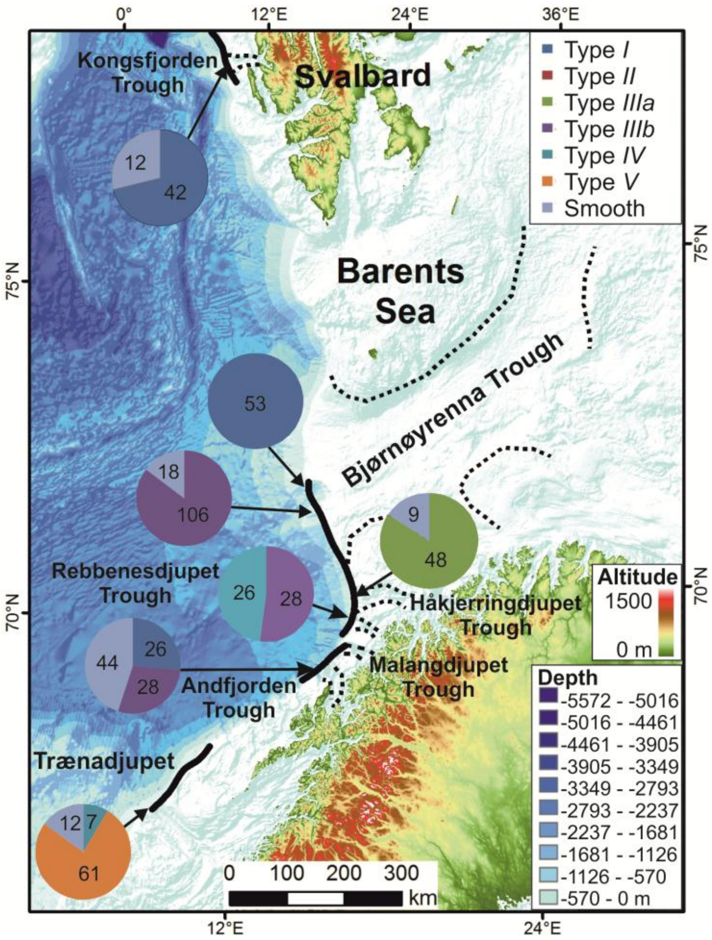

Figure 6. 6. Spatial distribution of Arctic gully types. Solid black lines indicate segments of shelf edge included in the analysis. Pie charts represent percentage of sea floor covered by different slope types. Numbers within the chart segments represent kilometres of seafloor covered by each gully type. Dashed lines mark cross-shelf troughs reaching the shelf edge in the areas analysed. Regional bathymetry is from IBCAO (Jakobsson et al., 2012).

6.5.1. Long-term versus short-term gully formation

Ice sheets are known to have advanced and retreated within cross-shelf troughs over many glacial cycles, and as a result of this, gullies may have formed during earlier glacial periods when ice was previously grounded at, or near to, the shelf edge. The significant differences in gully incision depth and shelf-indentation observed between some Arctic and Antarctic gullies may therefore reflect differences in the timing of initial glaciation and the number of subsequent glacial cycles. Ice has covered much of Antarctica for over 34 M.y. (Barrett, 2008), although onset of the west Antarctic glaciation is thought to have begun after 26 Ma (Barker et al., 2007) compared to 2.6 Ma on parts of the northwest European margin (Thiede et al., 1998). The difference in gully incision depth and shelf-incision may reflect a longer period of development over multiple glacial cycles for the larger gullies on the Antarctic margin. Recent studies using three-dimensional seismic data of submarine canyons on the northwest Mediterranean margin show that canyons may develop over longer timescales during continental margin construction (Amblas et al., 2012). Gullies present on mid-Pleistocene palaeocanyon flanks are maintained through to the modern canyon morphology (Amblas et al., 2012), demonstrating that the gullies can be preserved during slope progradation. If gullies on the Antarctic margin acted as conduits for glacigenic sediment, which was transported to the shelf edge during successive ice sheet readvances, and were reactivated in successive glacial stages, gully morphology may have been preserved and enhanced during slope progradation over multiple glacial cycles, leading to greater gully depths observed on parts of the Antarctic margin.

However, unlike the northwest Mediterranean margin, high latitude margins are influenced by huge quantities of glacigenic sediment transported by ice sheets to the shelf edge during glacial periods. This may have filled any shelf-edge depressions present, masking palaeogullies, and resulting in new gullies forming with each new glacial cycle. Subbottom data from some high latitude margins (e.g. Storfjorden Trough Mouth Fan, northern Barents Sea; Belgica Fan, Bellingshausen Sea, Antarctica) show that submarine gullies incise glacial debris flow deposits, suggesting that erosion of some gullies likely post-dates the last glacial advance to the shelf edge (Vorren et al., 1989; Laberg et al., 2007; Pedrosa et al., 2011; Ó Cofaigh., 2012). As the most recent episodes of gully erosion (since LGM) may obscure previously developed gullies, it is difficult to rule out gully formation over longer timescales without high resolution three-dimensional seismic datasets from glacial margins.

6.5.2. Timing and rate of deglaciation

The length of time that ice was grounded at the shelf edge during the LGM and the timing of particular glacial stages may affect gully size and shelf-incision. If the dominant mechanism forming V-shaped and deeply incised high latitude gullies is erosive flows initiated by sediment-laden subglacial meltwater (Lowe and Anderson, 2002; Ó Cofaigh et al., 2003; Dowdeswell et al., 2004; 2006; 2008; Heroy and Anderson, 2005; Noormets et al., 2009; Ó Cofaigh, 2012; Gales et al., 2013), this suggests that either these processes were operating over longer time periods in Antarctica, were more frequent, or occurred in larger events, for example becuase of larger drainage basin sizes or from the outer shelf morphology facilitating flow toward the shelf edge.

Gullies with the greatest incision depth and shelf-incision (types *IIIa* and *IIIb*) are located along the Bellingshausen and Amundsen Sea margin and at the mouths of the Pine Island West (PIW) and Belgica troughs, Antarctica (Figs. 6.1C and D; Table 6.3). The difference between the closely spaced and smaller type *I* gullies observed on some Arctic margins and the larger, more widely spaced type *III* gullies may reflect a difference in the length of time that the gullies developed over, perhaps becuase of differences in the length of time that ice was present near to the shelf edge and the effective drainage areas associated with the palaeo-ice streams. A larger drainage basin size and/or longer presence of ice near the shelf edge may increase the size and frequency of flows down the gully thalwegs.

Table 6. 3. Comparison of submarine gully coverage (km) along high latitude continental margins.

	Western Antarctic Penin-sula		Bellings-hausen Sea		Amundsen Sea		Filchner Trough (Weddell Sea)	Kongsfjorden Trough (Western Svalbard)	South-west Barents Sea			Northern Norway		
	a	*b*	*c*	*d*	*e*	*f*			*g*	*h*	*i*	*j*	*k*	*m*
Type *I*	75	281		8		35		42	53				26	
Type *II*				98	3	64	68							
Type *IIIa*				127		163					48			
Type *IIIb*			152	6	52	4				106		28	28	
Type *IV*		31										26		7
Type *V*														61
Smooth	11	68		145		467	63	12		18	9		44	12
Inside trough	X	-	X	-	X	-	X	X	X	-	X	X	X	-
Inter-trough	-	X	-	X	-	X	-	-	-	X	-	-	-	X
Slide scars present	-	-	-	-	-	-	X	X	X	-	-	-	X	X
Canyon present	-	-	-	-	-	-		-	-	-	-	-	X	-
Gradient > 5°	X	X	-	-	-	X	-	-	-	-	-	-	X	-
Density* (gully/km) > 1.3	X	X	-	X	-	X	-	X	-	-	-	X	X	-

a = Marguerite Trough; b = west Antarctic Peninsula inter-trough; c = Belgica Trough; d = Bellingshausen Sea margin; e = Pine Island West Trough; f = Amundsen Sea margin; g = Bjørnøyrenna Trough; h = SW Barents Sea margin; i = Håkjerringdjupet Trough; j = Rebbenesdjupet Trough; k= Andfjorden (Andøya) Trough; l = Trænadjupet. Values for Antarctic gully coverage taken from Gales et al. (2013). *Of predominant gully type.

Although only limited sediment cores are available from the outer shelf of Belgica and PIW troughs, ice streams within these troughs are suggested to have undergone slow and episodic retreat, with mean retreat rates of ~15 m y^{-1} for Belgica Trough and 18 m y^{-1} for the Amundsen Sea Embayment outer shelf (Lowe and Anderson, 2002; Ó Cofaigh et al., 2008; Graham et al., 2010; Hillenbrand et al., 2010a; Smith et al., 2011; Kirshner et al., 2012; Livingstone et al., 2012). Conversely, some Arctic ice streams underwent significantly more rapid and later retreat. Studies suggest that deglaciation commenced around ~17.5 cal. ky BP. on the western Barents Sea margin and Norwegian continental shelf and 20.5 ± 5 cal. ky BP. on the outer western Svalbard shelf (Jessen et al., 2010), with core evidence showing that deglaciation was relatively rapid, occurring over a period of ~2000 years, with retreat rates of 60 m y^{-1}, increasing to 275 m y^{-1} for the Bjørnøyrenna ice stream (Vorren and Plassen, 2002; Landvik et al., 2005; 2012; Winsborrow et al., 2012). A longer duration of grounded ice at the shelf edge and/or slower initial ice-sheet retreat history would have exposed the shelf edge to more continuous turbidity-current or subglacial meltwater activity for longer time periods, facilitating the development of larger features over time.

Belgica and PIW troughs are characterised by large drainage basin sizes, with areas of 417,000 km^2 for Pine Island Trough (Rignot et al., 2008) and 217,000-256,000 km^2 for Belgica Trough (Livingstone et al., 2012). Although Bjørnøyrenna Ice Stream is also associated with a large drainage basin (576,000 km^2; Elverhøi et al., 1998), other Arctic margins analysed are characterised by significantly smaller drainage basin sizes, e.g., Kongsfjorden Trough (1426 km^2; Elverhøi et al., 1998) and by narrow continental shelves along the northern Norwegian margin. Ice streams within larger drainage basins are associated with slower or more episodic retreat rates in Antarctica (Livingstone et al., 2012) and greater sediment yields (Elverhøi et al., 1998). The difference in drainage basin size may therefore influence the rate of ice retreat as well as potentially influencing the abundance or frequency of subglacial meltwater discharged from beneath an ice sheet. Mechanisms for the production of subglacial meltwater include strain heating and geothermal heat flux at the base of the ice sheet (Joughin et al., 2004). The generally longer ice stream flow paths in larger drainage basins may have resulted in warmer basal ice and therefore higher abundances of subglacial meltwater, although studies of modern systems have shown that basal meltwater generated is in the range of millimetres per year (e.g. Beem et al., 2010; Pattyn, 2010).

The PIW and Belgica troughs have relatively unusual outer shelf morphologies (Figs. 6.7A and B). The PIW Trough is characterised by a 're-entrant' feature where the outer shelf is slightly seaward-sloping compared to most other Antarctic outer shelves that slope landward as a result of erosional over-deepening of the inner shelves by palaeo-ice sheets and lithospheric flexure from ice sheet loading (ten Brink and Cooper, 1992; Bart and Iwai, 2012). This may encourage the gravity-driven flow of subglacial meltwater toward the shelf edge, leading to larger and more deeply incised gullies located at the mouths of these troughs. The outer shelf of Belgica Trough is also slightly seaward sloping and asymmetric, with a 'subtrough' to the west of the trough axis (Graham et al., 2011). The unusual trough bathymetry may have enabled slower rates of glacial retreat and may have funnelled flows of subglacial meltwater toward the shelf edge during early stages of deglaciation.

6.5.3. Environmental controls

Although large-scale variations in high latitude gully morphology may result from varied glacial histories, local differences in gully morphology may result from different environmental controls, including slope gradient, oceanographic, and geotechnical differences. Within the data analysed, the most deeply incised gullies (type *IIIa*; mean incision depth of 81 m at 50 m below the shelf edge) occur on relatively low slope gradients of ~4-5.5° in Antarctica. The deepest gullies on Arctic margins (type *IIIb*; mean incision depth of 21.5 m at 50 m below the shelf edge) occur on higher slope gradients of ~7.5º (Fig. 6.8). No relationship is observed between mean gully incision depth or shelf-indentation and slope gradient, suggesting that gradient is unlikely to be influencing the differences observed between high latitude margins.

The occurrence of a previously uncharacterised gully type identified on the Arctic margin (type *V;* Baeten et al., 2013) suggests that different processes are operating in this region, that are absent from the Antarctic margin. Type *V* covers 61 km of the northern Norwegian continental shelf and upper slope and is found on relatively low slope gradients (~2°). Although no shelf edge data is available for this region at present, backscatter data show that the gullies are filled with sediments that are characterised by high backscatter (Baeten et al., 2013), suggesting that sediment within the gullies is coarser grained than the surrounding sediment. Seismic data from this region show that a glacigenic wedge formed of debris flow deposits is present on the upper slope, with gullies initiating within, or at the boundary of the wedge (Baeten et al., 2013., Submitted). The type *V* gullies may therefore have been formed by deposition on to the wedge during past glaciations that initiated sediment gravity flows. The local oceanographic regime may also influence the unique gully morphology, for example, by strong contour current activity activating sediment flows over the shelf edge and potentially allowing sediment gravity flows to develop. More detailed oceanographic data is needed from this region to confirm this.

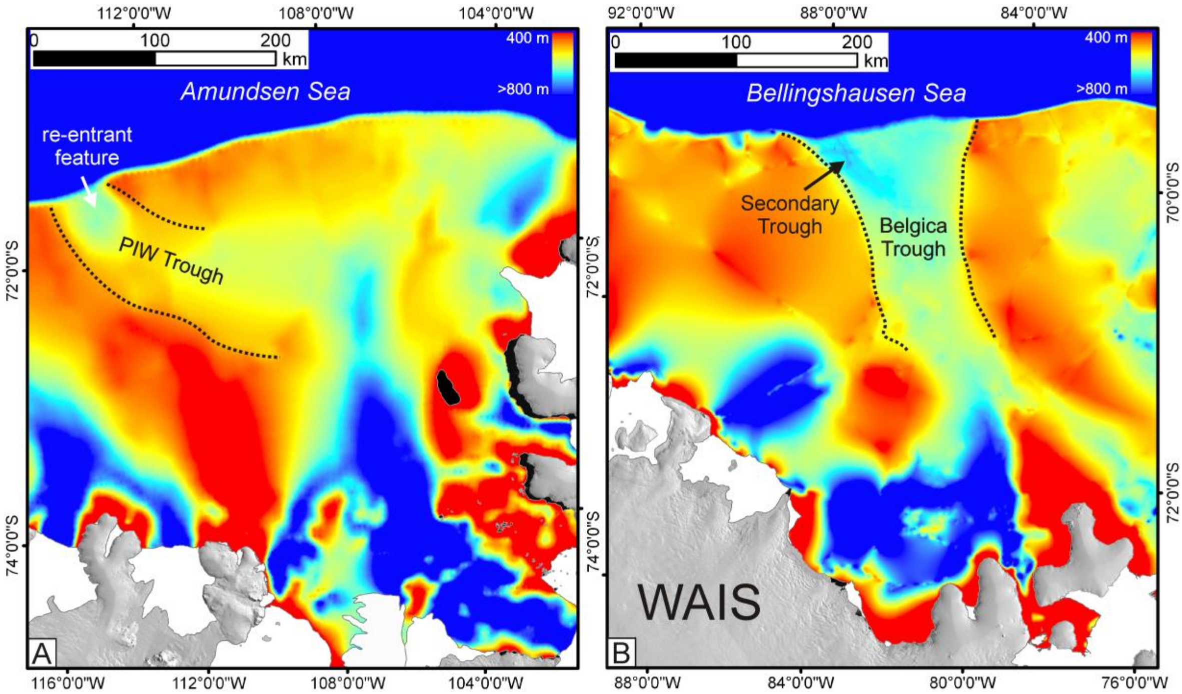

Figure 6. 7. A. Pine Island West (PIW) Trough, Amundsen Sea. **B.** Belgica Trough, Bellingshausen Sea. Bathymetry data is from Bedmap2 (Fretwell et al., 2013). Antarctic continent is shaded Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007). WAIS is West Antarctic Ice Sheet. Dashed lines mark boundaries of cross-shelf troughs.

Along the northern Norwegian margin, variation in along-slope sedimentation may influence differences in local sedimentation rates (Laberg et al., 2002). This may affect gully infilling leading to a reduction in the apparent incision depths on Arctic margins and may be one factor influencing the large discrepancy in high latitude gully incision depths. On the outer shelf offshore from Vesterålen (Fig. 6.3), bottom current velocities of 0.7-0.8 m s^{-1} have been suggested from modelling studies, reaching maximum velocities of 1 m s^{-1} (Bøe et al., 2009). The presence of strong bottom currents influences local sedimentation, with erosion occurring in high-energy environments and fine-grained sediment accumulating within sheltered environments (Bøe et al., 2009). However, even if some Arctic gullies did undergo significant infilling, this would not explain the distinct difference in shelf-incision. Subbottom acoustic data from the Arctic and Antarctic margins show that gullies are not significantly influenced by sediment infilling and largely maintain their original gully morphologies (Vorren et al., 1989; 1998; Gales et al., 2012; 2013). Additional subbottom data from Arctic margins are needed to rule out this possibility.

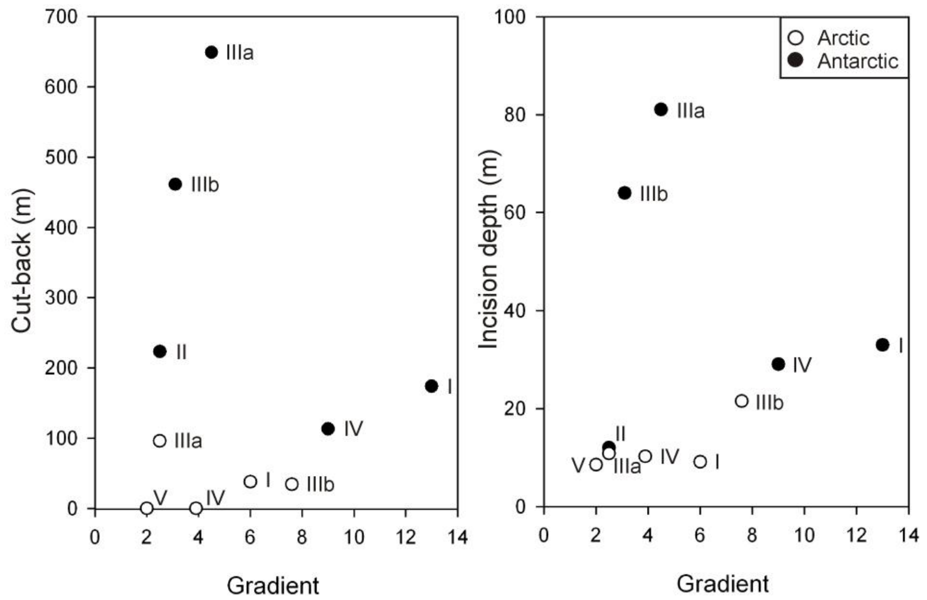

Figure 6. 8. Mean cut-back (m) and gully incision depth (m) versus slope gradient (º) for Arctic (hollow circles) and Antarctic (filled circles) gully types (*I*, *II*, *IIIa*, *IIIb*, *IV*, *V*).

6.5.4. Continental slope instability

Type *IV* gullies are relatively rare on high latitude continental margins, covering 6% of the shelf edge of Arctic margins analysed and 2% of Antarctic margins analysed. In Antarctica, type *IV* gullies are only observed in an inter-trough region and are absent from the mouths of cross-shelf troughs. New data from the Arctic show that similar gully morphologies occur on inter-trough margins and at the mouths of a cross-shelf trough (i.e. Rebbenesdjupet Trough; Fig. 6.6). Here, type *IV* gullies occur adjacent to an area of deeply incised and dendritic type *IIIb* gullies. Type *IV* gullies may indicate a continental slope that has undergone subsequent gradual modification through mass-wasting processes (Gales et al., 2013) and show characteristics similar to small-scale slides, exhibiting small-scale escarpments and an absence of a well-defined gully thalweg (Kenyon., 1987). The greater occurrence of type *IV* gullies on the Arctic margins may indicate a higher susceptibility to small-scale mass-wasting processes, for example, owing to the presence of weak geological layers on parts of the Arctic upper slope.

6.6. Conclusion

The mechanisms involved in gully formation and the factors that influence this are highly complex and poorly constrained. This study identifies key differences between high latitude submarine gullies and provides insight into the causes behind the different morphologies observed and the processes operating in these environments. The major findings of this study are as follows:

- Quantitative analysis of over 1450 gullies from multibeam bathymetric data along 2441 km of high latitude continental shelf and upper slope identifies six gully types from a common gully classification scheme.

- Two distinct differences exist between some Arctic and Antarctic continental slopes. There is a lack of deeply incised (> 30 m gully incision depth at 50 m below the shelf edge) and shelf-incising gullies in the Arctic continental margin datasets that we examined. Secondly, unique gully signatures are observed on Arctic and Antarctic margins, with type *II* gullies absent from Arctic margins and type *V* gullies absent from Antarctic margins analysed.

- We suggest two hypotheses for the distinct differences in gully incision depth and shelf-incision: (i) some Antarctic gullies were formed over multiple glacial cycles, dating back to before the LGM; and (ii) the stages in which gullies were formed prevailed over longer time periods in Antarctica than on Arctic margins, during and since the LGM. A longer period of grounded ice at the shelf edge and/or slower and more episodic retreat history of ice streams within Belgica and PIW troughs may have allowed the deeply incised type *IIIb* gullies to develop over longer timescales. Environmental controls, such as the oceanographic regime and geotechnical differences, likely influenced particular styles of gully erosion (i.e. type *IV* and *V*).

- The unusual outer shelf morphology of both the Belgica and PIW troughs may have encouraged the gravity-driven flow of subglacial meltwater toward the shelf edge as the ice sheet retreated. Further research is needed in reconstructing detailed ice sheet history, as flow switching within ice streams and glacial readvance may also have influenced gully morphology.

- Our study highlights the need for three-dimensional seismic datasets from the Antarctic continental margin, in particular for the outer shelf and upper slope regions. These data would constrain the evolutionary stages of high latitude gullies and the timescales they develop over, as well as providing insight into the mechanisms influencing submarine erosion and slope instability on high latitude continental margins.

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Chapter 7.

Southern Weddell Sea shelf edge geomorphology: Implications for gully formation by the overflow of high-salinity water.

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7. Abstract:

Submarine gullies are the most common morphological features observed on Antarctic continental slopes. The processes forming these gullies, however, remain poorly constrained. In some areas, gully heads incise the continental shelf edge, and one hypothesis proposed is erosion by overflow of cold, dense water masses formed on the continental shelf. We examined new multibeam echo-sounder bathymetric data from the Weddell Sea continental slope, the region that has the highest rate of cold, dense water overflow in Antarctica. Ice Shelf Water (ISW) cascades down-slope with an average transport rate of 1.6 Sverdrups (Sv) in the southern Weddell Sea. Our new data show that within this region, ISW overflow does not deeply incise the shelf edge. The absence of gullies extending deeply into the glacial sediments at the shelf edge implies that cold, high salinity water overflow is unlikely to have caused the extensive shelf edge erosion observed on other parts of the Antarctic continental margin. Instead, the gullies observed in the southern Weddell Sea are relatively small and their characteristics indicative of small-scale slides, probably resulting from the rapid accumulation and subsequent failure of proglacial sediment during glacial maxima.

7.1. Introduction

Submarine gullies are distinct, small-scale, confined channels, forming the most common morphological features observed on the Antarctic continental slope. Gullies exist along the shelf edge and upper continental slope to the west and northeast of the Antarctic Peninsula, and in the Bellingshausen, Amundsen, Ross, and Weddell seas (Vanneste and Larter, 1995; Shipp et al., 1999; Lowe and Anderson, 2002; Michels et al., 2002; Dowdeswell et al., 2004, 2006, 2008; Heroy and Anderson, 2005; Noormets et al., 2009). Along most of the length of these margins seismic reflection and coring studies show that the gullies form in sediments deposited in glacially influenced environments (e.g. Cooper et al., 2008). Many authors have described the lithology, physical properties, grain-size composition and mineralogical composition of continental slope sediments incised by gullies (e.g. Anderson and Andrews, 1999; Michels et al., 2002; Dowdeswell et al., 2004, 2006; 2008). These properties are often homogenous and show a similar range of characteristics to glaciomarine and subglacial diamictons on the adjacent shelf (Hillenbrand et al., 2005, 2009). Variation in substrate type is therefore unlikely to be an important factor controlling gully geomorphological expression. The abrupt and angular shape of the gully interfluves in cross-section suggests that they are formed by erosion of the channels, rather than by aggradation of the interfluves (Dowdeswell et al., 2008). Numerous morphological gully styles are observed on Antarctic continental slopes (Fig. 7.1), varying in gully size (length, width and depth), branching order, sinuosity, extension into the shelf edge, cross-sectional shape and spatial density.

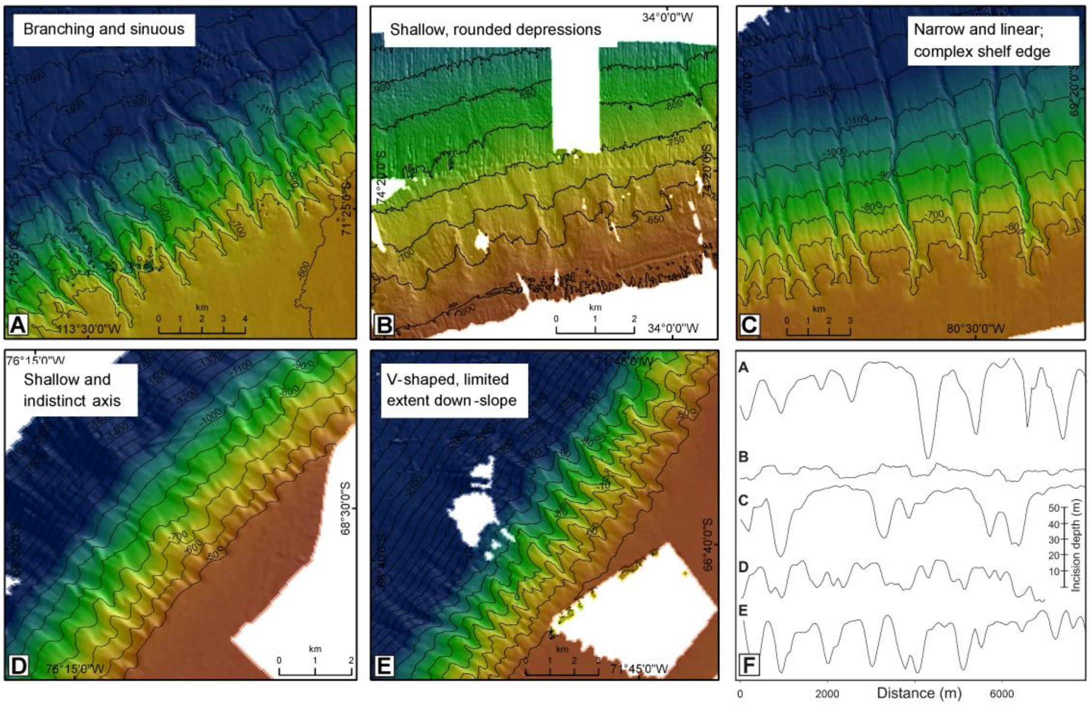

Figure 7. 1. Antarctic slope geomorphology: Different morphological gully styles observed across Antarctic continental margins. **A**. Branching and sinuous gullies in the Amundsen Sea. **B.** Shallow, rounded depressions in the southern Weddell Sea. **C.** Narrow and linear gullies with complex shelf edge in the Bellingshausen Sea. **D.** Shallow gullies with indistinct gully axes on the western Antarctic Peninsula margin. **E**. V-shaped gully with limited expression down-slope on the western Antarctic Peninsula margin. **F.** Cross shelf profiles at 50 m below shelf edge for gullies A-E. See Table 7.1 for data sources.

The variability in gully morphologies reflects the complexity of erosional processes occurring on high latitude continental margins where it is likely that different processes, or a different balance of processes, have resulted in different morphological styles. By constraining these gully forming mechanisms, we increase our understanding of continental slope processes, seafloor erosion patterns and continental margin evolution, which will help to interpret sediment core records from the continental slope and rise. This will also enable us to gain a better understanding of how gullies develop as precursors to the more major features of continental slopes, such as canyons. Antarctic gully formation has been attributed to erosion by: a) mass flows, such as sediment slides, slumps, debris flows and turbidity currents, with triggering mechanisms including resuspension by shelf and contour currents, gas hydrate dissociation, tidal pumping beneath large icebergs and near ice shelf grounding lines, iceberg scouring, tectonic disturbances and rapid accumulations of glaciogenic debris at the shelf edge during glacial maxima (Larter and Cunningham, 1993; Vanneste and Larter, 1995; Shipp et al., 1999; Michels et al., 2002; Dowdeswell et al., 2006, 2008); b) subglacial meltwater discharge from ice-sheet grounding lines during glacial maxima or deglaciations, whether by constant release (Wellner et al., 2001; Dowdeswell et al., 2006; 2008, Noormets et al., 2009), or more episodic and large scale release (Wellner et al., 2006) possibly by meltwater evacuation from subglacial lakes (Goodwin, 1988; Bell, 2008); and c) dense water overflow (Kuvaas and Kristoffersen, 1991; Dowdeswell et al., 2006, 2008; Noormets et al., 2009).

The potential for cascading, dense bottom water to erode gullies is not well documented or understood; however such dense water overflow has been proposed as one potential mechanism for gully erosion in the Bellingshausen, Amundsen and Weddell seas (Kuvaas and Kristoffersen, 1991; Dowdeswell et al., 2006; Noormets et al., 2009). Most authors, however, favour the hypothesis of ice-sheet derived basal meltwater and related mass-wasting on the upper slope (Wellner et al., 2001; Lowe and Anderson, 2002; Dowdeswell et al., 2006, 2008; Noormets et al., 2009). Although dense water overflow is not an active process off the western Antarctic Peninsula or in the Bellingshausen and Amundsen Sea today, the possibility that it may have been active during previous stages of glaciation cannot be dismissed. This mechanism for gully erosion is also suggested on Norwegian margins, such as Bear Island Trough (Vorren et al., 1989; Laberg and Vorren, 1995, 1996) and at mid-latitude margins (Micallef and Mountjoy, 2011).

In the terrestrial environment, fluvial erosion typically produces V-shaped incisions (Simons and Sentürk, 1992), and submarine fluid flow is widely thought to generate similarly-shaped gullies (e.g. Micallef and Mountjoy, 2011) and channels (e.g. Lonsdale, 1977). If dense water overflow was the mechanism responsible for forming the highly incisional and V-shaped gullies found on other Antarctic continental margins (Fig. 7.1), we would expect a similar seafloor signature on the southern Weddell Sea margin, where dense bottom water overflow is well documented (e.g. Nicholls et al., 2009).

In this paper, we present new morphological data from the shelf edge and continental slope in the southern Weddell Sea. We present a quantitative analysis of the gully morphology present and discuss the potential for cascading cold, dense water overflow to have eroded these features.

7.2. Study Area

The Weddell Sea lies to the east of the Antarctic Peninsula (Fig. 7.2). The Filchner and Ronne Ice Shelves form floating extensions of the East and West Antarctic Ice Sheet, respectively, covering 450,000 km^2 (Nicholls et al., 2009). The continental shelf edge lies mostly between 500-600 m water depth, increasing to around 630 m within the Filchner Trough (GEBCO, 2003). Filchner Trough is a major bathymetric cross-shelf feature, extending seaward of the Filchner Ice Shelf and reaching a width of 125 km at the shelf edge. The continental slope seaward of the Filchner Trough mouth is characterised by outward-bulging contours, marking the presence of a Trough Mouth Fan (Crary Fan) (Kuvaas and Kristoffersen, 1991).

The Weddell Sea is a major region of bottom water formation, contributing ca. 50-70% of the 10 Sv (1 Sv = 10^6 m^3 s^{-1}) of Antarctic Bottom Water (AABW) which is exported from the Southern Ocean (Naveira Garabato et al., 2002; Nicholls et al., 2009). AABW forms the southern component of the global thermohaline circulation and is responsible for cooling and ventilating the abyssal world ocean (Foldvik et al., 2004).

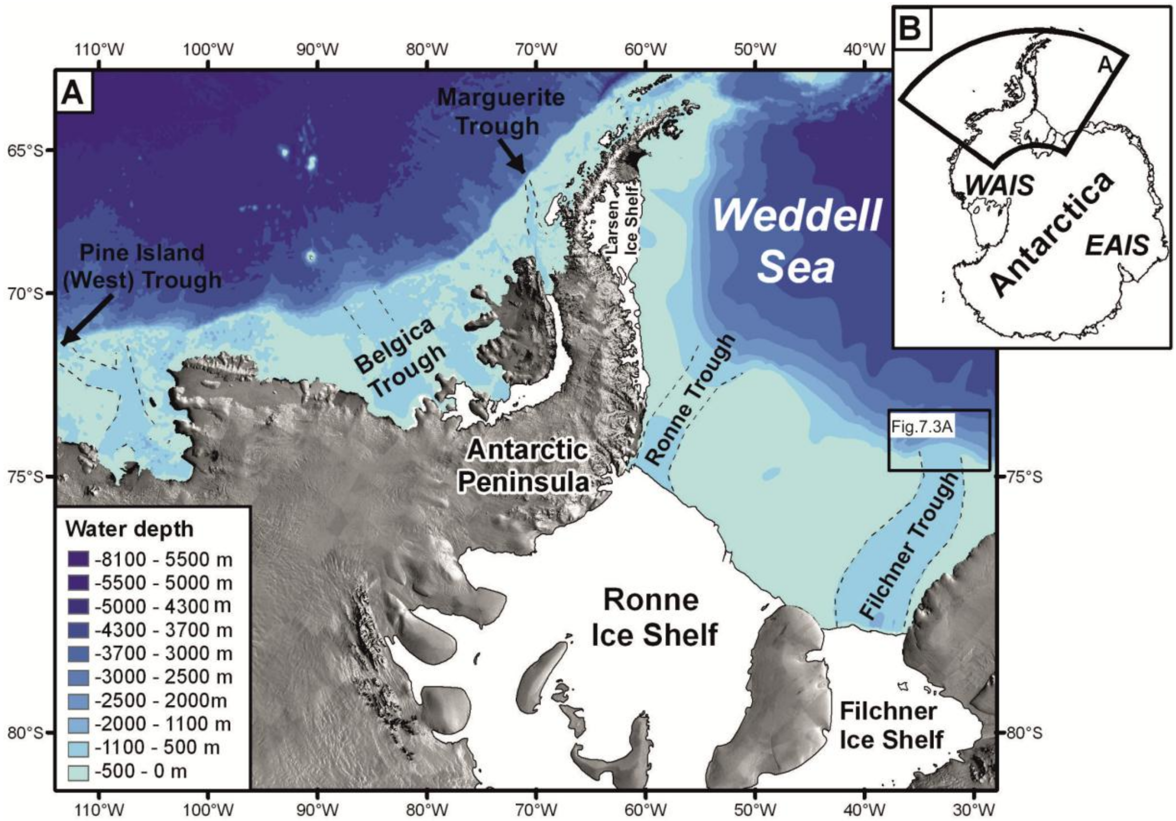

Figure 7. 2. A. Location map of the study area on the Antarctic continental margin, showing locations of major cross-shelf troughs mentioned in text, including: Pine Island (West) Trough, Belgica Trough, Marguerite Trough, Ronne Trough and Filchner Trough. **B**. Inset shows the spatial extent of Fig. 7.2A. Regional bathymetry is from GEBCO dataset (GEBCO, 2003). Antarctic continent is shaded Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007). WAIS is West Antarctic Ice Sheet; EAIS is East Antarctic Ice Sheet.

High Salinity Shelf Water (HSSW) is produced during sea ice production through brine rejection. HSSW is subsequently supercooled and slightly freshened by circulation beneath the ice shelves, producing cold and dense Ice Shelf Water (ISW) (Nicholls et al., 2009).

Although past volume fluxes of ISW are difficult to estimate, this must have been closely related to circulation changes in the clockwise flowing Weddell Gyre. These, in turn, were closely related to circulation changes within the west-wind driven Antarctic Circumpolar Current. Studies of marine and terrestrial palaeo-climate archives (e.g. Bianchi and Gersonde, 2004; McGlone et al., 2010) did not reveal any significant changes in circulation of the Weddell Gyre or the westerly wind system for the time after ca. 8 ka before present, implying that any major changes in ISW production during the middle and late Holocene are unlikely. During the Last Glacial Maximum (LGM), when the ice sheet is thought to have advanced across the Weddell Sea shelf (Hillenbrand et al., 2012), ISW production likely ceased as ice shelf cavities, needed to super-cool HSSW, would not have existed if the ice-sheet grounding line had reached the shelf edge. The maximum extent of grounded ice during the LGM is under debate but a recent review concluded that ice streams had reached the outer shelf of the Weddell Sea and grounded within the deepest parts of the Filchner and Ronne Troughs (Hillenbrand et al., 2012). Hillenbrand et al.

(2012) interpreted sediments recovered in cores from the southern Weddell Sea shelf to be of subglacial origin and likely to be of LGM age. The latter conclusion is based on: (1) radiocarbon ages of glaciomarine sediments overlying the subglacial sediments giving predominantly ages younger than the LGM; (2) current velocity measurements of bottom currents on the shelf, which are deemed unlikely to have eroded a widespread unconformity separating subglacial deposits from the overlying Holocene glaciomarine deposits; and (3) radiocarbon dates of post-LGM age obtained from glaciomarine sediments overlying terrigenous deposits on the continental slope, which indicate that glaciogenic detritus was supplied directly to the Weddell Sea shelf edge during the LGM (Hillenbrand et al., 2012).

ISW flows toward the shelf edge through the Filchner Trough, where it cascades down slope. Volume transport increases down slope from an estimated 1.6 Sv to 4.3 Sv due to mixing with surrounding Weddell Deep Water (WDW) (Foldvik et al., 2004; Wilchinsky and Feltham, 2009). Mean flow velocities of 0.38 m s^{-1} have been measured on the upper slope (10 m above seabed) with maximum current velocities of 0.8 m s^{-1} recorded, increasing to 1 m s^{-1} downslope (Foldvik et al., 2004). Calculating the bed shear stress (τ_0) from the mean flow velocity according to:

$$\tau_0 = C_d \rho |u|^2 \qquad (7.1)$$

where C_d is a friction factor, found typically to be 0.0025 (dimensionless), ρ is water density (kg m^{-3}) and $|u|$ is depth-averaged flow speed (m s^{-1}), the resulting τ_0 value (0.37 Pa) is within the general threshold for erosion of marine muds (0.05 – 2 Pa), although at the lower end of the erosion threshold for sandy muds (0.1 – 1.5 Pa) from experiments reviewed by McCave (1984) and Jacobs et al. (2011). According to flume experiments by Singer and Anderson (1984), the mean current velocities measured on the shelf may be sufficiently high to winnow clay and silt particles resuspended by bioturbation from the seabed. Calculations are however based on mean flow velocities and high flow speeds associated with documented bursts within the flow (Foldvik et al., 2004) will increases shear stress and may cause short-term erosion of the seabed.

Surface seafloor sediments on the outer shelf consist largely of sand and gravelly sand due to winnowing of finer grained particles by ISW (Melles et al., 1994). Cores from the outermost shelf (Fig. 7.3A) recovered glaciomarine and subglacial sediments, including over-consolidated deposits and diamictons (e.g. Elverhøi, 1984; Melles and Kuhn, 1993; Hillenbrand et al., 2012).

7.3. Methods

During cruise JR244 with RRS *James Clark Ross* in early 2011, a 177 km stretch of the Weddell Sea continental shelf edge and upper slope was imaged using a Kongsberg EM120 multibeam swath bathymetry system, with a frequency range of 11.75-12.75 kHz and swath width of up to 150° (Fig. 7.3A). A TOPAS PS 018 subbottom acoustic profiling system was used to image the sub-surface. The TOPAS system transmits two primary frequencies at around 18 kHz, from which 10 to 15 ms-long secondary chirp pulses containing frequencies ranging from 1300 to 5000 Hz were generated. In this configuration the system can image acoustic layering in unconsolidated, fine grained sediments to a depth of more than 50 m below the sea-floor with a resolution better than 1 m. Data were digitally recorded at a sampling rate of 20 kHz. Traces were cross-correlated with the secondary transmission pulse signature and instantaneous amplitude records derived from the correlated output were displayed as variable density traces. Navigation data were acquired using a Seatex Global Positioning System receiver.

Table 7. 1. Data sets used in Figure 7.1[a].

Data set			Reference / PI (Principal Investigator)
Figure	**Cruise / ID**	**Year**	
1A	JR84	2003	Evans et al. (2006); Dowdeswell et al. (2006)
	JR141	2006	Larter et al. (2007); Graham et al. (2010)
	NBP0001	2000	Nitsche et al. (2007)
	NBP0702	2007	Nitsche et al. (2007)
	ANT-XXIII/4	2006	Gohl (2006); Nitsche et al. (2007)
1B	JR97	2005	K. Nicholls (PI)
	JR244	2011	R. Larter (PI)
1C	JR104	2004	Ó Cofaigh et al. (2005b); Dowdeswell et al. (2008)
	JR141	2006	Noormets et al. (2009)
1D	NBP0103	2008	Bolmer (2008)
	NBP0202	2008	Bolmer (2008)
1E	JR59	2001	Ó Cofaigh et al. (2002)
	JR71	2002	Ó Cofaigh et al. (2005a); Dowdeswell et al. (2004)
	JR157	2007	Noormets et al. (2009)
	JR179	2008	R. Larter and P. Enderlein (PIs)
	NBP0103	2001	P. Wiebe (PI); Bolmer (2008)
	NBP0104	2001	P. Wiebe (PI); Bolmer (2008)
	NBP0202	2002	P. Wiebe (PI); Bolmer (2008)

[a]Cruise IDs beginning JR indicate RRS *James Clark Ross*, those beginning NBP indicate RVIB *Nathanial B. Palmer* and ANT-XXIII/4 was on RV *Polarstern*.

Multibeam data were processed and gridded to 20 m cell size using public-access MB-System software (Caress and Chayes, 1996). These data were analysed and compared with existing swath bathymetry and TOPAS data from previous cruises (Table 7.1) from other Antarctic continental margin areas, including the western Antarctic Peninsula, and the Bellingshausen and Amundsen seas. The slope morphology was quantitatively analysed using standard geographic information system (GIS) tools, by extracting cross-sectional profiles parallel to the continental shelf edge, along which gully parameters were measured. Measured parameters include gully depth, width, length, gully density at 50 m below the shelf edge (gullies/km), gully cross-sectional area (gully width times gully depth on the uppermost slope) and the general gradient of the continental slope. The cross-sectional shape of the gullies was analysed using the General Power Law ($^{G}P_{L}$) program (Pattyn and Van Huele, 1998), which approximates gully cross-sectional shape according to the general power law equation:

$$y - y_0 = a \mid x - x_0 \mid^b \qquad (7.2)$$

where x and y are horizontal and vertical coordinates along a cross-sectional profile, x_0 and y_0 are the coordinates of the minimum of the fitted profile (automatically determined), and a and b are constants. The program carries out a least squares analysis for a range of b values and identifies the value giving the minimum RMS misfit between the observed profile and the idealised shape defined by the above equation. The resulting b value provides a measure of cross-sectional shape (U/V index), with 1 being ‘perfect V-shape’ and 2 being parabolic shape, commonly referred to as ‘U-shape’ by geomorphologists. Values <1 express a gully with convex-upwards sides, and values > 2 express a more box-shaped gully, where steepness increases away from the axis more rapidly than a parabolic curve.

7.4. Results

Our new data show that 76 gullies incise the mouth of the Filchner Trough within the extent of the multibeam bathymetry data. We observe two separate gully types: (i) seventy two small-scale U-shaped gullies with rounded gully heads (e.g. Fig. 7.3C) and (ii) four small-scale V-shaped gullies. The gully distribution is not uniform with the highest density found at the centre of the trough, corresponding to the deepest section of the trough (Fig. 7.4). This is the opposite of the pattern of gully distribution observed on the western Antarctic Peninsula, Amundsen and Bellingshausen Sea margins, where gully density increases towards the trough margins (Noormets et al., 2009). Gully cross-sectional area does not change significantly with distance from the trough axis. Measurements taken along a profile that is parallel to, and 50 m below the shelf edge show that gullies are on average 630 m wide, 12.5 m deep and 2.7 km long. Gully length measurements are, however, constrained in places by the limited down-slope data extent. The gullies cut back on average 220 m into the shelf edge and are found on slope gradients of 2-3°. The frequency distribution of gullies shows that most gullies (60%) at the shelf edge are between 5 and 15 m deep (Fig. 7.5B).

The V-shaped gullies are characterised by greater lengths, and lower widths (average 283 m) and depths (9.5 m) (e.g. Fig. 7.3D). Large, deeply incised V-shaped gullies as found on the western margin of the Antarctic Peninsula (Fig. 7.1E), Bellingshausen (Fig. 7.1C) and Amundsen seas (Fig. 7.1A) are not observed in the study area. Regional data from the southeastern Weddell Sea do however show large channel systems further down-slope, which do not initiate at the shelf edge (Michels et al., 2002). Quantitative analysis we have carried out on gullied slopes offshore from the mouths of other palaeo-ice stream troughs, such as Marguerite (western Antarctic Peninsula), Belgica (Bellingshausen Sea) and Pine Island West (Amundsen Sea) Troughs (Table 7.2),

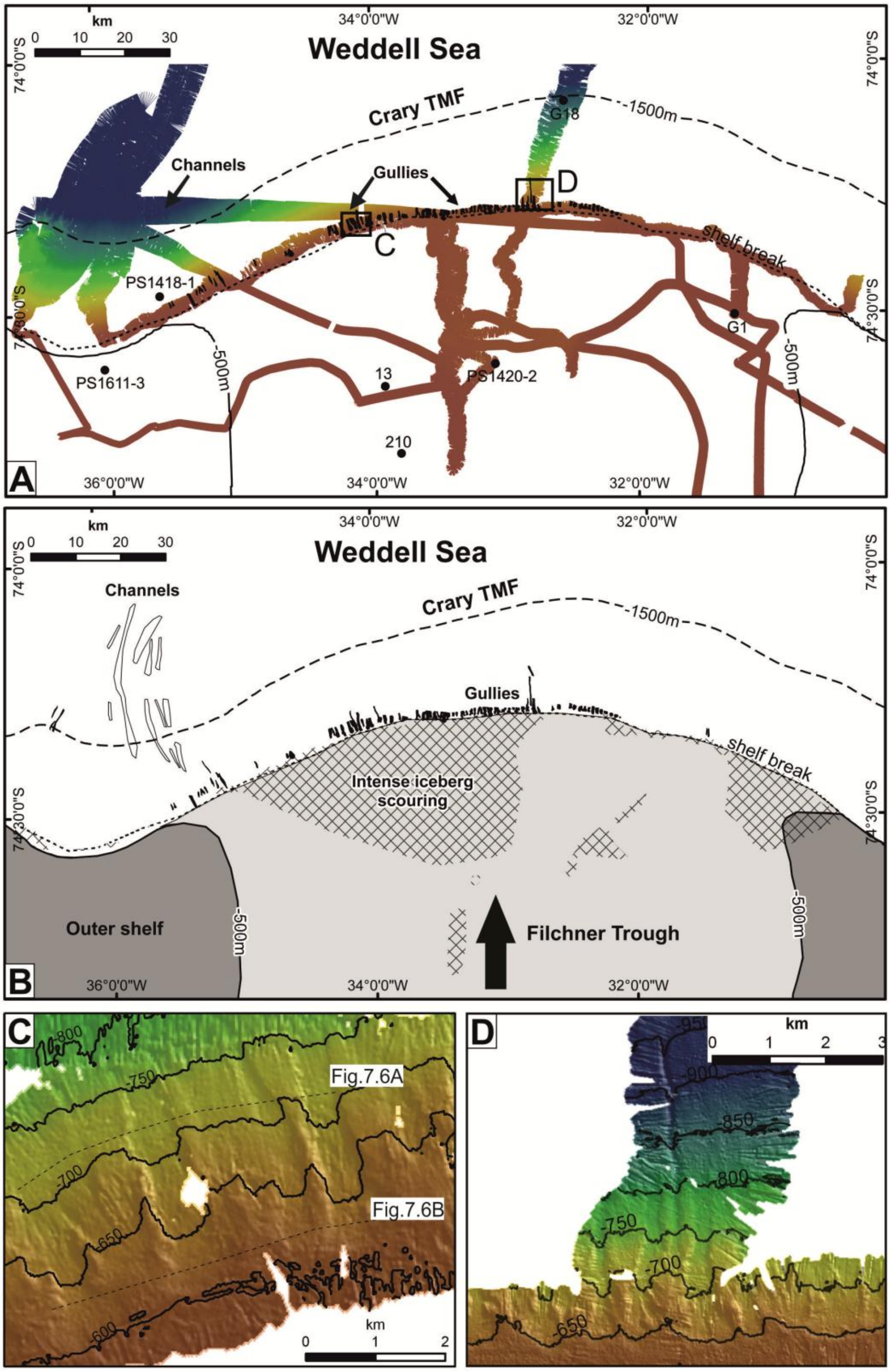

Figure 7. 3. The major morphological features of the outer shelf and slope offshore from Filchner Trough, Weddell Sea. **A**. Shaded relief swath bathymetry image of the study area. Boxes indicate locations of Figs. C and D. Black circles mark sediment core locations: cores PS1418-1, PS1420-2, PS1611-3 (Melles, 1991; Melles and Kuhn, 1993); G1, G18 (Anderson et al., 1980); 13, 210 (Elverhøi and Maisey, 1983). **B**. Schematic diagram of Crary Trough Mouth Fan (TMF). Black arrow represents direction of palaeo-ice stream flow. **C.** Slope morphology offshore from Filchner Trough. Dashed lines indicate location of TOPAS profiles in Figs. 7.6A and 7.6B. **D**. Slope morphology offshore from Filchner Trough.

shows that they are geomorphically different from the U-shaped gullies offshore from Filchner Trough (Fig. 7.5). The latter gullies measure close to ‘2’ (U-shape) according to the general power law equation, display no branching or sinuosity and have low depth:width and length:width ratios (Figs. 7.5A, C, D).

TOPAS data show that the gullies are formed in poorly stratified or acoustically impenetrable layers (Figs. 7.6A, B) indicating their erosion into the seabed, contrasting good TOPAS penetration into gully interfluves which would be expected if they were formed by interfluve aggradation. Headwall scars are not observed, but in some instances the TOPAS data show an acoustically transparent layer of variable thickness that covers the underlying topography and may subdue the expression of any headwall scars present along the Filchner Trough mouth (Fig. 7.6B).

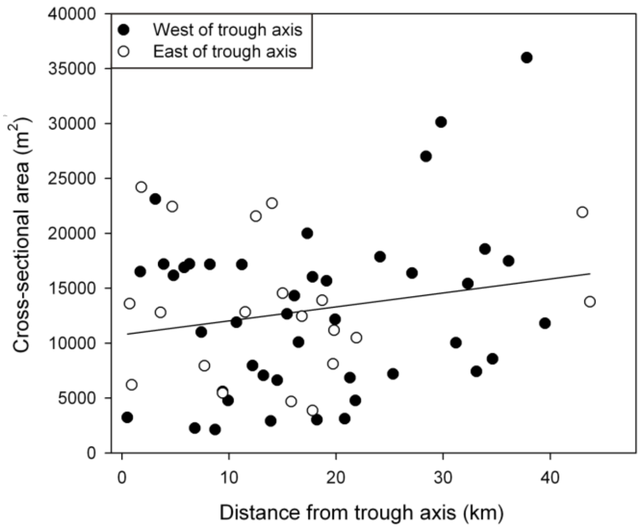

Figure 7. 4. Gully distribution and cross-sectional area at Filchner Trough mouth, Weddell Sea. Solid circles are gullies west of the trough axis; white circles are gullies east of the trough axis. Solid line is the least-squares regression fit to both datasets (r^2 is 0.039)

7.5. Discussion

Along the Filchner Trough shelf edge, the predominant morphological signature is small-scale U-shaped gullies. Morphologically, the gullies resemble small-scale slide scars, displaying relatively flat-floored cross-sections and steep walls (Kenyon, 1987). TOPAS data show an acoustically transparent layer of variable thickness overlying an acoustically impenetrable layer which the gully morphologies are formed in. The acoustically transparent layer was not resolved by earlier 3.5 kHz acoustic surveys, but may correspond to post-glacial, bioturbated sands that overlie diamictons interpreted as gravitational slide deposits in cores (PS1494 and PS1612) recovered from the upper slope in the southern Weddell Sea to the west of our study area (Melles and Kuhn, 1993). The presence of an acoustically transparent layer that drapes over the gully morphologies and interfluves suggests that the seafloor topography is a relict with gullies likely formed under full glacial conditions/early stages of deglaciation rather than during the postglacial interval in which there have been seasonally open water conditions. Under full glacial conditions during the Last Glacial Maximum, the West Antarctic Ice Sheet extended to, or close to, the shelf edge on the western margin of the Antarctic Peninsula (Vanneste and Larter, 1995; Heroy and Anderson, 2005) and in the Bellingshausen (Hillenbrand et al., 2010) and Amundsen seas (Lowe and Anderson, 2002; Graham et al., 2010), while the East Antarctic Ice Sheet is also considered to have extended across the Weddell Sea shelf during this period (Hillenbrand et al., 2012). If the ice-sheet grounding line reached the shelf edge under full glacial conditions, ISW production would have been limited. No ice-shelf cavities would have existed to super-cool HSSW, a precursor to dense bottom water formation. Under full glacial conditions it is also likely that thicker and more permanent sea ice was present, restricting the potential for the production of new sea ice and therefore the amount of HSSW produced.

The slide scars are likely the result of small-scale slope failure. 60% of the U-shaped gullies incise the seabed to a depth of 5-15 m (Fig. 7.5B) suggesting that the sediment slides may be retrogressive, with sediment failure occurring along a plane of weakness in the sedimentary structure (Laberg and Vorren, 1995; Canals et al., 2004). This plane of weakness may result from the change from glacial to interglacial sedimentation, as hemipelagic sediments, which are common during deglacial and interglacial periods, have higher porosity than sediments deposited during glacial periods. This change may create a weakened layer which is more susceptible to failure (Long et al., 2003). Alternatively, planes of weakness may form during periods of stronger current flow when winnowing of finer grained sediment creates instabilities. Acoustic and seismic evidence from further down the continental slope also suggest that extensive mass-wasting occurred on the Weddell Sea continental slope in the past (e.g. Melles and Kuhn, 1993; Bart et al., 1999).

Table 7. 2. Average gully parameters observed at the mouths of cross shelf troughs around Antarctica and in the presence / absence of Trough Mouth Fans (TMF).

Location	Gully density* (gully/km)	Mean Gully cross- sectional area* (m^2)	Slope gradient (°)	Within Trough	TMF present
Filchner Trough (U-shape), Weddell Sea (e.g. Fig. 7.1B) (35°40'W - 31°00'W)	0.60	14720	2.5	yes	yes
Filchner Trough (V-shape), Weddell Sea (e.g. Fig. 7.3D) (35°40'W - 31°00'W)	< 0.10	3787	2.5	yes	yes
Belgica Trough, Bellingshausen Sea (88°W - 84°30'W)	0.55	11518	1.7	yes	yes
Marguerite Trough, western Antarctic Peninsula (e.g. Fig. 7.1E) (72°W - 70°48'W)	0.76	18184	9.0	yes	no
Pine Island (West)Trough, Amundsen Sea (e.g. Fig. 7.1A) (114°30'W - 113°2'W)	1.40	32442	4.5	yes	no

*Measurements taken at 50 m below the shelf edge.

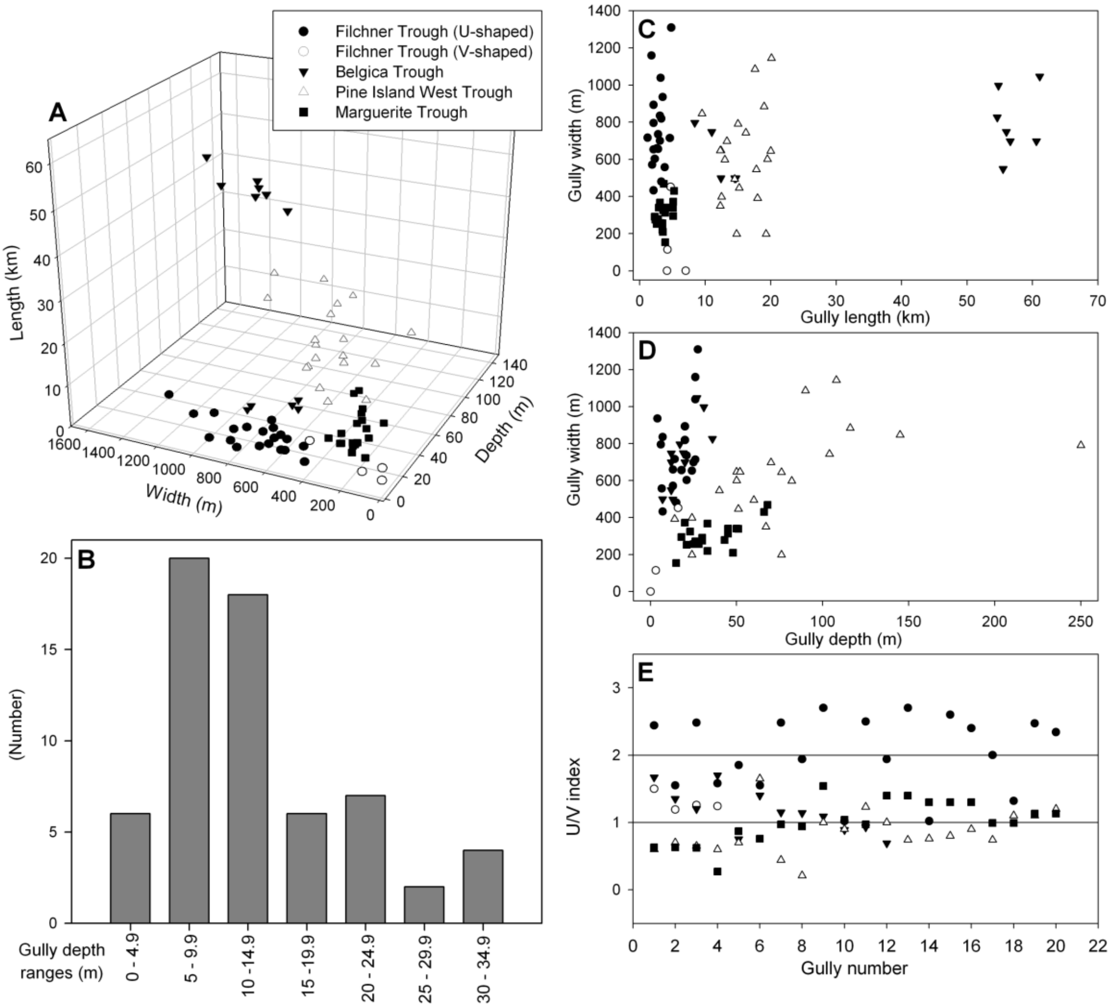

Figure 7. 5. A. Gully morphological parameters (length, width and depth) offshore from four palaeo-ice stream troughs around Antarctica: Filchner Trough (circles), Marguerite Trough (black squares), Belgica Trough (black triangles) and Pine Island Trough (West) (white triangles). For additional data on gully locations, see Table 7.2. **B.** Frequency analysis of gully incision depth ranges across the shelf edge offshore from Filchner Trough. **C.** Gully width vs gully length; **D.** Gully width vs gully depth; **E.** Gully cross-sectional shape (U/V index) from general power law analysis. U/V index values near '1' are distinctly V-shaped; '2' are distinctly U-shaped.

Marine slope instability is influenced by factors including oversteepening, seismic activity, rapid sediment accumulation, gas charging, gas hydrate dissociation, glacial loading, slope angle and mass movement history (Locat and Lee, 2002). The Weddell Sea is a passive margin with no evidence for the presence of gas hydrates (Bart et al., 1999) and with a relatively low slope gradient (2-3°) compared to most other Antarctic margins (Table 7.2). Small-scale slope failure along the Filchner Trough margin is likely the result of rapid accumulation of sediment due to glacial transport (cf. Larter and Cunningham, 1993; Melles and Kuhn, 1993). Bottom currents may also have enhanced sediment transport toward the shelf edge during interglacials as the present ISW flow velocities are capable of eroding medium-sand particles and transporting gravel grains (Melles et al., 1994). Current-winnowing of fine grained particles from the shelf edge and redeposition on the slope may influence slope stability by interleaving fine-grained, high water content sediment layers with denser, coarse-grained layers (cf. Melles and Kuhn, 1993). This process may also explain the local occurrence of winnowed, coarse-grained surface sediments on the slope (Melles et al., 1994), while formation of a similar winnowed lag at the shelf edge might have inhibited further erosion and explain the lack of large erosional features at the shelf edge when compared to some other Antarctic continental margins.

Along the southern Weddell Sea margin, highly erosional and V-shaped gullies like those observed on other parts of the Antarctic continental slope are not present. Assuming that fluid flow produces V-shaped incisions, as seen in the terrestrial environment (Simons and Sentürk, 1992), their absence in the southern Weddell Sea indicates that there is no fluid erosional process occurring here. The underlying geology of these areas (Table 7.2) shows a limited range of characteristics as the slopes are constructed largely of prograded sequences (Cooper et al., 2008) with sediment cores giving evidence that these consist of glaciogenic debris flows with similar lithology, physical properties, grain-size composition and mineralogical composition (e.g. Anderson and Andrews, 1999; Michels et al., 2002; Dowdeswell et al., 2004, 2006, 2008; Hillenbrand et al., 2005; 2009). The differences in the observed shelf edge morphologies are therefore unlikely to be controlled by the underlying substrate and instead are more likely due to differences in slope processes. ISW overflow may be exerting an influence further down-slope, as current velocities have been shown to increase down-slope (Foldvik et al., 2004) with large channel-levee systems also present toward the Weddell Sea continental rise (Michels et al., 2002). However, even on the lower continental slope of the southern Weddell Sea, erosional features are attributed to erosion by gravitational down-slope transport and contour currents during glacials, but not modern ISW overflow (Melles and Kuhn, 1993). This is fully consistent with our observation that there is no significant geomorphic signature of cold, high salinity water overflow at the shelf edge and upper slope.

Noormets et al. (2009) observed a clear pattern of gully size and density increasing toward the trough margins at the mouths of the Marguerite, Belgica and Pine Island West cross-shelf troughs. However, at the Filchner Trough mouth, highest gully densities are found at the centre of the trough and gully cross-sectional area does not change significantly. This difference suggests that the processes operating on other Antarctic continental slopes had a lesser effect on the Filchner Trough margin. The flow velocity of ice streams within cross-shelf troughs may explain the observed pattern of gullying, where in a horizontal sense, the velocity of an ice stream would be at a maximum at the trough axis and lower at the trough margins (Bindschadler and Scambos, 1991; Whillans and Van der Veen, 1997). Sediment delivery would therefore be higher at the trough axis, leading to increased slope instability and increased gully density.

A potential process forming the deeply incised and V-shaped gullies observed on other Antarctic continental margins is sediment-laden subglacial meltwater (Noormets et al., 2009). For subglacial meltwater to overcome the buoyancy of freshwater in normal seawater, 33 g l^{-1} of detritus must be entrained in order for it to remain at the seafloor (Syvitski, 1989). A possible mechanism for sufficient sediment to become entrained is high fluxes of meltwater associated with episodic discharge. Our observations suggest that if this was indeed the main process for eroding large V-shaped gullies, less episodic meltwater would have been discharged from the Filchner Trough mouth during glacial periods compared to troughs along the Pacific margin. A smaller amount of basal melt from this region is consistent with colder surface temperatures in the Weddell Sea embayment compared to the Antarctic coast in the Pacific sector (Dixon, 2008), a difference which probably persisted through glacial periods and would have resulted in colder ice. Basal ice temperatures are also affected by geothermal heat flux and strain heating, but as the southern Weddell Sea has been tectonically inactive since at least mid-Cretaceous times (DiVenere et al., 1996), geothermal heat flux is expected to be relatively low. Antarctic heat fluxes inferred from a global seismic model support this suggestion (Shapiro and Ritzwoller, 2004). Strain heating is an important factor affecting basal temperatures, but it is a feedback effect, not a primary cause of basal melting and ice flow (Hindmarsh, 2009).

Our results suggest that cold, high salinity water overflow is an unlikely formation mechanism for the deeply incised and V-shaped gullies observed at the shelf break along other parts of the Antarctic continental margin. Other possible explanations for such features include sediment-laden subglacial meltwater discharge from the base of an ice sheet grounded at the shelf edge during glacial maxima through either episodic release, possibly by subglacial lake water outbursts, or more continuous release.

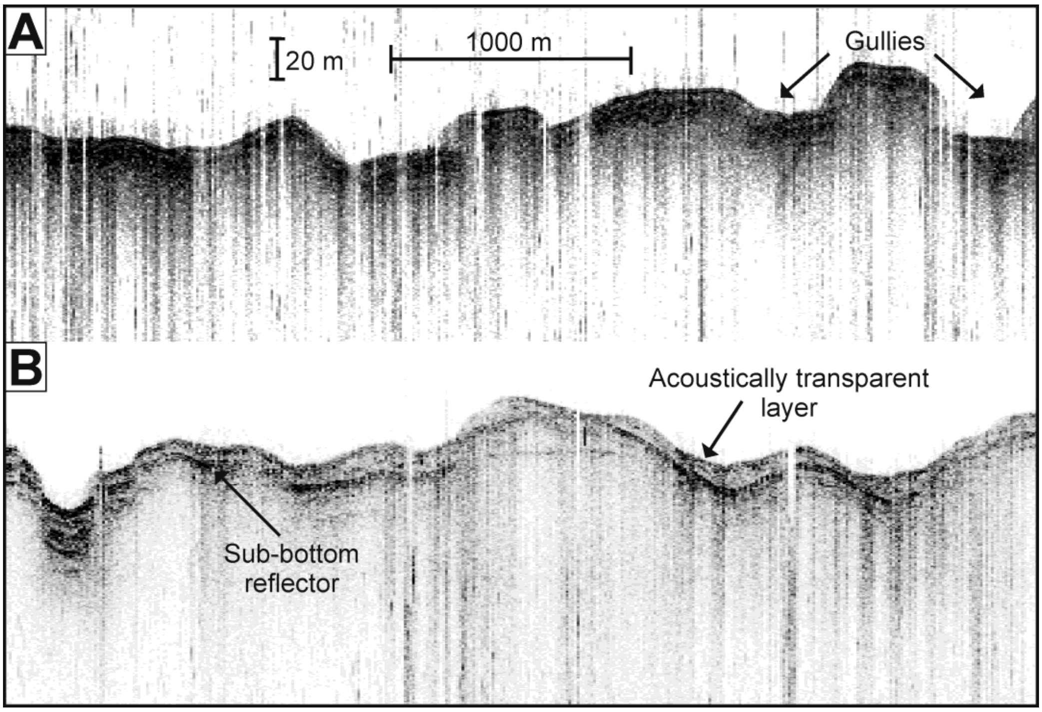

Figure 7. 6. A. TOPAS profile through lower section of continental slope along profile ‘Fig.7.6A’ in Fig.7.3C (dashed black line). **B.** TOPAS profile through upper section of continental slope along profile ‘Fig.7.6B’ in Fig.7.3C (dashed black line).

7.6. Conclusions

We have presented geomorphological analyses of new bathymetric data from an area of active cold, dense water overflow along the Weddell Sea continental margin. The analyses show that U-shaped gullies offshore from the Filchner Trough are geomorphologically distinct from gullies observed elsewhere on the Antarctic continental margin. These gullies are likely produced by small-scale slides, probably resulting from the rapid accumulation and subsequent failure of proglacial sediment during glacial maxima. The features are quantitatively different from the highly incisional and V-shaped gullies which dominate some other Antarctic continental margins. The distinctly different geometry of the gullies in the southern Weddell Sea will have a significant impact on the calculation of dense water outflow and will enhance the ability of models to predict the flow and entrainment of dense water as it passes over the shelf break (Muench et al., 2009).

Our findings indicate that past overflow of cold, high salinity water was unlikely to be the dominant mechanism for the extensive gully erosion observed in other areas of the Antarctic, confirming the speculation of earlier workers. We hypothesise that other processes, such as mass flows or subglacial meltwater discharge, likely played a greater role in gully formation elsewhere along Antarctic continental margins.

7.7. Acknowledgements

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Chapter 8.

Slides, gully and channel systems of the southern Weddell Sea, Antarctica.

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8. Abstract

New multibeam bathymetric data from the southeastern Weddell Sea show significant differences in surface morphology of the outer continental shelf and slope between two adjacent cross-shelf troughs. These are the Filchner Trough and a smaller trough to the east we refer to as 'Halley Trough'. Multibeam bathymetric data, acoustic subbottom profiler and seismic data show major differences in the incidence and morphologies of submarine gullies, channel systems, submarine slides and iceberg scours, and in sediment deposition. These large-scale differences suggest significant variation in slope and sedimentary processes between the two troughs, leading to much greater deposition at the mouth of the Filchner Trough. Bedforms, including a terminal moraine and scalloped embayments on the outer shelf of Halley Trough, provide insight into the timing and extent of past ice and point to grounded ice near to the shelf edge during the late Quaternary.

New data reveal two large-scale submarine slides on the upper slope of the eastern Crary Fan, a trough mouth fan offshore from Filchner Trough. Both slides head at the shelf edge (~500 m water depth), with the largest slide measuring 20 km wide and with an incision depth of 60 m. Multibeam and seismic data show elongate slide blocks on the seafloor surface of the mid-slope. The lack of a discernible sedimentary cover suggests that they were generated after the Last Glacial Maximum (LGM). This is unusual because post-LGM submarine slides are very rare on the Antarctic continental margin, and to our knowledge, no other post-LGM slides have been documented on an Antarctic trough mouth fan. Because the slides occur on a part of the continental slope where the deposition of glacial debris was highest, we speculate that weaker, unconsolidated sedimentary layers within the subsurface are important for slide initiation here.

Keywords: slope processes; Antarctica; continental slope; slide; geomorphology; trough mouth fan.

8.1. Introduction

The Antarctic continental margin has been influenced by the advance and retreat of grounded ice since 34 Ma (Barrett, 2008), which has led to a diverse range of continental slope morphologies. These include trough mouth fans, formed at the mouths of some glacially carved cross-shelf troughs; iceberg keel marks, gullies, channels, mass-wasting features (slides, slumps), ridges, furrows and mounds (e.g. Ó Cofaigh et al., 2003; Dowdeswell et al., 2004; 2006; 2008; Noormets et al., 2009; Gales et al., 2013a). These bedforms, such as gullies, vary in size (width, incision depth and length), shelf incision, sinuosity, branching order, density and cross-sectional shape (Noormets et al., 2009; Gales et al., 2012).

Palaeo-morphologies influenced by past glacial activity are difficult to decipher from more recent slope processes due to overprinting the expression of past glacial processes (i.e. iceberg scouring). Therefore, processes forming the complex continental slope morphologies and the factors influencing these processes are not well constrained.

Processes which have been suggested to influence slope morphology include: (1) oceanographic processes such as geostrophic currents, tides and cascading dense water, formed during sea-ice formation through brine rejection (Kuvaas and Kristoffersen, 1991; Dowdeswell et al., 2006; Noormets et al., 2009; Muench et al., 2009); (2) sedimentary processes, such as mass flows (slides, slumps, debris flows and turbidity currents) influenced by a range of triggering mechanisms, including tectonic influences, gas hydrate dissociation, sediment loading, presence of weak sedimentary layers within the seabed, re-suspension by iceberg scouring, currents, tidal activity and changes in sea level; and (3) glacigenic processes, such as subglacial meltwater discharged from beneath an ice sheet (e.g. released by sub-glacial lake discharge or subglacial volcanic eruptions), iceberg grounding, and high accumulations of glacigenic debris arising from the rapid transport by ice-streams to the shelf edge (Goodwin, 1988; Larter and Cunningham, 1993; Vanneste and Larter, 1995; Wellner et al., 2006; Long et al., 2003; Imbo et al., 2003; Hillenbrand et al., 2005; Dowdeswell et al., 2006, 2008; Dowdeswell and Bamber, 2007; Fricker et al., 2007; Piper et al., 2012). These processes may be influenced by environmental controls, such as local slope character (slope geometry, gradient), debris content of ice, large-scale spatial characteristics (e.g. size of drainage basins, location of cross-shelf troughs) and ice-sheet history (Noormets et al., 2009; Peakall et al., 2012; Gales et al., 2013a). Deciphering the extent that these processes and environmental controls influence slope morphology and the timescales over which they occur remains a major challenge.

Post-Last Glacial Maximum (LGM) submarine mass flows are rare on the Antarctic continental margin (Barker et al., 1998; Dowdeswell and Ó Cofaigh, 2002; Nielsen et al., 2005), with no major slides documented on Antarctic trough mouth fans during the Quaternary. One of the few Quaternary slide examples from the Antarctic margin is the Gebra Slide, located on the lower continental slope of the Trinity Peninsula, Antarctica, in water depths between 1500 and 2000 m (Imbo et al., 2003; Casas et al., 2013). Early Poliocene slides and major erosional channels with chaotic infills have been documented on the Crary Trough Mouth Fan and in its distal part in the Weddell Sea basin (Bart et al., 1999), while late Pliocene mega-scale debris flow deposits were observed on the western Antarctic Peninsula continental margin (Diviacco et al., 2006). Widespread Miocene mass-wasting events have been documented off western Wilkes Land (East Antarctica) (Donda et al., 2008). Submarine slides are common on other northern Hemisphere high latitude continental margins e.g. Storegga Slide (Bugge et al., 1987; Evans et al., 1996; Bryn et al., 2003), Trænadjuptet Slide (Laberg and Vorren, 2000; Laberg et al., 2002), Andøya Slide (Kenyon, 1987; Dowdeswell et al., 1996) and the Bjørnøyrenna Slide (Laberg and Vorren, 1993), located on the Norwegian and southwest Barents Sea margins.

In this paper we present a quantitative analysis of the outer continental shelf and slope morphology of the southern and southeastern Weddell Sea. We examine differences in slope morphology observed between a trough mouth fan and the neighbouring part of the continental slope and discuss factors influencing the observed large-scale differences and the implications for past ice-sheet history and dynamics. We describe the morphology of two relatively young and large-scale submarine slides on the eastern flank of the Crary Fan and discuss possible slide initiation mechanisms.

8.2. Study area

8.2.1. Physiographic setting

The study area includes the shelf edge and upper slope of the southern Weddell Sea and the outer shelf and slope of the southeastern Weddell Sea, Antarctica, covering an area of ~56,300 km^2 (Fig. 8.1). The morphology of parts of the southern and southeastern Weddell Sea has been described previously (Kuvaas and Kristoffersen, 1991; Melles and Kuhn., 1993; Kuhn and Weber, 1993; Weber et al., 1994; Melles et al., 1995; Michels et al., 2002; Weber et al., 2011; Larter et al., 2012; Gales et al., 2012). Within the study area, the southern Weddell Sea shelf is dissected by the glacially-carved Filchner Trough which extends to the shelf edge and is associated with a fan offshore from its trough mouth (Crary Fan). The fan is characterised by convex-outward contours, thought to have been formed by repeated advances of ice to the shelf edge causing the shelf to prograde 70-80 km from its pre-glacial location (Kuvaas and Kristoffersen, 1991). At the mouth of the Filchner Trough, small-scale and U-shaped gullies occur which are interpreted as small slide scars (Gales et al., 2012). Further down-slope (below ~2400 m water depth) and to the east, large asymmetric channels are present, orientated southwest-northeast (Kuvaas and Kristoffersen, 1991; Kuhn and Weber, 1993; Weber et al., 1994). The channels feed into a large channel-ridge system at around 3400 m depth (Michels et al., 2002).

Helmert Bank forms the eastern margin of the Filchner Trough and the western boundary of a smaller trough to the east which we refer to as ‘Halley Trough’. Filchner Trough extends ~450 km from the Filchner Ice Shelf and has a width of 125 km and an axial depth of 630 m at the shelf edge. Halley Trough is considerably smaller, with a width of 62 km, measured at the 400 m contour and a maximum depth of ~530 m at the shelf edge. International Bathymetric Chart of the Southern Ocean data (IBCSO; Arndt et al., 2013) show that Halley Trough extends > 200 km inshore. Brunt Basin, a depression seaward of the Brunt Ice Shelf, lies to the east of Halley Trough (Fig. 8.1).

Table 8. 1. Geomorphic features of Filchner Trough and Halley Trough mouths.

Geomorphic feature	Filchner Trough	Halley Trough
Upper slope gradient (º)	2.5	3.7
Width (km)[a]	125	62
Maximum depth (m)	630	540[b]/515[c]
Length (km)	> 450	> 200
No. of Gullies	76	48
Gullies/km	0.61	0.80

[a]Measured at shelf edge; [b]west of Trough; [c]east of Trough.

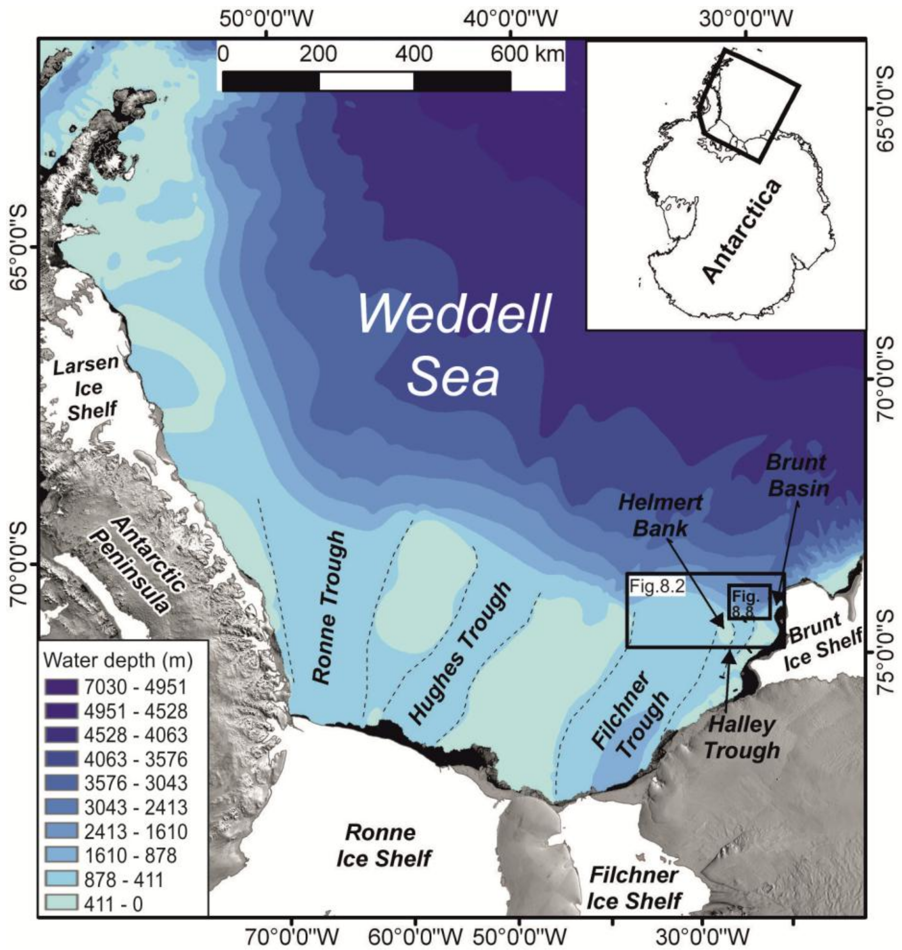

Figure 8. 1. Study area. Bathymetric data is from IBCSO (Arndt et al., 2013). Antarctic Ice Sheet is shaded Landsat Image Mosaic of Antarctica (LIMA) data (U.S Geological Survey, 2007). Dashed lines mark cross-shelf trough margins. Black box marks study area and locates Fig. 8.2 and Fig. 8.8A.

8.2.2. Glaciological, oceanographic and geological setting

The history of grounded ice extent on the southern and southeastern Weddell Sea has been widely debated, with disparities between marine and onshore ice-sheet reconstructions (Hillenbrand et al., 2012). Marine geological and geophysical data suggest that ice advanced across the southern Weddell Sea shelf, grounding at, or near to, the shelf edge of the Filchner and Ronne troughs during the Last Glacial Maximum (LGM) (Elverhøi, 1981; Bentley and Anderson, 1998; Hillenbrand et al., 2012; Larter et al., 2012; Stolldorf et al., 2012). Hillenbrand et al. (2012) concluded that ice was grounded even within the deepest sections of the Filchner and the Ronne troughs during the LGM. However, available onshore cosmogenic exposure ages (Bentley et al., 2010; Hein et al., 2011) and glaciological modelling studies (LeBrocq et al., 2010) as well as some radiocarbon dates of pre-LGM age obtained from foraminifera in shelf sediment cores (Stolldorf et al., 2012), suggest that ice may not have been thick enough to ground at the shelf edge during the LGM.

The clockwise-flowing Weddell Gyre dominates the Weddell Sea oceanographic circulation. The gyre transports Circumpolar Deep Water southward where it becomes

Warm Deep Water (WDW) and, after further modification in the southern Weddell Sea, Modified Warm Deep Water (MWDW) (Orsi et al., 1993). High Salinity Shelf Water (HSSW) is produced during sea ice formation through brine rejection. Cold and dense Ice Shelf Water (ISW) forms when HSSW is super-cooled and freshened beneath ice shelves (Nicholls et al., 2009). Current meter measurements installed 10 m above the seafloor show that ISW flows toward the shelf edge within the Filchner Trough and cascades down the continental slope with a mean flow velocity of 0.38 m s^{-1} and a maximum velocity of 1 m s^{-1} (Foldvik et al., 2004). ISW mixes with MWDW to form a major component of Weddell Sea Deep Water and Weddell Sea Bottom Water, which in turn contributes to Antarctic Bottom Water (Foldvik et al., 2004). The latter is exported from the Southern Ocean and forms the deep southern branch of the global thermohaline circulation (Orsi et al., 1993; Naveira Garabato et al., 2002; Nicholls et al., 2009).

The near-surface sedimentary architecture of the southern and southeastern Weddell Sea shelf and slope has been well documented (e.g. Elverhøi and Roaldset, 1983; Haase, 1986; Ehrmann et al., 1992; Kuhn et al., 1993; Melles et al., 1994; Weber et al., 2011; Hillenbrand et al., 2012). Within the Filchner Trough, the seafloor surface sediments on the inner shelf are predominantly gravelly and sandy mud, with the trough floor on the mid-shelf largely covered by muds and sandy muds (Melles et al., 1994). The surface sediments on the outer shelf and shelf edge consist of sand and gravelly sand, while the underlying sediments recovered in cores were interpreted as glaciomarine and subglacial deposits (Elverhøi, 1984; Melles and Kuhn, 1993; Melles et al., 1994; Hillenbrand et al., 2012).

8.3. Data and methodology

We use multibeam bathymetric data collected on RRS *James Clark Ross* during cruises JR259 in 2012, JR244 in 2011 and JR97 in 2005 and bathymetric data collected on ten RV *Polarstern* cruises during expeditions ANT-IV/3, V/4, VI/3, VIII/5, IX/3, X/2, XII/3, XIV/3, XV/3 and XVI/2 in the 1990s. The extent of the datasets is shown in Fig. 8.2. Data were collected on RRS *James Clark Ross* using a hull-mounted Kongsberg EM120 (JR244 and JR97) and EM122 (JR259) multibeam bathymetric system with a swath width of up to 150º (for the EM120 system) and 134° (for the EM122 system), with maximum angle variation depending on water depth, sea state (i.e. swell, sea-ice cover) and bathymetry. Both systems have a 191 beam array with real-time beam steering and active pitch and roll compensation and a frequency range of 11.75-12.75 kHz. The data aboard RV *Polarstern* were collected using a Hydrosweep DS1/2 with a frequency of 15 kHz. A Kongsberg TOPAS PS 018 acoustic subbottom profiler was used during cruise JR244. The TOPAS system transmits two primary frequencies at around 18 kHz from which secondary frequencies are generated ranging from 1300 to 5000 Hz by the parametric effect. The system is able to resolve sedimentary layers to < 1 m and can penetrate $\geq$ 50 m below the seafloor in fine-grained sediment. The seismic line AWI 90060 was collected by RV *Polarstern* on cruise ANT-VIII/5 (1989-1990) using an 8-liter array of PRAKLA-SEISMOS

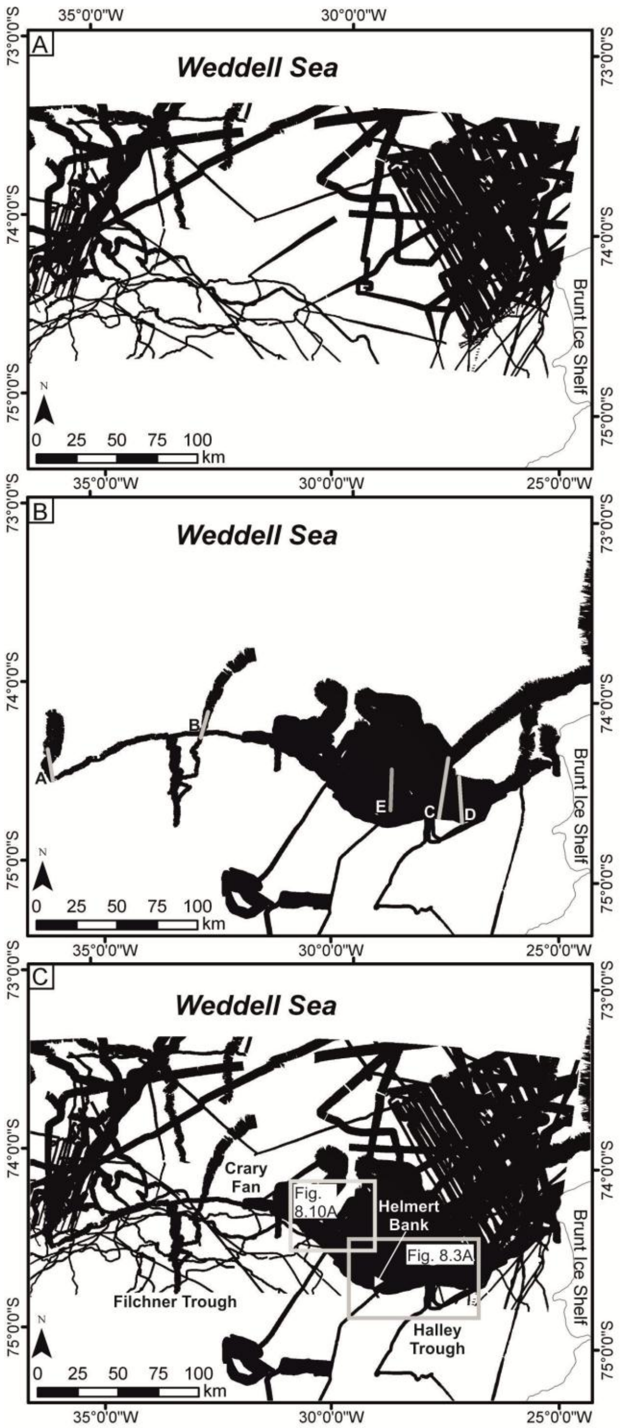

Figure 8. 2. Data extent for the southern Weddell Sea. **A**. AWI data from expeditions ANT-IV/3, V/4, VI/3, VIII/5, IX/3, X/2, XII/3, XIV/3, XV/3 and XVI/2. **B.** Data from cruises JR244, JR97 and JR259. Grey profiles (A-E) locate down-slope profiles in Fig. 8.6. **C.** Combined multibeam data sets. Grey squares locate Fig. 8.3A and Fig. 8.10A.

airguns which operated at 140 bars and with an active hydrophone streamer length of 600 m and a shot spacing of 30 m recorded with a 2 ms sample interval.

The cleaned multibeam data were processed using public access MB-system software (Caress and Chayes, 1996) and grids with cell sizes of 20 m and 50 m were produced. Vertical resolution for the multibeam data is < 1 m at 500 m water depth (for soundings between nadir and $\pm$60°), increasing to > 10 m for soundings at $\pm$70° (De

Moustier, 2001). Horizontal resolution varies with ship speed, water depth, beam angle, track spacing and seabed topography, with typical values of 10-20 m at 1000 m water depth. Cleaned hydrosweep data, also used for the IBCSO map (Arndt et al., 2013), were gridded to a cell size of 50 m due to lower data resolution. The reflection seismic data were processed on R/V *Polarstern* and at the Alfred Wegener Institute, Bremerhaven and displayed using a Landmark Promax system. The data were processed to a Common-Mid-Point (CMP) stack using standard procedures, with a CMP spacing of 25 m. A Stolt f-k migration was carried out on the data with a constant velocity of 1480 m s^{-1}. Backscatter maps were produced using FM Geocoder (Fonseca and Calder, 2005), the strength of backscattering being dependent upon sediment type, grain size, survey conditions, bed roughness, compaction, and slope (Blondel and Murton, 1997).

The surface morphology was analysed quantitatively by extracting profiles parallel to the shelf edge along which bedform parameters were measured. We identify a gully as a small channel with a depth of > 5 m that initiates from the upper slope or shelf edge. In this study, we define the point where a gully becomes a channel as the region where channel interfluve growth begins (in this study typically at 1600 m water depth). Measured parameters include bedform width, incision depth, length, sinuosity, branching order, cross-sectional shape, mean spacing and slope gradient.

Cross-sectional shape was calculated using the General Power Law ($^{G}P_{L}$) programme (Pattyn and Van Huele, 1998) which provides a measure ranging from V-shape (1) to parabolic or U-shape (2) (*b* value; referred to as U/V index). The calculation is based on the power law equation (8.1), and the programme calculates a measure of cross-sectional shape (*b* value) by finding the minimum RMS misfit between the observed cross-section and a large set of symmetrical shapes defined by the equation. In equation (8.1), *a* and *b* are constants, χ and y are the horizontal and vertical coordinates taken from the cross-sectional bedform profile and χ_0 and y_0 are automatically determined coordinates of maximum and minimum elevation on the bedform profile.

$$y - y_0 = a\,|\chi - \chi_0|^b \qquad (8.1)$$

8.4. Results

From new multibeam bathymetric data and IBCSO data (Arndt et al., 2013) we have identified ‘Halley Trough’, located to the east of the Filchner Trough (Fig. 8.1). The Halley Trough mouth is 62 km wide at the 400 m bathymetric contour with the following asymmetric trough floor morphology (Fig. 8.3A; Table 8.1): water depths to the west of the trough axis reach 540 m at the shelf edge, compared to ~515 m to the east. The difference in depth is associated with an abrupt west facing scarp boundary to the east of the trough axis (near the 500 m bathymetric contour; Fig. 8.3, inset profile B-B’). The eastern flank of Halley Trough is scalloped, with six concave NW-facing embayments (Fig. 8.3C; Fig. 8.4). The individual arcuate embayments have a mean length (measured along the NW-SE axis) of 2.1 km, a mean width (measured along the NE-SW axis) of 2.3 km and a mean height of

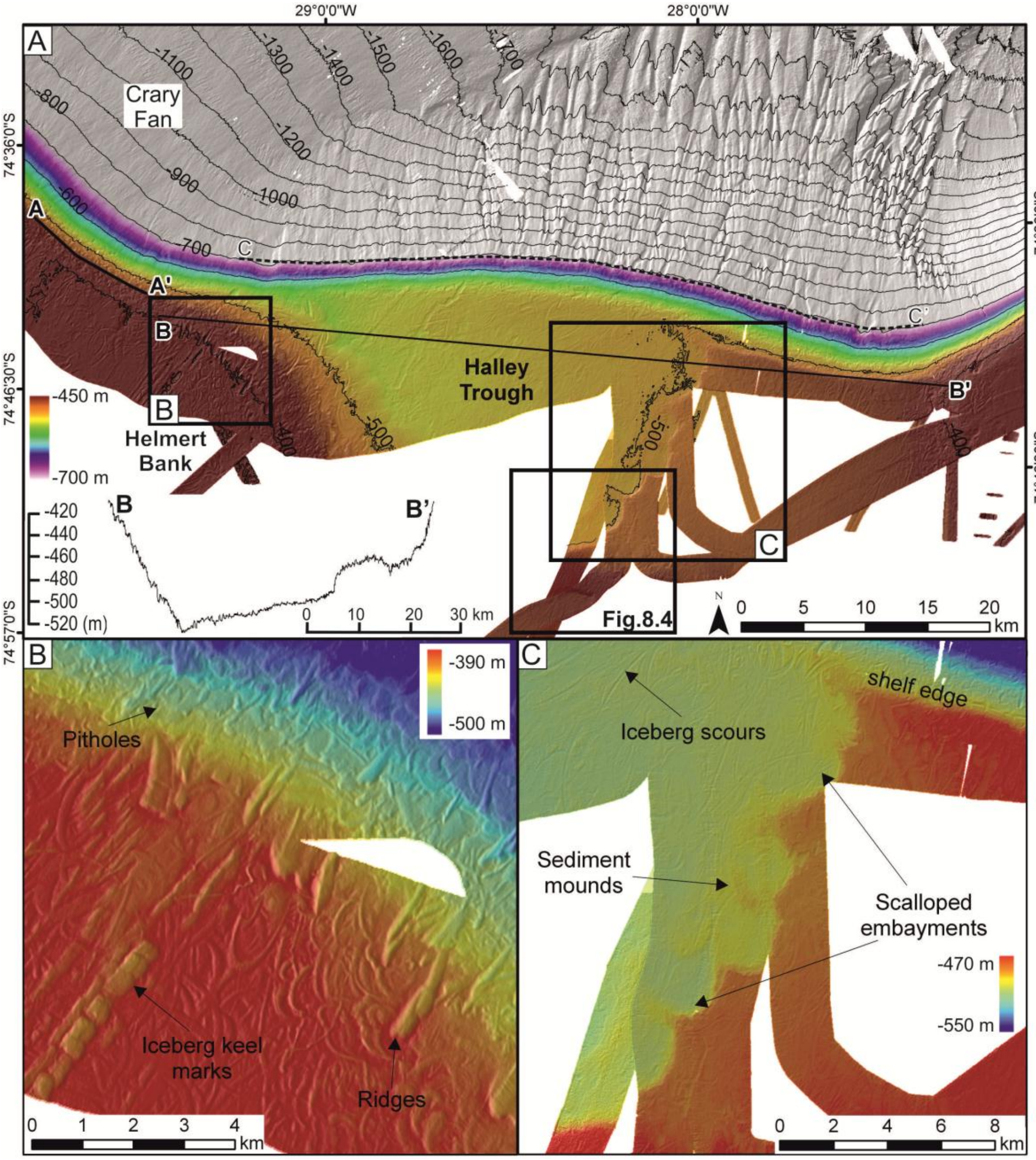

Figure 8. 3. A. Outer shelf and upper slope morphology of the Halley Trough, southeastern Weddell Sea. Profile A-A' highlights location of cross-shelf profile in Fig. 8.9. Solid line B-B' is cross-shelf profile in bottom left. Dashed line C-C' highlights transect that 48 gullies were measured along. **B.** Iceberg keel marks, sediment ridges and pitholes on the outer shelf of Halley Trough. **C.** Scalloped embayments within Halley Trough. Location of Fig. 8.3A is shown in Fig. 8.2C.

20 m. Directly to the south of the scalloped features, a ridge runs parallel to the shelf edge, ~22 km landward of it (Fig. 8.4). From the available data, the ridge is 45 m high, 2.9 km wide with a minimum length of 4.5 km. The ridge has an asymmetric shape measured along profile C-C' (Fig. 8.4) with a slightly steeper landward flank (1.1°) compared to its seaward flank (0.5°).

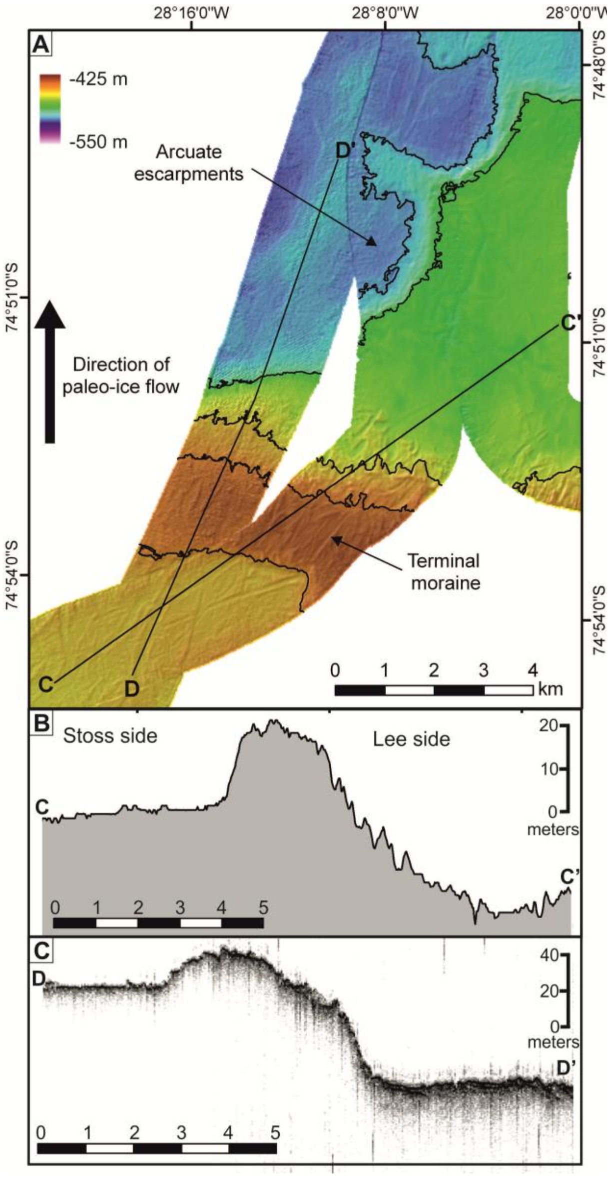

Figure 8. 4. A. Terminal moraine. Location of Fig. 8.4A is marked in Fig. 8.3. **B**. C-C' profile marked on part A. **C**. D-D' is TOPAS profile marked on part A.

8.4.1. Southeastern Weddell Sea morphology

8.4.1.1. Outer continental shelf (Halley Trough)

The outer shelf is characterised by intense iceberg scouring, and the uppermost slope is dissected by iceberg keel marks to water depths of ~720 m (Fig. 8.3B, 8.5, 8.8B). In some instances, the keel marks on the uppermost slope incise the seabed by up to 24 m, with small arcuate ridges observed around the landward terminations of the scours.

8.4.1.2. Shelf edge and upper slope

Eastward of 30º41'W (eastern flank of Crary Fan), the shelf edge and uppermost slope are characterised by small elongate pitholes (Fig. 8.3B). These depressions give the shelf edge a rough morphology and continue 100 m below the shelf edge to a water depth of ~720 m. The pitholes are between 5 and 22 m deep and become less pronounced at the mouth of Halley Trough and further eastward. Between ~74º41'S, 27º19'W and 74°39'S, 27°04’W, furrows that are ~9 km long and orientated parallel to the shelf break are observed 150 m below the shelf edge (Fig. 8.8B).

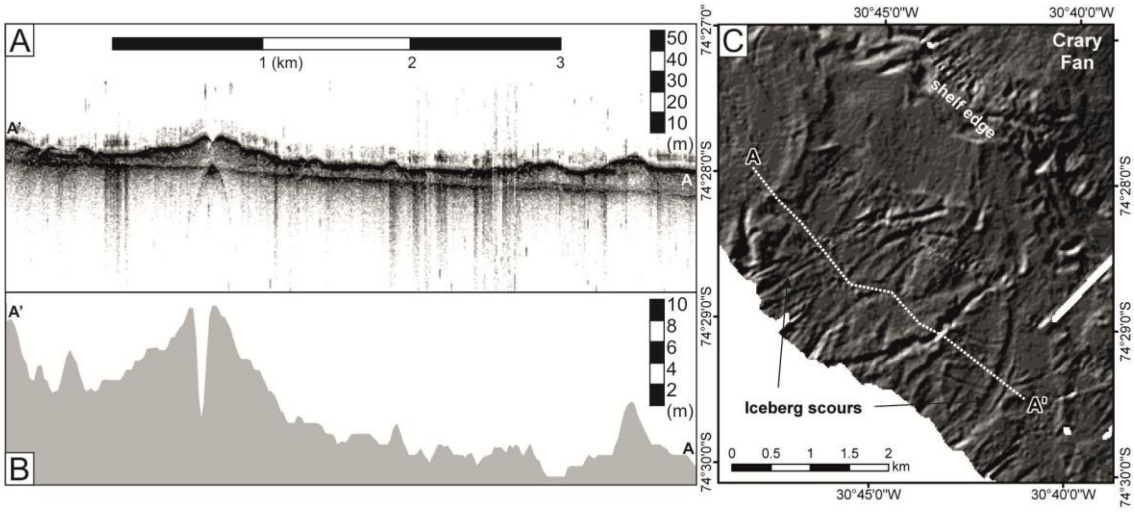

Figure 8. 5. Iceberg scour on the southern Weddell Sea outer shelf. **A.** TOPAS profile through A-A’ on C. **B.** Shelf profile through A-A’ on C. **C.** Hillshaded bathymetry data of the southern Weddell Sea outer shelf. Location of 8.5C is shown in Fig. 8.10.

To the west of Halley Trough, the mean gradient of the continental slope decreases from 3.9º to 0.5º at 1400 m water depth, with a concave downward lower slope profile (Fig. 8.6, Profile E). This change in gradient corresponds to a change in acoustic backscattering and texture observed on the hillshaded bathymetry data (Fig. 8.7C). Higher backscatter occurs at the mouth of Halley Trough, with lower backscatter observed to the west of the trough mouth on the Crary Fan. To the east of the trough mouth, the mean upper slope gradient is 3.7º, with the lower slope displaying a concave profile (Fig. 8.6, profile D).

8.4.1.3. Gully systems

Forty eight gullies occur at the mouth of the Halley Trough (Fig. 8.3A, transect C-C’). The gullies do not incise the shelf edge and initiate ~150 m below the shelf edge (Fig. 8.9). The gullies have a mean length of 12.9 km, an average width of 560 m and a mean incision depth of 11.8 m with low sinuosities and V-shaped cross-sections (Table 8.2). At ~1300 m water depth, a boundary occurs where small-scale mass wasting features are observed, similar to ‘type *IV*’ gullies observed on other high latitude continental margins (Gales et al., 2013a; 2013b). The small-scale mass-wasting features cross cut the submarine gullies (Fig. 8.8C).

Table 8. 2. Mean gully parameters from Filchner Trough and Halley Trough.

Location	**Gully parameters**					
	Length (km)	Width (m)	Depth (m)	Sinuosity	U/V	Branching
Filchner Trough[a]	2.7	630	12.5	1.01	1.88	No
Halley Trough[b]	12.9	560	11.8	1.02	1.13	No

[a]Measurements taken at 50 m below the shelf edge; [b]measurements taken at 200 m below the shelf edge.

East of the mouth of Halley Trough (i.e. eastward of 26°59’W), 70 small-scale and low sinuosity gullies occur on the upper-mid slope offshore from the Brunt Basin. The gullies do not significantly incise the shelf edge and have a mean length of 3.2 km, a mean width of 303 m and a mean incision depth of 13.8 m. The gullies become progressively deeper and wider towards Brunt Basin. A relatively smooth upper slope (depressions < 5 m) occurs eastward of 25º58’W and also along the continental margin panning the 15 km distance between the mouth of the Halley Trough and the Brunt Basin.

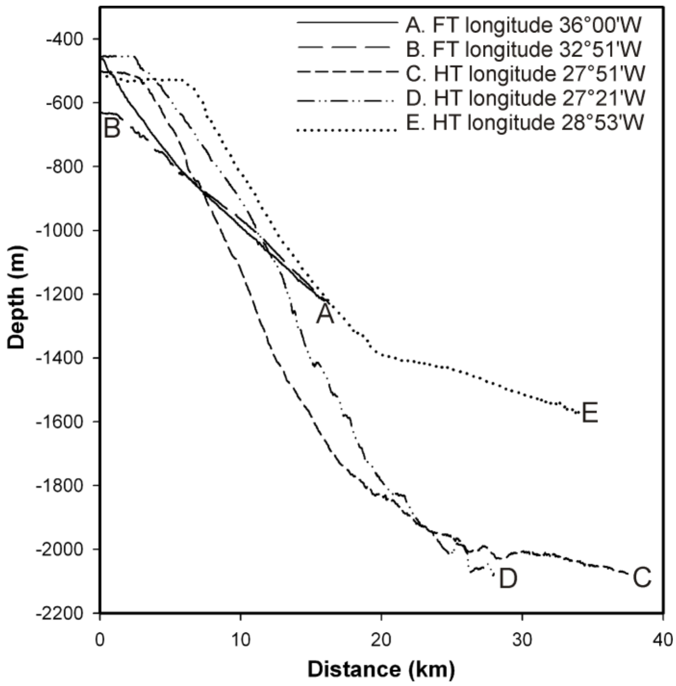

Figure 8. 6. Slope profiles from Filchner Trough (FT) and Halley Trough (HT), Weddell Sea. The locations of profiles A-E are highlighted in Fig. 8.2B.

8.4.1.4. Channel systems

Down slope of Halley Trough, gullies merge with branching and sinuous channels at ~1600 m depth (Fig. 8.8). The channels are V-shaped (U/V = 0.93), with a mean width of 1.2 km and a relief of 46 m taken along a profile at 1800 m water depth. Two sub-systems are identified. The first, between 28º42’W and 27º36’W, merges down-slope into a large channel-ridge system trending northeastward (Fig. 8.8). This previously documented ‘mega-scale channel’ (e.g. Kuvaas and Kristoffersen, 1991; Michels et al., 2002) is 18 km wide and 190 m deep at the 2800 m bathymetric contour.

The second sub-system occurs eastward of 27º36’W. The channels merge with a ~2 km wide larger channel at 2600 m water depth that joins the mega-scale channel at ~3200 m water depth (Fig. 8.8) and proceeds as the Polarstern Canyon (Arndt et al., 2013). Within the sub-system, some channels display a braided morphology (Fig. 8.8D). Acoustic backscattering is higher within the channels than on the adjacent channel interfluves (Fig. 8.7C). Further channel-ridge systems are observed down-slope of the Brunt Basin (Fig. 8.8) (cf. Michels et al., 2002).

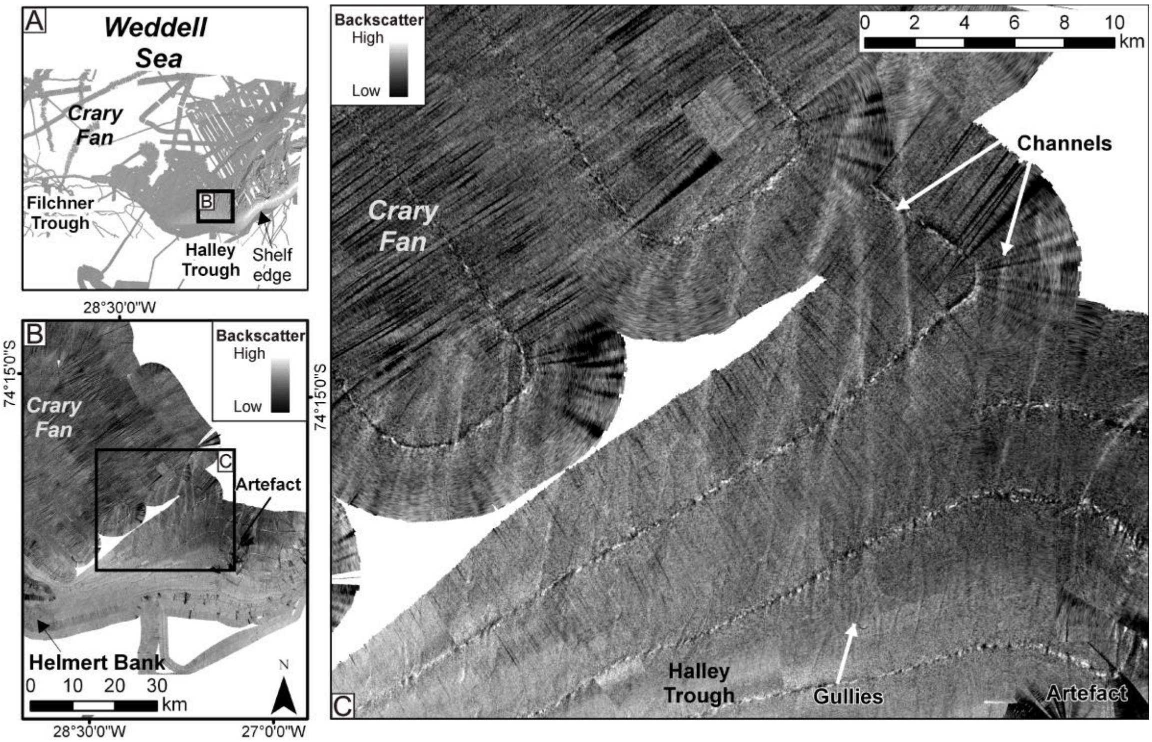

Figure 8. 7. Backscatter analysis from the southeastern Weddell Sea. **A.** Inset figure showing location of part B on hillshaded bathymetry data. **B**. Backscatter data of the mouth of the Halley Trough and slope. Black box locates part C. **C**. Backscatter data of slope seaward of Halley Trough. Darker shading indicates lower backscatter.

8.4.1.5. Mass-wasting features

Two large slides occur on the upper slope of the eastern flank of the Crary Fan, with the headwall of the largest slide (Slide *I*) between 74º33'S, 30º06'W and 74°27'S, 30°41'W (Figs. 8.10, 8.11; Table 8.3). This headwall has a width of 19.6 km and a relief of 60 m near the shelf edge. The slide can be traced for > 20 km down slope. At the shelf edge, the relief of the slide is symmetric (Fig. 8.10, profile A-A'), with depressions at the margins and a shallow relief toward the centre. The headwall of the slide scar has a scalloped and stepped appearance to the west (Figs. 8.10B, 8.10E; Fig. 8.11) with smaller scarps along the headwall to the east, creating an irregular shelf edge (Fig. 8.10). Pitholes are observed on the multibeam data of the headwall of Slide *I* down to a water depth of 570 m (Fig. 8.10E). Seismic data show small scarps on the upper slope surface at 700, 725 and 800 ms TWT (Fig. 8.12). Smaller scarps also occur along the margins of Slide *I* (i.e. 30º1'W; 74º31S; Fig. 8.10). These are ~500 m wide with a mean incision depth of < 20 m (Fig. 8.10). TOPAS data show semi-transparent subsurface bodies within the scar of Slide *I* that appear lens-shaped in down-slope profiles (Fig. 8.10C). In some areas, stacks of lens-shaped bodies occur (Fig. 8.10D). Mean slope gradient within the slide scar is 2º.

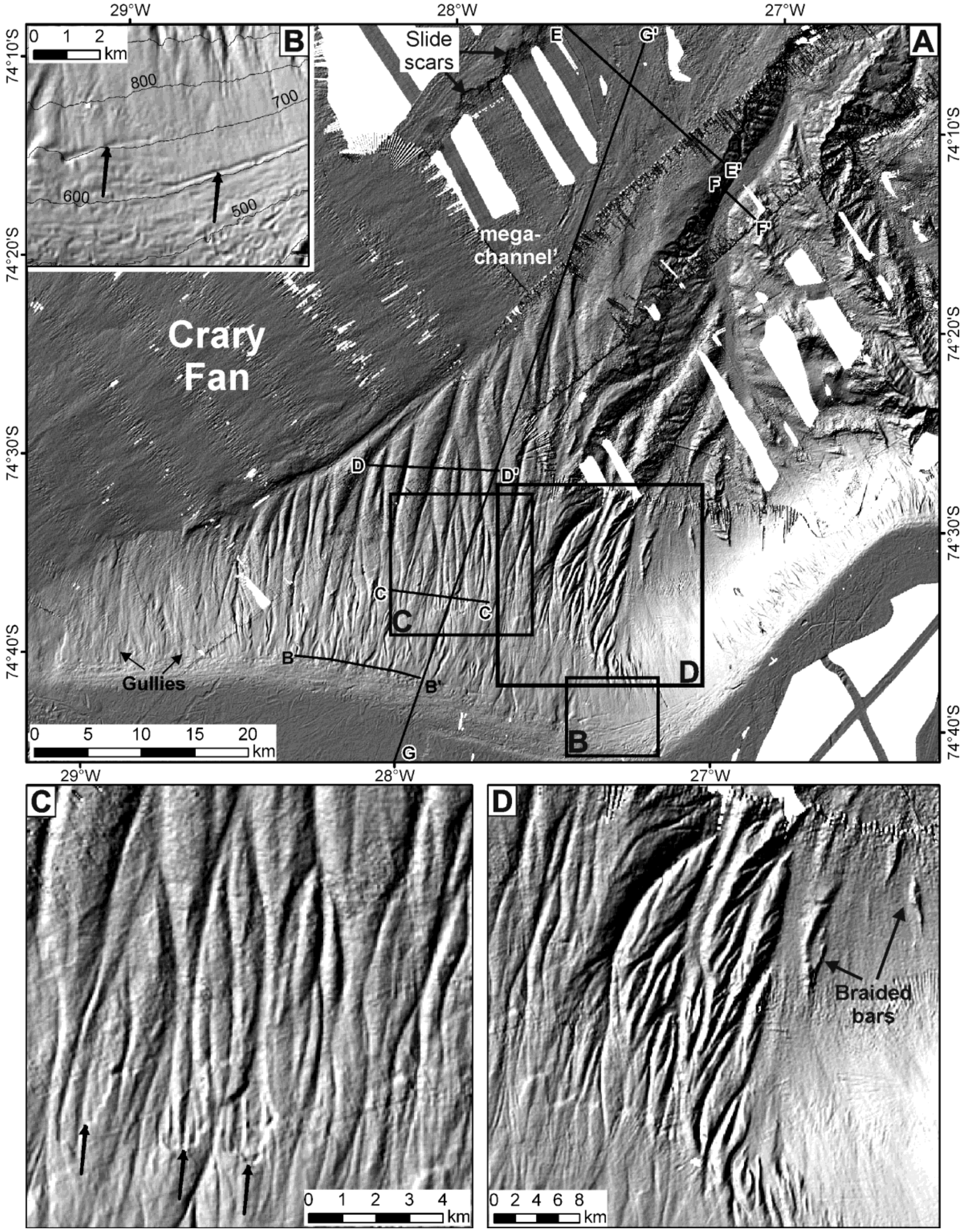

Figure 8. 8. Hillshaded bathymetry data with sun illumination from NW. **A.** Southeastern Weddell Sea continental shelf edge and upper slope. **B**. Iceberg scouring and shelf-parallel furrows (location marked by black arrows). **C.** Small-scale mass-wasting features (location marked by black arrows). **D.** Braided channel system. Lines B-B', C-C', D-D', E-E' and F-F' mark location of cross-shelf profiles in Fig. 8.9. Black down-slope line (G-G') marks down-slope profile in Fig. 8.9. Location of Fig. 8.8A is shown in Fig. 8.1.

Table 8. 3. Dimensions of the slides observed on the eastern Crary Fan, southern Weddell Sea.

Slide	Slope gradient (º)	Length (km)	Width[b] (km)	Incision depth[c] (m)	Direction of movement	Headwall	Segments within deposit
I 30º21W, 74º31S	2.0	> 20	19.6	60	SW to NE	Complex, multiple small steps down-slope of larger scar. Small scars within larger scar.	Elongate rafted slide blocks (> 4).
II 30º59'W, 74º23S	2.3	21[a]	3.0	20	SW to NE	Small, vertical headwall.	No slide blocks observed.

[a]Limited by data extent; [b]maximum width; [c]maximum incision depth.

At least four elongate slabs occur down-slope and to the east of Slide *I* (Fig. 8.10; 8.12A), at 1350 m, 1500 m, 1600 m and 1700 m water depth. Here, the gradient is reduced to 1.4º. The slabs have widths of < 1.8 km, heights of < 25 m and lengths of < 12 km. Seismic data across the largest slab at ~2200 ms TWT (Fig. 8.12A) show reflectors downlapping onto a surface that is continuous with the surrounding seafloor. Below the slab, reflectors are sub-parallel and continuous.

The head of Slide *II* is located at 74º23'S, 30º59'W, and it increases in width with distance down-slope, reaching a maximum width of 3 km and an incision depth of 25 m at 1240 m water depth. It has a steep, nearly vertical scarp of 45 m which initiates at, but does not cut back significantly into, the shelf edge (Fig. 8.11). Slide *II* can be traced for 21 km down-slope with a mean slope gradient of 2.3º.

Small-scale slide scars are also observed along the SE-ridge side of the western bank of the mega-scale channel down-slope of Halley Trough (Fig. 8.8), initiating at ~2350 m water depth. The small-scale slides have a mean width (measured along SW-NE axis) of 2.48 km, a mean incision depth of 39 m and a mean length (measured along NW-SE axis) of 2.16 km.

8.5. Interpretation and discussion

8.5.1. Weddell Sea slope morphology

A strong contrast in the morphology of the seafloor surface is observed between the slopes offshore from the mouths of the Filcher Trough and the Halley Trough, with the former being characterised by deposition and the latter by erosion. Halley Trough is smaller, shallower and shorter than Filchner Trough, and its adjacent continental slope is characterised by a simple slope ramp. In contrast, convex-outward contours at the mouth of the Filchner Trough which remain evenly spaced down the slope, indicate the presence of a trough mouth fan. Slope gradient is on average ~2.5º in front of the Filchner Trough, whereas at the Halley Trough, the slope is steeper, with a mean upper slope gradient of 3.7º (Fig. 8.6). A progression of bedforms is observed down-slope of Halley Trough, with the bedforms increasing in both relief and width down-slope (Fig. 8.9) and ranging from small-scale iceberg scours on the upper slope to large-scale channels on the lower slope. These differences suggest that dominant slope processes vary across this region, probably due to differences in the glacial histories, ice-sheet drainage basin sizes, ice dynamic processes, geology of the sediment source areas and/or the time when these systems were active.

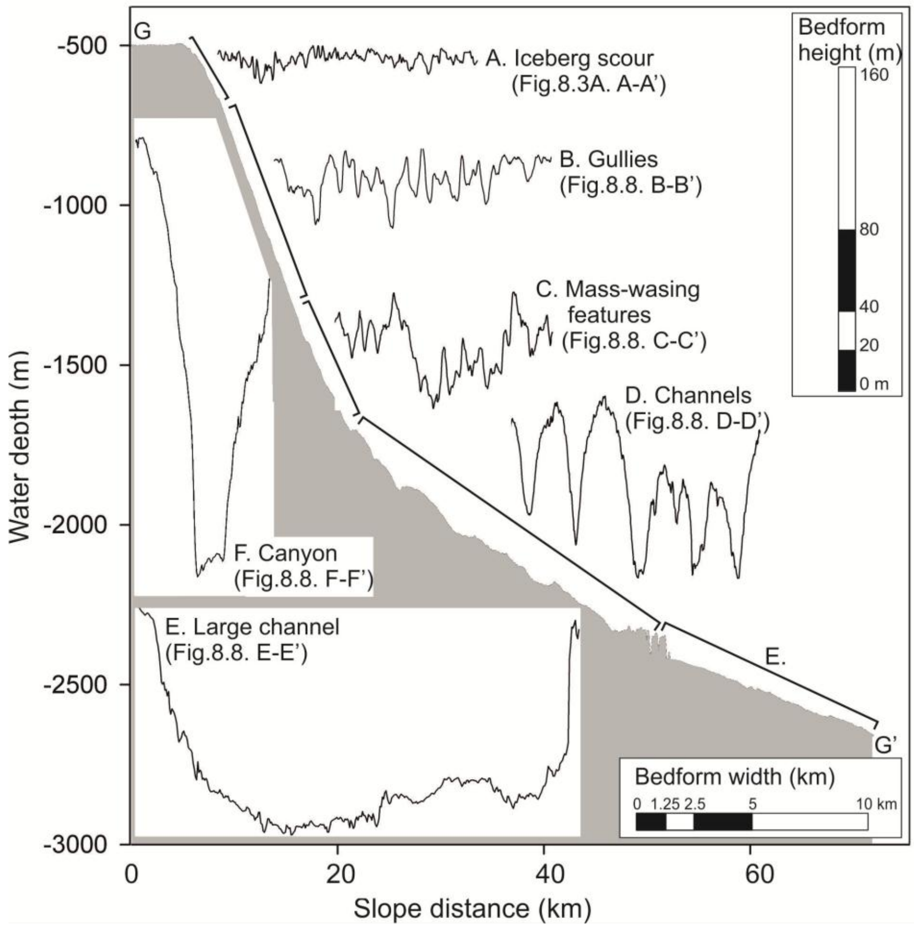

Figure 8. 9. Down-slope bedform progression at Halley Trough mouth. Cross-shelf profile A-A' is marked by black profile A-A' parallel to the shelf in Fig. 8.3A. Cross-shelf profiles B-B', C-C', D-D', E-E' and F-F' are marked by black profiles parallel to the shelf edge in Fig. 8.8. The down-slope profile is marked by the black down-slope profile (G-G') in Fig. 8.8.

8.5.1.1. Iceberg scours

The upper slope seaward of the Halley Trough is significantly affected by pitholes and scours, interpreted to result from scouring by the keels of impacting icebergs. Iceberg keel marks scour the seabed up to 200 m down-slope of the shelf edge (~720 m water depth) (Fig. 8.8B; 8.9). Although the uppermost slope on the Crary Fan is also influenced by iceberg scouring, scours are less continuous there and the seafloor less deeply incised. The maximum water depth of 630 m within the outermost Filchner Trough implies that the icebergs that ploughed the deepest scours did not originate from this region. Icebergs potentially originated from farther along the margin of East Antarctica, which is consistent with observations that show icebergs drifting within the southern limb of the Weddell Gyre in this area, i.e. from NE to SW (Weber et al., 1994; Stuart and Long, 2011). These icebergs were probably injected via the westward flowing Antarctic Coastal Current into the Weddell Gyre (e.g. Gladstone et al., 2001).

Small arcuate sediment ridges are present at the landward terminations of some iceberg scours (Fig. 8.3B). These ridges are formed by sediment pushed by icebergs, suggesting that the shelf edge has been subjected to icebergs impacting from offshore. At present day, Antarctic iceberg scouring in water depths > 500 m are rare, with the deepest scours restricted to the inner and middle shelf. This is because most icebergs calve from ice shelves, whose thickness usually decreases downstream of the grounding line due to creep thinning and basal melt (Dowdeswell and Bamber, 2007), and due to the landward dipping profile of the continental shelf (e.g. Livingstone et al., 2012; Arndt et al., 2013).

8.5.1.2. Gullies

Forty-eight gullies occur along the upper-slope offshore from Halley Trough and eastward towards the Brunt Basin with a relatively high mean gully spacing (average 0.8 gully/km). Although gully widths and depths are similar to those at the mouth of the Filchner Trough (Table 8.2), the Halley Trough gullies have distinct V-shaped cross-sections, whereas the Filchner Trough gullies are predominantly U-shaped. Down-slope of the Halley Trough, the gullies merge with a large channel system. The gullies initiate below a region of intense iceberg scouring at 720 m water depth (Fig. 8.9), and do not incise the shelf-edge. This contrasts with the surface morphology offshore from the Filchner Trough, where gullies incise the shelf edge by ~220 m (Gales et al., 2012). The gullies on the slope offshore from Filchner Trough have short lengths, a shallow mean depth of 12.5 m and U-shaped cross-sections and thus are characteristic of small-scale slide scars (Gales et al., 2012).

The gullies offshore from the Halley Trough mouth are characteristic of 'type *I*' gullies, displaying a non-branching, V-shaped and low sinuosity morphology (Gales et al., 2013a). Such gullies may have formed by suspended sediment flows, such as turbidity currents. Iceberg scouring is one mechanism which can initiate turbidity currents, as intense scouring may resuspend sediment deposited at the mouth of the trough, thus initiating dense turbid flows. This may explain why a large gully-channel system is present at the mouth of the Halley Trough, while sediment at the mouth of the Filchner Trough remain largely unaffected by intense iceberg scouring, with only small-scale gullies incising the shelf edge. These morphological differences suggest that during the period of intense iceberg ploughing at the mouth of the Halley Trough, ice was discharged through the mouth of Filchner Trough, either as an ice shelf protruding over the uppermost slope, or in the form of a continuous flux of icebergs, and thus protected the upper-slope from iceberg impact. Even if no ice shelf protruded over the shelf break during this time, a steady supply of icebergs calving from the Filchner palaeo-ice stream front would have prevented large incoming icebergs from impacting the upper slope, as long as its grounding line position was located on the outermost slope. Alternatively, grounding-line advance to the shelf break of the Filchner Trough may have occurred significantly later than in Halley Trough, so that any deep iceberg scours on the slope offshore from the Filchner Trough were subsequently buried by glacigenic sediment supplied to the shelf edge. Conversely, intense and persistent iceberg scouring on the uppermost slope offshore from the Halley Trough may suggest a small local flux of icebergs calving from the Halley palaeo-ice stream at the time that large icebergs were arriving from offshore, or that the grounding line of the Halley ice stream was located significantly landward of the shelf edge at the time.

The significant iceberg scouring within Halley Trough suggests that intense iceberg activity occurred either after or during retreat of the Halley palaeo-ice stream, when grounded or floating ice was absent from the vicinity of the shelf edge. This is consistent with increased iceberg activity during interglacial periods documented by the high content of iceberg rafted debris in interglacial sediments on the continental rise of the southeastern Weddell Sea (Weber et al., 1994).

Another initiator of turbidity currents is meltwater released at the grounding line of an ice stream. Such flows require a sediment concentration of 1-5 kg m^3, taking into account the effects of fine-scale convective instability (Parsons et al., 2001; Mulder et al., 2003) in order to initiate a freshwater flow that is dense enough to remain at the seafloor. Cascading flows of cold, dense water produced during sea ice formation through brine rejection, may also influence continental slope morphology. However, V-shaped and deeply incised gullies are absent from the mouth of the Filchner Trough, where energetic flows of cold, dense water have been documented flowing down-slope with maximum velocities of 1 m s^{-1} (Foldvik et al., 2004). Studies also suggest that bottom water production is limited on the southeastern Weddell Sea shelf due to a narrower continental shelf (Fahrbach et al., 1994). Like the gullies at the mouth of the Filchner Trough (Gales et al., 2012), it is unlikely that gullies at the mouth of the Halley Trough were formed by cold, dense water overflow.

8.5.1.3. Small-scale mass-wasting features

Small-scale mass-wasting features are observed along a boundary at ~1300 m depth on the slope in front of the Halley Trough but are not observed on the Crary Fan, for which data coverage is more limited (Fig. 8.2). The mass-wasting features incise the gullies, suggesting that they formed later. The features are similar to 'type *IV*' gullies observed on other high latitude margins (Gales et al., 2013a; 2013b), which are characteristic of small-scale mass-wasting. As all the small-scale mass wasting features initiate along a common boundary at ~1300 m water depth, this suggests that they formed due to a change in substrate strength within the shallow subsurface. A sediment core recovered from a channel-levee down-slope of the mass-wasting features (PS1789, see location in Fig. 8.13), retrieved 1.5 m of post-glacial bioturbated muds deposited during the last 16 kyr overlying a sequence of siliciclastic laminated and bioturbated biogenic-bearing sediments deposited from ~24 cal. ka BP to 16 cal. ka BP (Weber et al., 1994; 2011). It is likely that the small-scale slides were caused by failure within weak interglacial substrate sediments.

8.5.1.4. Channels

A large channel system is present down-slope of the Halley Trough, between 28º42'W and 27º16'W (Fig. 8.8). The channels display a sinuous and branching morphology and merge with smaller and shallower gullies further up-slope and a larger mega-scale channel down-slope (Fig. 8.8). There is a clear down-slope increase in both relief and channel spacing. The relief development may originate from increased erosion by the convergence of flows from adjacent tributaries, which is expected to result in increasing flow discharge and bed shear stress (Mitchell, 2004). Alternatively, the increase in relief may reflect the deposition of fine-grained detritus on the channel levees and interfluves. The channels are likely formed by turbidity currents initiated on the upper-mid slope (cf. Kuhn and Weber, 1993; Weber et al., 1994; Michels et al., 2002). At Halley Trough, the slope gradient is higher (~3.5º) which may have allowed sediment transport to largely bypass the slope, with flows becoming confined within the erosional channels (Wynn et al., 2012). No channels systems are observed on the mid-slope offshore from Filchner Trough, although large channel systems do occur on the lower slope (Michels et al., 2002).

In some areas, the channels have a braided morphology (e.g. 74º33'S, 27º15'W). The presence of this braided network suggests a change in flow power, associated with changes in sediment or flow discharge (Ercilla et al., 1998; Hesse et al., 2001). As the slope gradient increases slightly down-slope of the Halley Trough (Fig. 8.6, Profile D), the existence of the braided channel network suggests that suspended load must also increase for sediment to be deposited, as flow velocity otherwise increases with increasing slope gradient. The braided network may be associated with mass-wasting features further up-slope which provide increased detritus supply during periods of small-scale slope failure.

8.5.1.5. Slides

Two large slides occur on the eastern flank of the Crary Fan. Both slides have complex headwalls initiating at the shelf edge. Down-slope bathymetric profiles and seismic data show a 'stepped' appearance from the shelf edge down to ca. 600 m water depth (Figs. 8.10B; 8.11; 8.12B) indicating that sections of the upper slope have crept down-slope, perhaps by retrogressive erosion. Several smaller scarps are present on the headwall of Slide *I*, demonstrating that multiple stages of instability have occurred. The seismic data show that sub-parallel and continuous reflectors occur below the stepped shelf edge and below one of the elongate slabs observed further down-slope. The slabs on the slope below Slide *I* which are visible on both seismic and bathymetry data (Fig. 8.10; 8.12A), are either slide blocks rafted from further up-slope or resemble remnant blocks. Seismic data show that the slabs rest on a palaeo-seafloor surface covered by a sediment drape, indicating that they are relatively young features.

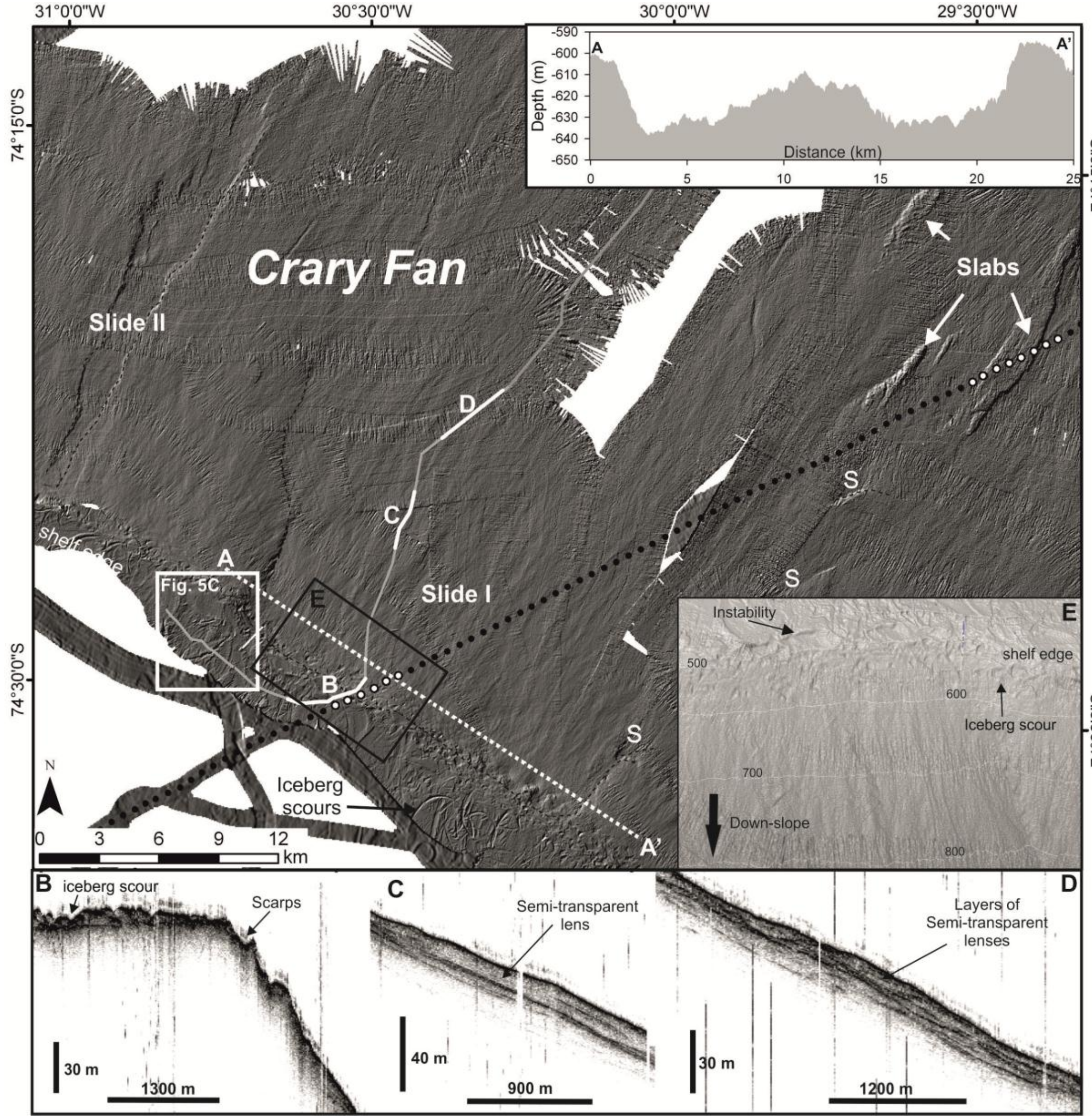

Figure 8. 10. A. Submarine slides and mounds on the eastern Crary Fan flank, southern Weddell Sea. White dotted line is cross-shelf profile A-A' in upper right inset. Grey solid line highlights TOPAS profile (JR244) with white solid lines locating TOPAS profile sections shown in Fig. 8.10B, 8.10C and 8.10D. Bold, black dots highlight seismic profile AWI90060 (ANT VIII/5) with white dots locating sections displayed in Fig. 8.12A (eastern section) and Fig. 8.12B (western section). Black dashed lines illustrate extent of Slide *I* and *II* and scarps. Letters 'S' indicate scarps of slide scars. Location of Fig. 8.10 is shown in Fig. 8.2C. White square locates Fig. 8.5C. **B.** TOPAS profile section through shelf edge and upper-slope. **C.** TOPAS profile section through upper slope. **D.** TOPAS profile section through mid-slope. **E.** Hillshaded multibeam bathymetry of shelf edge and upper slope.

Recent submarine slides are rare on the Antarctic continental margin, with few other documented examples (e.g. Barker et al., 1998; Imbo et al., 2003). Although the relief is not as large as the 175 m deep Gebra Slide on the Trinity Peninsula margin (Imbo et al., 2003), Slide *I* has a greater width and can be traced for a similar length down-slope. Seismic data from the axis of Slide *I* show its location in a region of the Crary Trough Mouth Fan, where large-scale mass wasting affected the upper continental slope during the early Pliocene (Bart et al., 1999). This collapse resulted in the erosion of major channels by sediment gravity flows, which were later infilled. This early Pliocene collapse of the Crary Fan may have resulted from isostatic rebound and sea-level rise associated with a major reduction in Antarctic ice-sheet volume (Bart et al., 1999).

Other factors which influence slope instability on high latitude continental margins include: (1) gas hydrate dissociation or methane-gas generation; (2) tectonic influences (earthquakes, tsunamis); (3) rapid accumulation of sediment at the shelf edge and uppermost slope resulting in under-compaction and excess pore pressure and/or slope over steepening; (4) loading or unloading of ice at the shelf break; and (5) presence of 'weak' sediment layers within the seabed (Prior and Coleman, 1984; Bugge et al., 1987; Larter and Barker, 1991; Dowdeswell and Ó Cofaigh., 2002; Long et al., 2003; Nielsen et al., 2005; Laberg and Camerlenghi, 2008). The Weddell Sea is a passive margin, with no evidence for methane gas or gas-hydrates (e.g. Bart et al., 1999). The low slope gradients (~2º) suggest that slope over-steepening is not likely to have triggered the slides as previous studies have shown that high latitude margins are able to maintain high slope gradients due to the generally poor sorting of glacially-transported sediments (Larter and Barker, 1989).

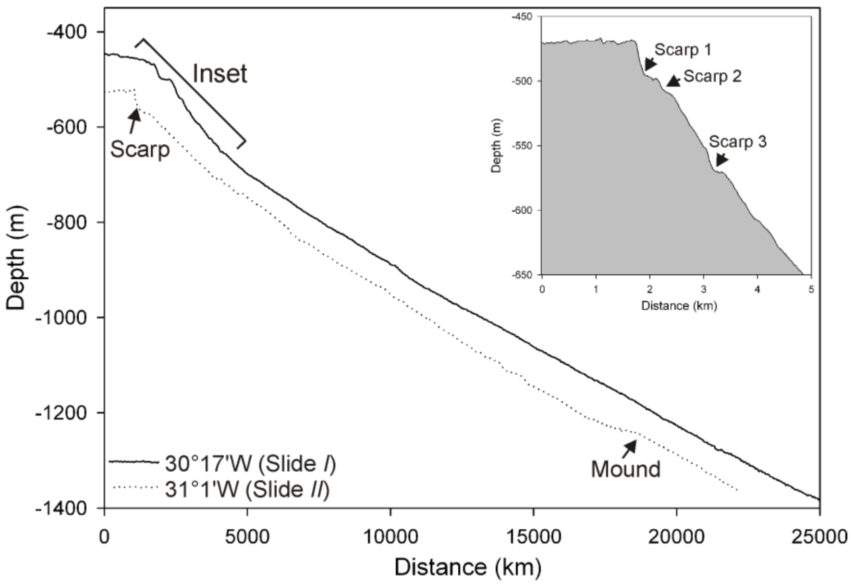

Figure 8. 11. Down-slope profiles within Slide I (solid black line) and Slide II (dashed line). **B**. Inset figure showing down-slope profile of shelf edge and upper slope.

One mechanism which may influence the occurrence of slides on the Crary Fan is the presence of weak sediment layers interbedded with unsorted glacigenic detritus at the seafloor and within the seabed. Weak layers (i.e. fine grained, saturated, high clay content and/or underconsolidated) are common on Arctic continental margins and may have been essential in the initiation of the Storegga, Trænadjupet and Afen slides (Bugge et al., 1983; Wilson et al., 2003; Canals et al., 2004). On the Antarctic upper slope, weak layers are generally thought to be less common, with contouritic layers mainly controlled by ocean circulation, and hemipelagic layers formed during interglacial or deglacial periods (Long et al., 2003; Nielsen et al., 2005; Laberg and Camerlenghi, 2008). However, the southern

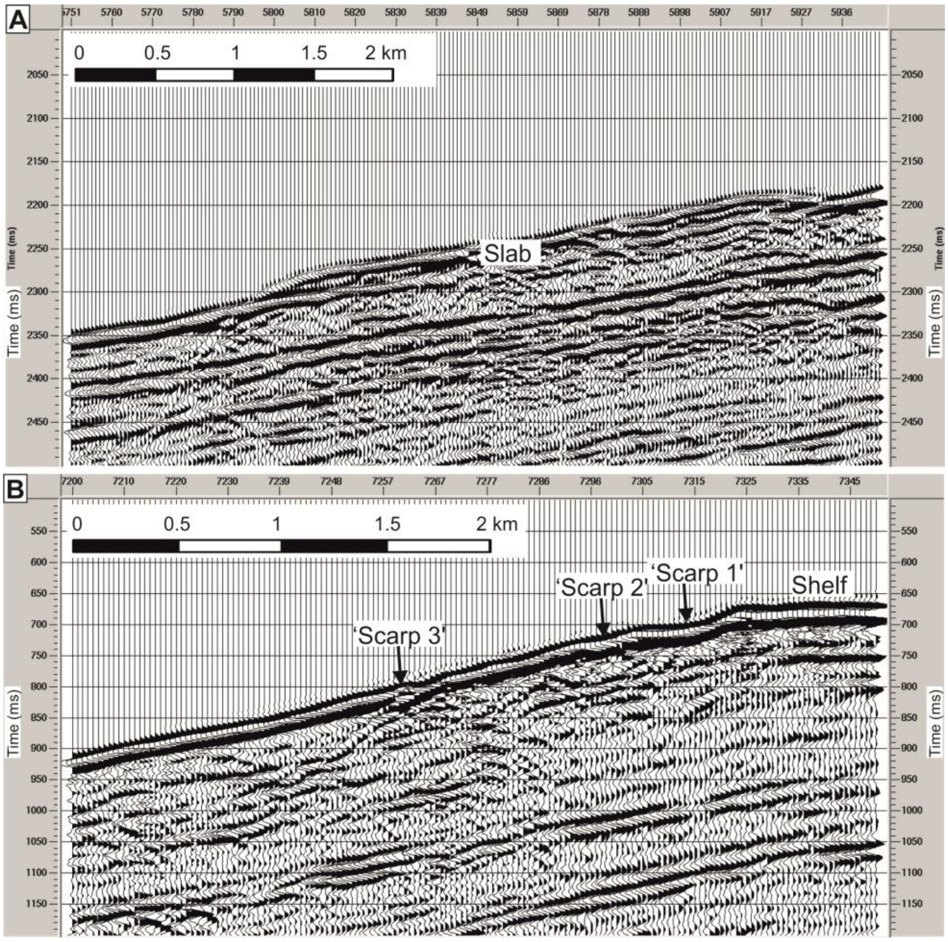

Figure 8. 12. Seismic data from seismic profile AWI90060 (Cruise ANT VIII/5) located in Fig. 8.10. **A**. Slabs on the lower slope. **B.** Scarps within slide *I* on the upper slope.

Weddell Sea shelf is one of the most important regions for dense bottom water formation and cascading flows may winnow fine particles from the shelf, depositing finer sediments further down-slope (Melles and Kuhn, 1993; Weber et al., 2011). These sediments may become mixed with glaciomarine contouritic, potentially hemipelagic and glacigenic sediments on the slope. These finer sediments may form porous layers with a high-water content, low density, low strength and higher susceptibility to liquefaction under further loading (Kuhn and Weber, 1993; Long et al., 2003). Compaction of the substrate with further loading may expel water along the more permeable and weaker layers (Dugan and Flemings, 2000) reducing the stability of the slope. This may explain why recent slides occur on the Crary Fan, in a region of energetic and cascading dense water overflow, but are largely absent from many other Antarctic slope areas, where the oceanographic setting precludes cold, dense water formation.

TOPAS data from the axis of Slide *I* show lenticular bodies of semi-transparent sediment, interpreted to be debris flow deposits. Stacked debris flow deposits occur down slope (Fig. 8.10C; 8.10D), indicating that the debris flows post-date the slide event and possibly originate from retrogressive failures of the slide headwall. Rapid sediment transfer to the slope may be a further mechanism influencing the instability of the Crary Fan, especially if sediment is deposited rapidly on top of weaker layers, increasing pore pressures within the seabed and thus slope instability. Additional sediment cores from this region are needed to constrain both the timing of the events and the nature of the underlying sediment, which may shed light on slide initiation mechanisms.

8.5.2. Past glacial history

The glacial history of the Filchner Trough is widely debated (Hillenbrand et al., 2013). In contrast, there is little knowledge about the glacial history of the Halley Trough; however, our new bathymetric data show arcuate escarpments, which may mark the limit of a till sheet, and a terminal moraine suggests that ice was grounded near to the shelf edge during previous glaciations.

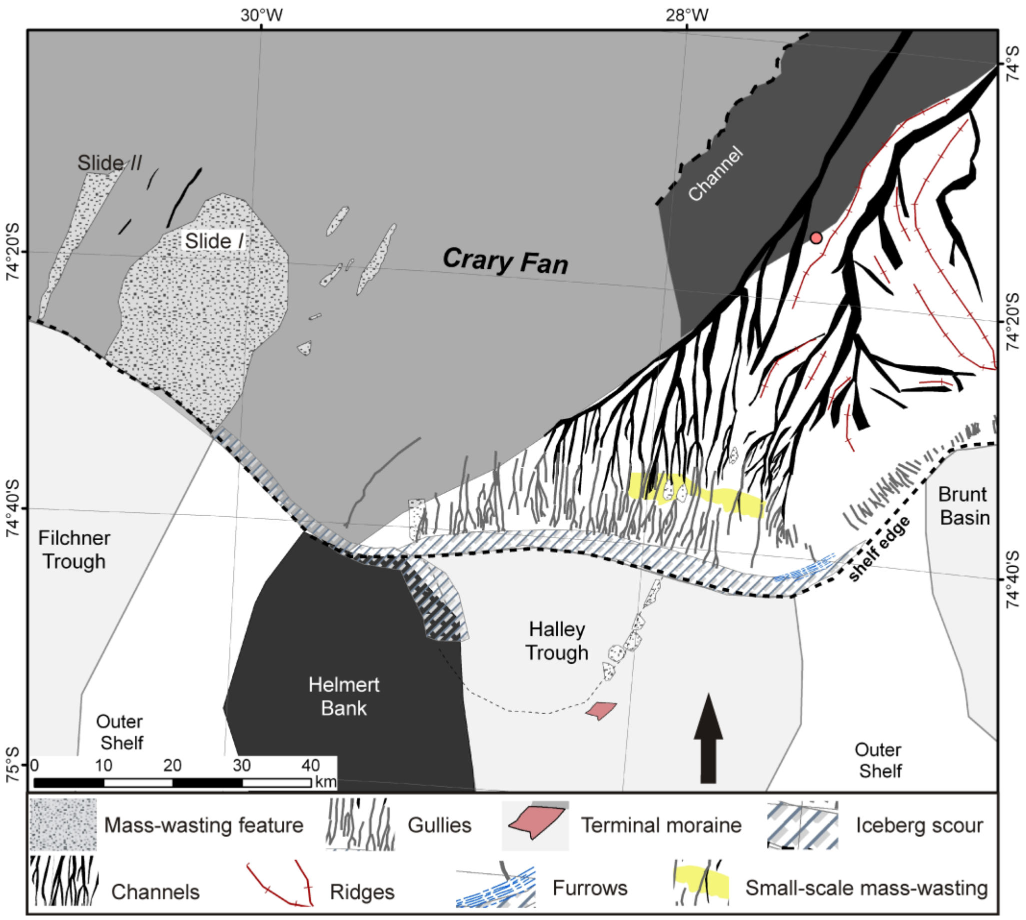

Figure 8. 13. Interpretation of geomorphological features in the southeastern Weddell Sea. Dashed line marks shelf edge. Dotted line marks boundary of deeper section of seafloor within Halley Trough. Black arrow marks direction of palaeo-ice flow. Red point marks location of sediment core PS1789 (Weber et al., 1994, 2011).

8.5.2.1. Evidence for grounded ice

Geomorphic features from the mid to outer shelf suggest that ice was present near the shelf edge during previous glacial periods in the Filchner Trough (Larter et al., 2012; Hillenbrand et al., 2012). We observe a terminal moraine (Fig. 8.4) within the Halley Trough (22 km landward of the shelf edge), documenting that ice advanced to the outermost shelf here as well. The moraine has a 45 m high lee side, displaying a characteristic asymmetric shape toward the trough mouth. The presence of this terminal moraine probably marks the maximum extent of a grounding line advance. Mega-scale glacial lineations are not observed landward of the terminal moraine, but these may have been eradicated by subsequent iceberg scouring.

A NE-SW trending escarpment is characterised by six scalloped embayments (Fig. 8.3C), which incise the outer shelf between the shelf-edge and the terminal moraine. The features are characteristic of small-scale slope failures, with mounds at the mouths of the scars that are probably slump or debris deposits. This unique morphology has not been observed along other cross-shelf troughs in Antarctica, and the formation process for the embayments is unclear. We suggest that they formed as a result of slope failures in unconsolidated mud deposits which were overridden by a till sheet. As the region is also heavily iceberg scoured, prolonged iceberg furrowing may also have eroded the margin of the till sheet. A tentative explanation is that ice extended towards the outer shelf from the southeast during the last glacial period, but did not ground on the deeper western side of the trough. The embayments may therefore have formed on the western boundary of the till sheet deposited along the eastern side of the trough and may mark the westernmost extent of grounded ice during the last ice advance. A more detailed study of the inner and mid-shelf morphology and information from sediment cores are needed to further constrain the past glacial history of the Halley Trough.

8.5.2.2. Drainage basin size and source area

The significant differences in sedimentation and surface morphology along the continental margin from the Filchner Trough to the Halley Trough are likely to reflect differences in drainage basin size and basal conditions of the ice streams draining through them and probably also differences in the timing of their advance and retreat. Regional bathymetry data (IBCSO; Arndt et al., 2013) suggest that the Filchner and the Halley troughs were fed from different sources, with the Filchner Trough fed by a palaeo-ice stream with a drainage basin extent into the interior of both West and East Antarctica (Larter et al., 2012) and Halley Trough fed by glaciers draining a smaller basin along the Caird Coast in East Antarctica.

A clear boundary is observed in the backscatter data between the eastern Crary Fan flank and the slope west of Halley Trough (Fig. 8.7). Coincidentally, the slope gradient on the eastern Crary Fan (Fig. 8.6, profile E) is lower than further west (Fig. 8.6, Profile A-D). The lower backscatter on the Crary Fan probably indicates deposition of finer grained sediments than on the upper slope offshore from the Halley Trough and suggests a predominantly depositional environment. This variation in sediment texture may reflect differences in the source rock areas and composition, glacial transport distances, oceanographic circulation (e.g. fine-grained sediment transport along the pathways of cold, dense water flow) or the size of the drainage basins. The extent of the glacier and drainage basins may also influence the amount of meltwater discharged down-slope. The Filchner palaeo-ice stream would have drained a large area in the interior of the Antarctic Ice Sheet. In the vicinity of the onset of this palaeo-ice stream, snow would have accumulated at higher elevations and in colder temperatures, resulting in colder ice, compared to the palaeo-ice stream that drained over a relatively short distance through the Halley Trough (cf. Dixon, 2008; Arndt et al., 2013). Conversely, high elevation ratios of glacial lineations mapped on the shelf within the Filchner Trough indicate that there, the palaeo-ice stream flowed relatively fast (Larter et al., 2012). Therefore, strain heating at the base of the Filchner palaeo-ice stream may alternatively have resulted in warmer basal ice. The balance between these two effects has implications for the amount of subglacial meltwater produced, and numerical ice sheet modelling may provide insight into which effect dominates.

8.6. Conclusions

We observe significant differences in continental slope morphology at the mouths of two Antarctic cross-shelf troughs in the southern Weddell Sea. Although the troughs are geographically close to each other, we identify large-scale differences in sediment depositional and erosional features.

- Large-scale differences between the Filchner Trough and the Halley Trough are observed in the outer shelf and upper slope morphology. Offshore from the mouth of the Filchner Trough, a large trough mouth fan (Crary Fan) is present, with small-scale and U-shaped gullies incising the shelf edge and two large slides observed on the slope. The mouth of the Halley Trough has a distinct erosional morphology, with deeply entrenched and sometimes braided channel systems and small-scale and upper-slope V-shaped gullies which do not incise the shelf edge. Bedforms (iceberg scours, gullies, small-scale mass-wasting features, channels and mega-scale channels) all increase in size down-slope.
- Two slides are observed on the eastern flank of the Crary Fan, with slide blocks rafted down-slope to the northeast. Slide initiation here may have involved weak sedimentary layers and/or cascading flows of cold, dense water through the Filchner Trough. This is the first example of a relatively young slide on an Antarctic trough mouth fan and provides evidence for large-scale mass-wasting processes on the Antarctic margin during the late Quaternary.
- Halley Trough mouth is significantly affected by iceberg plough marks with the icebergs originating from the shelf further east. Filchner Trough is less affected by intense iceberg impacts, which suggests that during the time when large icebergs scoured the Halley Trough mouth, the mouth of Filchner Trough was covered by an ice shelf protruding over the shelf break; the trough was protected from the impacting icebergs by icebergs being discharged from the front of the Filchner palaeo-ice stream; or that the grounding line of this ice-stream was located further inshore and advanced to the shelf break much later than the Halley palaeo-ice stream, resulting in the burial of deep iceberg scours by glacigenic debris.
- The differences in slope morphology observed between the Filchner Trough and the Halley Trough may reflect differences in the drainage basin size, basal ice conditions and in the histories of advance and retreat of the palaeo-ice-streams

that drained through them. Backscatter data suggest that the sediment offshore from the mouth of Halley Trough is coarser grained than offshore from the mouth of Filchner Trough, potentially indicating differences in the source areas and source rock composition of these sediments.

- This study highlights the need for a more detailed study of the inner and mid-shelf of the southern and southeastern Weddell Sea in order to better constrain the chronology and behaviour of past ice-sheet history, and the need for further marine geological, geotechnical and geophysical investigations to improve understanding of how differences in ice-stream systems are manifested in slope processes and morphology.

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Section 3:

Synthesis and conclusions

Chapter 9.

Synthesis, conclusions and recommendations for further study.

In this final chapter, the findings of the previous chapters (chapters 4-8) are summarised and the wider implications of the results in terms of new contributions to science, unanswered questions and future research, are discussed.

9.1. Summary of results

The primary objective of this research was to constrain processes operating on Antarctic continental slopes with the aim of better understanding what processes produced the slope morphology present. This was achieved through addressing the following aims:

Aim 1: What are the main morphological features of Antarctic continental slopes and how do these morphologies vary?

The work presented in this thesis includes the analysis of over 2670 km of multibeam bathymetric data along high latitude continental shelves and upper slopes of the Arctic and Antarctic and over 1568 submarine gullies. Antarctic gully characteristics are summarised in chapter 4, including a detailed description of gully morphology from the southern Weddell Sea. In chapter 5, five quantitatively distinct gullies types were identified from the Antarctic continental margin. This analysis included measuring a range of parameters for > 1100 separate gullies on the Antarctic continental slope, with parameters including gully width, depth, length, branching order, sinuosity, shelf-incision, gully spacing and slope gradient. Statistical analysis showed that the five gully types identified were statistically significant. Based on this analysis, identification criterion were proposed to quantitatively categorize gullies on Antarctic continental margins. Using these identification criterion, the spatial distribution of Antarctic gullies was mapped for over 1921 km of the Antarctic continental margin.

In chapter 6, using a similar methodology to chapter 5, gully morphologies were analysed over a 520 km stretch of the Arctic continental shelf and upper slope, including the northern Norwegian margin, southwest Barents Sea and western Svalbard margin. Arctic gullies were identified according to the identification criteria proposed in chapter 4 and their spatial distribution mapped. The parameters of over 1500 Arctic and Antarctic gullies were compared, including length, width, incision depth, cross-sectional shape, density, branching order and sinuosity. Antarctic gullies were found to display significantly deeper incision depths (mean depth of 48 m compared to 12 m in the Arctic) and significantly greater shelf-incisions (325 m compared to 35 m in the Arctic). Statistical analysis showed that the differences between Arctic and Antarctic gullies were significant.

In chapters 7 and 8, new multibeam bathymetric data collected during 2011 and 2012 from the southern and southeastern Weddell Sea were analysed. The shelf edge and upper-slope morphology of these areas were previously unknown due to the isolated nature of the area and difficult surveying conditions (i.e. extensive sea-ice cover). In chapter 7, a detailed examination of the southern Weddell Sea seafloor morphology revealed the presence of small-scale and U-shaped gullies at the mouth of the Filchner Trough. Large, V-shaped and deeply incised gullies, as observed on other Antarctic continental margins were absent.

In chapter 8, regional bathymetric data and new multibeam bathymetric data collected in 2012 showed a small and shallow cross-shelf trough to the east of the Filchner Trough in the southeastern Weddell Sea, referred to as ‘Halley Trough’. At the mouth of Halley Trough, a down-slope progression of bedforms is observed, including iceberg scours, gullies, small-scale mass-wasting features, channels and mega-scale channels. The features increase in incision depth down-slope. The outer shelf and upper slope is significantly iceberg scoured with small-scale and V-shaped gullies occurring on the upper-slope. The gullies do not incise the shelf edge but initiate ~150 m below the shelf edge. Small-scale mass-wasting features cross-cut the gullies, with branching and sinuous channels observed down-slope of the gullies. In some areas, the channels display a braided morphology. At ~3200 m water depth, the channels merge into a mega-scale channel. On the eastern flank of Filchner Trough, two large submarine slides are observed. These are the first relatively recent slides to be documented on an Antarctic trough mouth fan and provide evidence for geologically recent large-scale mass-wasting.

Aim 2: What factors or processes are responsible for the different morphologies observed on Antarctic continental margins?

In chapters 5 and 6, the large-scale slope characteristics, oceanographic regime and underlying geology were analysed and used to infer the environmental controls influencing differences in gully type and gully spatial distribution. The results of chapter 5 show that Antarctic gully morphology reflects a balance between local slope character (slope geometry, slope gradient), large-scale spatial characteristics (i.e. subglacial meltwater production rates, drainage basin size, location of cross-shelf troughs, regional heat flow and strain heating distribution), ice-sheet history and sediment yield. In chapter 6, differences in the environmental controls (i.e. oceanographic regime, past glacial history, geotechnical differences) were used to infer factors influencing the differences observed between Arctic and Antarctic gullies. These differences suggest that either Antarctic gullies formed over significantly longer time-periods compared to gullies on the Arctic continental margin, or the process forming the more deeply incised and shelf-incising Antarctic gullies were more prevalent in Antarctica, since the Last Glacial Maximum.

In Chapter 7, the hypothesis that gullies were eroded by cascading flows of cold, dense water was tested. New multibeam bathymetric data were acquired in 2011 from the southern Weddell Sea, in a region of active and highly energetic cold, dense water overflow. Deeply incised and V-shaped gullies, which are common along other parts of the Antarctic continental margin (chapter 5), were shown to be absent from the mouth of the Filchner Trough. Instead, the shelf edge and upper slope is incised by small-scale and U-shaped gullies. This suggests that cascading flows of cold, dense water are not likely to have formed the V-shaped and deeply incised gullies observed on the Antarctic continental margin.

In chapter 8, the slope morphology at the mouths of two adjacent Antarctic cross-shelf troughs was analysed. Distinct differences in slope morphology were observed between the Filchner (southern Weddell Sea) and Halley (southeastern Weddell Sea) Trough. The Filchner Trough morphology, analysed in chapter 7, is interpreted as a largely depositional environment, with a large trough mouth fan present. At Halley Trough, the slope gradient is steeper and large gully-channel systems occur. The outer shelf and upper slope of Halley Trough is significantly iceberg scoured, with gullies initiating 150 m below the shelf edge, compared to the Filchner Trough, where gullies initiate at the shelf break and incise back ~220 m into the shelf. These variations suggest that different environmental controls are influencing processes operating in the southern Weddell Sea. These variations likely result from differences in the glacial systems and histories of the ice streams that fed them, including differences in drainage basin size, source areas and probably different basal conditions, dynamic behaviours and phases of ice-sheet advance and retreat.

The differences in gully morphology observed between the Filchner and Halley Troughs (chapter 8) suggests differences in gully-forming mechanisms on the continental slope. The U-shaped gullies at the mouth of Filchner Trough are likely small-scale slides. The small-scale and V-shaped gullies observed at the mouth of the Halley Trough initiate at the maximum extent of iceberg scouring down-slope (~720 m water depth). One potential gully-forming mechanism at the mouth of the Halley Trough is the initiation of turbidity currents by iceberg scouring, where intense scouring resuspends sediment deposited at the mouth of the trough initiating dense gravity flows.

Aim 3: What can the slope processes inferred from the Antarctic continental margin contribute to the wider field of continental slope research?

The study provides an extensive analysis of high latitude submarine gully geomorphology and presents an identification scheme to categorize high latitude gully types. Through comparing the spatial distributions of high latitude slope gullies along with local slope characteristics and environmental controls, this study provides a step forward in constraining gully-forming mechanisms and processes occurring on glaciated continental margins. The study finds that cascading flows of cold, dense water are not likely to have formed the deeply eroded and V-shaped gullies observed over much of the Antarctic continental margin (chapter 5 and 7). Instead, the V-shaped and deeply incised gullies are likely formed by turbidity currents initiated by processes including release of sediment-laden subglacial meltwater from beneath an ice sheet grounded near to the shelf edge (chapter 5). Other initiation mechanisms include the resuspension of sediment by intense iceberg scouring on the shelf edge and upper slope (chapter 8). Environmental controls, including local slope characteristics (slope gradient, geometry), drainage basin size, geotechnical properties and ice dynamic histories are shown to modify the effect slope processes have on the seafloor morphology. These may influence gully parameters including gully length, shelf-incision, sinuosity, branching order, incision depth and spatial density (chapter 5 and 6).

Chapter 6 provides the first systematic comparison of Arctic and Antarctic submarine gullies. Five quantitatively distinct gully types are observed on Arctic continental margins. Four of the gully signatures are similar to Antarctic gully types and one type is unique to the Arctic continental margins analysed in this study. Two distinct differences are observed between Arctic and Antarctic submarine gullies, with Antarctic gullies displaying significantly deeper incision depths, and gullies that incise the shelf-edge further. This suggests that either Antarctic gullies were formed over longer time-periods than gullies on the Arctic continental margin, or the processes which formed the gullies were more abundant in Antarctica, since the Last Glacial Maximum.

Chapter 7 and 8 provide the first analysis of the continental shelf and upper-slope geomorphology at the mouths of the Filchner and Halley Troughs, southern Weddell Sea. Seafloor features observed landward of the Halley Trough mouth (e.g. terminal moraine, scalloped escarpments) provide insight into past glacial histories and ice dynamic behaviours. The presence of a terminal moraine probably marks the maximum extent of an ice grounding line advance, suggesting that ice was grounded at least near to the shelf edge during previous ice advance.

Two large and geologically recent (Late Quaternary) submarine landslides are described in chapter 8 on the eastern flank of Filchner Trough, southern Weddell Sea. This is unusual because recent submarine slides are largely absent from the Antarctic continental margin and no other recent slides have been documented on an Antarctic trough mouth fan. The occurrence of Antarctic slides provides evidence for recent, large scale mass-wasting which is thought to be rare on the Antarctic continental margin. Studying the geotechnical properties in more detail in this area may provide insight into controls influencing large-scale slope instability and factors influencing slide initiation. This is particularly important for better understanding geohazards and predicting future risks associated with slope instability.

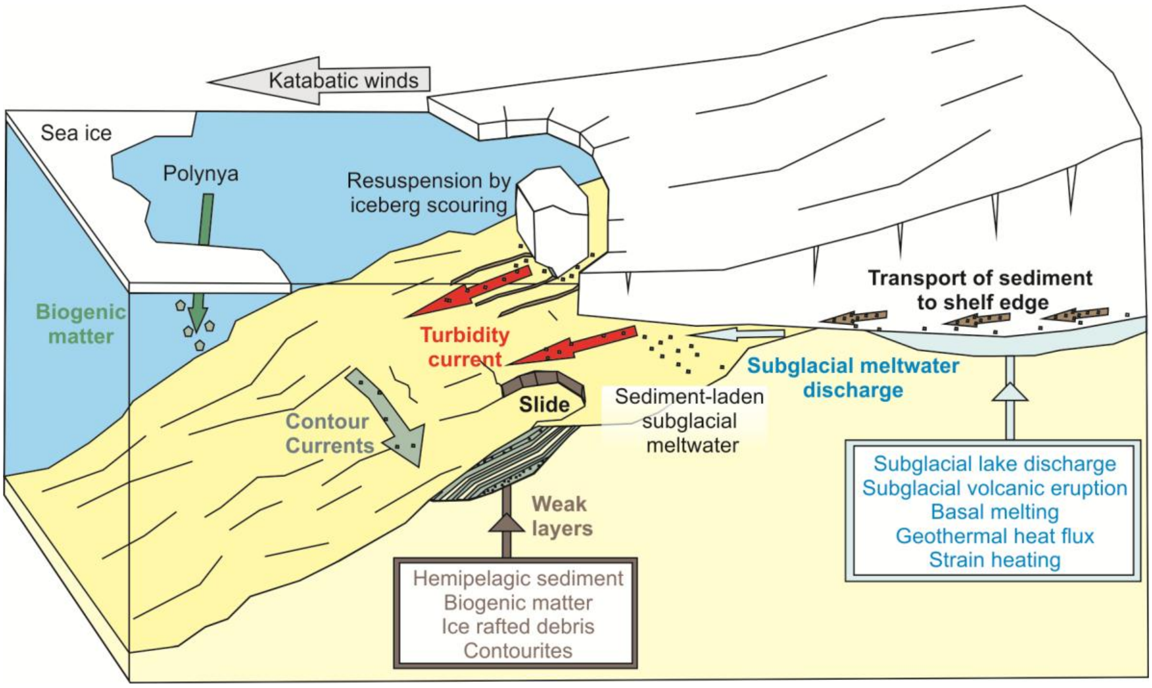

Figure 9. 1. Synthesis of gully-forming mechanisms operating on Antarctic continental margins, including turbidity currents initiated by sediment resuspended by iceberg scouring and/or sediment-laden subglacial meltwater; and small-scale slides influenced by weak layers in the shallow subsurface. Adapted from: http://www.spp-antarktisforschung.de/projects/Bernhard-Diekmann/Bernhard-Diekmann-DI-655-3.html.

9.2. Unresolved questions and recommendations for future research

The study has shown that cascading flows of cold, dense water overflow are unlikely to have formed the V-shaped and deeply eroded gullies observed over much of the Antarctic continental margin. Instead, these gullies are likely formed by turbidity currents initiated by either sediment-laden subglacial meltwater released from beneath an ice sheet grounded near to the shelf edge, or by sediment resuspension by intense iceberg scouring. However, there remain unresolved questions concerning the ability of subglacial meltwater to erode the seafloor. Studies have documented extensive palaeo-meltwater channels on the Antarctic and Arctic shelf (e.g. Lowe and Anderson, 2002; Graham et al., 2009; Hogan et al., 2010; Nitsche et al., 2013); it is therefore clear that subglacial meltwater was present beneath the ice. However, modern studies show that the volume of subglacial meltwater produced through processes such as basal melting and geothermal heat flux is in the range of mm/yr (Pattyn, 2010). Mechanisms suggested to increase this volume include subglacial lake discharge and volcanic eruptions beneath the ice (e.g. Nitsche et al., 2013). Studies of hyperpycnal flows have shown that 1-5 kg m^3 of sediment is needed to initiate a flow which is dense enough to overcome the buoyancy effects of freshwater in seawater (Parsons et al., 2001; Mulder et al., 2003). However, whether small volumes of continuous meltwater released from beneath an ice-sheet can initiate turbidity currents, or whether bedforms are the result of episodic flows of greater volumes of meltwater remain unknown. Further research into subglacial meltwater and the effects this has on sediment dynamics is needed. Measurements taken at the grounding line, for example using an Autonomous Underwater Vehicle (AUV) to measure turbidity, temperature, salinity, density, depth and video footage, would allow properties of the water column at the grounding line to be analysed. Secondly, push-cores taken by an AUV may provide insight into the effects on shallow sediment dynamics at the grounding line.

A further unresolved matter is the timing over which gullies form. The question of whether gullies develop with each new glacial cycle, or are able to maintain their morphology and develop over longer time periods and perhaps over multiple glacial cycles, can only be resolved by employing a 3D seismic survey over the Antarctic outer shelf and upper slope (e.g. perhaps using the ‘P-cable’ 3D single-channel system). This would allow palaeo-gully morphologies to be observed in the subsurface and would allow gully evolution to analysed.

Unresolved questions remain regarding slide initiation mechanisms. Sediment coring and further seismic surveying within the two slides observed on the Crary Fan, southern Weddell Sea, and the surrounding area may provide insight into the geotechnical properties influencing slide initiation including whether active gas or gas hydrates are present in the sediment, which may influence slope instability and slide initiation.

Finally, the thesis has focussed on submarine gully morphology from glaciated high latitude continental margins. However, gullies are also observed on low and mid-latitude margins, on hillslopes in the terrestrial environment, on the flanks of deep-sea sediment drifts and even on Martian landscapes (Malin and Edgett, 2000; Hartmann et al., 2003; Micallef and Mountjoy, 2011; Vachtman et al., 2012). The similarities or differences between these different gully morphologies may provide insight into gully-forming mechanisms and environmental controls influencing their morphology.

9.3 Concluding remarks

Submarine gullies are the most common features of high latitude submarines slopes. They influence sediment transport and deposition down-slope, have a key role in continental margin evolution and contribute to sediment deposits on the continental rise. Gullies have been suggested to be the first features to develop on steeply dipping slopes, similar to terrestrial settings, where gullies are known to represent the first step in the fluvial dissection of landscapes (Bloom, 1991). Although this study provides a step forward in constraining the processes operating on high latitude continental margins, and the factors influencing gully morphology, many unresolved questions still remain, with our ability to interpret and understand the submarine environment limited by data availability, resolution and type.

A better understanding of processes operating on continental margins is essential for geomorphologists, geologists interpreting sediment cores and glaciologists modelling ice-sheet change. Improved knowledge of the past extents of ice-sheets, past ice-dynamic histories, and rates of past ice-sheet change can contribute to improved ice-sheet modelling which aim to inform and address future risks associated with ice-sheet change and sea-level. A better knowledge of continental slope processes will also increase our understanding of submarine geohazards, the factors that influencing slope instability and the risks posed by future instability.

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