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Past Ice-Sheet Behaviour: Retreat Scenarios and Changing Controls in the Ross Sea, Antarctica

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Abstract. Studying the history of ice-sheet behaviour in Antarctica's largest drainage basin, the Ross Sea, can improve our understanding of patterns, timing, and controls on marine-based ice-sheet dynamics, and provide constraints on numerical ice-sheet models. Newly collected high-resolution multibeam swath bathymetry data, combined with two decades of legacy multibeam and seismic data, are used to map glacial landforms and reconstruct paleodrainage.

Last Glacial Maximum grounded ice reached the continental shelf edge in the eastern but not western Ross Sea. Recessional geomorphic features in the western Ross Sea indicate virtually continuous retreat of the ice sheet in contact with the bed. In the eastern Ross Sea, well-preserved linear features and a lack of small-scale recessional landforms record rapid lift-off of grounded ice from the bed. Physiography exerted a first-order control on ice behaviour, while seafloor geology played an important subsidiary role.

Previously published grounding-line retreat scenarios are based on terrestrial observations; however, this study uses Ross Sea-wide geomorphology to constrain marine deglaciation. Our analysis of retreat patterns suggests that: (1) a large embayment formed in the eastern Ross Sea; (2) retreat was complex and asynchronous between troughs; and (3) the eastern Ross Sea largely deglaciated prior to the western Ross Sea.

1 Introduction

With a catchment area encompassing ~25% of the entire Antarctic Ice Sheet, the Ross Embayment is the largest ice drainage basin in Antarctica and receives drainage from multiple ice streams sourced from the East Antarctic (EAIS) and West Antarctic (WAIS) ice sheets (Fig. 1). The nature of ice-sheet paleodrainage and retreat in the Ross Sea has significant implications for understanding the dynamics of the WAIS and EAIS, and their respective sensitivities to climate change. Recent paleo-ice flow models indicate complex ice behaviour in the Ross Sea, particularly during deglaciations (e.g. Pollard and DeConto, 2009; Golledge et al., 2014). Geologic reconstructions of ice dynamics from the Ross Sea continental shelf can provide critical tests for these models.

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Results from marine geological research, including integrated seismic stratigraphy, geomorphology, and sediment core analyses, indicate that both the EAIS and WAIS advanced across the continental shelf during the Last Glacial Maximum (LGM; Licht et al., 1999; Shipp et al. 1999; Mosola and Anderson 2006; Anderson et al., 2014). The relative contributions of the EAIS and WAIS to ice flow and subsequent paleodrainage retreat behaviour in the Ross Sea remain controversial. Results from several land-based studies have led to the conclusion that the WAIS dominated ice flow during the LGM (e.g. Denton and Marchant, 2000; Hall et al., 2000, 2015). However, offshore till provenance analyses indicate that the EAIS and WAIS had roughly equal contributions to ice draining into the Ross Sea (Anderson et al., 1984; Licht et al., 2005; Farmer et al., 2006). Significant drainage of EAIS into the western Ross Sea is also supported by interpretations from seafloor glacial geomorphology (Shipp et al., 1999; Mosola and Anderson, 2006; Greenwood et al., 2012; Anderson et al., 2014) and exposure age dating in the Transantarctic Mountains (e.g. Jones et al., 2015).

Multibeam swath bathymetry produces a direct picture of the bed under the former ice sheet, revealing landforms associated with past ice flow. These landforms are powerful indicators of flow behaviour, and can be used to reconstruct paleo-flow dynamics of formerly grounded ice. Here we compile legacy multibeam swath bathymetry data from 41 cruises over the last 20 years (Table 1), combined with recently acquired high-resolution multibeam data, in order to characterize glacial geomorphic features across the Ross Sea. This unique, integrated dataset provides an opportunity to view the seafloor at a much higher resolution than is possible beneath the modern ice shelf and ice sheet. We can improve our understanding of factors that control regional ice-sheet dynamics and test existing ice-sheet retreat models by using this dataset to map the distribution of glacial geomorphic features and thus reconstruct ice-sheet paleodrainage across the Ross Sea during and following the LGM.

2 Study Area

The Ross Sea contains seven bathymetric troughs (Fig. 1), which are remnants of the extensional tectonic history of the region. Ice streams preferentially occupied these troughs and scoured the seafloor over multiple glacial cycles (Cooper et al., 1991; Anderson, 1999). The eastern Ross Sea (ERS) and the western Ross Sea (WRS) have distinctly different characteristics in terms of seafloor geology and physiography. The ERS is dominated by a single, large rift basin, bounded by Ross Bank and Marie Byrd Land, with near-surface stratigraphy dominated by unconsolidated Plio-Pleistocene sediments that thicken in a seaward direction (Alonso et al., 1992; De Santis et al., 1997). The WRS is geologically complex with older and more consolidated strata locally occurring at or near the seafloor (Cooper et al., 1991; Anderson and Bartek, 1992). The WRS contains high-relief banks and deep troughs and thus serves as an analogue to the modern Siple Coast grounding line where banks currently serve as ice rises. The ERS has more subdued physiography consisting of broad troughs separated by low-relief ridges and provides an analogue to the interior portions of the modern grounding line (Fig. 1).

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3 Methodology

This study synthesizes multibeam datasets from across the Ross Sea, combining legacy multibeam data (Table 1) with newly collected, high-resolution multibeam data collected in key areas for characterizing the nature of ice-sheet retreat (Fig. 2a). The combined multibeam tracklines cover over 250,000 km of the Ross Sea. New, high-resolution swath bathymetry data were acquired during an RV/IB *Nathaniel B. Palmer* NBP1502A cruise to the Ross Sea in the 2014-2015 austral summer. These data were collected with a Kongsberg EM-122 system in dual swath mode with a 1°x1° array, 12 kHz frequency, and gridded at 20 m. Vertical resolution varies from about 0.2% - 0.07% of water depth (Jakobsson et al., 2011); therefore, at water depths of 500 m, geomorphic features with sub-meter amplitudes can be resolved. Horizontal resolution is similarly depth-dependent and, in water depths of 500 m, is about 9 m. Ping editing was completed onboard using CARIS and imported into ArcGIS. In addition to multibeam data, newly acquired high-frequency seismic data (3.5 kHz sub-bottom data collected with a Knudsen CHIRP 3260 using a 0.25 ms pulse width) were interpreted along with legacy CHIRP data.

The seafloor geologic setting has been recorded in legacy seismic reflection data across the Ross Sea. We refer to seismic records as either 'high-frequency' denoting 3.5 kHz CHIRP data or 'low-frequency' referring to traditional seismic data (20-600 Hz). Low-frequency seismic lines from cruise PD-90, originally published in Anderson and Bartek (1992), were combined with ANTOSTRAT Project seismic lines compiled by Brancolini et al. (1995). Previously interpreted seismic units identify facies bounded by glacial unconformities, where each surface represents a glacial advance that cannibalized the previous substrate and deposited till and glacimarine sediments above the newly eroded surface.

4 Results

Glacial geomorphic features imprinting the Antarctic continental shelf are divided into three main categories, largely following the classification scheme of Benn and Evans (2010). These are: (1) subglacial features, such as mega-scale glacial lineations (MSGLs), grooves, drumlinoid features, megaflutes, and subglacial channels; (2) ice-marginal features, such as grounding zone wedges (GZWs), marginal moraines, and linear furrows; and (3) proglacial features, including gullies and arcuate iceberg furrows (Fig. 2). These features occur above the most recent shelf-wide glacial unconformity (with the exception of drumlinoids) and are covered by post-LGM sediments. They are, therefore, interpreted as LGM and post-LGM features (e.g. Shipp et al., 2002; Mosola and Anderson, 2006). We describe these landform classes here and present their Ross Sea-wide distribution in Fig. 3.

4.1 Subglacial Features

Subglacial features are formed under grounded ice, where the ice is thick enough to offset buoyant forces exerted by the ocean and remains permanently in contact with the bed. MSGLs (Fig. 2b) are the most common subglacial features on the

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Antarctic continental shelf. While the actual formation process for MSGLs is still debated (e.g. Tulaczyk et al., 2001; Shaw et al., 2008; Ó Cofaigh et al., 2008; Fowler, 2010), they are interpreted as having formed under streaming ice due to their association with modern ice streams (King et al., 2009) and their occurrence within paleo-glacial troughs (Anderson, 1999; Livingstone et al., 2012). The streamlined nature of MSGLs makes them excellent indicators of ice-flow direction (Clark, 1993; Stokes and Clark, 1999; Shipp et al., 1999; Ó Cofaigh et al., 2002; Dowdeswell et al. 2004; Spagnolo et al., 2014).

Previous studies have shown that MSGLs are associated with a massive seismic facies interpreted as the deforming till layer deposited above the latest glacial unconformity (Shipp et al., 1999; Ó Cofaigh et al., 2002, 2005; Heroy and Anderson, 2005). Most MSGLs in the Ross Sea have amplitudes of 1-9 m, lengths of about 1-10 km, and are characterized by extreme parallel conformity. As these features are strongly associated with actively deforming till, MSGL amplitudes should not be greater than the thickness of the deforming till layer. Grooves, in contrast, are erosional features characterized by generally larger and more variable amplitudes (e.g. Heroy and Anderson, 2005).

Drumlins and crag-and-tail forms are elongate, large-scale features (heights of tens of meters and lengths of hundreds of meters to a few kilometres) that taper in the direction of ice flow (e.g. Menzies, 1979; Benn and Evans, 2010). Crag-and-tail landforms are sculpted bedrock features, whereas drumlins are moulded sedimentary landforms (Benn and Evans, 2010). We group both features as drumlinoids given their similar morphology and implications for ice flow, as well as a lack of information on the internal structure of the observed Ross Sea features.

Megaflutes are elongate, small-scale features (heights of meters and lengths generally less than 200 meters) that taper in the direction of ice flow. They typically are associated with drumlinoid features and, because of their small size, are imaged only in high-resolution multibeam and side-scan sonar records. In Antarctica, drumlinoids and megaflutes are most often observed at the transition between crystalline bedrock and sedimentary deposits (Wellner et al., 2001, 2006; Graham et al., 2009). In the Ross Sea, drumlinoids and megaflutes are restricted to a single, prominent field in Glomar Challenger Basin and along the western flank of Drygalski Trough where crystalline bedrock occurs at or near the sea floor. Because these features are moulded from predominantly crystalline and sedimentary bedrock, they likely formed over multiple glacial cycles. They do, however, exhibit highly uniform orientations (Fig. 2c) that are consistent with MSGL orientations seaward of the drumlinoids, indicating that the most recent phase of ice flow was likely responsible for the final drumlinoid shape.

Channels on the Ross Sea continental shelf have been previously observed by Alonso et al. (1992), Wellner et al. (2006), and Greenwood et al. (2012), though their origin and link to subglacial meltwater has earlier been equivocal. Our newly acquired high-resolution multibeam data in the Ross Sea reveal subglacial channels that are clearly associated with ice-marginal features such as grounding zone wedges and recessional moraines (Simkins et al., in review; Fig. 2d). The channels are incised into till deposited above the LGM unconformity, and are sometimes overprinted by ice-marginal landforms; therefore, these are subglacial meltwater channels that were active during the most recent glacial recession. The implication

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of these channels with regard to ice-sheet dynamics and paleodrainage is still under investigation and not discussed in this paper.

4.2 Ice-Marginal Features

Ice-marginal features form at the ice margin, where ice transitions from being permanently grounded to the bed and decouples from the seafloor. They include GZWs, marginal moraines, and linear furrows.

GZWs are depositional features (Fig. 2f, 2g) formed during periods of stability of the grounding line. They grow as sediment is delivered to the grounding line through subglacial bed deformation and basal debris melt-out (Alley et al., 1986, 1989; Anderson, 1999; Anandakrishnan et al., 2007; Alley et al., 2007; Dowdeswell et al., 2008, Dowdeswell and Fugelli, 2012; Batchelor and Dowdeswell, 2015). GZWs are characterized by steep foreset slopes that result in asymmetrical morphologies, broadly indicating ice-flow direction during GZW deposition. Sedimentation at grounding lines, forming GZWs, can stabilize an ice sheet against small-scale sea-level rise and ice-sheet thinning (Alley et al., 2007). Large GZWs mark longer episodes of stability of the ice margin (Alley et al., 2007, Dowdeswell and Fugelli, 2012). The internal structure of large GZWs is occasionally detectable in low-frequency seismic data and includes distinct foreset beds indicative of wedge growth (e.g. Anderson, 1999; Heroy and Anderson, 2005), but more often internal stratification is not resolved in seismic data (e.g. Mosola and Anderson, 2006; Batchelor and Dowdeswell, 2015). Here, GZWs are grouped into three categories: small-scale, intermediate-scale, and large-scale. Small-scale GZWs have heights less than 10 m, cannot be traced across an entire trough width, and generally are only observable in high-resolution multibeam and side-scan sonar data (e.g. Shipp et al., 1999; Jakobsson et al., 2012; Simkins et al., in review). Intermediate-scale GZWs range from 10-50 m heights and often display very sinuous fronts. Large-scale GZWs (Fig. 2g) have heights exceeding 50 m and extend across the entire trough width.

Marginal moraines (Fig. 2e) are largely symmetric in cross section (Dowdeswell and Fugelli, 2012). They are generally believed to be formed by push-processes (Batchelor and Dowdeswell, 2015). In the Ross Sea, moraines are mostly observed as fields of small repeating ridges with heights of 1-4 m. Features with similar characteristics are sometimes interpreted as De Geer moraines, whose development is influenced by seasonal or cyclic processes (Hoppe, 1959; Lindén and Möller, 2005; Todd et al., 2007), or transverse ridges (Dowdeswell et al., 2008) which does not imply seasonal formation. Due to their limited amplitudes, these features are only resolvable with high-resolution bathymetric mapping techniques including the EM122 multibeam system and side-scan sonar (e.g. Shipp et al., 1999; Jakobsson et al., 2011; Simkins et al., in review).

Iceberg furrows form when deep-keeled icebergs contact the bed, causing scours on the seafloor (Lien et al., 1989). Arcuate iceberg furrows are classified as proglacial features (discussed in section 4.3), but we consider linear furrows to be ice-marginal features (Fig. 2h). The margins of marine-based ice sheets with low-slope profiles are particularly susceptible to tidal fluctuations, causing large areas of the ice sheet to intermittently contact the seafloor (Fricker and Padman, 2006; Brunt et al., 2010). We interpret linear furrows to form within this diffuse grounding zone, where ice is hovering at the buoyancy limit and cyclically contacting the seafloor. Small repeating corrugation ridges typically occur within fields of furrows or within individual furrows (Figs. 2h, 2k; Anderson, 1999; Jakobsson et al., 2011). Although the exact mechanism for their formation remains somewhat controversial, corrugation ridges are thought to form as icebergs move vertically with tides, causing iceberg keels to intermittently contact the bed (Jakobsson et al., 2011; Graham et al., 2013). Their tidal association is based on the occurrence of identical features in proglacial arcuate iceberg furrows (Anderson, 1999) and comparison of corrugation amplitude and spacing with tidal modelling (Jakobsson et al., 2011).

Here, we associate linear furrows with ice-shelf breakup events when icebergs near the grounding line are held upright within an iceberg armada (MacAyeal et al., 2003; Jakobsson et al., 2011). A deep keel capable of ploughing a linear furrow once ice has fractured could also have existed as an irregularity at the ice base prior to calving. Thus, linear furrows could also form by ploughing of ice-base irregularities in a diffusive grounding zone, as ice is intermittently in contact with the bed while still connected to fully grounded ice upstream. Both mechanisms for linear furrow formation signify that ice was still moving as a coherent body in contact with the seafloor. Linear furrows often display a geomorphic expression similar to MSGLs; however, they are erosional features whereas MSGLs are interpreted as either depositional or deformational features. Furrows should have more variable flow directions than MSGLs and exhibit more variable orientations in a seaward direction. However, linear furrows tend to exhibit high parallel conformity and are consistent with MSGL orientations. We argue that linear forward motion is propelled by the upstream ice flow; therefore, their significance for ice-flow direction is the same as MSGLs. For this reason, linear furrows and MSGLs are grouped into the inclusive term of 'linear features.'

4.3 Proglacial Features

Shelf-edge gullies occur where streaming ice reached the continental shelf break (Fig. 2i). Their origin remains uncertain; they have been attributed to point sources of sediment-dense meltwater from the grounding line when it was situated at the shelf break (Anderson, 1999; Evans et al., 2005), and also from small-scale slope failure due to accumulation of proglacial sediment (Gales et al., 2012). Both mechanisms imply proximity to the grounding line. Ross Sea shelf-edge gullies have not been extensively surveyed; however, a lack of significant sediment infilling suggests that they were active during the LGM (Shipp et al. 1999).

Arcuate iceberg furrows (Fig. 2j) are common features in the Ross Sea. They are generally clustered near the shelf margin and on bank tops, overprinting subglacial and ice-marginal features. These are clearly proglacial features formed by freely moving icebergs that drifted under the influence of ocean currents and winds. Corrugation ridges have been observed within arcuate iceberg furrows, which is the most compelling evidence that corrugation ridges result from tidal motion (Anderson, 1999).

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4.4 Western Ross Sea

Drygalski Trough is the deepest region of the Ross Sea with water depths over 1000 m. Within this trough, the most seaward geomorphic expression of the ice sheet grounding line is a large-scale GZW north of Coulman Island (D1, Fig. 3). This is consistent with previous interpretations of the maximum grounding-line location (Licht et al., 1999; Shipp et al., 1999). A prominent set of MSGLs extend continuously from the Drygalski Ice Tongue to the approximate latitude of Coulman Island. A few small GZWs occur along the flanks of the trough; otherwise the MSGLs are not overprinted by recessional features. North of Coulman Island, both linear and arcuate iceberg furrows overprint grounding zone wedge D1. The outermost shelf is covered by extensive arcuate iceberg furrows, which could have overprinted any older features.

Multibeam data is scarce in southern Drygalski Trough, a key area for reconstructing the final phase of deglaciation in the WRS. Available data show a field of closely spaced, small-scale GZWs that back-step up the southern margin of Crary Bank, a set of discrete intermediate-scale GZWs, and lineations offshore of Mackay Glacier that record westward grounding line retreat (Greenwood et al., 2012; Anderson et al., 2014).

JOIDES Trough is slightly foredeepened on the outer shelf, relatively flat on the middle shelf, and slopes steeply into the deep inner shelf Central Basin (Fig. 1, 3). The outer portion of JOIDES Trough is mostly devoid of linear features, with the exception of one group of linear furrows. High-frequency CHIRP data show a 4-8 m layer of acoustically laminated and draped glacimarine sediments on the outer shelf (Fig. 3a). LGM-age carbonates occur on outer shelf banks on both sides of JOIDES Trough (Taviani et al., 1993; Fig. 3), precluding the presence of grounded ice at those locations. A large-scale, mid-shelf GZW (J1, Fig. 3) is seismically resolved (Shipp et al., 1999), although the GZW crest lacks clear expression in multibeam data. This mid-shelf GZW is separated from the next intermediate-scale GZW near the southern end of Crary Bank (J2, Fig. 3) by a continuous field of marginal moraines and small-scale GZWs. Southern JOIDES Trough is characterized by a series of meltwater channels emanating from grounding lines, observed at GZW J2 and ice-marginal features south of J2.

Pennell Bank and Ross Bank are linked across Pennell Trough by a bathymetric saddle (referred to here as Pennell Saddle), separating the outer shelf Pennell Trough from the deep Central Basin (Fig. 3). A large-scale GZW (P1) occurs at the northern margin of the Pennell Saddle and a thick (up to 14 m) package of layered glacimarine sediments extends northward from beneath the toe of the P1 wedge (Figs. 3b, c). Small-scale sinuous wedges and relatively straight-crested moraines record the grounding line back-stepping from atop Pennell Saddle southward into Central Basin. These recessional features overprint a large subglacial meltwater channel within the saddle (Fig. 2d).

The Central Basin is a bathymetric low that reaches water depths of over 1000 m, situated south of all three WRS troughs. It contains multiple generations of poorly preserved linear features, suggesting phases of large-scale ice stream flow

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reorganization through the basin and McMurdo Sound (Greenwood et al 2012). Numerous pockets of small, marginal moraines are found throughout the Central Basin and do not seem to be oriented parallel to depth contours.

4.5 Eastern Ross Sea

The ERS contains three major troughs: Glomar Challenger Basin, Whales Deep, and Little America Basin, separated by low-relief ridges that are thought to have separated three paleo-ice streams (Mosola and Anderson, 2006). Linear features monopolize the ERS seafloor and extend to the continental shelf break (Fig. 3). They are associated with GZWs that are large enough to be identified in low-frequency seismic reflection data (Mosola and Anderson, 2006), (Fig. 2g; Fig. 3). Small- and intermediate-scale GZWs and moraines are confined to a few locations and no subglacial channels have been observed in the ERS. Shelf-edge gullies occur at the continental shelf break, implying the presence of basal meltwater outbursts from the grounding line.

Extensive linear features occur throughout Glomar Challenger Basin (Fig. 2b, 3). They exhibit both trough-parallel and trough-sub-parallel orientations. The only large field of drumlinoids observed in the Ross Sea occur on the inner shelf of Glomar Challenger Basin (Fig. 2c) and are associated with a near-surface occurrence of crystalline bedrock (Anderson, 1999). Legacy high-frequency CHIRP data in outer Glomar Challenger Basin show thin glacimarine sediments (Fig. 3d) and sediment cores sampled tills that typically occur within 1 to 2 meters of the seafloor. Two closely spaced large-scale GZWs exist at the continental shelf break, observed in low-frequency seismic lines (G1 and G2, Fig. 3). However, the morphologies of these GZWs are so subdued that they are not observable in multibeam bathymetric records. Two large-scale composite GZWs on the mid-to-inner shelf (G3, G4) are observed in both low-frequency seismic and multibeam data (Bart and Cone, 2011).

Whales Deep also contains a large-scale GZW at the continental shelf break (W1, Fig. 3), as well as a mid-shelf composite GZW observable in both low-frequency seismic and multibeam records (W2, Fig. 3). About 170 km landward, the front of another large-scale GZW extends only a short distance seaward of the modern Ross Ice Shelf calving line (W3, Fig. 3). A well-developed field of linear features extends from beneath the mid-shelf (W2) GZW to the continental shelf break. Linear features are notably absent south of W2. Little America Basin, like Glomar Challenger Basin, exhibits extensive linear features that extend across the entire trough to the shelf break. It is characterized by three GZWs (L1-3, Fig. 3). These large-scale GZWs are identified from low-frequency seismic data (Mosola and Anderson, 2006), but are not observed in the legacy multibeam swath bathymetry.

4.6 Flowsets

In the ERS, there are clearly multiple generations of linear features. Discrete flow episodes, corresponding to the formation of distinct sets of linear features, can thus be defined from the population of linear features in the ERS. Linear features were

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grouped based on their parallel concordance, close proximity, and similar morphometry (after Clark, 1999). Rose diagrams were constructed from each group of linear features (Fig. 4) to ensure that each flowset contains features of the same orientation. Linear features within each flowset are tightly clustered around a small range in orientation, generally within a standard deviation of less than 10°. Each flowset is assumed to represent a single flow configuration, formed isochronously (after Clark 1999). Assuming that all flowsets were shaped during the LGM and subsequent deglaciation, a relative chronology of flowsets was assessed based on the seaward-most extent of the individual flowsets, as well as any cross-cutting relationships with other flowsets. In order to reduce complexity and best represent large-scale regional flow patterns, flowsets with discrete, yet similar clusters of orientations, were assumed to reflect a similar ice-flow configuration.

Major flow patterns observed in the ERS often deviate from the trough-parallel drainage that has previously been assumed (Licht et al., 2005; Mosola and Anderson, 2006; Anderson et al., 2014). Flowsets in Glomar Challenger Basin partially exhibit evidence of trough-parallel flow (flowsets *a-c*, Fig. 4), and also display an extensive group of linear features extending across an inter-ice-stream ridge towards Whales Deep (flowsets *d-h*, Fig. 4). Flowset *g* contains the only curved flowlines observed. Rose diagrams of linear features within flowset *g* were used to ensure that the curvature was not comprised of two discrete flow events with very similar orientations, but instead populated by linear features that change orientation gradually and consistently. In Whales Deep Trough, only one flowset is observed, consisting of trough-parallel flow on the outermost shelf (flowset *i*, Fig. 4). Flow indicators in Little America Basin partially mirror the configuration in Glomar Challenger Basin. Some linear features in Little America Basin display generally trough-parallel flow (flowsets *j*, *k*, Fig. 4), while others are oriented oblique to the trough axis and directed towards Whales Deep (flowsets *l*, *m*, Fig. 4). A third group of linear features indicates flow out of Little America Basin into a neighboring outlet draining Marie Byrd Land to the east (flowset *n*, Fig. 4). Flowsets on the innermost shelf in all three ERS troughs are interpreted as late-stage deglacial flow configurations.

5 Discussion

5.1 Last Glacial Maximum ice extent and flow

In outer Drygalski Trough, we interpret the LGM grounding line to lie just north of Coulman Island at the seaward-most GZW, in accordance with Shipp et al. (1999) (D1, Fig. 3). Between Coulman Island and Drygalski Ice Tongue, a prominent field of MSGLs indicates trough-parallel flow, consistent with results from previous till provenance studies (Anderson et al., 1984). Although data coverage is sparse, no linear flow features are observed south of the ice tongue; however, small ice-marginal features indicate ice was grounded in the deepest parts of Drygalski Trough south of the ice tongue.

In JOIDES Trough, maximum ice extent is placed at the large-scale GZW (J1) on the mid-outer-shelf (Fig. 3), primarily due to the presence of up to 8 m of draped glacimarine sediments in the outer trough (Fig. 3a), and the observation of LGM-age carbonates on surrounding banks (Taviani et al., 1993). This is consistent with Licht et al. (1999), who concluded that ice did not reach the continental shelf break during the LGM based on the presence of LGM-age tephra layers in glacimarine sediments on the outer shelf. Linear features that occur seaward of this LGM limit are interpreted as iceberg furrows.

The LGM limit in Pennell Trough is placed at the large-scale GZW (P1, Fig. 3), located ~120 km landward of the shelf break in accordance with Howat and Domack (2003). High-frequency seismic data show that this wedge prograded across thick glacimarine sediments that fill the outer trough (Fig. 3b, c).

Large-scale GZWs at the shelf break in each ERS trough (Fig. 3), along with linear features that extend across the outer continental shelf within each trough, indicate that grounded ice reached the shelf break (Shipp et al, 1999; Mosola and Anderson 2006). Extensive shelf-edge gullies support the presence of grounded ice at the shelf break. Thin glacimarine sediments occur on the outer shelf and indicate a relatively short period of ice-free conditions.

Figure 5 shows the interpreted LGM grounding line and paleo-flow directions derived from linear features that occur near this LGM grounding line. Generally, linear features delineate trough-parallel flow, which is consistent with previous LGM flow reconstructions (Shipp et al., 1999; Mosola and Anderson, 2006; Anderson et al., 2014).

5.2 Western Ross Sea deglaciation

The WRS contains sparse and isolated patches of linear features, providing only glimpses of subglacial flow behaviour and direction. Therefore, most paleo-drainage interpretations in the WRS are based on ice-marginal features.

In Drygalski Trough, the ice sheet decoupled from the seafloor and back-stepped rapidly from its LGM position near Coulman Island to a mid-shelf position at Drygalski Ice Tongue. This interpretation is based on a lack of recessional features overprinting MSGL. South of Drygalski Ice Tongue, data are sparse and data quality is poor. The most prominent deglacial features are a series of intermediate-size GZWs that back-step westward towards Mackay Glacier from a location north of Ross Island (Greenwood et al., 2012).

In JOIDES and Pennell troughs, fields of closely spaced, small-scale, ice-marginal features (Figs. 2d-f) dominate the seascape, indicating that grounded ice remained in contact with the seafloor during retreat. This also implies that the overall retreat was punctuated by pauses that were long enough to form a small recessional feature before retreating and forming another recessional feature. Retreat slowed and the grounding line stabilized in the southernmost part of JOIDES Trough at an intermediate-scale GZW (J2, Fig. 3). A subglacial meltwater channel extending from GZW J2 to the south was likely linked to a large meltwater system that was active during deglaciation. Meltwater channels are observed in southern JOIDES and Pennell troughs and are associated with retreat of the grounding line from positions of stability (J2 and the Pennell

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Saddle), leading to final rapid deglaciation of grounded ice in the two troughs. The effect of channelized subglacial meltwater on grounding-line stability is still under investigation.

Ice in the deep Central Basin appears to have retreated quickly, leaving only scattered recessional moraines. Based on the orientations of these moraines, a grounding-line embayment formed over the Central Basin and the grounding line then receded to the east and the west (Fig. 6). Subglacial and ice-marginal features in the Central Basin indicate that ice remained grounded during deglaciation. Because ice did not lift off from the deep seafloor, we infer that retreat patterns were dictated by a steep ice profile rather than physiography, as the ice did not decouple concentrically according to depth contours.

North of Central Basin, extensive fields of small-scale GZWs and moraines back-step onto banks (Figs. 7; 8a-b), indicating that ice remained grounded on WRS banks during deglaciation. The presence of wedges and moraines implies that ice was actively flowing across the banks and mobilizing sediment in order to deposit these marginal features. Thus, these banks acted as independent ice rises during the late stages of ice sheet retreat from the WRS (Shipp et al., 1999; Anderson et al., 2014; Matsuoka et al., 2015). Modelling results indicate the presence of independent, detached ice rises on WRS banks late in deglaciation (Golledge et al., 2014). Additionally, Yokoyama et al. (in press) argue that a grounded ice shelf remained pinned on WRS banks until the late Holocene.

Reconstructed grounding-line steps (Fig. 7) illustrate retreat patterns across the Ross Sea. Observed grounding lines are linked together to form discrete time-steps of deglaciation, representing an interpretation of relative timing. These linkages are based on similar morphologies of observed GZWs, extension of grounding line orientations along bathymetric depth contours, and interpretation of local retreat rates from geomorphic features. Southern Drygalski Trough was the last area in the WRS to experience grounding-line retreat, as outlet glaciers (e.g. Mackay Glacier, and David Glacier flowing into the Drygalski Ice Tongue) receded toward the west and north, leaving fields of back-stepping moraines and wedges in their path (Greenwood et al., 2012; Anderson et al., 2014), (Fig. 7). Drainage from the EAIS played an active role in nourishing the ice sheet until the last stage of deglaciation (Fig. 7, steps 7-8). We infer a steep EAIS ice profile over the WRS throughout deglaciation, based on the contribution of EAIS ice through the Transantarctic Mountains (Fig. 6, Fig. 8) and grounding-line recession independent of physiography in the central WRS.

5.3 Eastern Ross Sea deglaciation

Linear features on the ERS seafloor are not overprinted by recessional bedforms, other than scattered large-scale GZWs (Fig. 3). This indicates episodes of ice sheet decoupling and grounding line retreat that back-stepped tens of kilometres in distance, punctuated by periods of long-term grounding line stability when large GZWs were formed.

We propose two alternative behaviour scenarios to explain the observed changes in flow orientation in the ERS. The first scenario (‘dynamic flow-switching model’) is characterized by alternating regional flow direction throughout the LGM,

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followed by north-south recession of the grounding line (Fig. 9a). In the second scenario ('embayment scenario'), the ice stream occupying Whales Deep experienced extensive retreat, forming a large grounding-line embayment in the ERS (Fig. 9b).

The dynamic flow-switching scenario entails significant flow reorganization with westward ice flow out of Marie Byrd Land (d1, Fig. 9a) followed by eastward flow across the inter-ice-stream ridge between Whales Deep and Glomar Challenger Basin (d2, Fig. 9a). Trough-parallel flow was then established (d3, Fig. 9a). In this scenario, ice then began to retreat north-to-south from the continental shelf in all ERS basins, depositing the large GZWs in Whales Deep and Glomar Challenger Basin (d4, Fig. 9a; G3, W2, then G4, W3, Fig. 3). MGSLs are preserved as the ice-margin retreats, but we would not expect them to be preserved if a re-advance or major new episode of streaming occurs. This scenario relies on preservation of at least three distinct flow orientations throughout three very different stages of flow across the outer shelf. Although there have been examples of preserved flow fabrics during events of flow-switching (Stokes et al., 2009; Winsborrow et al., 2012) or at localized patches of basal friction (Stokes et al., 2007; Kleman and Glasser, 2007), the preservation of such extensive flow fabrics throughout three different configurations is unlikely.

The embayment scenario proposes the formation of an embayment over Whales Deep, based on large flowsets in surrounding basins that curve across inter-ice-stream ridges into Whales Deep (flowsets g, k, Fig. 4). Trough-parallel flow likely occurred first (e1, Fig. 9b), as evidenced by the relatively undisturbed trough-parallel flowset in the outermost part of Whales Deep trough. An embayment in the grounding line then formed, drawing flow from outer Glomar Challenger Basin. Grounded ice stepped south in Whales Deep to a mid-shelf location long enough to deposit a large GZW (e2, Fig. 9b; W2, Fig. 3) and then stepped south again to deposit the inner-shelf GZW (e3, Fig. 9b; W3, Fig. 3). An absence of linear features between GZWs W2 and W3 implies that the ice sheet was not in contact with the bed as it retreated. This further implies a relatively thin, low profile ice sheet within Whales Deep. As a result, the Whales Deep embayment drew ice from Glomar Challenger Basin and Little America Basin, prompting flow across the inter-ice-stream ridges into Whales Deep. The ice stream feeding Whales Deep at the LGM may have experienced stagnation or outrun its inner-shelf ice source, destabilizing grounded ice on the outer shelf and causing an embayment to form.

Shipp et al. (1999) identify the inter-ice-stream ridges in the ERS as aggradational features, meaning that they were centres of focused sedimentation. Embayment grounding lines would have stabilized on the edges of the inter-ice-stream ridges on either side of Whales Deep, transporting sediment to these bathymetric features and aggrading the inter-ice-stream ridges. A large embayment over the ERS is also compatible with the interpreted WRS deglaciation pattern, where a steep EAIS profile is inferred. The formation of an embayment in the ERS is consistent with grounding-line recession in the ERS prior to the WRS followed by east-to-west deglaciation of the WRS (Fig. 7).

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The two retreat models described here each imply a sequence of events that can, in principle, be tested. Greater coverage of high-resolution multibeam data in outer Glomar Challenger Basin, illuminating cross-cutting relationships between flowsets, is crucial for establishing a relative chronology of cross-trough versus trough-parallel flow. Additional survey of inter-ice-stream ridges would also provide a better understanding of the relative chronology of flowset formation. Furthermore, reliable marine radiocarbon dates on the Whales Deep inner shelf might provide evidence of early retreat and the formation of a long-lived embayment in the grounding line. Based on the available data in this study, the embayment scenario is favoured, due to the preservation issues inherent to the dynamic flow-switching model. Thus, the embayment scenario is incorporated into our Ross Sea deglacial reconstruction (Fig. 7).

5.4 Comparison with existing deglacial models

Currently, there are two very different published retreat scenarios for the Ross Sea (Fig. 10). One of these models, the highly cited 'swinging gate' model, calls for a linear grounding line retreat across the Ross Sea, hinged just north of Roosevelt Island and extending to the Transantarctic Mountains (Conway et al., 1999). This model is constrained by one age from Roosevelt Island and ages of ice-free coastlines along the Transantarctic Mountains and indicates deglaciation of the WRS at a faster rate than the ERS. This model implies that controls on ice-sheet dynamics were the same throughout the Ross Sea and that physiography had little influence on ice retreat. The swinging gate model also implies that ice-sheet retreat from the Ross Sea was controlled mainly by changes in the WAIS catchment, suggesting very high rates of north-to-south retreat along the coast of the Transantarctic Mountains (Conway et al., 1999; Hall et al., 2013, 2015). The 'saloon door' model instead proposes early retreat in the ERS with a grounding-line embayment in the central Ross Sea (Ackert, 2008). The implied drainage pattern of the saloon door model requires significant inputs from both the EAIS and the WAIS. This model is supported by cosmogenic exposure ages indicating a thinner ice-sheet profile in the central Ross Sea than at the margins of the WAIS (Parizek and Alley, 2004; Waddington et al., 2005; Anderson et al., 2014).

This study and previous studies calling for an initial large back-step of the grounding line in northern Drygalski Trough are consistent with the swinging gate model, which assumes early and rapid recession of grounded ice along the Transantarctic Mountains. The observations of reconstructed grounding-line retreat in the rest of the Ross Sea (Fig. 7) contrasts with the swinging gate model. In particular, our marine-based reconstruction calls for the persistence of the EAIS in the WRS throughout deglaciation, as well as significant regional variations in grounding-line behavior. Our reconstruction supports the presence of an embayment in the ERS, similar to the saloon door model. Here, grounding-line recession in Glomar Challenger Basin is interpreted to precede retreat in the WRS (Fig. 8b), destabilizing grounded ice in southern Pennell Trough. Deglaciation of the ERS prior to the WRS supports the observation of the EAIS as a persistent feature in the WRS throughout deglaciation. Additionally, marine radiocarbon dates from previous studies suggest that the ERS deglaciated before the WRS (Licht and Andrews, 2002; Mosola and Anderson, 2006), despite complications dating Antarctic glacial-marine sediments.

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Neither the swinging gate nor the saloon door model have thus far incorporated marine data, and, as we show here, do not fully capture the complexity of grounding-line retreat across the Ross Sea. Our new marine-based model (Fig. 7), reconstructed from comprehensive mapping of seafloor geomorphic features directly recording grounding line retreat, can now be used to interpret more detailed Ross Sea paleo-ice sheet behavior and identify regional differences in deglacial behavior.

5.5 Physiographic and geological controls on deglaciation

Many cycles of glacial erosion and deposition have led to Antarctic continental shelves characterized by a fore-deepened shelf profile with exposed bedrock on the inner shelf and thicker sediments on the outer shelf (Anderson, 1999). Runaway grounding-line retreat can occur as ice retreats from the outer to inner shelf due to a lack of pinning points to stabilize the grounding line (e.g. Mercer, 1978; Jamieson et al., 2012). In the WRS, banks and volcanic seamounts provided stable pinning points during deglaciation (Anderson et al., 2014; Simkins et al., in review). Ice-marginal features are observed to back-step up onto WRS banks (Fig. 8), demonstrating a strong physiographic control on grounding line behaviour. These banks served as pinning points for retreating ice streams. WRS banks supported an extensive ice shelf that buttressed WRS grounding lines and contributed to the long-lived presence of the EAIS in the WRS (Anderson et al., 2014; Yokoyama et al., in press). While slight bottlenecking of ERS inter-ice-stream ridges may have played a role in determining positions of grounding-line stability and the formation of large-scale GZWs (Mosola and Anderson, 2006), the ERS seascape is much more topographically subdued than in the WRS. A lack of high-relief banks and troughs permitted more variable flow in the ERS, but did not allow for pinning and stabilization of ice streams as occurred in the WRS.

In addition to physiography, seafloor substrate has also been argued to exert a fundamental control on ice behaviour, as indicated by variations in geomorphic features across different substrates. In Antarctica, studies have shown that ice streams flowing across soft, deformable sedimentary beds are characterized by MSGLs (e.g. Wellner et al., 2001, 2006; Ó Cofaigh et al., 2002, 2005; Graham et al., 2009). Ice flowing over unconsolidated beds can mobilize subglacial sediments and develop a thick layer of pervasive deformation till, facilitating faster ice flow than is possible by internal ice deformation (Alley et al., 1989). By contrast, crystalline bedrock or older and more consolidated strata outcropping on the seafloor would be more resistant to glacial erosion, preventing the development of deforming till underneath a flowing ice stream, and are associated with bedrock erosional features such as drumlinoids that indicate slower ice-flow velocities and stick-slip motion. An excellent example is the field of drumlinoids in inner Glomar Challenger Basin that corresponds to a localized area of outcropping bedrock (Figs. 2c, Fig. 11). At the point where sedimentary deposits lap onto bedrock, these features transition seaward into MSGLs (Anderson, 1999).

The compilation of geological data in Fig. 11 shows the strata beneath the most recent glacial erosional surface, representing the substrate that ice flowing across the continental shelf at the LGM would have encountered. The ages and degree of

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lithification of these strata are derived from information obtained from drill cores collected during Deep Sea Drilling Project Leg 28 extrapolated to high-resolution seismic stratigraphic correlations across the Ross Sea (Anderson and Bartek, 1992; Alonso et al., 1992; Anderson, 1999; Bart et al., 2000). The WRS is characterized by more variable geology and by older substrate below the LGM unconformity, while mostly unconsolidated Plio-Pleistocene sediments blanket the ERS shelf. Thick and extensive unconsolidated sediments likely contributed to a pervasive layer of deformation till in the ERS. This condition facilitated fast flowing ice and a classic till conveyor-belt mechanism transporting sediment to large-scale GZWs, which then contributed to a low profile ice sheet that episodically decoupled from the seafloor during retreat from the continental shelf (Mosola and Anderson, 2006). Complex, more consolidated strata outcropping in the WRS may have limited such pervasive subglacial deformation, potentially causing slower ice stream velocities in WRS troughs. This characteristic seafloor geology, coupled with numerous pinning points, was conducive to a higher profile ice sheet that remained in contact with the seafloor throughout much of its retreat from the continental shelf.

Part of flowset h (Fig. 4) in Little America Basin is routed eastward away from the Ross Sea (Fig. 11). Ice in Little America Basin flowed over the eastern-most ridge and converged with an outlet glacier draining Marie Byrd Land. This implies that Little America Basin at its maximum configuration was not able to drain all of the ice flowing into it from its ice stream sources. The substrate that Little America Basin ice streams flowed across during the LGM was composed of late Oligocene and Miocene sediments (Fig. 11). Thus, it was more resistant to ductile subglacial deformation than the substrates encountered by other ice streams flowing across the ERS. Resulting flow velocities were therefore not high enough to transport all of the ice entering the Little America Basin outlet, some of which was captured and funnelled into the neighbouring outlet.

Numerous processes affect glacial dynamics, such as ice-shelf buttressing, tidal amplification, sediment shear strength and ice-bed coupling, and subglacial meltwater. Ongoing work on characterizing Ross Sea glacial geomorphology highlights the effect of these processes on local grounding-line stability. Physiography exerts a first-order control on regional ice stream flow and retreat dynamics. Seafloor geology plays an important subsidiary role in controlling ice behaviour.

6 Conclusion

During the LGM, grounded ice reached the continental shelf break in the ERS, but not in the WRS. The WRS seafloor is characterized by recessional geomorphic features that signify episodic, rapid recession following the LGM, and indicate a persistent presence of a steep-profiled EAIS in the WRS throughout deglaciation. Retreat in the ERS was likely initiated by the formation of a large grounding-line embayment over Whales Deep trough. Based on our interpretation of glacial geomorphic indicators, Glomar Challenger Basin in the ERS was completely deglaciated prior to retreat of grounded ice from the deep Central Basin in the WRS. Retreat was asynchronous between Ross Sea troughs.

Considering the complex glacial geomorphic assemblages across the entire Ross Sea, the swinging gate and saloon door models both fail to fully capture the style of deglaciation in the Ross Sea. The saloon door model is more consistent with glacial geomorphic observations on the Ross Sea continental shelf, describing a mode of deglaciation that may have occurred in more than one sector as the Ross Sea retreated into its component sub-catchments. Based on this study, we conclude that it is eminently clear that deglaciation across the Ross Sea did not involve a linear grounding line across the multiple troughs and banks.

Major differences between regional retreat characteristics can be attributed to physiography. Ice was pinned on the high-relief banks in the WRS, whereas the lack of comparable features in the ERS indicates that the WAIS wasn't stabilized by pinning points. Seafloor geology played a secondary role in influencing paleodrainage patterns. Younger and relatively unconsolidated Plio-Pleistocene sediments in the ERS, with the exception of Little America Trough, are associated with fast ice flow, whereas the older and more consolidated strata that characterized the WRS seafloor may have hindered pervasive till deformation and contributed to slower ice-stream velocities. The controls on flow behaviour and deglaciation patterns revealed in our new Ross Sea deglacial reconstruction can now be incorporated into future work on understanding marine ice-sheet behaviour.

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Table 1. Multibeam dataset compilation.

Cruise Number	**Vessel**	**Multibeam System**	**Date**	**PI**
NBP9801	*Nathaniel B. Palmer*	*SeaBeam Instruments 2112*	1/16/98	Anderson
			2/18/98	
NBP9802	*Nathaniel B. Palmer*	*SeaBeam Instruments*	2/22/98	Honjo

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		2112	4/2/98	
NBP9803	*Nathaniel B. Palmer*	*SeaBeam Instruments 2112*	5/1/98 6/17/98	Jeffries
NBP9807	*Nathaniel B. Palmer*	*SeaBeam Instruments 2112*	11/1/98 12/12/98	Dunbar
NBP9901	*Nathaniel B. Palmer*	*SeaBeam Instruments 2112*	12/26/98 2/4/99	Jeffries
NBP9902	*Nathaniel B. Palmer*	*SeaBeam Instruments 2112*	2/12/99 3/22/99	Anderson
NBP9909	*Nathaniel B. Palmer*	*SeaBeam Instruments 2112*	12/20/99 2/9/00	Bengtson
NBP0001	*Nathaniel B. Palmer*	*SeaBeam Instruments 2112*	2/14/00 3/30/00	Jacobs
NBP0209	*Nathaniel B. Palmer*	*Kongsberg EM120*	12/11/02 12/30/02	Cande
NBP0301	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/5/03 1/29/03	Bartek
NBP0301A	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/1/03 2/18/03	Bart
NBP0301B	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/20/03 2/22/03	Smith
NBP0302	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/24/03 4/4/03	Gordon
NBP0305A	*Nathaniel B. Palmer*	*Kongsberg EM120*	12/20/03 12/30/03	Smith
NBP0306	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/4/04 1/15/04	Luyendyk
NBP0401	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/19/04 2/17/04	Wilson
NBP0402	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/21/04 4/6/04	Visbeck
NBP0408	*Nathaniel B. Palmer*	*Kongsberg EM120*	10/12/04 12/6/04	Jacobs
NBP0409	*Nathaniel B. Palmer*	*Kongsberg EM120*	12/18/04 1/21/05	Kiene

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NBP0501	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/28/05 2/13/05	Gordon
NBP0508	*Nathaniel B. Palmer*	*Kongsberg EM120*	10/26/05 12/3/05	Neale
NBP0601	*Nathaniel B. Palmer*	*Kongsberg EM120*	12/17/05 1/24/06	DiTullio
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NBP0602	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/30/06 2/21/06	Stock
NBP0608	*Nathaniel B. Palmer*	*Kongsberg EM120*	11/3/06 12/11/06	DiTullio
NBP0701	*Nathaniel B. Palmer*	*Kongsberg EM120*	12/22/06 1/28/07	Cande
NBP0702	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/2/07 3/23/07	Jacobs
OSO0708	*Oden*	*Kongsberg EM122*	11/29/07 1/07/2008	Jakobsson
NBP0801	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/9/08 1/26/08	Caron
NBP0802	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/30/08 2/20/08	Caron
NBP0803	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/22/08 3/13/08	Bart
NBP1005A	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/13/10 1/16/11	Yager
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NBP1005	*Nathaniel B. Palmer*	*Kongsberg EM120*	11/26/10 1/16/11	Yager
OSO1011	*Oden*	*Kongsberg EM122*	12/08/10 1/16/11	Jakobsson
NBP1101	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/19/11 2/15/11	Kohut
NBP1102	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/19/11	Swift

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			4/23/11	
NBP1201	*Nathaniel B. Palmer*	*Kongsberg EM120*	12/24/11 2/11/12	McGillicuddy
NBP1202	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/11/12 2/27/12	Owen
NBP1210	*Nathaniel B. Palmer*	*Kongsberg EM120*	1/6/13 2/9/13	Halanych
NBP1302	*Nathaniel B. Palmer*	*Kongsberg EM120*	2/12/13 4/5/13	Hansell
NBP1310B	*Nathaniel B. Palmer*	*Kongsberg EM120*	12/3/13 1/23/14	Arrigo
NBP1502	*Nathaniel B. Palmer*	*Kongsberg EM122*	1/23/15 3/20/15	Bart, Anderson

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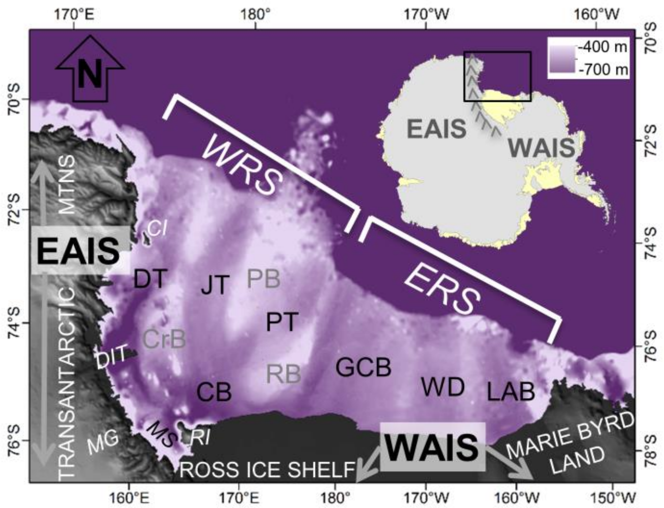

Figure 1. Regional bathymetry of the Ross Sea continental shelf, with bed topography data acquired from BEDMAP2 (Fretwell, 2013). Inset shows the West and East Antarctic ice sheets (WAIS and EAIS, respectively), separated by the Transantarctic Mountains (cross-hatched area) with the Ross Sea study area outlined. WRS (Western Ross Sea), ERS (Eastern Ross Sea), EAIS (East Antarctic Ice Sheet), WAIS (West Antarctic Ice Sheet), DT (Drygalski Trough), JT (JOIDES Trough), PT (Pennell Trough), CB (Central Basin), CrB (Crary Bank), PB (Pennell Bank), RB (Ross Bank), GCB (Glomar Challenger Basin), WD (Whales Deep), LAB (Little America Basin), CI (Coulman Island), DIT (Drygalski Ice Tongue), MG (Mackay Glacier), MS (McMurdo Sound).

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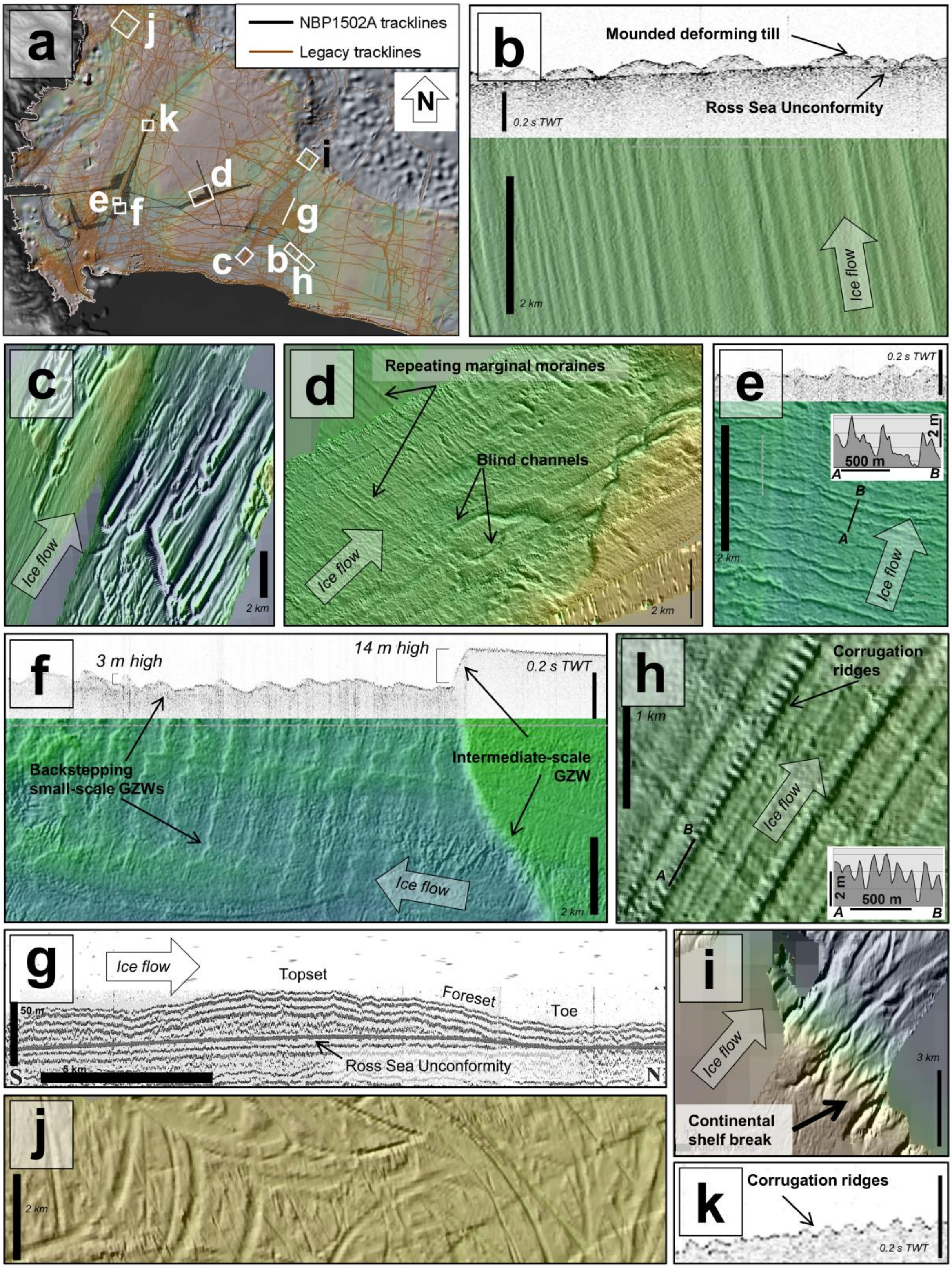

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Figure 2. Glacial geomorphic features of the continental shelf. (a) Ross Sea tracklines from cruise NBP1502A (black lines) and legacy (brown) cruises and locations of (b-k). (b) MSGLs (3-5 m in amplitude) on the inner shelf of Glomar-Challenger Basin are composed of relatively soft deformation till above a glacial erosional surface, imaged by the high-frequency seismic. (c) Drumlinoids on the inner shelf of Glomar Challenger Basin. (d) A subglacial meltwater channel in Pennell Trough with complex channel morphology and associated with small-scale recessional ice-marginal features. (e) Marginal moraines in JOIDES Trough. (f) Small-scale GZWs in JOIDES Trough (~3 m high). (g) GZW (4b) in Glomar-Challenger Basin modified from Mosola & Anderson (2006). (h) Linear iceberg furrows with average depth of 14 m; corrugation ridges inside the furrows have heights of 0.5-2 m. (i) Shelf-edge gullies on the eastern Ross Sea continental shelf break. (j) Arcuate cross-cutting iceberg furrows on the outer shelf of Drygalski Trough. (k) Corrugation ridges in outer JOIDES trough. These features have heights ranging from 0.5-2 m. Dashed lines on the multibeam images indicates the location of the CHIRP profiles.

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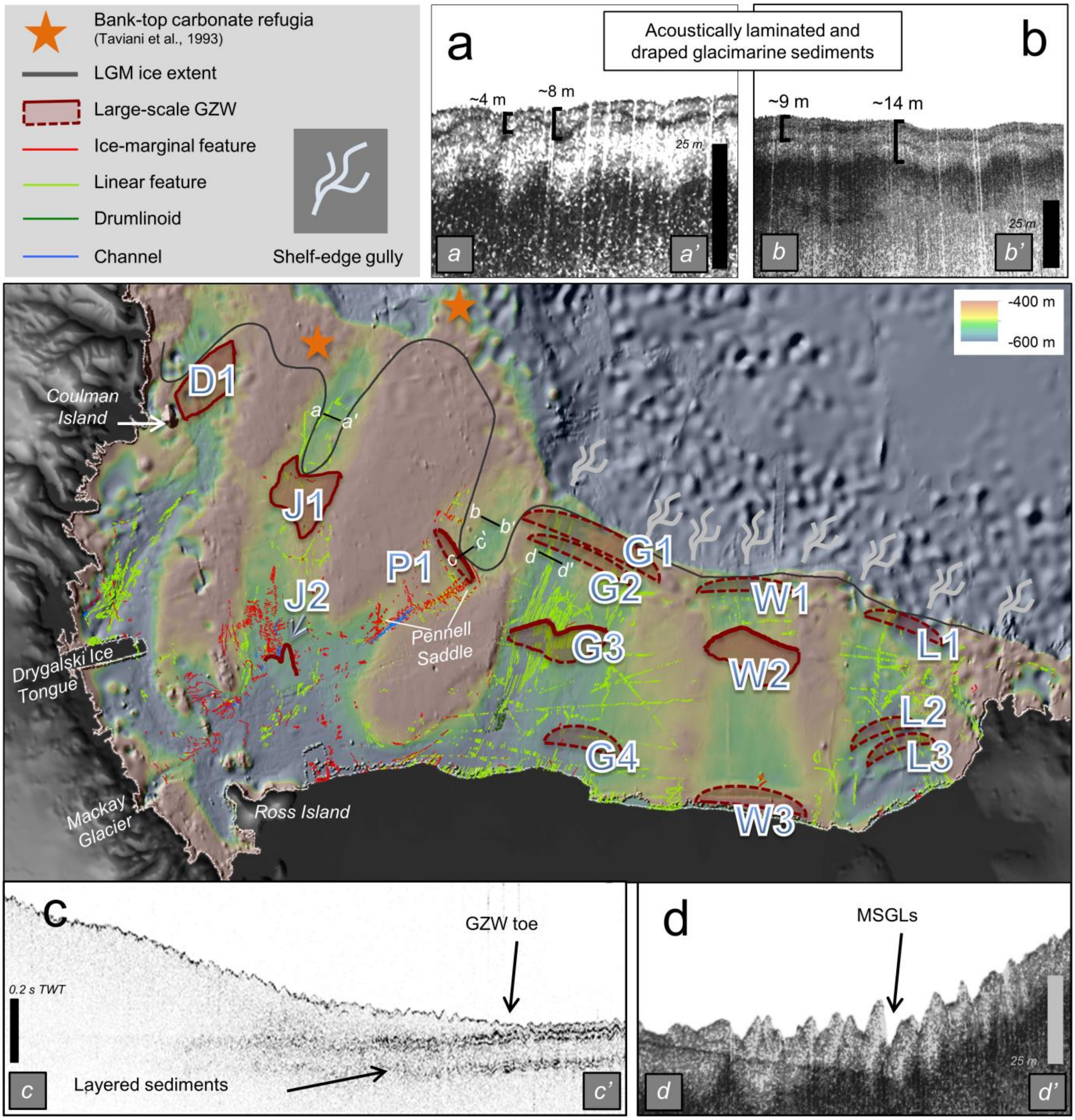

Figure 3. Distribution of geomorphic features and evidence for LGM extent. Large-scale GZWs are outlined with a solid line where the GZW boundary is known, and a dotted line where the boundary is inferred based on depth contours. GZWs that are only identified in seismic lines are symbolized with a dotted lens shape. High-frequency seismic profiles showing thick, draped glacimarine sediments in (a) JOIDES and (b) Pennell troughs. (c) The LGM GZW foreset in Pennell Trough

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prograded over thick pre-LGM glacial marine sediments. (d) MSGLs with no appreciable post-glacial sediments in outer Glomar Challenger Basin.

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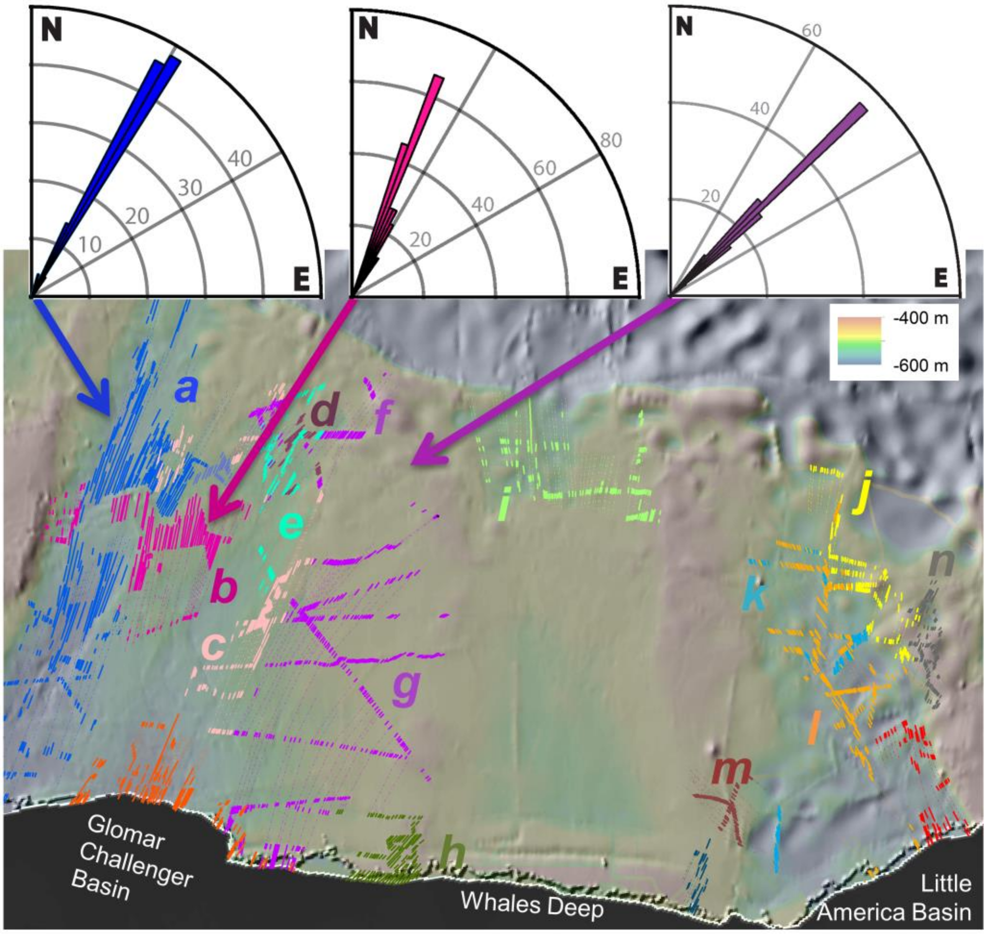

Figure 4. Linear features in the eastern Ross Sea were grouped into flowsets based on features mapped using multibeam data (solid lines) and interpolation between multibeam lines (dotted lines). Major flowsets are labeled for reference in text. Flowsets were placed in a relative chronology partially based on maximum seaward extent; each orientation represents a different vintage of flow. Three example flowsets and corresponding Rose diagrams are shown in outer Glomar Challenger Basin.

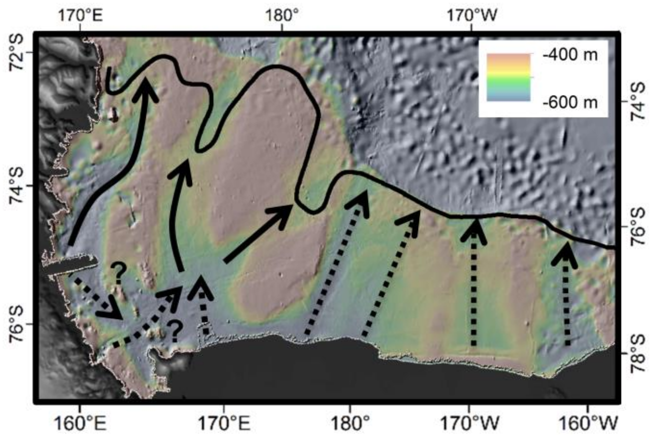

Figure 5. LGM flow based on geomorphic flow indicators. Dotted lines in south-western Ross Sea denote flow patterns based on a geomorphic record of local ice flow out of EAIS outlet glaciers during deglaciation. It remains uncertain whether those flow patterns were also active during the LGM. In the eastern Ross Sea, lineations corresponding to LGM flow (dotted arrows) are also unclear. We assume that LGM ice streams flowed roughly parallel to trough axes, based on the most seaward flowsets.

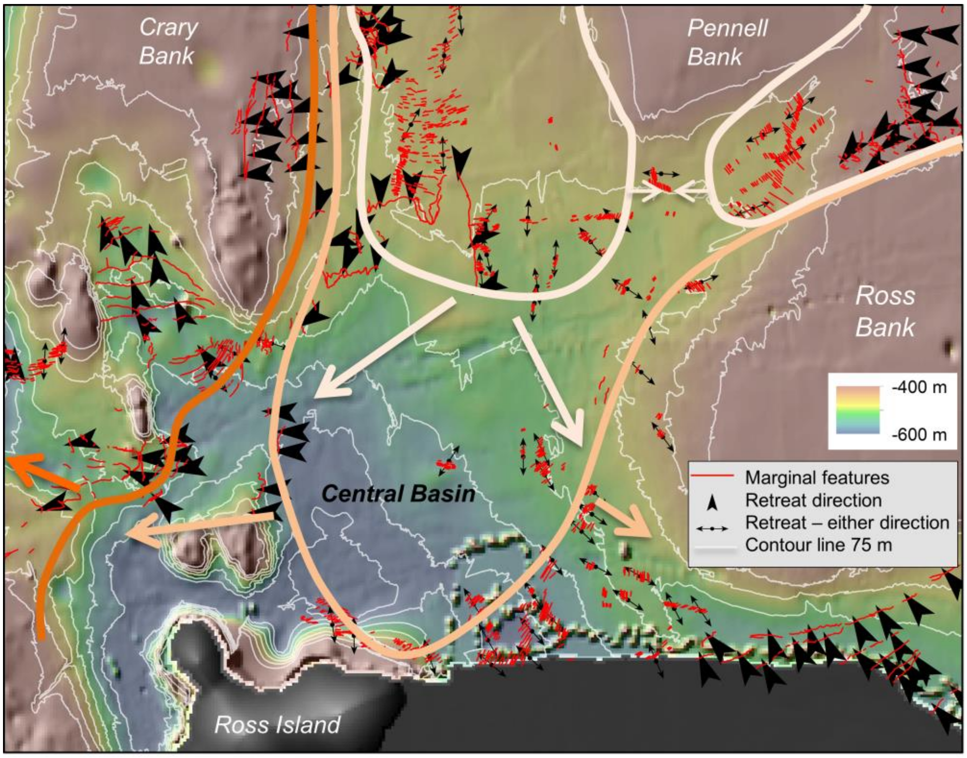

Figure 6. Retreat patterns in the western Ross Sea are noted by retreat direction (inferred from GZWs, arrowheads) and symmetric marginal moraines (double-sided arrows). Reconstructed grounding lines (solid lines) are accompanied by large arrows indicating regional retreat. Thin white lines are depth contours at 75-m increments. Deglaciation in the deep Central Basin did not follow depth contours, implying a steep deglacial EAIS ice profile in order to remain grounded.

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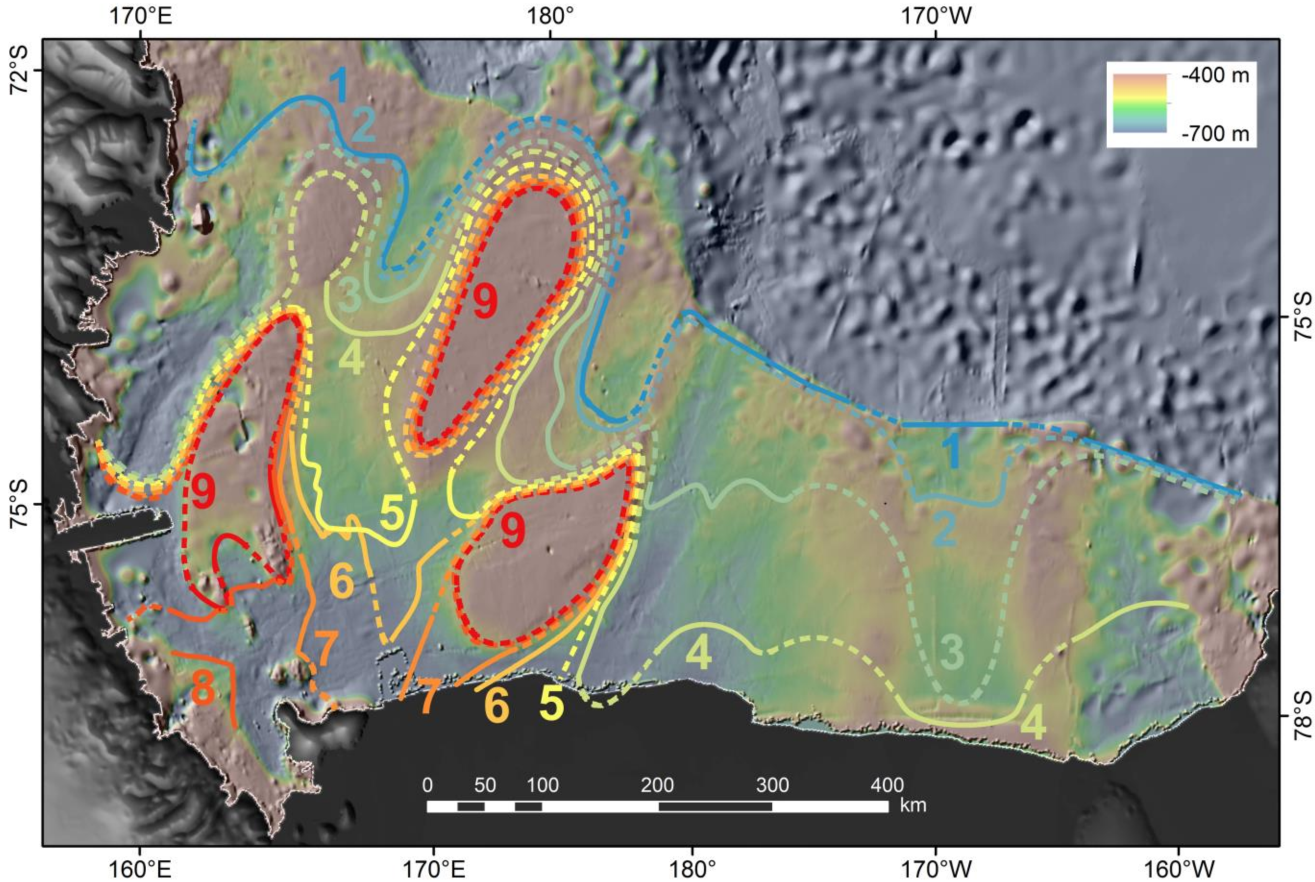

Figure 7. Reconstructed grounding-line retreat across the Ross Sea based on geomorphic indicators of grounding lines (solid lines) and inferred grounding-line locations (dashed). Each line marks a relative step in grounding-line retreat starting with step 1 at the LGM grounding line and ending with step 9 with ice pinned on banks.

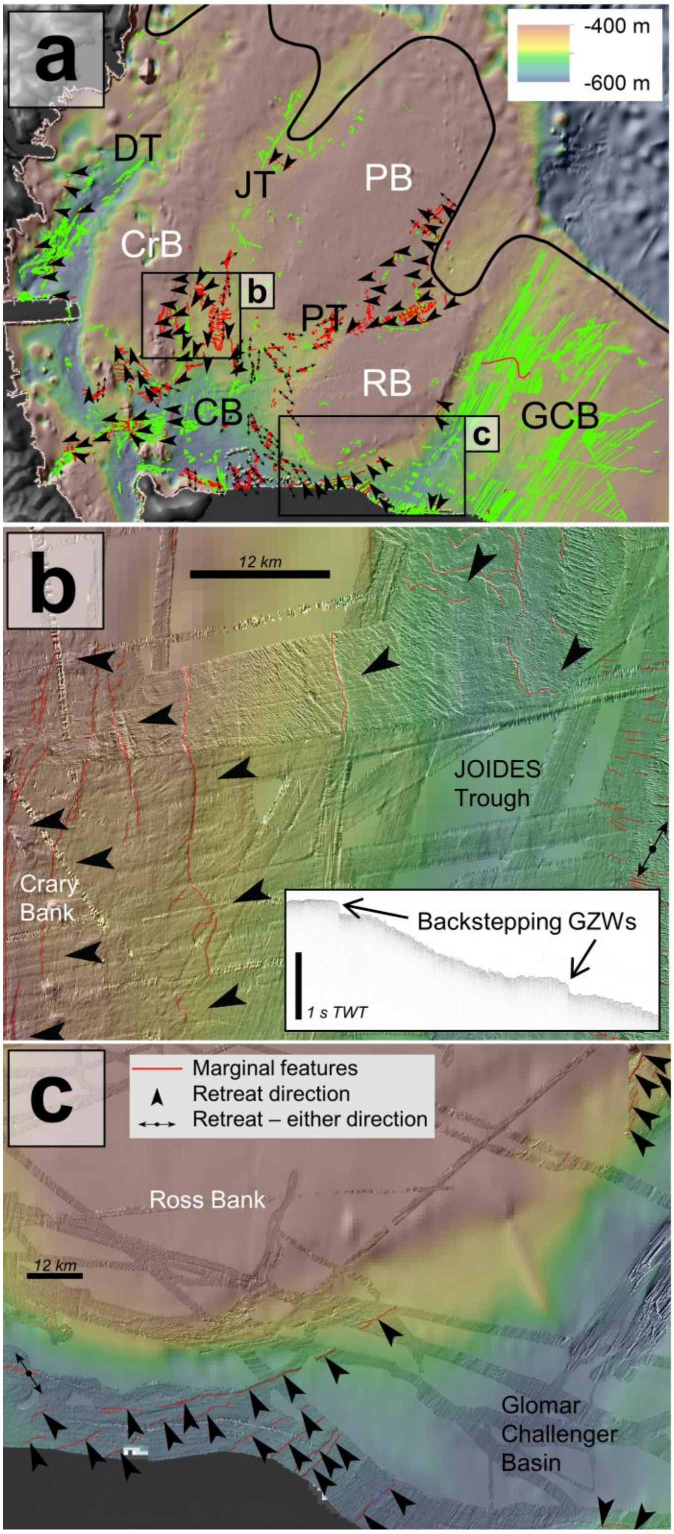

Figure 8. (a-b) Grounding lines are observed to retreat up onto banks, as shown by back-stepping wedges and marginal moraines. Arrowheads denote retreat direction. (c) Back-stepping grounding lines in southwestern Glomar Challenger Basin imply that ice had decoupled there before retreating westward into the WRS.

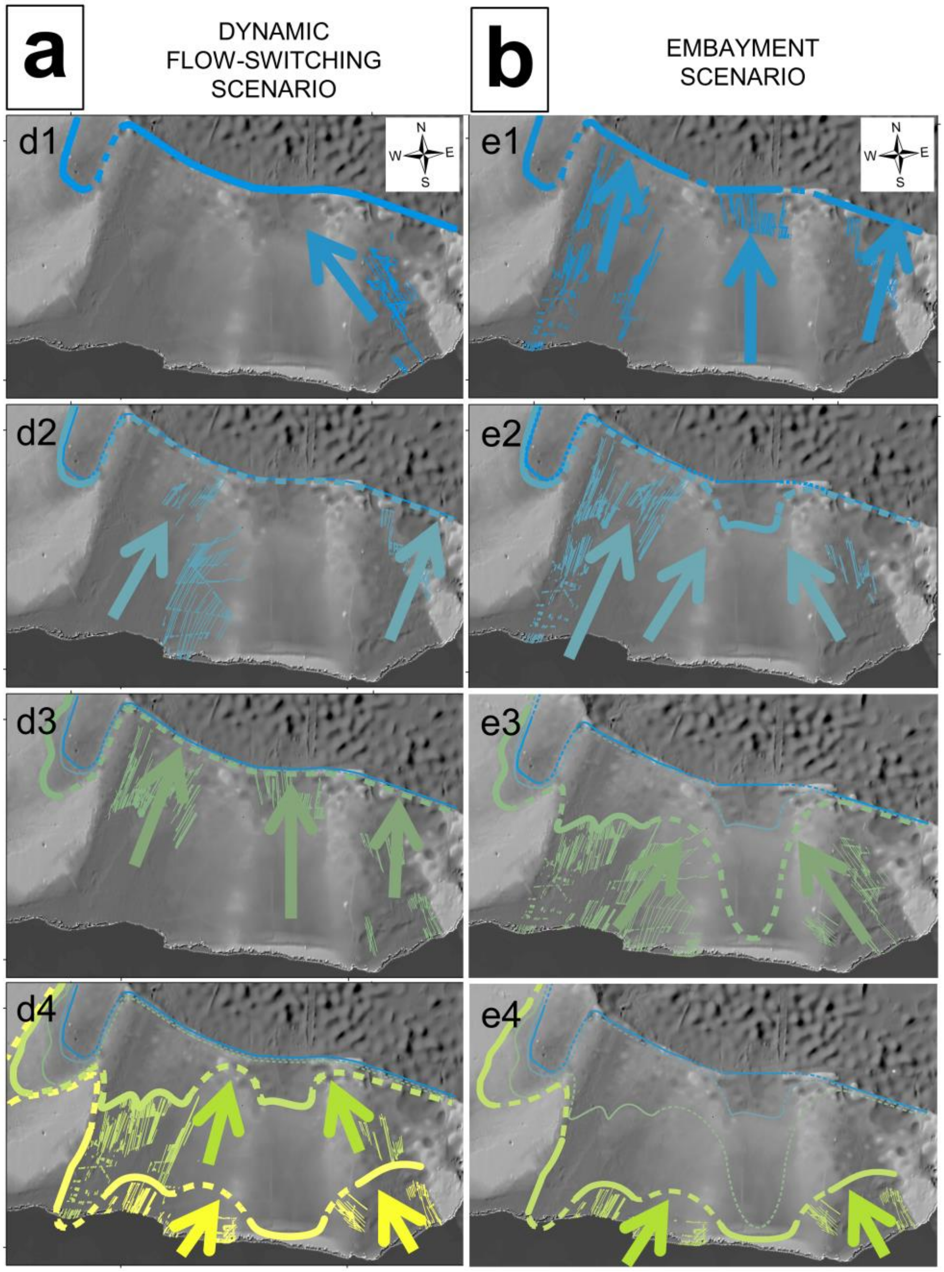

Figure 9. Possible retreat scenarios for the eastern Ross Sea interpreted from flowsets. (a) The dynamic flow-switching scenario calls for alternating regional flow direction throughout the LGM, followed by north-south recession of the grounding line. This model requires preservation of at least three different flow fabrics as ice remains grounded on the outer continental shelf. (b) In the embayment scenario, a large grounding-line embayment in the eastern Ross Sea forms over Whales Deep. The embayment scenario is independently more consistent with inland paleo-ice thickness reconstructions and seafloor seismic observations.

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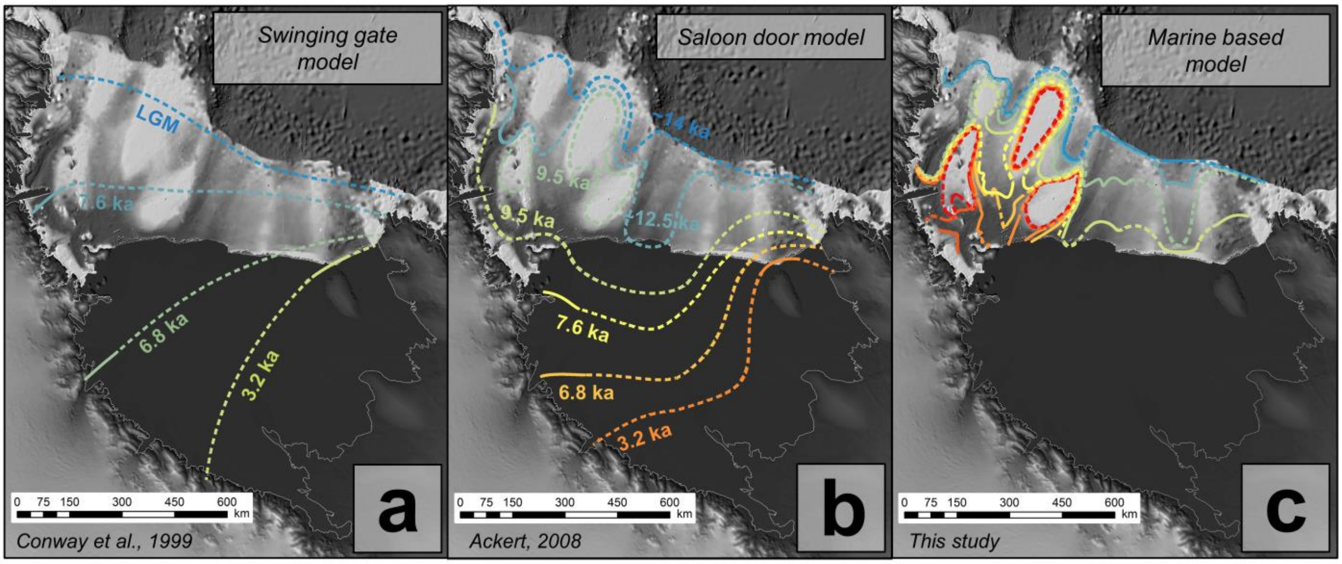

Figure 10. Comparison of existing models of Ross Sea deglaciation. (a) The 'swinging gate model' (Conway et al., 1999) assumes a linear grounding line swinging across the Ross Sea, implying that controls on ice-sheet dynamics are the same throughout the Ross Sea and that physiography has little influence on ice retreat. This model indicates deglaciation of the WRS prior to the ERS, and implies that the Ross Sea was filled with WAIS ice during LGM and throughout deglaciation. (b) The 'saloon door' model of deglaciation suggests early retreat in the ERS with a potential grounding-line embayment in the central Ross Sea (Ackert, 2008), requiring significant inputs from both the EAIS and the WAIS. (c) The marine-based reconstruction presented here uses glacial geomorphology to interpret paleo-grounding-line retreat.

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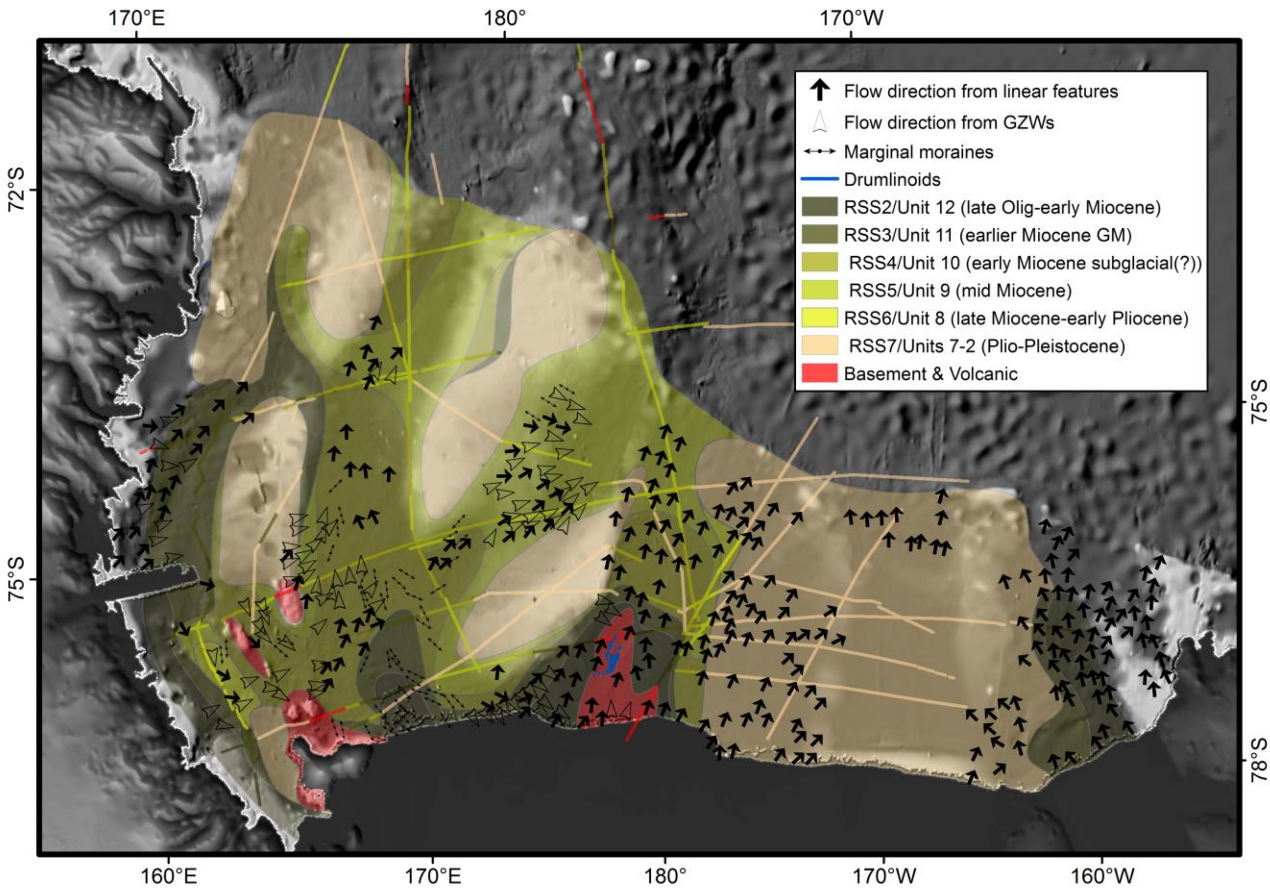

Figure 11. Control of seafloor geology on ice dynamics. Geologic boundaries were interpolated from legacy seismic lines (shown here) with pre-interpreted seismic units by Anderson and Bartek (1992) and Brancolini et al. (1995). The WRS is characterized by complex, older and more consolidated strata, where ice streams have eroded down to Oligocene-age strata. Volcanic islands and seamounts outcrop in the southern portion of the WRS. The western side of Glomar Challenger Basin, bordering Ross Bank, contains older and more variable geologic strata outcropping at the seafloor, including a patch of basement outcrop on the inner shelf. In general, thick unconsolidated Plio-Pleistocene strata fill most of the ERS and increase in thickness in an offshore direction (Alonso et al., 1992). Plio-Pleistocene sediments are thin in southern Whales Deep, overlying older Miocene strata. Farther east in Little America Basin, lithified late Oligocene through Miocene deposits occur beneath the LGM unconformity. Arrows indicating flow direction are based on geomorphic features.