Flow and retreat of the Late Quaternary Pine Island-Thwaites palaeo-ice stream, West Antarctica

Alastair G.C. Graham[1,*], Robert D. Larter[1], Karsten Gohl[2], Julian A. Dowdeswell[3], Claus-Dieter Hillenbrand[1], James A. Smith[1], Jeffrey Evans[4], Gerhard Kuhn[2], Tara Deen[1]

[1]*British Antarctic Survey, Cambridge. UK.*

[2]*Alfred Wegener Institute for Polar and Marine Research, Bremerhavn, Germany.*

[3]*Scott Polar Research Institute, Cambridge. UK.*

[4]*Dept. of Geography, Loughborough University, Loughborough. UK.*

*Correspondence to: alah@bas.ac.uk; 01223221604.

Revised manuscript #2, 29th March 2010.

Abstract

Multibeam swath bathymetry and sub-bottom profiler data are used to establish constraints on the flow and retreat history of a major palaeo-ice stream that carried the combined discharge from the parts of the West Antarctic Ice Sheet now occupied by the Pine Island and Thwaites glacier basins. Sets of highly-elongated bedforms show that, at the last glacial maximum, the route of the Pine Island-Thwaites palaeo-ice stream arced north-northeast following a prominent cross-shelf trough. In this area the grounding line advanced to within ~68 km of, and probably reached, the shelf edge. Minimum ice-thickness is estimated at 715 m on the outer shelf, and we estimate a minimum ice discharge of ~108 km^3 yr^{-1} assuming velocities similar to today's Pine Island glacier (~2.5 km yr^{-1}). Additional bedforms observed in a trough northwest of Pine Island Bay likely formed via diachronous ice-flows across the outer shelf, and demonstrate switching ice-stream behaviour. The 'style' of ice-retreat is also evident in five grounding zone wedges, which suggest episodic deglaciation characterised by halts in grounding-line migration up-trough. Stillstands occurred in association with changes in ice-bed gradient, and phases of inferred rapid retreat correlate to higher bed-slopes, supporting theoretical studies that show bed geometry as a control on ice-margin recession. However, estimates that individual wedges could have formed within several centuries still imply a relatively rapid overall retreat. Our findings show that the ice stream channelled a substantial fraction of West Antarctica's discharge in the past, just as the Pine Island and Thwaites glaciers do today.

3045 Seafloor morphology, geology, and geophysics
3002 Continental shelf and slope processes
1621 Cryospheric change
0730 Ice streams

1. Introduction and background

Pine Island and Thwaites glaciers account for most of the ice discharge from the Amundsen Sea drainage sector of the West Antarctic Ice Sheet (WAIS) (Fig. 1). They presently exhibit the most rapidly decreasing surface elevation within the WAIS [Shepherd et al., 2001; Pritchard et al., 2009], and increasing discharge over the past 35 years through flow acceleration (Pine Island) and widening (Thwaites) indicates that this is dynamic thinning [Rignot, 2006, 2008; Scott et al., 2009]. Grounding line retreat associated with this thinning has also been demonstrated for Pine Island Glacier [Rignot, 1998]. The dynamic thinning has been widely attributed to intrusion of relatively warm Circumpolar Deep Water onto the continental shelf, melting the bases of glacier tongues and ice shelves and reducing their "buttressing" effect [e.g., Jacobs et al., 1996; Rignot and Jacobs, 2002; Payne et al., 2004; Shepherd et al., 2004; Walker et al., 2007; Thoma et al., 2008]. Deep bathymetric cross-shelf troughs, which are thought to have been incised by glaciers through the Miocene-Quaternary (Fig. 1) may be important for this process, channelling warm water towards the ice fronts. Enhanced geothermal heat flux beneath the Amundsen Sea glacier catchments might be another driver of flow changes [Rignot et al., 2002], enhancing melting at the glacier beds and thus increasing lubrication. Alternatively the recent imbalance may simply reflect a short-term deviation from the ice sheet's long-term retreat trajectory. In order to test this hypothesis it is necessary to establish the maximum extent of ice during the last glaciation, and the timing and pattern of its subsequent retreat. An important question is whether or not there may have been previous periods of rapid thinning and retreat interrupted by intervals when the grounding line and flow stabilized.

This history of flow stability is particularly pertinent to the Amundsen Sea sector today because the ice-bed currently lies below sea level beneath its main glacier trunks, and slopes inland monotonically from the grounding line in the Thwaites catchment [Holt et al., 2006; Vaughan et al., 2006]. According to some theoretical studies this geometric configuration is one of the key criterion for a marine ice-sheet collapse [Vaughan and Arthern, 2007; Schoof, 2007] and has contributed to the Amundsen Sea sector often being referred to as 'the weak underbelly' of the WAIS [Hughes, 1981], with suggestions that the region may be prone to future rapid deglaciation [Vaughan, 2008].

However, confidence in models that purport to simulate future changes in the ice sheet can only be established by verifying those models against past behaviour; in particular flow history and changes since the last glacial maximum (LGM) at ~20 cal. kyrs before present (BP) – here referring to the last period of regional maximum ice-sheet extent around Antarctica [Anderson et al., 2002] when grounded ice had advanced onto the adjacent continental shelf – and during the deglaciation between ~20-10 cal. kyrs BP [Anderson et al., 2002; Lowe and Anderson, 2002]. The most readily available and detailed information on the former footprint of these ice streams, and thus their flow history, comes from the morphological record preserved at the sea floor today [Shipp et al., 1999; Canals et al., 2000; Evans et al., 2005; Ó Cofaigh et al., 2005a, 2005b; Dowdeswell et al., 2008]. Previous studies have used the presence of geomorphic ice-flow indicators such as preserved streamlined bedforms, to suggest that an ice stream flowed out of Pine Island Bay towards the continental shelf edge in the past. Geophysical data and sediment cores from the large cross-shelf trough that extends offshore from Pine Island and Thwaites glaciers were interpreted by Lowe and Anderson [2002, 2003] as showing that grounded ice had extended at least as far as the middle shelf (as a minimum possible extent), and probably to the shelf edge at the LGM (as a maximum scenario). On the basis of radiocarbon dates on calcareous microfossils extracted from cores, Lowe and Anderson [2002] concluded that ice subsequently retreated from the middle shelf prior to ~16 ^{14}C ka BP (uncorrected) and retreated to the inner shelf by 10 ^{14}C ka BP (uncorrected). Evans et al. [2006] presented multibeam data showing subglacial bedforms that extend to the shelf edge in an outer shelf trough at 114°W, and argued that their sea-floor position and the lack of overlying sediment drape indicate that the WAIS was grounded at the shelf edge during the LGM. However, coverage of multibeam swath bathymetry data remains sparse on large parts of the outer Amundsen Sea shelf, especially north of 72˚ 30'S, where the path of the main Pine Island cross-shelf trough is less clear (Fig. 1) [Nitsche et al., 2007]. In addition, only two radiocarbon dates constrain the 'timing' of post-LGM deglaciation and even less information exists about the 'style' of ice retreat from the eastern Amundsen Sea shelf [Lowe and Anderson, 2002], the latter of which constitutes a key parameter for validating ice-sheet numerical simulations.

In this paper we describe streamlined subglacial bedforms that record the flow and retreat style of a major palaeo-ice stream – the Pine Island-Thwaites palaeo-ice stream (PITIS) – across the Amundsen Sea continental shelf. This ice stream received input from the ancestral Pine Island, Thwaites and Smith glacier systems (Fig. 1). The bedforms are located in a trough which extends to the shelf edge, at 106˚W (Fig. 1). Using new geophysical datasets we test the following hypotheses: (1) the LGM WAIS extended to the outer shelf in the eastern Amundsen Sea; (2) the major drainage pathway and main outlet of the PITIS can be traced from the present grounding line through the trough in the eastern Amundsen Sea Embayment (ASE); and (3) sea-floor geomorphic evidence can be used to constrain the style and course of grounding-line retreat to its present-day configuration.

2. Methods

We used marine geophysical data acquired using a Kongsberg EM120 (191 beams at 11.25–12.75 kHz) and an Atlas Hydrosweep DS-2 (59 beams at 15.5 kHz) multibeam swath bathymetry system, together with TOPAS parametric sub-bottom profiles ('burst' pulse at secondary frequency of 2.8 kHz and 'chirp' pulse with secondary frequencies of 1.5–5 kHz), to map the topography and distribution of relict subglacial bedforms on the continental shelf of the Amundsen Sea. Geophysical data were collected during cruises JR84, JR141 and JR179 of the RRS *James Clark Ross* (*JCR*; 2003, 2006 and 2008), and during cruise ANT-XXIII/4 of the RV *Polarstern* (2006). These data were combined with existing swath bathymetric data from the Lamont-Doherty Earth Observatory, Marine Geosciences Data System (http://www.marine-geo.org/; Fig. 1). Navigation data were acquired using GPS receivers. Ping-edited swath data were gridded at a 30-m cell size.

3. Subglacial bedforms and features

A recent regional bathymetric data compilation for the Amundsen Sea Embayment [Nitsche et al., 2007] shows a continuous cross-shelf trough that arcs north-westwards across the shelf, and which has two possible outlets, as depicted by bathymetric depressions: one to the NW [Evans et al., 2006], another to the NE (Fig. 1). Hereafter, we refer to these outlets as Pine Island Trough West (PITW) and Pine Island Trough East (PITE).

On cruise JR179, we followed the axis of the largely unsurveyed PITE, from its mouth at the shelf edge to the main cross-shelf trough, and then continued onwards to the deepest part of the trough on the inner shelf, near the March 2008 fast-ice edge (to within ~100 km of the modern Pine Island ice shelf front). Water depths generally increase inshore along this 400 km-long track, from ~500 m at the shelf edge, to >1600 m on the inner shelf (Fig. 1).

Figure 2a shows grey-scale shaded-relief swath bathymetry from the outer shelf in the PITE axis. A suite of crudely-aligned to parallel lineations are observed on the sea floor. Bedforms have a general NNE alignment (25-35˚; north azimuth), and occasionally cross cut. They are up to 500 m wide, 16 km long, and have amplitudes of 2-6 m (Fig. 2a inset). The maximum elongation (length: width) ratio for lineations is ~64:1, and most have ratios greater than 10:1. The main set of imaged lineations are formed on the landward flank of a bathymetric high and terminate at a prominent seaward-facing (ice leeward) ramp (Fig. 2a, 3a, 4a). The ramp is up to 20 m high, and imaged over a distance of 1700 m, producing a ‘wedge’ transverse to the trough long-axis at the seabed (GZW1) (Fig. 2a, 3a, 4a). Curvilinear furrows characterise the shallower portions of the surrounding bathymetry, and in places cut across more parallel lineations (Fig. 2a). A narrow corridor of multibeam data extending to the northeast of GZW1 shows that lineations are also present further north. A second ramp, with similar dimensions to that forming the seaward limit of GZW1, lies ~20 km to the south of the first wedge (GZW2) (Fig. 2a, 3a, 4a), and there are subtle changes in lineation orientations across each of the wedge fronts (on the order of several degrees).

Following the trough south, additional sets of highly-parallel bedforms comprising undulating grooves and ridges are apparent at the sea floor, lying within the deepest parts of the trough (Fig. 2b, 5b). Bedforms are 5-14 km in length, 250-490 m in width, and have amplitudes of 3-20 m; most ~5 m (Fig. 2b inset). These dimensions give rise to high elongation ratios, up to 52:1 (minimum 18:1). The trend of the bedforms is again uniformly NNE-SSW (31-33˚), suggesting they may simply be the up-trough extension (albeit, a slightly younger generation) of the bedforms imaged in Figure 2a. The bedforms shown in Figs 2a and 2b both lie within the outer shelf region defined by Lowe and Anderson (2002) gemorphic Zone 4, in which they suggested the only bedforms are randomly oriented iceberg furrows.

Farther landward along the trough, additional sets of parallel and attenuated bedforms on the middle shelf are interrupted at the sea floor by continuous seabed ramps that form prominent escarpments, striking transverse to the trough long-axis, and with gently dipping back-slopes (Fig. 2c). At least three separate ramps are imaged in the middle shelf area: one at 72˚ 33'S (GZW3 front) (Fig. 3b), another at 72˚ 42'S (GZW4 front) (Fig. 3c), and a third at 72˚ 51'S (GZW5 front) (Fig. 1, 2c, 3c). In TOPAS profiles and from multibeam analysis, it is apparent that individual ramps are the seaward flanks of sea floor wedges (Fig. 2c inset, 3, 4b). GZW3 is a small ridge ~15 m high, with a series of more subtle, low amplitude ridges formed behind it (Fig. 3b, 4b). GZW4 is ~35 m high at its crest, has an along-trough extent of ~16 km, and is imaged within a 2.5 km-wide corridor of multibeam data. Small, lobate mounds characterise the lower slope of the sea floor ramp in front of GZW4 (Fig. 4b). GZW5 is the largest of the three mid-shelf 'wedge' landforms with an estimated volume of at least 6 km^3 (50 m average height, by 20 km along-trough extent, by a minimum of 6 km across the wedge front, Fig. 3c; note also the similar slope angles of its northern and southern flanks, Fig. 4b). All the wedges are likely formed of sediment, as shown by an existing seismic profile through the landforms

[Figure 4b in Lowe and Anderson, 2002], and they form a stacked, back-stepping complex on the middle shelf (Fig. 4).

Individual streamlined bedform sets are associated with each of the imaged wedges, with lineations terminating either at the crest of the seabed ramp (Fig. 2c inset, 3c) or further landward of the wedge back-slope, as a result of ploughing by iceberg keels near the crest (Fig. 2c inset, 3c). The orientation of lineations changes progressively up-trough, from a NNE-SSW orientation (28-32˚) to an NNW-SSE orientation (353˚) either side of GZW4 (Fig. 2c inset, Fig. 3b). On the up-trough flank of GZW5, further large and elongated lineations are well developed at the seabed, and these trend NNW-SSE (353˚) (Fig. 2c, 3c). These bedforms extend southwards along the trough floor to the transition of the inner to the middle shelf (Fig. 2d) [cf. Lowe and Anderson, 2002]. The lineations have dimensions of 4-12 km length, ~250-500 m width, and amplitudes of 2-10 m (average ~5 m) with length: width ratios up to ~48:1 (Fig. 2d), and thus are similar to those in Figs. 2a and 2b.

On the inner shelf, more than 280 km from the shelf edge, elongated drumlins, less elongate streamlined grooves, meltwater channels and tunnel valleys were reported by Lowe and Anderson [2002, 2003]. The authors showed that these features were ubiquitously formed as erosional bedforms on substrate that they identified as acoustic basement in seismic profiles (which they interpreted as crystalline bedrock). By contrast, all the bedforms and GZWs described north of ~73˚ 30’ S are formed over a seaward-thickening sedimentary substratum, as shown in seismic records [Lowe and Anderson, 2002, 2003].

4. Palaeoglaciological interpretations

Large, straight to curvi-linear lineations on the outermost shelf in the ASE, at the crests of the two sea-floor wedges closest to the shelf edge (GZW1 and GZW2; Figs. 2a, 3a, 4a), are interpreted as a subglacial or grounding-line proximal bedform signature. We interpret their formation at, or close to, a laterally-restricted, grounded, and calving ice-sheet margin (such as an ice-stream front) based on their: (1) high elongation ratios and long lengths; (2) few cross cuts, and grouping into general parallel to sub-parallel alignments that both suggest local restriction to movement of grounded ice or icebergs, which are just afloat; (3) distinct, rounded ridge-crest surfaces in crossing TOPAS profiles, reminiscent of subglacial lineations rather than typically ‘v’-shaped iceberg ploughmarks; and (4) larger, straighter, and more elongate morphology than neighbouring iceberg furrows. The varying types of bedform geometry indicate that the sea floor in this location probably records a mixture of subglacial and iceberg-keel ploughed morphologies. Similar mixed morphologies have been observed on multibeam data in the Vega Trough, and in the Robertson Trough, offshore of the Antarctic Peninsula [Fig. 4 in Heroy and Anderson, 2005].

To the south on the outer shelf, highly parallel, pervasive sets of ridge-groove streamlined bedforms, measuring several to >15 km in length and with elongation ratios between 18->60:1, are interpreted, more typically, as subglacial mega-scale glacial lineations [MSGLs; Clark, 1993; Stokes and Clark, 1999]. MSGLs, analogous in both geometry and form to those imaged in this study, are found as a relict expression of fast-flow on many palaeo-ice stream beds [e.g., Wellner et al., 2001; Stokes and Clark, 2002; Shipp et al., 1999, 2002; Evans et al., 2005; Ó Cofaigh et al. 2005a, 2005b], and have recently been imaged at the bed of the Rutford Ice Stream in a downstream zone where it is fast-flowing (>300 m yr^{-1}) confirming, through direct observation, an ice stream-MSGL connection [King et al., 2009]. Like flutings on glacial forelands, the MSGLs are aligned parallel to former glacier flow [Clark, 1993]. Therefore, the MSGLs in the PITE indicate a dominant flow direction arcing to the N and NNE.

Importantly, MSGLs are also found in association with gravely diamictons in sediment cores recovered from the Pine Island trough, which have been interpreted as subglacial deformation tills [Lowe and Anderson, 2002]. Although some studies have prescribed catastrophic outbursts of meltwater to explain Antarctic bedform formation [Shaw et al., 2008], overwhelming evidence appears to lie in support of the theory that MSGLs form, and ice-streams operate upon and within layers of dilatant deforming till, similar if not identical to those recovered from the ASE [King et al., 2009; Ó Cofaigh et al., in press]. In many other places around Antarctica geophysical evidence for MSGLs combined with geological data supporting a deforming sedimentary substrate have been shown to be clear diagnostic criteria for palaeo-ice stream activity [Wellner et al., 2001; Shipp et al., 2002; Anderson et al., 2002; Dowdeswell et al., 2004; Evans et al., 2005; Heroy and Anderson, 2005; Ó Cofaigh et al. 2005a, 2005b; Mosola and Anderson, 2006]. Our data from the ASE, in combination with previous core studies [Lowe and Anderson, 2002], support these conclusions. The elongate MSGLs also form part of a larger convergent down-flow landform suite including drumlins and bedrock grooves, which are generally considered typical of a palaeo-ice stream landform continuum [Wellner et al., 2001; Graham et al., 2009].

The bedforms in the eastern ASE therefore record the former presence of a fast-flowing grounded ice stream (the PITIS) that drained through Pine Island Bay and its seaward extension, Pine Island Trough East. The bedforms clearly lie in the centre and deepest parts of the main cross-shelf trough as depicted by bathymetric profiles, and as expected for the locale of a major palaeo-ice stream (Fig. 5a and 5b). The pristine preservation of the bedforms, combined with the published deglacial chronology [Lowe and Anderson, 2002], indicates that this ice stream existed at and shortly after the LGM and that ice must have grounded to within 68 km of the shelf edge. Beyond the most northeasterly streamlined bedforms, any further grounded-ice indicators have been removed by ice keel ploughmarks. Thus, the PITIS must have occupied a minimum of ~87% of the currently ice-free trough, and probably covered extensive parts of the shelf outside of it at the LGM.

Sets of bedforms on the middle shelf, aligned at different orientations to the flowset on the outer shelf, record the back-stepping subglacial imprint of the ice stream during its retreat across the continental shelf. Bedform sets are commonly associated with pronounced sedimentary wedges, which we interpret as grounding zone wedges (GZWs 1-5) (Fig. 3, 4): depositional accumulations formed at ice-stream mouths during stillstands of the grounding line where deformable till sheets at the ice-bed act as sediment-conveyors [Alley et al., 1989; Anderson, 2007; Anandakrishnan et al., 2007]. The fact that bedforms terminate at wedge fronts and are well-preserved in ‘soft’ sedimentary strata [Lowe and Anderson, 2002] indicate that they are features formed during recent ice recession rather than older lineations that have been subsequently overridden [cf. Ó Cofaigh et al., 2005a; Dowdeswell et al., 2008]. Smaller mounds and flow-transverse ridges are interpreted as proglacial debris flows or minor recessional moraines, both associated with ice-marginal sediment delivery (Fig. 3b); the scale and corrugated, sub-parallel morphology of the latter compare well to ‘corrugation moraines’ that formed during the last deglaciation and were mapped in cross-shelf troughs in the Ross Sea [Shipp et al., 2002; Dowdeswell et al., 2008]. The presence of GZWs interspersed by distinct MSGL flowsets identifies an *episodic* style of deglaciation [e.g., Larter and Vanneste, 1995; Dowdeswell et al. 2008; Ó Cofaigh et al. 2008], with stepped grounding line retreat across the continental shelf. These GZWs afford a preliminary reconstruction of the retreat pattern of the PITIS, where former stationary or readvanced positions of the grounding line can be traced (Fig. 1, marked by purple lines).

On the inner shelf, bathymetric sills were likely to provide additional pinning points for the grounding line in a zone where sediments are sparse and the topography is highly pronounced (Fig. 5c). Bedforms scoured into acoustic basement elsewhere on the inner shelf are interpreted as the upstream signature of palaeo-ice flow in the former Pine Island and

Thwaites glacier systems [Lowe and Anderson, 2002, 2003], though they may not all relate to ice streaming at the LGM [cf. Graham et al., 2009].

5. Flow outlet of the Pine Island-Thwaites Ice Stream

Using an earlier, more limited multibeam dataset, Evans et al. [2006] suggested that the main outlet from Pine Island Trough at the LGM drained to the NW via a trough that reaches the shelf edge at 114˚ W (PITW) (Fig. 6). The presence of subdued streamlined bedforms within this NW trough does support the idea that ice drained through this outlet at some point in the past. However, MSGLs on the outer shelf along the PITE at 106˚ W are more pristine and have greater elongation ratios than those in the PITW, which suggests that major discharge and outflow of the PITIS at its last maximum extent was directed to the northeast along the PITE (cf. Figs 2b and 6). We rule out variation in post-glacial sediment coverage to explain differences in bedform preservation between PITE and PITW. Evans et al. [2006] showed only a thin post-glacial sediment drape in the PITW from acoustic profiles, and thin post-glacial sediments were also recovered in cores from the PITE [Lowe and Anderson, 2002]. Line A-A’ is the only continuous west-east bathymetric profile to cross the PITE on the outer shelf, near the NE Amundsen Sea shelf break (Fig 5a). In conjunction with our multibeam datasets, it shows, for the first time, that the PITE actually continues to the shelf edge supporting a main flow outlet in this location (Fig. 1, shown by the regions of bathymetry shaded in orange; Fig. 5a).

In the Pine Island cross-shelf trough, Lowe and Anderson [2002, 2003] also defined a number of geomorphic zones forming a downflow progressive model across the inner and middle shelf. Much of the sea bed on the Amundsen Sea outer shelf has been pervasively scoured by icebergs so our discovery of subglacial bedforms in the PITE is significant for showing ice extent beyond the middle shelf, where streamlined features were previously thought to terminate [Lowe and Anderson, 2002, 2003]. These bedform patterns show that there was streaming ice flow, at various time intervals, over an along-trough distance of at least ~200 km on the middle-to-outer shelf, demonstrating fast-flow at the scale of or even larger than other major ice-stream systems in West Antarctica during the LGM, such as those in Marguerite Trough [Ó Cofaigh et al., 2005b], Belgica Trough [Ó Cofaigh et al., 2005a], Boyd Strait [Canals et al., 2000], Anvers Trough [Pudsey et al., 1994; Domack et al., 2005], and the Ross Sea [Shipp et al., 1999; Mosola and Anderson, 2006]. Our study indicates an extensive palaeo-ice stream in the eastern ASE and strongly supports previous interpretations that the WAIS in this sector reached the shelf edge at the LGM [Lowe and Anderson, 2002; Evans et al., 2006].

6. Outer shelf ice dynamics

While we consider it likely that the PITE was the main outlet for the PITIS, the presence of two well-preserved subglacial bedform sets, in separate outer-shelf troughs, points to complex ice dynamics. Figure 7 illustrates three possible interpretations, which could be used to reconcile the two datasets.

Our favoured interpretation is that phases of ice streaming within the PITE and PITW were asynchronous (Fig. 7a). Switching ice-stream flow between the two adjacent troughs probably produced the bedforms at different times [cf. Dowdeswell et al., 2006a], and we suggest that flow through PITW preceded the flow through PITE. This flow-switching scenario is supported by: (1) the better preservation and exceptional clarity of the bedform signature in the PITE compared with that in the PITW, suggesting a younger age of the MSGLs to the northeast; (2) the presence of only a single major trough crossing the inner and middle shelf, but two well-defined outer-shelf outlets; (3) the deeper incision of the PITE outlet into the shelf when compared to the PITW outlet (i.e. incision to different base levels;

Fig. 5b), which increases the likelihood of earlier ice flow being captured by more dominant ice-stream flow draining through the PITE; (4) the existence of slope gullies at the mouth of both the PITE and PITW which indicate that slope processes have probably been diachronous across the Amundsen Sea margin. Although there is contention over their mode of formation [Noormerts et al., 2009], gullies at the mouth of the PITW were interpreted as derived from sediment-laden meltwater or slope instability flows [Dowdeswell et al. 2006b; Noormets et al. 2009], whereas slope debris flows more typical of ice marginal sediment delivery are found near the mouth of the PITE [Dowdeswell et al., 2006b]. This disparity would appear to support a younger age for the PITE.

Clear evidence of flow switching of this kind has been found in 3D seismic data from the North Norwegian margin, linked to sedimentary infill of accommodation space within cross-shelf troughs [Dowdeswell et al., 2006a]. Flow-switching in ice streams has not been directly observed in ice masses, but is suspected to be possible driven by changing subglacial water drainage [Vaughan et al., 2008] and has almost certainly occurred in the recent past [Conway et al., 2002].

An alternative interpretation is that two separate ice-flow limbs were synchronous features within the Late Quaternary WAIS and existed as two bifurcating ice streams on the outer shelf (Fig. 7b). However, very few bifurcating analogues are observed in contemporary ice sheets, and most reconstructions of Quaternary palaeo-ice sheets in the northern hemisphere lack evidence for ice-stream bifurcation in their downstream regions [Ottesen et al., 2005, 2008; Stokes et al., 2009].

A third hypothesis is that the two outlets and their bedform signatures result from one large fast-flow sector across the outer shelf, with ice flow in the form of a divergent 'sheet' (Fig. 7c). To test this idea, we estimated a potential discharge for sheet flow across the area of outer shelf from the PITW to the PITE, assuming a minimum cross-sectional flow area of ~400 km x 550 m. The PITIS basin, including the now-submerged shelf areas and modern Pine Island-Thwaites catchment, would cover an accumulation area of ~500,000 km^2 at the LGM; approximately three times the size of the modern Pine Island Glacier catchment. By scaling the modern net surface accumulation over Pine Island Glacier catchment (~69-71 km^3 ice yr^{-1}) [Rignot, 2006] to this catchment area (~207-213 km^3 ice yr^{-1}), and assuming significantly lower precipitation rates during the LGM (approximately half; ~105 km^3 ice yr^{-1}) we show it would be possible to sustain sheet discharge through this gate at a velocity of ~480 m yr^{-1}: a fast-flow rate comparable to some contemporary Antarctic ice streams. However, while some models of ice sheets predict 'sheet'-style flows at marine margins [Hubbard et al., 2009], modern glacial analogues remain absent, at least at similar scales (but see Malaspina Glacier, Alaska, for a possible scaled-down comparison) [Sharpe, 1958]. Therefore, while conceivable, we do not favour this interpretation.

Regardless of the specific scenario, our data document a major discharge of the ancient Pine Island-Thwaites ice stream along the PITE. Taking into account a lowered sea level at the LGM of ~120 m, a minimum ice thickness of 715 m is estimated if ice grounded in the trough at 72° 30'S. This estimate neglects any glacio-isostatic depression of the shelf by the ice that extended onto it, which would make the minimum ice thickness even greater. This cross-section would, in turn, accommodate a minimum flux of ~108 km^3 ice yr^{-1} (based on an assigned flow velocity of 2.5 km yr^{-1}) exceeding the modern grounding line flux of 75-80 km^3 ice yr^{-1} from Pine Island Glacier today (where velocity is measured at ~ 2.5 km yr^{-1}). Thus, we deduce the PITIS to have been a key drainage feature of the WAIS, since at least the last glacial period.

7. Retreat of the Pine Island-Thwaites Ice Stream

Our data appear to show that multiple GZWs formed at the grounding line of the PITIS as it retreated from the outer and middle shelf. Previously, only a single wedge had been identified in the PITE [Lowe and Anderson, 2002] with a second tentatively suggested for the PITW at ~108° W by Evans et al. [2006]. Our geophysical data show that the mid-shelf wedge of Lowe and Anderson (2002) is in fact a complex of back-stepping wedges, formed at a number of ice marginal positions (Fig. 4). This geometry suggests that the ice margin periodically stabilised during retreat, when stillstands of the grounding line may have occurred. The presence of large GZWs (minimum volume of ~6 km^3 for GZW5) indicates significant subglacial erosion - of the upstream portion of the ice stream trunk and the Antarctic interior - and shows that the ice stream was also a major conduit for basal sediment transport, at least during the early deglacial phase. The lack of comparatively thick sediment cover on parts of the inner shelf may be due to the PITIS stripping sediment from its bed during ice retreat [Smith and Murray, 2009].

Despite new constraints on palaeo-ice margin location, the deglacial history of the PITIS since the LGM remains poorly constrained. The chronology of PITIS retreat is restricted to a radiocarbon age of ~15.8 ± 3.9 ^{14}C ka BP obtained from deglacial transitional sediments on the mid-shelf near Burke Island and a minimum age of ~10.2 ± 0.4 ^{14}C ka BP for grounding-line retreat from the inner shelf (at 73° 55’ S and 106° 39’ W) of the PITE [Lowe and Anderson, 2002]. However, these ages give little insight into the duration of stillstands during grounding line migration up the trough. If we adopt sediment flux for an ice stream of similar size to the PITIS [e.g., 8 x 10^3 m^3 yr^{-1} per metre width, Norwegian Channel Palaeo-Ice Stream; Nygård et al., 2007], we can calculate an equivalent total sediment flux from its grounding line, across a GZW of minimum 6 km width (and 6 km^3 volume; similar to GZW5), of ~0.05 km^3 yr^{-1}. Under these constant flux conditions, it is feasible for GZWs in the eastern Amundsen Sea to have formed in ~120 years (i.e., on centennial timescales). This rough estimate differs with previous interpretations, which suggest that GZW formation may stabilise an ice margin for thousands of years or more [Anderson, 2007]. Consequently, landforms of "episodic" retreat in the ASE may not necessarily indicate prolonged ice-margin recession, but may instead be manifestations of effective sediment delivery during a punctuated but otherwise rapid deglaciation [Larter and Vanneste, 1995; Dowdeswell et al., 2008].

The observed landform configuration (presence of GZWs and the lack of recessional moraines) on the outer shelf generally supports our interpretations of a rapid, though episodic PITIS retreat [cf. Dowdeswell et al. 2008; Ó Cofaigh et al., 2008]. Runaway collapse, of the modern ice sheet has been suggested where the ice-bed slopes continuously into the ice-sheet interior [Vaughan and Arthern, 2007]. For the PITE and PITIS, the palaeo-ice bed from the outer shelf to the inner 50 km of the shelf consists of a broadly continual reverse slope towards the coast (Fig. 5c). Notably, the only breaks in this gradient are locations where the sea floor levels out to form "benches". These benches occur in close association with grounding zone wedges (Fig. 5c), and the dimensions of the GZWs themselves are too small to account for the changes in sea-floor slope. Thus, periods of relative stability in ice retreat appear to have coincided with ice grounding on lower gradient beds (average slope measured at 0.015°), while well-preserved bedforms occur on beds with higher gradients (average slope measured at 0.149°), with no evidence for grounding line stabilization in these areas. This relationship suggests phases of rapid retreat over beds with greater slopes. Therefore, our observations lend support to model experiments indicating that stepped patterns of ice-stream retreat are highly sensitive to the gradient of the bed on which they are grounded [Weertman, 1974; Schoof, 2007].

Between the wedges, the presence of MSGLs indicates that either: (1) ice streamed consistently during the retreat, or (2) renewed onset of streaming occurred prior to each phase of ice margin recession. The feasibility of these dynamic patterns of ice-stream retreat may be tested by future coupled glaciological-sedimentary models, which attempt to reconstruct the retreat history of major West Antarctic ice stream systems. Such models should seek to replicate periodic stillstands of the PITIS grounding line.

8. Conclusions

We have identified a major LGM-outlet of the Pine Island and Thwaites ice streams in the ASE, based on observations from new sea floor and sub-bottom imagery. Addressing the hypotheses introduced at the start of the paper, we conclude the following:

(1) Streamlined glacial bedforms show that the West Antarctic Ice Sheet extended to within 68 km of, and probably right up to, the continental shelf edge in the northeastern Amundsen Sea Embayment at the LGM, supporting previous studies farther west in the embayment, which suggested extensive LGM ice-sheet limits [Evans et al., 2006; Larter et al., 2009];

(2) The major drainage pathway of the PITIS can be traced along a cross-shelf trough that connects to a trough-mouth on the outer shelf at ~106º W. This mouth was likely the main outlet for the PITIS. Earlier phases of ice flow may have diverted along an outer-shelf trough to the west, with an outlet at ~114º W. Flow-switching is our favoured explanation for the presence of the two outlets, and the two sets of bedforms preserved within each outer-shelf trough.

(3) Sea-floor geomorphic evidence constrains a stepped, episodic style of deglaciation for the PITIS. The course of ice retreat from the outer and middle shelf in the trough of the eastern ASE is shown by five grounding zone wedges, which formed during retreat. Stillstands occurred in association with changes in subglacial bed gradient,

and we suggest that more rapid phases of retreat correlated to higher bed slopes. Prominent bedrock pinning points on the inner shelf probably served to halt this retreat even further, as the grounding line receded to its present-day configuration. Further chronological data from sediment cores are urgently required to better verify the timing of stepwise retreat during the last deglaciation.

Acknowledgements

We thank the captains and crews of the RRS *James Clark Ross* and RV *Polarstern*, and shipboard parties of numerous cruises, for collecting data. This work forms a contribution to the Ice Sheets component of the British Antarctic Survey Polar Science for Planet Earth Programme. It was funded by The Natural Environment Research Council.

Figure Captions

Figure 1. Location map of Pine Island Bay and the eastern Amundsen Sea Embayment. Inset shows the study area (red box) and delineates the Amundsen Sea drainage sector at the present day (thick black line). (main figure) Background bathymetry is from the regional compilation of Nitsche et al. [2007], and is overlain by a shaded-relief swath bathymetry grid (150-m cell size) compiled from BAS and AWI cruises, supplemented by other data from the Lamont-Doherty multibeam synthesis. Thick purple lines indicate crests of grounding zone wedges formed during ice retreat – labelled as referred to in the text. Coastline, ice sheet and ice shelves are drawn from the Antarctic Digital Database. Profiles A-A’, B-B’ and start and end points of C-C’ are illustrated in white (see Fig. 5). Note that profile C-C’ is not drawn in full in order to avoid masking sea-floor features. The TOPAS profiles in Figure 4 are also shown in white. Note that all multibeam bathymetry lines were collected in conjunction with sub-bottom profiler data, and both data types were used in our interpretations. PIG: Pine Island Glacier, GZW: grounding zone wedge, PITW/PITE: Pine Island Troughs West/East. See online version for colour figure.

Figure 2. Four regions of grey-scale shaded multibeam swath bathymetry along the Pine Island Trough East and Pine Island cross-shelf trough, from the outer shelf to the mid-shelf. Subglacial bedforms are imaged right along the cross-shelf trough, as are grounding zone wedges (labelled GZWs 1-5). Images located on Figure 1. Hillshade from WNW in all cases. Grid cell size 30 m.

Figure 3. Three-dimensional perspective images of multibeam swath bathymetry illustrating five grounding zone wedges (GZWs) along the Pine Island cross-shelf trough, seaward of

Pine Island Bay. Note ice flow direction arrows, and that Figure 3c is rotated so that palaeo-ice flow is towards the reader. Thick black arrows mark the crests of each wedge. For location of GZWs see the labels in Figure 1. Grid cell size 40 m. See online version for colour figure.

Figure 4. TOPAS sub-bottom acoustic profile showing back-stepping grounding zone wedges: (a) on the outer shelf (GZW 1 and GZW 2); dashed line shows splice of two profile parts; (b) on the middle-shelf within the Pine Island cross-shelf trough (GZW3, GZW4 and GZW5). The crests of the wedges are arrowed, and indicate palaeo-grounding lines as illustrated in Figure 1 (purple lines). Profiles located on Figure 1 as thick white lines, and as boxes on Figure 5c.

Figure 5. Bathymetric profiles A-A' and B-B' illustrating the physiographic setting of the two outer shelf troughs derived from the grid presented in Nitsche et al. [2007]. Note the continuation of the PITE towards the shelf edge. C-C' shows the long-axis profile of the main Pine Island cross-shelf trough and the PITE, derived from the compiled swath bathymetry grid shown in Figure 1. Locations see Figure 1.

Figure 6. EM120 multibeam swath bathymetry in Pine Island Trough West (PITW), showing subtle and subdued streamlined landforms at the mouth of the cross-shelf trough. Grid cell size 50 m. Location see arrow on Figure 1. Note cross-track artefacts in the dataset. See online version for colour figure.

Figure 7. Three conceptual models for the flow of the extended Pine Island-Thwaites palaeo-ice stream at its last maximum extent, based on new and existing datasets. A shelf-edge ice extent is interpreted in all scenarios (thin black line). See text for full descriptions.

References

Alley, R. B., D. D. Blankenship, S. T. Rooney, and C. R. Bentley (1989), Sedimentation beneath ice shelves–the view from ice stream B, *Mar. Geol., 85,* 101-120.

Anderson, J.B. (2007), Ice sheet stability and sea level rise, *Science, 315*, 1803-1804.

Anderson, J.B., S.S. Shipp, A.L. Lowe, J.S. Wellner, and A.B. Mosola (2002), The Antarctic Ice Sheet during the Last Glacial Maximum and its subsequent retreat history - a review, *Quat. Sci. Revs., 21*, 49-70.

Anandakrishnan, S., G. A. Catania, R. B. Alley, and H. J. Horgan (2007), Discovery of Till Deposition at the Grounding Line of Whillans Ice Stream, *Science, 315* (5820), 1835. DOI: 10.1126/science.1138393.

Canals, M., R. Urgeles, and A.M. Calafat (2000), Deep sea-floor evidence of past ice streams off the Antarctic Peninsula, *Geology, 28*, 31-34.

Clark, C.D. (1993), Mega-scale glacial lineations and cross-cutting iceflow landforms, *Earth Surf. Proc. Land., 18*, 1-19.

Conway H., G. Catania, C.F. Raymond, A.M. Gades, T.A. Scambos, and H. Engelhardt (2002), Switch of flow direction in an Antarctic Ice Stream, *Nature, 419*, 465-467.

Domack, E. W., D. Amblas, R. Gilbert, S. Brachfeld, A. Camerlenghi, M. Rebesco, M. Canals, and R. Urgeles (2005), Subglacial morphology and glacial evolution of the Palmer deep outlet system, Antarctic Peninsula, *Geomorph., 75*, 125-142.

Dowdeswell, J.A., C. Ó Cofaigh, and C.J. Pudsey (2004), Thickness and extent of the subglacial till layer beneath an Antarctic paleo-ice stream, *Geology, 32*, 13-16.

Dowdeswell, J.A., D. Ottesen, and L. Rise (2006a), Flow-switching and large-scale deposition by ice streams draining former ice sheets, *Geology, 34*, 313-316.

Dowdeswell, J.A., J. Evans, C. Ó Cofaigh, J.B. Anderson (2006b), Morphology and sedimentary processes on the continental slope off Pine Island Bay, Amundsen Sea, West Antarctica, *Geol. Soc. Am., Bull., 118*, 606-619.

Dowdeswell, J.A., D. Ottesen, J. Evans, C. Ó Cofaigh, J.B. Anderson (2008), Submarine glacial landforms and rates of ice-stream collapse, *Geology, 36*, 819-822.

Evans, J., C.J. Pudsey, C. Ó Cofaigh, P.W. Morris and E.W. Domack (2005), Late Quaternary glacial history, dynamics and sedimentation of the eastern margin of the Antarctic Peninsula Ice Sheet. *Quat. Sci. Rev., 24*, 741-774.

Evans, J., J.A. Dowdeswell, C. Ó Cofaigh, T.J. Benham, and J.B. Anderson (2006), Extent and dynamics of the West Antarctic Ice Sheet on the outer continental shelf of Pine Island Bay during the last glaciation, *Mar. Geol., 230*, 53-72.

Graham, A.G.C., R.D. Larter., K. Gohl, C.-D. Hillenbrand, J.A. Smith, and G. Kuhn (2009), Bedform signature of a West Antarctic ice stream reveals a multi-temporal record of flow and substrate control, *Quat. Sci. Rev., 28*, 2774–2793 doi:10.1016/j.quascirev.2009.07.003.

Heroy, D.C. and J.B. Anderson (2005), Ice-sheet extent of the Antarctica Peninsula region during the Last Glacial Maximum (LGM) - Insights from glacial geomorphology. *Geol. Soc. Am. Bull., 117*, 1497-1512.

Holt, J.W., D.D. Blankenship, D.L. Morse, D.A. Young, M.E. Peters, S.D. Kempf, T.G. Richter, D.G. Vaughan, and H.F.J. Corr (2006), New boundary conditions for the West Antarctic Ice Sheet: Subglacial topography of the Thwaites and Smith Glacier catchments, *Geophys. Res. Lett., 33*, L09502, doi:10.1029/2005GL025561.

Hubbard, A.L., T. Bradwell, N.R. Golledge, A. Hall, H. Patton, D.E. Sugden, R.M. Cooper, and M.S. Stoker (2009), Dynamic binge-purge cycles, ice streams and their impact on the extent and chronology of the last British-Irish Ice Sheet, *Quat. Sci. Revs., 28*, 759-777.

Hughes, T.J. (1981), The weak underbelly of the West Antarctic ice sheet, *J. Glaciol., 27,* 518-525.

Jacobs, S.S., H.H. Hellmer, and A. Jenkins (1996), Antarctic ice sheet melting in the Southeast Pacific, *Geophys. Res. Lett., 23*, 957-960.

King, E.C., R.C.A. Hindmarsh, and C.R. Stokes (2009), Formation of mega-scale glacial lineations observed beneath a West Antarctic ice stream, *Nat. Geosci., 2,* 585-588. doi:10.1038/ngeo581.

Larter, R. D., and L.E. Vanneste (1995), Relict subglacial deltas on the Antarctic Peninsula outer shelf, *Geology, 23*, 33-36.

Lowe, A.L., and J.B. Anderson (2002), Reconstruction of the West Antarctic ice sheet in Pine Island Bay during the Last Glacial maximum and its subsequent retreat history, *Quat. Sci. Revs., 21*, 1879-1897.

Lowe, A.L., and J.B. Anderson (2003), Evidence for abundant subglacial meltwater beneath the paleo-ice sheet in Pine Island Bay, Antarctica, *J. Glaciol.*, *49*, 125-138.

Mosola, A.B., and J.B. Anderson (2006). Expansion and rapid retreat of the West Antarctic Ice Sheet in eastern Ross Sea: Possible consequence of over extended ice streams? *Quat. Sci. Rev., 25*, 2177-2196.

Nitsche, F.O., S.S. Jacobs, R.D. Larter, and K. Gohl (2007), Bathymetry of the Amundsen Sea continental shelf: Implications for, geology, oceanography and glaciology, *Geochem., Geophys., Geosys., 8*, Q10009, doi: 10.1029/2007GC001694.

Noormets, R., J.A. Dowdeswell, R.D. Larter, C. Ó Cofaigh, and J. Evans (2009), Morphology of the upper continental slope in the Amundsen and Bellingshausen seas – implications for sedimentary processes at the shelf edge of West Antarctica, *Mar. Geol., 258*, 100-114.

Nygård, A., H.P. Sejrup, H. Haflidason, W.A.H. Lekens, C.D. Clark, and G.R. Bigg (2007), Extreme sediment and ice discharge from marine-based ice streams: New evidence from the North Sea, *Geology, 35*, 395-398.

Ó Cofaigh, C., R.D. Larter, J.A. Dowdeswell, C.-D. Hillenbrand, C.J. Pudsey, J. Evans, and P. Morris (2005a), Flow of the West Antarctic Ice Sheet on the continental margin of the Bellingshausern Sea at the Last Glacial Maximum, *J. Geophys. Res.*, *110*, B11103, doi: 10.1029/2005JB003619.

Ó Cofaigh, C., J.A. Dowdeswell, C.S. Allen, J. Hiemstra, C.J. Pudsey, J. Evans, and D.J.A. Evans (2005b), Flow dynamics and till genesis associated with a marine-based Antarctic palaeo-ice stream, *Quat. Sci. Revs., 24*, 709-740.

Ó Cofaigh, C., J.A. Dowdeswell, J. Evans, and R.D. Larter (2008), Geological constraints on Antarctic palaeo-ice stream retreat, *Earth Surf. Proc. Land., 33*, 513-525.

Ó Cofaigh, C., J.A. Dowdeswell, E.C. King, J.B. Anderson, C.D. Clark, D.J.A. Evans, J. Evans, R.C.A. Hindmarsh, R.D. Larter, and C. R. Stokes (2010), Comment on Shaw J., Pugin, A. and Young, R. (2008): "A meltwater origin for Antarctic shelf bedforms with special attention to megalineations", Geomorphology 102, 364–375, *Geomorphology,* in press.

Ottesen, D., J.A. Dowdeswell, and L. Rise (2005), Submarine landforms and the reconstruction of fast-flowing ice streams within a large Quaternary ice sheet: the 2,500 km-long Norwegian-Svalbard margin (57º to 80ºN), *Geol. Soc. Am., Bull., 117*, 1033-1050.

Ottesen, D., C.R. Stokes, L. Rise, and L. Olsen (2008), Ice sheet dynamics and ice streaming along the coastal parts of northern Norway, *Quat. Sci. Revs., 27*, 922–940.

Payne, A.J., A. Vieli, A. Shepherd, D.J. Wingham, and E. Rignot (2004), Recent dramatic thinning of largest West-Antarctic ice stream triggered by oceans, *Geophysical Research Letters*, *31*, L23401.

Pudsey, C.J., P.F. Barker, and R.D. Larter (1994), Ice sheet retreat from the Antarctic Peninsula shelf, *Cont. Shelf Res., 14*, 1647-1675. doi: 10.1016/0278-4343(94)90041-8.

Pritchard, H.D., R.J. Arthern, D.G. Vaughan, and L.A. Edwards (2009) Extensive dynamic thinning on the margins of the Greenland and Antarctic ice sheets, *Nature, 461*, 971-975 doi:10.1038/nature08471

Rignot, E. (1998), Fast recession of a West Antarctic glacier, *Science, 281*, 549–551.

Rignot, E. (2006), Changes in ice dynamics and mass balance of the Antarctic ice sheet, *Phil. Trans. R. Soc. A., 364* (1844), 1637–1655.

Rignot, E. (2008), Changes in West Antarctic ice stream dynamics observed with ALOS PALSAR data, *Geophys. Res. Lett., 35*, L12505, 10.1029/2008GL033365.

Rignot, E., and S.S. Jacobs (2002), Rapid bottom melting widespread near Antarctic Ice Sheet grounding lines, *Science, 296*, 2020-2023.

Rignot, E., D.G. Vaughan, M. Schmeltz, T. Dupont, and D. MacAyeal (2002) Acceleration of Pine Island and Thwaites Glaciers, West Antarctica, *Ann. Glacio., 34*, 189–194.

Schoof, C. (2007), Ice sheet grounding line dynamics: Steady states, stability and hysteresis, *J. Geophys. Res., 112*, F03S28, doi:10.1029/2006JF000664.

Scott, J. B. T., G.H. Gudmundsson, A.M. Smith, R.G. Bingham, H.D. Pritchard, and D.G Vaughan (2009), Increased rate of acceleration on Pine Island Glacier strongly coupled to changes in gravitational driving stress, *The Cryosphere, 3*, 125-131.

Sharpe, B.M. (1958), Malaspina Glacier, Alaska, *Bull. Geol. Soc. Am., 69*, 617–646.

Shaw, J., A. Pugin, and R.R. Young (2008) A meltwater origin for Antarctic shelf bedforms with special attention to megalineations, *Geomorphology*, *102*, 364–375.

Shepherd, A., D.J. Wingham, J.A.D. Mansley, and H.F.J. Corr (2001), Inland Thinning of Pine Island Glacier, West Antarctica, *Science, 291*, 862-864. doi: 10.1126/science.291.5505.862.

Shepherd, A., D. J. Wingham, and E. Rignot (2004), Warm ocean is eroding West Antarctic Ice Sheet, *Geophys. Res. Lett., 31*, L23402, doi:10.1029/2004GL021106.

Shipp, S., J.B. Anderson, and E. Domack (1999), Late Pleistocene–Holocene retreat of the West Antarctic Ice-Sheet system in the Ross Sea: Part 1—Geophysical results, *Geol. Soc. Am. Bull., 111*, 1486-1516.

Shipp S.S., J.S. Wellner, and J.B. Anderson (2002), Retreat signature of a polar ice stream: sub-glacial geomorphic features and sediments from the Ross Sea, Antarctica, in *Glacier-influenced Sedimentation on High-Latitude Continental Margins*, edited by J.A. Dowdeswell and C. Ó Cofaigh, pp. 277-304, Geological Society, London, Special Publication.

Smith, A.M., and T. Murray (2009), Bedform topography and basal conditions beneath a fast-flowing West Antarctic ice stream, *Quat. Sci. Revs., 28*, 584-596. doi:10.1016/j.quascirev.2008.05.010.

Stokes, C.R., and C.D. Clark (1999), Geomorphological criteria for identifying Pleistocene ice streams, *Annals of Glaciology*, *28*, 67-74.

Stokes, C.R., and C.D. Clark (2002), Are long subglacial bedforms indicative of fast ice flow?, *Boreas, 31*, 239-249.

Stokes, C.R., C.D. Clark, and R. Storrar (2009), Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide Ice Sheet, *Quat. Sci. Revs., 28*, 721-738.

Thoma, M., A. Jenkins, D. Holland, and S.S. Jacobs (2008), Modelling Circumpolar Deep Water intrusions on the Amundsen Sea continental shelf, Antarctica, *Geophys. Res. Letters, 35*, L18602, doi:10.1029/2008GL034939.

Vaughan, D.G. (2008), West Antarctic Ice Sheet collapse – the fall and rise of a paradigm, *Clim. Change, 91*, 65-79.

Vaughan, D.G., and R. Arthern (2007), Why Is It Hard to Predict the Future of Ice Sheets?, *Science, 315*, 1503-1504. doi: 10.1126/science.1141111.

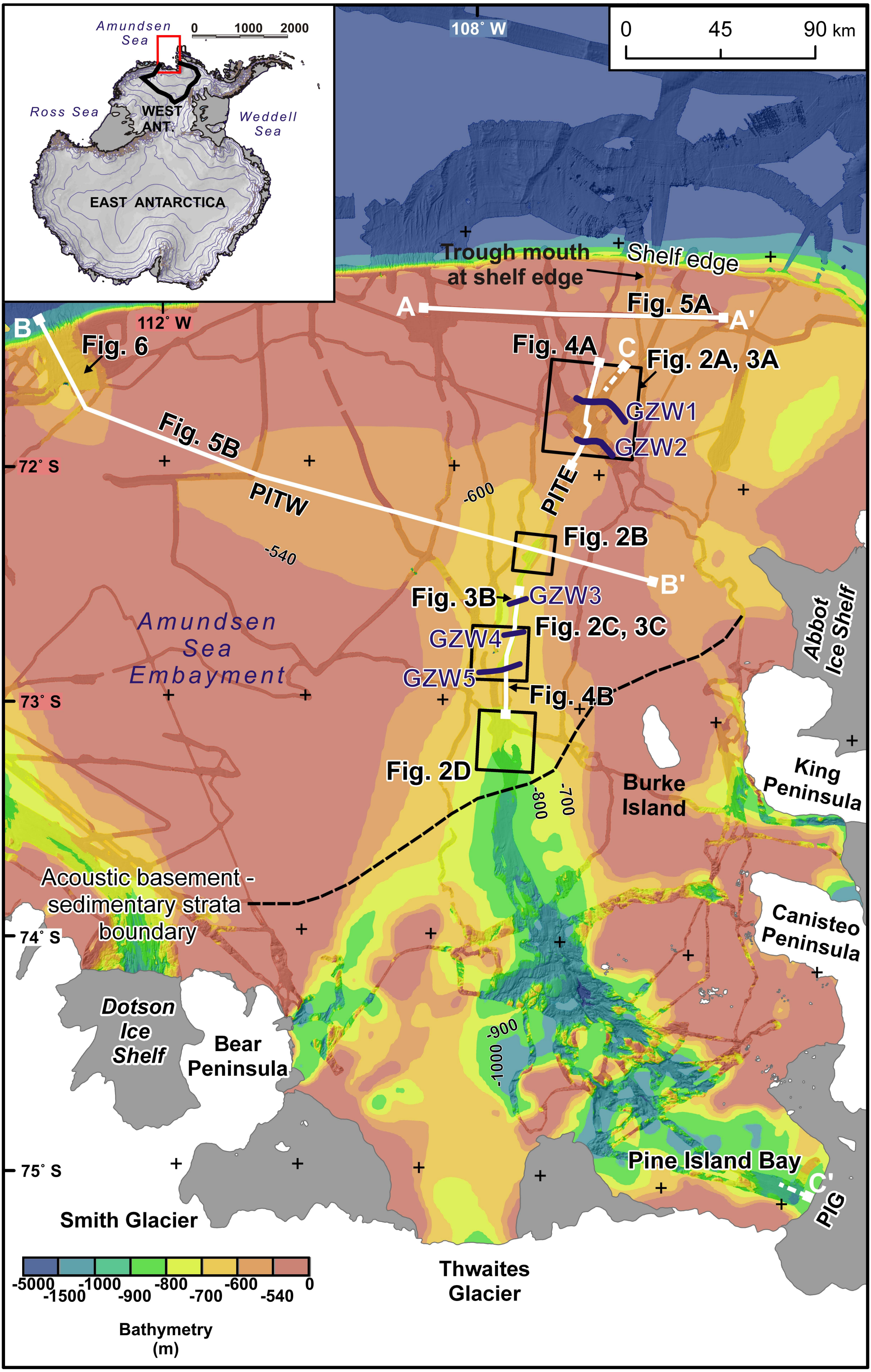
Vaughan, D.G., H.F.J. Corr, F. Ferraccioli, N. Frearson, A. O'Hare, D. Mach, J.W. Holt, D.D. Blankenship, D. Morse, and D.A. Young (2006), New boundary conditions for the West Antarctic ice sheet: Subglacial topography beneath Pine Island Glacier, *Geophys. Res. Letters, 33*, L09501, doi:09510.01029/02005GL025588.

Vaughan, D.G., H.F.J. Corr, H. Pritchard, A. Shepherd, and A.M. Smith (2008), Flow-switching and water-piracy between Rutford Ice Stream and Carlson Inlet, West Antarctica, *J. Glaciol*., *54*, 41-48.

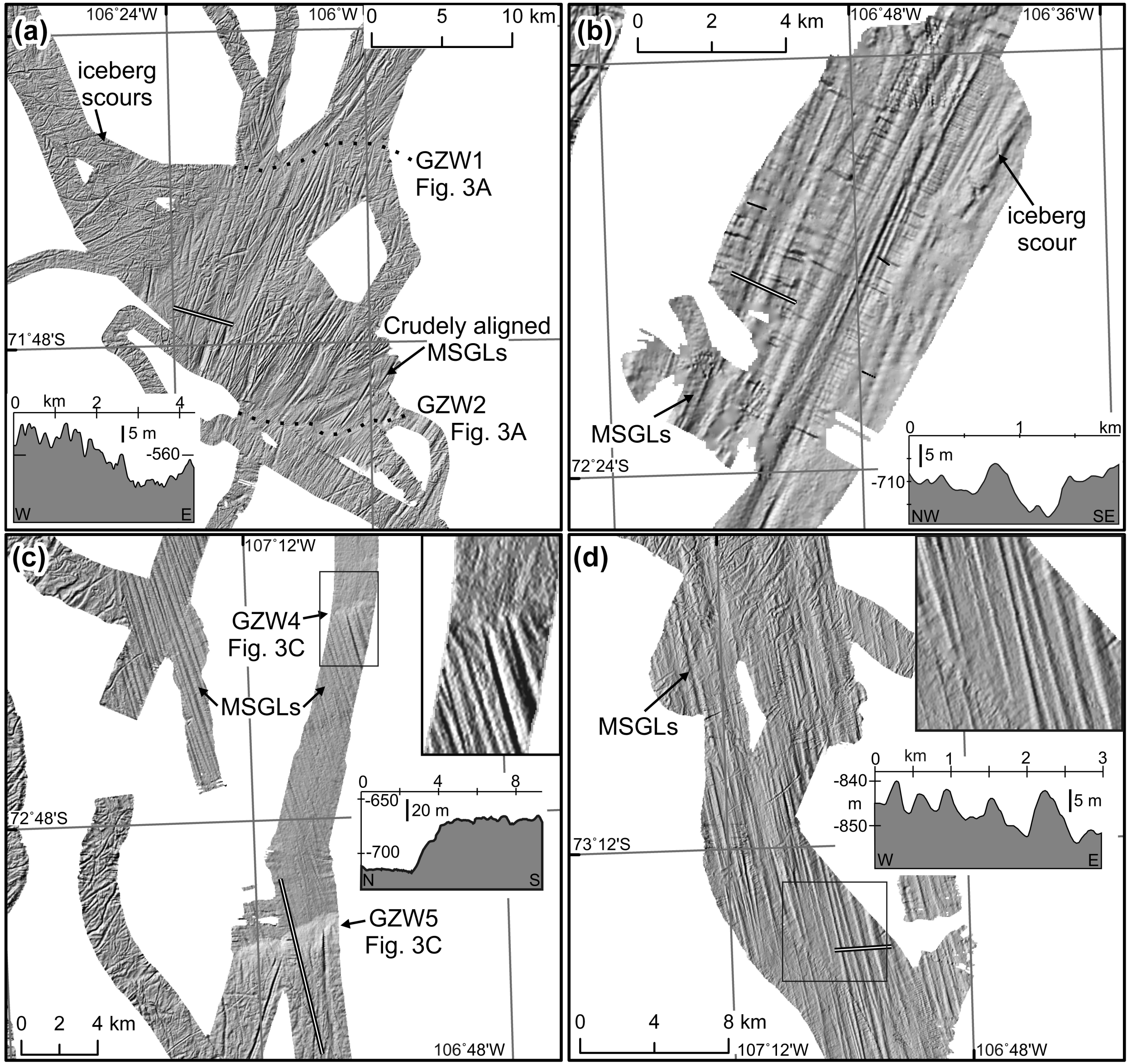
Walker, D.P., M.A. Brandon, A. Jenkins, J.T. Allen, J.A. Dowdeswell, and J. Evans (2007), Oceanic heat transport onto the Amundsen Sea shelf through a submarine glacial trough, *Geophys. Res. Lett., 34*, L02602, doi:10.1029/2006GL028154.

Weertman, J. (1974), Stability at the junction of an ice sheet and an ice shelf, *J. Glaciol., 13*, 3-11.

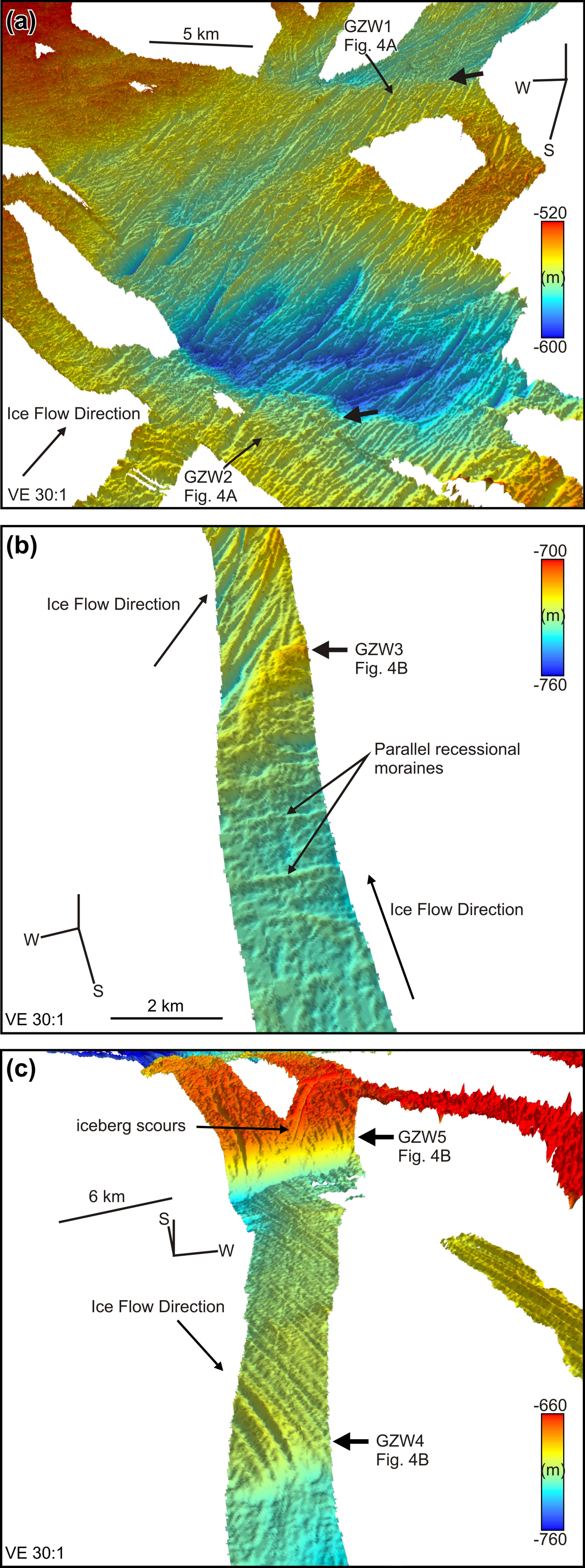
Wellner, J.S., A.L. Lowe, S.S. Shipp, and J.B. Anderson (2001), Distribution of glacial geomorphic features on the Antarctic continental shelf and correlation with substrate: implications for ice behaviour, *J. Glaciol., 47*, 397-411.



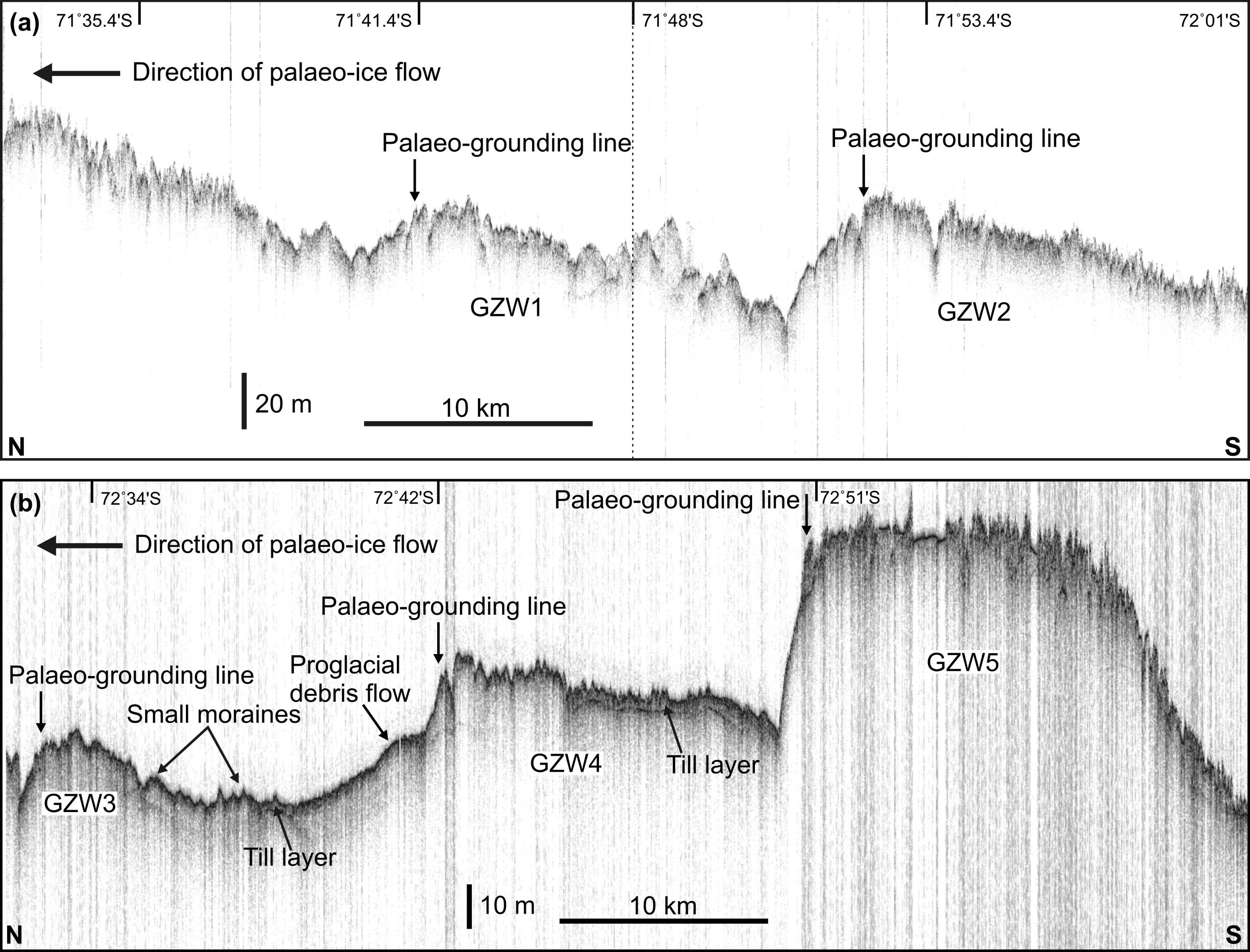
Amundsen Sea
0 1000 2000
Ross Sea
WEST ANT.
Weddell Sea
EAST ANTARCTICA
108° W
0 45 90 km
Trough mouth at shelf edge
Shelf edge
A
Fig. 5A
A'
B
112° W
Fig. 6
Fig. 4A
C
Fig. 2A, 3A
GZW1
GZW2
Fig. 5B
72° S
PITW
PITE
-600
-540
Fig. 2B
B'
Fig. 3B
GZW3
GZW4
Fig. 2C, 3C
GZW5
Fig. 4B
Amundsen Sea Embayment
73° S
Abbot Ice Shelf
Fig. 2D
-800
-700
Burke Island
King Peninsula
Acoustic basement - sedimentary strata boundary
74° S
Canisteo Peninsula
Dotson Ice Shelf
Bear Peninsula
-900
-1000
75° S
Pine Island Bay
C'
PIG
Smith Glacier
Thwaites Glacier
-5000 -1500 -1000 -900 -800 -700 -600 -540 0
Bathymetry (m)



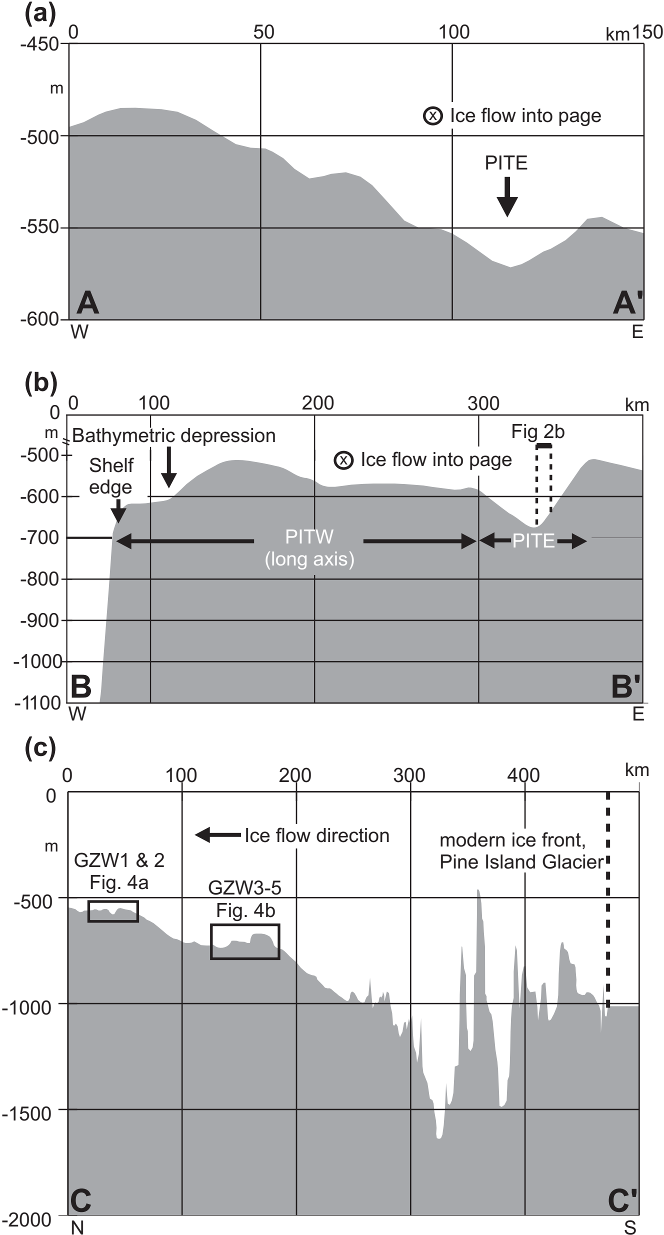
(a)
106°24'W
106°W
0 5 10 km
iceberg scours
GZW1
Fig. 3A
Crudely aligned
MSGLs
71°48'S
0 km 2 4
5 m
-560
W E
GZW2
Fig. 3A
(b)
0 2 4 km
106°48'W
106°36'W
iceberg scour
MSGLs
72°24'S
0 1 km
5 m
-710
NW SE
(c)
107°12'W
GZW4
Fig. 3C
MSGLs
0 4 8
-650
20 m
-700
N S
72°48'S
GZW5
Fig. 3C
0 2 4 km
106°48'W
(d)
MSGLs
0 km 1 2 3
-840
m
5 m
-850
W E
73°12'S
0 4 8 km
107°12'W
106°48'W



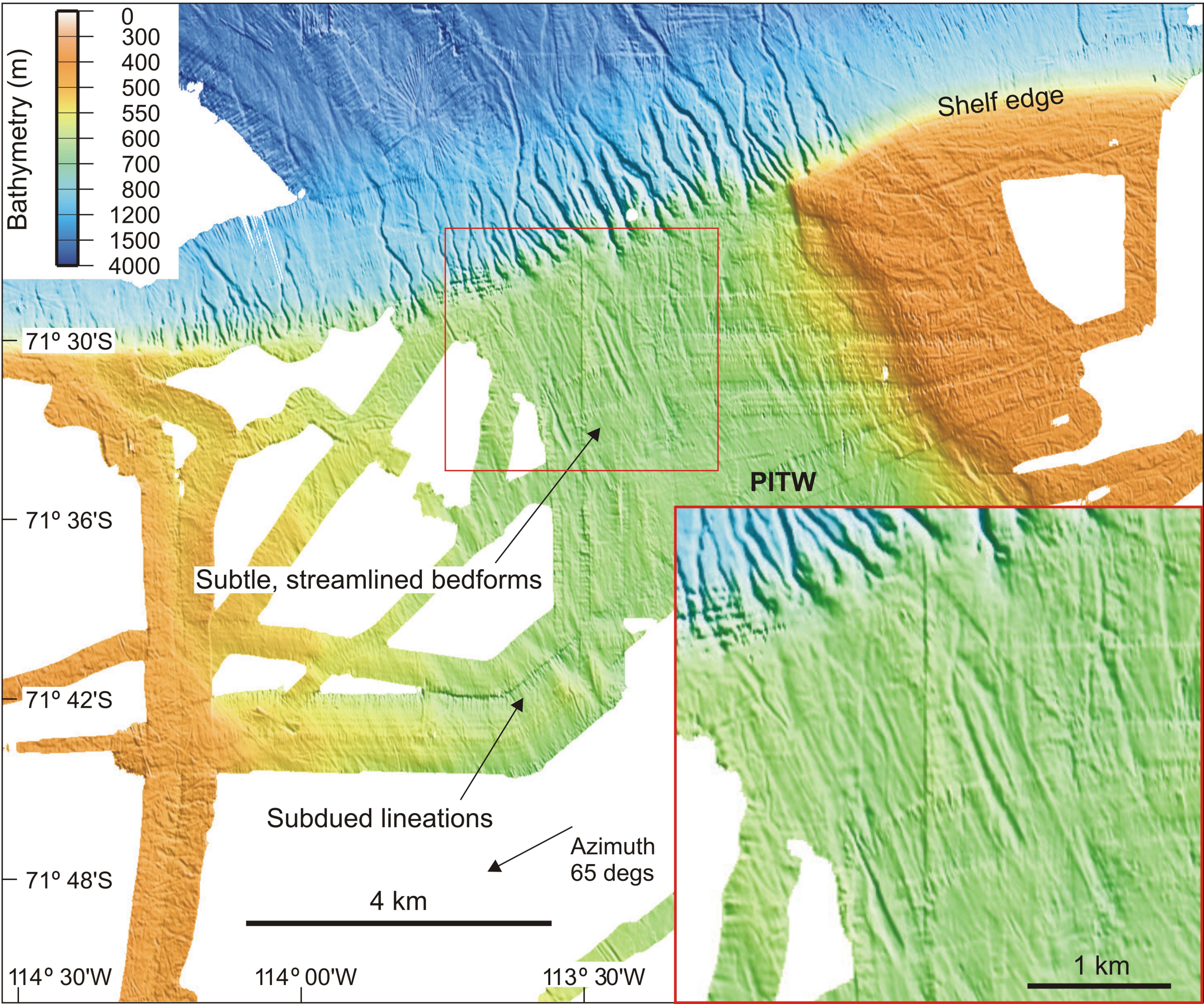
(a)
5 km
GZW1
Fig. 4A
W
S
-520
(m)
-600
Ice Flow Direction
VE 30:1
GZW2
Fig. 4A
(b)
-700
(m)
-760
Ice Flow Direction
GZW3
Fig. 4B
Parallel recessional
moraines
Ice Flow Direction
W
S
2 km
VE 30:1
(c)
iceberg scours
GZW5
Fig. 4B
6 km
S
W
Ice Flow Direction
-660
(m)
-760
GZW4
Fig. 4B
VE 30:1



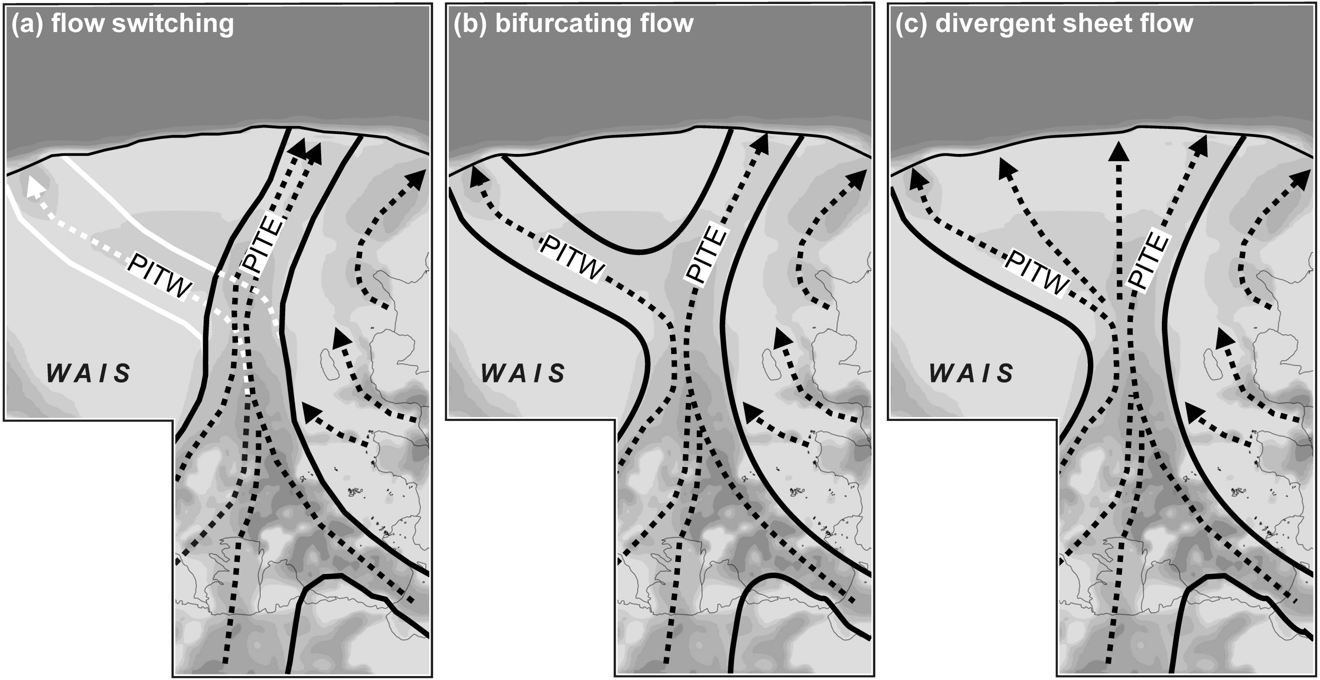
(a)
71°35.4'S
71°41.4'S
71°48'S
71°53.4'S
72°01'S
Direction of palaeo-ice flow
Palaeo-grounding line
Palaeo-grounding line
GZW1
GZW2
20 m
10 km
N
S
(b)
72°34'S
72°42'S
Palaeo-grounding line
72°51'S
Direction of palaeo-ice flow
Palaeo-grounding line
Proglacial debris flow
Palaeo-grounding line
Small moraines
GZW3
Till layer
GZW4
Till layer
GZW5
10 m
10 km
N
S



(a)
0
50
100
km 150
-450
m
-500
-550
-600
Ice flow into page
PITE
A
A'
W
E
(b)
0
100
200
300
km
0
m
-500
-600
-700
-800
-900
-1000
-1100
Bathymetric depression
Shelf edge
Ice flow into page
Fig 2b
PITW (long axis)
PITE
B
B'
W
E
(c)
0
100
200
300
400
km
0
m
-500
-1000
-1500
-2000
Ice flow direction
modern ice front, Pine Island Glacier
GZW1 & 2 Fig. 4a
GZW3-5 Fig. 4b
C
C'
N
S



Bathymetry (m)
0
300
400
500
550
600
700
800
1200
1500
4000
Shelf edge
71° 30'S
71° 36'S
71° 42'S
71° 48'S
PITW
Subtle, streamlined bedforms
Subdued lineations
Azimuth
65 degs
4 km
1 km
114° 30'W
114° 00'W
113° 30'W


(a) flow switching
PITW
PITE
WAIS
(b) bifurcating flow
PITW
PITE
WAIS
(c) divergent sheet flow
PITW
PITE
WAIS