Antarctic palaeo-ice streams

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ABSTRACT

We review the geomorphological, sedimentological and chronological evidence for palaeo-ice streams on the continental shelf of Antarctica and use this information to investigate basal conditions and processes, and to identify factors controlling grounding-line retreat. A comprehensive circum-Antarctic inventory of known palaeo-ice streams, their basal characteristics and minimum ages for their retreat following the Last Glacial Maximum (LGM) is also provided. Antarctic palaeo-ice streams are identified by a set of diagnostic landforms that, nonetheless, display considerable spatial variability due to the influence of substrate, flow velocity and subglacial processes. During the LGM, palaeo-ice streams extended, via bathymetric troughs, to the shelf edge of the Antarctic Peninsula and West Antarctica, and typically, to the mid-outer shelf of East Antarctica. The retreat history of the Antarctic Ice Sheet since the LGM is characterised by considerable asynchroneity, with individual ice streams exhibiting different retreat histories. This variability allows Antarctic palaeo-ice streams to be classified into discrete retreat styles and the controls on grounding-line retreat to be investigated. Such analysis highlights the important impact of internal factors on ice stream dynamics, such as bed characteristics and slope, and drainage basin size. Whilst grounding-line retreat may be triggered, and to some extent paced, by external (atmospheric and oceanic) forcing, the individual characteristics of each ice stream will modulate the precise timing and rate of retreat through time.

Antarctica; ice stream; grounding-line retreat; glacial geomorphology; deglacial history

1. INTRODUCTION

Ice streams are corridors of fast-flowing ice within an ice-sheet and are typically hundreds of kilometres long and tens of kilometres wide (Bennett, 2003). Their high velocities enable them to drain a disproportionate volume of ice and they exert an important influence on the geometry, mass balance and stability of ice sheets (e.g. Bamber et al. 2000; Stokes & Clark, 2001). Recent observations of ice streams in Antarctica and Greenland have highlighted their considerable spatial and temporal variability at short (sub-decadal) time-scales and include acceleration and thinning, deceleration, lateral migration and stagnation (Stephenson & Bindschadler, 1988; Retzlaff & Bentley, 1993; Anandakrishnan & Alley, 1997; Conway et al. 2002; Joughin et al. 2003; Shepherd et al. 2004; Truffer & Fahnestock, 2007; Rignot, 2008; Wingham et al. 2009). The mechanisms controlling the fast and variable flow of ice streams and the advance and retreat of their grounding lines are, however, complex (Vaughan and Arthern, 2007) and a number of potential forcings and factors have been proposed. These include: (i) oceanic temperature (Payne et al. 2004; Shepherd et al. 2004; Holland et al. 2008; Jenkins et al. 2010); (ii) sea-level changes (e.g. Hollin, 1962); (iii) air temperatures (Sohn et al. 1998; Zwally et al. 2002; Parizek & Alley, 2004; Howat et al. 2007; Joughin et al. 2008); (iv) ocean tides (Gudmundsson, 2007; Griffiths & Peltier, 2008, 2009); (v) subglacial bathymetry (Schoof, 2007); (vi) the formation of grounding zone wedges (Alley et al. 2007); (vii) the availability of topographic pinning points (Echelmeyer et al. 1991); (viii) the routing of water at the base of the ice sheet (Anandakrishnan and Alley, 1997; Fricker et al. 2007; Stearns et al. 2008; Fricker & Scambos, 2009); (ix) the ice stream's thermodynamics (Christoffersen and Tulaczyk, 2003a; b); and (x) the size of the drainage basin (Ó Cofaigh et al. 2008). Resolving the influence of each of these controls on any given ice stream represents a major scientific challenge and it is for this reason that there are inherent uncertainties in predictions of future ice sheet mass balance (IPCC, 2007; Vaughan and Arthern, 2007).

An important context for assessing recent and future changes in ice streams and the controls on their behaviour is provided by reconstructions of past ice stream activity. It has been recognised that ice streams leave behind a diagnostic geomorphic signature in the geologic record (cf. Dyke & Morris, 1988; Stokes & Clark, 1999) and this has resulted in a large number of palaeo-ice streams being identified, mostly dating from the last glacial cycle and from both marine (e.g. Shipp et al. 1999; Canals et al. 2000; Evans et al. 2005, 2006; Ó Cofaigh et al. 2002, 2005a; Ottesen et al. 2005; Mosola & Anderson, 2006; Dowdeswell et al. 2008a; Graham et al. 2009) and terrestrial settings (e.g. Clark & Stokes, 2001; Stokes & Clark, 2003, Winsborrow et al. 2004; De Angelis & Kleman, 2005, 2007; Ó Cofaigh et al. 2010a). The ability to directly observe the beds of palaeo-ice streams has also allowed scientists to glean important spatial and temporal information on the processes that occurred at the ice-bed interface and on the evolution of palaeo-ice streams throughout their glacial history.

Over the last two decades, there has been a burgeoning interest in marine palaeo-ice streams, particularly off the coast of West Antarctica and around the Antarctic Peninsula. This has focused primarily on identifying individual ice stream tracks in the geologic record and deciphering their geomorphic and sedimentary signatures to reconstruct their ice-flow history and the timing and rate of deglaciation (e.g. Wellner et al. 2001; Canals et al., 2000; Lowe & Anderson, 2002; Ó Cofaigh et al. 2002; Graham et al. 2009). In this paper, we aim to collate this information and provide a new and complete inventory of published accounts of Antarctic palaeo-ice streams. In synthesising the literature, we present an up-to-date chronology of the retreat histories of various ice streams and use the geomorphic evidence to elucidate the various 'styles' of ice-stream retreat (e.g. Dowdeswell et al. 2008; Ó Cofaigh et al. 2008). The processes that trigger and control the retreat of marine palaeo-ice streams remains a key research question in glaciology and one that has important implications for constraining future modelling predictions of contemporary ice-stream retreat and contributions to sea-level. By summarising the key characteristics of each Antarctic palaeo-ice stream, including their bathymetry and drainage basin area, geology and geomorphology, relationships between the inferred/dated retreat styles and the factors that control ice stream retreat are investigated. A further aim, therefore, is to provide a long term context for many present-day ice streams, which previously extended onto the outer continental shelf (e.g. Conway et al. 1999) and, crucially, provide spatial and temporal information on ice stream history for initialising and/or testing ice sheet modelling experiments (cf. Stokes & Tarasov, 2010).

2. PALAEO ICE-STREAM INVENTORY

Ice streams can be simply classified according to their terminus environment. Terrestrial ice streams terminate on land and typically result in a large lobate ice margin whereas marine-terminating ice streams flow into ice shelves or terminate in open water, where calving results in the rapid removal of ice and the maintenance of rapid velocities (cf. Stokes and Clark, 2001). With this classification in mind, all palaeo-ice streams in Antarctica (and, indeed, their contemporary cousins) were marine-terminating and, at the LGM, extended across the continental shelf with most of their main trunks below present sea level. Therefore, Antarctic palaeo-ice streams could be viewed as a sub-population of ice streams with specific characteristics.

Evidence for such marine palaeo-ice streams is based on the geomorphology of glacial landforms preserved in bathymetric troughs on the modern Antarctic shelf, which are identified from multibeam swath bathymetry and side-scan sonar data. This seafloor geomorphological data has been complemented by high resolution seismic studies of acoustic stratigraphy as well as sediment cores from which subglacial and glacimarine lithofacies have been both identified and dated. These techniques have enabled diagnostic geomorphological, sedimentological and geotechnical criteria of ice streaming to be identified (see Stokes & Clark, 1999, 2001). They include the presence of mega-scale glacial lineations (MSGL), abrupt lateral margins, evidence of extensively deformed till, focused sediment delivery to the ice stream terminus and characteristic shape and dimensions. Antarctic marine palaeo-ice streams are also located in cross-shelf bathymetric troughs (Wellner et al. 2006), often associated with grounding zone wedges (GZW) within the troughs (e.g. Mosola & Anderson, 2006) and occasionally associated with voluminous sediment accumulations, i.e. trough mouth fans (TMF), on the adjacent continental slope (e.g. Ó Cofaigh et al. 2003; Dowdeswell et al. 2008b).

The first glacimarine investigations in Antarctica utilised echosounder data, till petrographic studies and seismic data to reconstruct the expansion of grounded ice across the continental shelf during the last glaciation (e.g. Kellogg et al. 1979; Anderson et al. 1980; Orheim & Elverhøi, 1981; Domack, 1982; Haase, 1986; Kennedy & Anderson, 1989). The advent of hull-mounted and deep-tow side-scan sonar and especially multibeam swath bathymetry was a critical development for reconstructing palaeo-ice sheets because, for the first time, marine glacial geomorphic features could be easily observed and palaeo-ice streams identified (Pudsey et al. 1994; Larter & Vanneste, 1995; O'Brien et al. 1999; Shipp et al. 1999; Canals et al. 2000, 2002, 2003; Anderson & Shipp, 2001; Wellner et al. 2001; Ó Cofaigh et al. 2002, 2003, 2005a,b; Lowe & Anderson, 2003; Dowdeswell et al. 2004a,b; Evans et al. 2004, 2005; Heroy & Anderson, 2005). More recently, these datasets have culminated in the release of regional, high resolution (~1 km) bathymetric grids aggregated from existing depth soundings along the continental shelf (Nitsche et al. 2007; Bolmer, 2008; Graham et al. 2009, in press; Beaman et al. 2010). They provide an important morphological context and can be utilised as boundary conditions in numerical modelling experiments. Additionally, in order to improve and augment existing databases, a novel method of using mammal dive-depth data has recently been demonstrated (Padman et al., 2010).

In Table 1, we present the first comprehensive inventory of Antarctic palaeo-ice streams and the main lines of evidence that have been used in their identification. Figure 1 shows the location of each of these ice streams. This inventory includes palaeo-ice streams whose existence has been proposed in the literature on the basis of several lines of evidence, and which are fairly robust, but also more speculative palaeo-ice streams where there are distinctive cross-shelf bathymetric troughs. The majority of palaeo-ice streams are located in West Antarctica and the Antarctic Peninsula region, where most research on this topic has been conducted; and the associated geological evidence suggests that the ice sheet extended at least close to the continental shelf edge at the LGM (cf. Heroy & Anderson, 2005; Sugden et al. 2006) (Fig. 1). The western Ross Sea may be considered as an exception to this, because here the geological evidence indicates that the grounding lines of the former Drygalski and JOIDES-Central Basin ice streams only reached the outer shelf (Licht, 1999; Shipp et al. 1999; Anderson et al. 2002). It has to be kept in mind, however, that ice feeding into these two palaeo-ice streams was mainly derived from the East Antarctic Ice Sheet (e.g., Farmer et al. 2006). A paucity of marine geological data, from the southern Weddell Sea shelf specifically, means the ice extent at the LGM in that region is poorly-defined (e.g. Bentley & Anderson 1998). Diamictons recovered from cores and interpreted as tills (Fütterer & Melles, 1990; Anderson & Andrews 1999), in conjunction with terrestrial data constraining palaeo-ice-sheet elevation (Bentley et al. 2010), suggest that grounded ice extended locally onto the outer Weddell Sea shelf during the last glacial cycle. However, it is unclear whether the WAIS grounded in Ronne Trough at the LGM (Anderson et al. 2002), and there are conflicting conclusions about grounding of ice in Crary Trough (Fütterer & Melles, 1990; Anderson et al. 2002; Bentley et al. 2010).

The picture in East Antarctica is less clear, although from current evidence it is thought that ice expanded only as far as the mid to outer shelf (see Anderson et al. 2002 for a detailed overview). This is best demonstrated in Prydz Channel, where sea-floor topography in conjunction with sediment core stratigraphy constrain the maximum extent of the grounding line of the Lambert Glacier during the last glaciation to ca. 130 km landward of the shelf edge (Table 1) (Domack et al. 1998; O'Brien et al. 1999, 2007).

3. BASAL CHARACTERISTICS OF ANTARCTIC PALAEO ICE STREAMS

The basal conditions beneath ice streams are critical in controlling both the location and the flow variability of ice streams. By studying the former flow paths of ice streams, we can directly observe the ice stream bed at a variety of scales and can therefore acquire important information on basal conditions of the ice sheet, such as basal topography, bed roughness, geological substrate and sediment erosion, transport and deposition. In this section, we review the basal characteristics of Antarctic palaeo-ice streams in order to investigate possible substrate controls on ice stream flow and grounding line retreat. The following section then assesses the timing and rate of palaeo-ice stream retreat, using a new compilation of deglacial dates from around the Antarctic continental shelf.

3.1 Bathymetry and Drainage Basin

3.1.1 Theoretical and modelling studies

In the 1970s, the paradigm of marine ice sheet instability emerged with a number of theoretical studies. These studies identified the 'buttressing' effect of ice shelves as a critical control on the stability of ice-stream grounding lines. It was argued that removal of ice shelves from around the largely marine-based West Antarctic Ice Sheet (WAIS) could trigger catastrophic grounding-line retreat (Mercer, 1978; Thomas, 1979). Further retreat might, theoretically, be irreversible because the bed of the WAIS deepens inland (Weertman, 1974; Thomas & Bentley, 1978; Thomas, 1979). There are two elements to this theory and it is therefore useful to distinguish between the roles of: (i) ice-shelf butressing as a (de)stabilizing mechanism; and (ii) marine ice-sheet instability *sensu stricto*. Supporting evidence for both the 'marine ice-sheet instability hypothesis' and the critical importance of ice-shelf buttressing has been reported from the Amundsen Sea sector of the West Antarctic Ice Sheet (Shepherd et al. 2004), smaller glaciers on the Antarctic Peninsula (De Angelis & Skvarca, 2003; Rignot et al. 2004) and Jakobshavns Isbrae in the Greenland Ice Sheet (Joughin et al. 2004), all of which have accelerated and/or thinned following significant melting or collapse of buttressing ice shelves. Recent numerical ice-sheet modelling studies have also suggested that ice streams on reverse slopes are inherently unstable and can propagate the rapid collapse of an ice sheet (e.g. Schoof, 2007; Nick et al. 2009; Katz & Worster, 2010). Basal topography is, therefore, thought to exert a fundamental control on ice-stream and tidewater glacier stability (Vieli et al. 2001; Schoof, 2007; Nick et al. 2009; Katz & Worster, 2010). However, well-documented examples from the palaeo-record indicate that rapid grounding-line retreat does not necessarily occur on reverse slopes (Shipp et al. 2002; Ó Cofaigh et al. 2008; Dowdeswell et al. 2008a). This dichotomy indicates that existing models of ice stream retreat may be failing to capture the full complexity of grounding line behaviour, perhaps as a result of oversimplified boundary conditions (e.g. basal or lateral geometry), or as a result of limitations in the physical processes incorporated in the models (e.g. ice shelf buttressing or lateral and longitudinal stress components). One suggestion is that in the case of some ice streams that are grounded on reverse-sloping beds, resistive 'back stresses' afforded by pinning points, and 'side drag' by trough width and relief may exert a supplementary modulating effect on ice-stream stability (Echelmeyer et al. 1991, 1994; Whillans & van der Veen, 1997; Joughin et al. 2004). Indeed, modelling studies have demonstrated that ice shelves can act to stabilise the grounding line on a reverse slope (Weertman, 1974; Dupont, 2005; Walker, 2008; Goldberg, 2009). In addition, Gomez et al. (2010) demonstrate that gravity and deformation-induced sea-level changes local to the grounding-line can act to stabilize ice sheets grounded on reverse bed slopes. Basal friction is also an important component in the force balance of an ice stream (Alley, 1993a; MacAyeal et al. 1995; Siegert et al. 2004; Rippin et al. 2006) and bed roughness and the presence of 'sticky-spots', such as bedrock bumps, can exert a strong influence on ice-sheet dynamics (see Stokes et al. 2007 for a review).

3.1.2 Empirical evidence

Table 2 provides a synthesis of the key physiographic data of each palaeo-ice stream in our new inventory (see Fig. 1 and Table 1). All of the Antarctic palaeo-ice streams identified in the literature are topographically controlled, with landforms pertaining to fast-flow restricted to cross-shelf bathymetric troughs (e.g. Evans et al. 2005). This 'control' on ice stream location exposes a classic 'chicken-and-egg' situation, whereby it is hard to discern whether the palaeo-ice streams preferentially occupied pre-existing troughs, or whether the troughs formed as a consequence of focused erosion during streaming (cf. Winsborrow et al. 2010). Certainly, some palaeo-ice streams exhibit a strong tectonic control, such as the Gerlache-Boyd palaeo-ice stream, which flowed SW-NE along the Bransfield rift through the Gerlache Strait before turning sharply west into the Hero Fracture Zone across Boyd Strait (cf. Canals et al. 2000). However, it is clear that the cross-shelf bathymetric troughs were repeatedly occupied by ice streams over multiple glacial cycles (e.g. Larter & Barker, 1989, 1991; Barker, 1995; Bart et al. 2005) and would certainly have predisposed ice stream location in more recent glacial periods (ten Brink & Schneider, 1995).

It is also apparent from Table 2 that there is considerable spatial variation in physiography between Antarctic palaeo-ice streams. Lengths range between 70 and 400 km, widths from 5 to 240 km and drainage basin areas from 23,000 km^2 to 1.6 million km^2. This variability is demonstrated by the difference between the eastern Ross Sea palaeo-ice streams, which occupy very broad troughs (100-240 km) with low-relief intervening ridges (>500 m deep) (Mosola & Anderson, 2006) and the Gerlache-Boyd palaeo-ice stream, which is controlled by a deep (up to 1200 m) and narrow (5-40 km) trough, heavily influenced by the underlying geological structure (Canals et al. 2000, 2003; Evans et al. 2004; Heroy & Anderson, 2005). Thus, the Gerlache-Boyd palaeo-ice stream may expect to be influenced more by 'drag' from its lateral margins and topographic 'pinning points'. Indeed, this is supported by the geomorphic evidence, with Smith Island on the outer-shelf interpreted to have acted as a barrier to ice flow (Canals et al. 2003), while large bedrock fault scarps and changes of relief within the main trough are associated with thick wedges of till, which are therefore thought to have acted as pinning points (Heroy & Anderson, 2005). On the Pacific margin of the Antarctic Peninsula, Biscoe (Amblas et al. 2006), Anvers-Hugo Island (Pudsey et al. 1994;

Domack et al. 2006) and Smith (Pudsey et al. 1994) troughs are also disrupted by a narrow, elongate structural ridge (at ~300 m water depth) known as the "Mid-Shelf High" (Larter & Barker, 1991). A number of East Antarctic troughs, such as Astrolabe-Français, Mertz-Ninnes and Mertz troughs along Adelié Land, are also characterised by a shallower sill at the continental shelf edge (Beaman et al. 2010).

The majority of the Antarctic palaeo-ice streams retreated across reverse slopes (Table 2; Fig. 2) probably created by repeated overdeepening of the inner shelf by glacial erosion over successive glacial cycles (ten Brink & Schneider, 1995). The obvious exceptions to this are the central Bransfield Basin palaeo-ice streams (Lafond, Laclavere and Mott Snowfield), which exhibit steep normal slopes on the inner shelf and then dip gently towards the shelf-break (650-900 m) (Canals et al. 2002), whilst a number of the troughs have a seaward dipping outer shelf, such as Belgica Trough (Fig. 2c) (Hillenbrand et al. 2005; Graham et al. in press). On the outer shelf in Pine Island Bay, Graham et al. (2010) correlated phases of rapid retreat with steeper reverse bed-slopes (local average of -0.149°), while lower angled slopes (local average of -0.015°) have been associated with temporary still-stands and GZW formation. This observation lends credence to model experiments proposing sensitivity of ice streams to bed gradients (e.g. Schoof, 2007). However, and as noted above, the inferred slow retreat of some of the ice-streams since the LGM (e.g. JOIDES-Central Basin: Shipp et al. 1999; Ó Cofaigh et al. 2008) suggests that additional complexity exists.

In a comparison of four Antarctic palaeo-ice streams, Ó Cofaigh et al. (2008) proposed that drainage basin size could be a key control on ice-stream dynamics. Geomorphic evidence for slow retreat from the outer shelf of JOIDES-Central Basin is reconciled with two large drainage basins (1.6 million km^2 and 265,000 km^2) (Table 2) feeding the palaeo-ice stream from East Antarctica (Farmer et al. 2006; Ó Cofaigh et al. 2008). In contrast, rapid retreat of the Marguerite Bay palaeo-ice stream (Ó Cofaigh et al. 2002, 2005b, 2008; Kilfeather et al. 2010) is suggested to relate to the much smaller size of its drainage basin (10,000-100,000 km^2), which is likely to have been much more sensitive to external and internal forcing. Additionally, it is also likely that basal conditions, such as basal melting and freezing rates (e.g. Tulaczyk & Hossainzadeh, 2011), and climatic conditions, such as precipitation (e.g. Werner et al. 2001), were quite different between the Antarctic Peninsula and Ross Sea sectors, and therefore may have contributed to the different retreat histories.

While some palaeo-ice streams consist of just one central trunk (e.g. Lafond, Laclavere and Mott Snowfield: Canals et al. 2002), others have multiple tributaries (in an onset zone) that converge into a central trough on the mid-outer shelf (e.g. Robertson palaeo-ice stream: Evans et al. 2005; Getz-Dotson Trough: Graham et al. 2009, Larter et al. 2009; Gerlache-Boyd palaeo-ice stream: Canals et al. 2000, 2003; Evans et al. 2004; Biscoe palaeo-ice stream: Canals et al. 2003) (see Fig. 1; Table 2). In Robertson Trough, competing ice-flows from multiple tributaries (Prince Gustav channel, Greenpeace trough, Larsen-A & -B and BDE trough) have left behind a palimpsest geomorphic signature of up to four generations of cross-cutting MSGL, indicating switches in ice-flow direction (Camerlenghi et al. 2001; Gilbert et al. 2003; Evans et al. 2005; Heroy & Anderson 2005). Clearly, the characteristics of the ice stream's catchment area are likely to influence its behaviour in that an ice stream with several tributaries with different characteristics (e.g. bathymetry) might retreat in a fundamentally different way from one which has a single tributary. Such differences are an important consideration when attempting to predict the future behaviour of ice streams in Greenland and Antarctica and, undoubtedly, add considerable complexity when attempting to model the behaviour of ice streams and resolve subglacial topography in ice sheet models.

3.2 Geology/Substrate

Many contemporary ice streams have been shown to be underlain by a soft, dilatant deformable sediment layer (Alley et al. 1987; Blankenship et al. 1987; Engelhardt et al. 1990; Smith, 1997; Anandakrishnan et al. 1998; Engelhardt & Kamb, 1998; Kamb, 2001; Studinger et al. 2001; Bamber et al. 2006; King et al., 2009). However, there is still uncertainty surrounding the exact contribution of the deforming layer to ice stream motion (i.e. basal sliding vs. sediment deformation), the thickness of the deforming layer, and the till rheology (i.e. viscous or plastic) (Alley et al. 2001). This is complicated by the spatial and temporal variability in bed properties that can characterise ice stream beds and has led both palaeo and contemporary scientists to propose a 'mosaic' of basal sliding and deformation to reconcile the often conflicting sedimentary evidence (Alley, 1993b; Piotrowski & Kraus, 1997; Clark et al. 2003; Piotrowski, 2004; D.J.A. Evans et al. 2006; Smith & Murray, 2008; King et al. 2009; Reinardy et al. 2011b). Crucially, Antarctic palaeo-ice streams present a useful opportunity to integrate bed properties over large spatial scales, enabling more complete descriptions of substrate characteristics beneath ice streams and its importance in controlling ice stream flow and landform development.

The majority of palaeo-ice streams in West Antarctica are characterised by a transition from crystalline bedrock on the inner shelf to unconsolidated sedimentary strata on the middle and outer shelf (Shipp et al. 1999; Wellner et al. 2001, 2006; Lowe & Anderson, 2002, 2003; Ó Cofaigh et al. 2002, 2005a; Canals et al. 2002, 2003; Evans et al. 2004, 2005, 2006; Anderson & Oakes-Fretwell, 2008; Graham et al. 2009; Weigelt et al. 2009). It has been suggested that this transition is crucial in modulating the inland extent of ice streams (Anandakrishnan et al. 1998; Bell et al. 1998; Studinger et al. 2001; Peters et al. 2006) and this is supported by palaeo-landform models that show a geomorphic transition from inferred slow flow over bedrock, to drumlins at the zone of acceleration (corresponding to the crystalline bedrock-unconsolidated sediment transition) and then into the high velocities of the main ice stream trunk, as recorded by MSGL (Canals et al. 2002; Ó Cofaigh et al. 2002, 2005a; Shipp et al. 1999; Wellner et al. 2001; Evans et al. 2006; and see section 3.3.8). However, this relationship between substrate and ice velocities is complicated by observations of highly elongate bedforms within the zone of crystalline bedrock in the Marguerite Bay and Getz-Dotson troughs (Ó Cofaigh et al. 2002; Graham et al. 2009). Furthermore, the substrates of the palaeo-ice streams offshore of the Sulzberger Coast, in Smith Trough and in the upstream section of the Gerlache-Boyd palaeo-ice stream are primarily composed of crystalline bedrock, and spectacular parallel grooves (up to 40 km long) are incised into the bedrock (Canals et al. 2000; Wellner et al. 2001, 2006; Heroy & Anderson 2005). Indeed, a transition from stiff till on the inner shelf to deformation till on the outer shelf in Robertson Trough, East Antarctic Peninsula, has also been associated with a change in basal processes (from basal sliding to deformation) and an increase in ice velocity (Reinardy et al. 2011b). The time-dependent changes in freezing-melting and thermo-mechanical coupling between the ice and the underlying sediment will play an important role in modulating ice flow, bedform genesis and retreat rates and yet we only see a time-integrated subglacial imprint. Thus, given the limited information about subglacial sediments (i.e. in sediment cores), we can only really speculate about these processes from the palaeo-ice stream records.

Clearly, the underlying geology exerts an important control on the macro-scale roughness of an ice-stream bed, which influences the frictional resistance to ice flow, with rougher areas likely to act as 'sticky-spots' and reduce flow velocities. As a result, bed roughness of Antarctic palaeo-ice streams tends to increase inland, i.e. upstream towards the onset zone (Fig. 2; and also see Graham et al. 2009, 2010), and is therefore in accordance with radio-echo sounding evidence from below contemporary ice streams (Siegert et al. 2004). This change in roughness is typically driven by a transition from bedrock (inner shelf) to unconsolidated sediment (outer shelf) and therefore supports the notion that ice stream flow may be controlled by underlying geology and its roughness (Siegert et al. 2004, 2005; Bingham & Siegert, 2009; Smith & Murray, 2009; Winsborrow et al. 2010). Incidentally, it is surprising that so few studies have taken advantage of the now-exposed palaeo-ice stream beds to provide a more detailed assessment of subglacial roughness, similar to those that have been undertaken from sparse radio-echo-sounding flight-lines and localised studies from beneath the ice (e.g. Siegert at al., 2004).

3.2.1. Till characteristics and associated deposits

One of the recurrent features identified from geophysical investigations on the Antarctic continental shelf is an acoustically transparent sedimentary unit (Fig. 3) that is confined to cross-shelf troughs previously occupied by palaeo-ice streams and that underlies the post-glacial sedimentary drape (Ó Cofaigh et al. 2002, 2005a,b, 2007; Dowdeswell et al. 2004; Evans et al. 2005, 2006; Mosola & Anderson, 2006; Graham et al. 2009). This unit is underlain by a prominent subbottom reflector that ranges in thickness from 1-30 m, is typically associated with MSGL, and consists of soft (shear strengths typically <20 kPa), massive, matrix-supported diamicton (cf. Ó Cofaigh et al. 2007). The acoustically transparent unit comprises diamicton that has been interpreted as both a subglacial deformation till (Anderson et al. 1999; Shipp et al. 2002; Dowdeswell et al. 2004; Hillenbrand et al. 2005, 2009, 2010a; Ó Cofaigh et al. 2005a,b; Evans et al. 2005, 2006; Heroy & Anderson, 2005; Mosola & Anderson 2006; Graham et al. 2009; Smith et al. 2011) and as a "hybrid" till formed by a combination of subglacial sediment deformation and lodgement (Ó Cofaigh et al. 2007).

The geometry of the basal reflector underlying the acoustically transparent unit ranges from smooth and flat to irregular and undulating (Fig. 3) (Ó Cofaigh et al. 2005b, 2007; Evans et al. 2005, 2006). It has been hypothesised that an undulating basal reflector is indicative of an origin by grooving (Evans et al. 2006; Ó Cofaigh et al. 2007), whereby keels at the ice-sheet base (consisting of ice or rock) mobilise, erode and deform the underlying sediment (cf. Canals et al. 2000; Tulaczyk et al. 2001; Clark et al. 2003). In contrast, a smooth, flat, basal reflector is thought to result from the mobilization of underlying stiff till into a traction carpet of soft till and its advection downstream (Ó Cofaigh et al. 2005b, 2007). Penetration of the subbottom reflector by sediment cores reveals a much stiffer (>98 kPa in Marguerite Bay) and less porous, massive and matrix-supported diamicton (Shipp et al. 2002; Dowdeswell et al. 2004; Ó Cofaigh et al. 2005b, 2007; Evans et al. 2005, 2006; Mosola & Anderson, 2006; Graham et al. 2009), which has been either interpreted as a lodgement till (Wellner et al. 2001; Shipp et al. 2002) or a 'hybrid' lodgement-deformation till (Ó Cofaigh et al. 2005b, 2007; Evans J. et al. 2005; Evans D.J.A. et al. 2006; Reinardy et al. 2011a). The genesis of the overlying soft till is thought to result from reworking of underlying stiff till and pre-existing sediments (Evans et al. 2005; Ó Cofaigh et al. 2005b, 2007; Hillenbrand et al. 2009; Reinardy et al. 2009, 2011a).

Geotechnical and micromorphological evidence (Ó Cofaigh et al. 2005b, 2007; Reinardy et al. 2011a) from troughs on the shelf east and west of the Antarctic Peninsula indicates that shear is concentrated within discrete zones between the stiff and soft till, up to 1.0 m thick. This implies that deformation is not pervasive throughout the soft till (Ó Cofaigh et al. 2005b, 2007; Reinardy et al. 2011a). Nonetheless, geophysical evidence for large-scale advection of the soft till implies that these localised shear zones can integrate to transport significant volumes of sediment beneath palaeo-ice streams (Ó Cofaigh et al. 2007; cf. Hindmarsh, 1997, 1998). Such transport is also manifest in the formation of substantial depocentres, both in the form of grounding-zone wedges on the shelf and trough mouth fans on the continental slope (Larter & Vanneste, 1995; Bart et al. 1999; Shipp et al. 2002; Canals et al. 2003; Ó Cofaigh et al. 2003; Mosola & Anderson, 2006; Dowdeswell et al. 2008b).

Deglacial sediment facies can provide important information on the style of retreat and depositional processes occurring at the grounding line. The thickness of the deglacial sediment unit has been used as a crude proxy for the retreat rate, with its absence or thin units, such as those from Marguerite Trough (typically <0.7 m) and troughs in the central and eastern Ross Sea (<1.0 m), suggesting rapid retreat of the palaeo-ice streams (Ó Cofaigh et al. 2005b, 2008; Mosola & Anderson, 2008). However, these authors acknowledge that sediment supply and bathymetric configuration also play important roles in controlling the deposition of deglacial sediments (cf. Leventer et al. 2006). Thus, a 'deglacial unit' may appear thick in a trough area, where an ice-shelf could be sustained for a long time (e.g. because of available pinning points or in an embayment) which may be completely unrelated to the retreat rate of the grounding-line. Conversely, deglacial (and open-marine) sediments in Belgica Trough are extremely thin, suggesting a rapid retreat, but this is contradicted by the bedform evidence and radiocarbon chronology (Ó Cofaigh et al. 2005a; Hillenbrand et al. 2010a). There, current-induced winnowing is apparently responsible for a relatively thin postglacial sediment drape on the outer shelf (Hillenbrand et al. 2010a).

Many Antarctic palaeo-ice streams may have terminated in ice shelves during deglaciation (e.g. Pope & Anderson, 1992; Domack et al. 1999; Pudsey et al. 1994, 2006; Kilfeather et al. 2010). These include the Gerlache-Boyd system, Marguerite Trough, Belgica Trough, the Ross Sea troughs, Robertson Trough, Anvers Trough, Nielsen Basin and Prydz Channel (e.g. see Willmott et al. 2003). The corresponding sediments comprise glacimarine diamictons and/or a granulated facies (consisting of pelletized sandy-muddy gravel) typically overlain by mud (sometimes laminated), interpreted to record rainout of sediment from the base of an ice-shelf (Pudsey et al. 1994, 2006; Licht et al. 1996, 1998, 1999; Domack et al. 1998, 1999, 2005; Harris & O'Brien, 1998; Evans & Pudsey, 2002; Brachfeld et al. 2003; Evans et al. 2005; Ó Cofaigh et al. 2005, Hillenbrand et al. 2005, 2009, 2010a,b; Kilfeather et al. 2010; Smith et al. 2011). Ice shelves may play an important role in buttressing and therefore stabilizing the flow of marine palaeo-ice streams on a foredeepened bed (e.g. Dupont & Alley, 2005, 2006; Goldberg et al. 2009), with their reduction or loss capable of inducing rapid acceleration and collapse of the grounded ice (e.g. Rignot et al. 2004; Scambos et al. 2004). It is, therefore, important to identify the former presence of ice shelves when reconstructing the history of palaeo-ice streams and constraining modelling experiments. In Gerlache-Boyd Strait, for example, Willmott et al. (2003) used the great thickness (6-70 m) of deglacial and post-glacial sediment to infer the retreat history of the ice stream. They assumed that sedimentation rates are uniform along the trough and argued that the thickest deposits in Western Bransfield Basin are thought to record the earliest decoupling of ice. In contrast, the confined setting of the Gerlache Strait, where the postglacial sediment drape is negligible, is thought to have helped sustain the presence of an ice stream for a longer period (Willmott et al. 2003). This reconstruction was based on the assumption that sedimentation rates are uniform along the trough.

3.3 Geomorphology

As noted above, Antarctic palaeo-ice streams exhibit a number of characteristic landforms, some of which have recently been observed beneath a modern-day West Antarctic ice stream (e.g. Smith & Murray, 2008; King et al. 2009). These features are summarised in Table 3 and described in detail below in order to illustrate the range of landforms that are associated with ice stream flow and their implications for subglacial processes.

3.3.1 *Mega-scale glacial lineations*

Mega-scale glacial lineations (MSGLs) are present on all Antarctic palaeo-ice stream beds with the exception of Smith Trough and Sulzberger Bay Trough, which are dominated by parallel bedrock grooves (see Tables 1 and 3) (e.g. Shipp et al. 1999, 2002; Canals et al. 2000, 2002, 2003; Anderson et al. 2001; Wellner et al. 2001; Ó Cofaigh et al. 2002, 2005a,b; Lowe & Anderson, 2002, 2003; Dowdeswell et al. 2004; Evans et al. 2004, 2005, 2006; Graham et al. 2009, 2010). It is hard to discern whether MSGLs are diagnostic of ice streams or whether they are only preserved in the troughs and overprinted on the shallower shelf beyond the trough margins by iceberg scours. MSGLs comprise parallel sets of grooves and ridges, with elongation ratios >10:1 (and up to ~90:1), formed in soft, dilatant till (Fig. 4a) (cf. Wellner et al. 2006 for a review). Crest-to-crest spacings are typically 200-600 m (mode: 300 m), with widths and lengths up to 500 m and 100 km respectively and amplitudes of 2-20 m (cf. Heroy & Anderson, 2005; Wellner et al. 2006), although considerable intra-ice stream variability exists and a 'single' set of MSGL may have individual lineations of varying sizes, although it is often not clear whether lineations of different age are preserved. Heroy & Anderson (2005) discuss two outliers, which do not fit this morphometric categorisation of MSGLs: Biscoe Trough has MSGLs with crest spacings of >1 km and the 'bundle structures' in the Gerlache-Boyd palaeo-ice stream have crest spacings of 1-5 km and amplitudes of up to 75 m (Canals et al. 2000, 2003; Heroy & Anderson, 2005). According to Heroy & Anderson (2005), the unusual size of the Gerlache-Boyd palaeo-ice stream MSGLs are thought to have resulted from groove-ploughing (Clark et al. 2003) associated with the bedrock structure of the trough and large changes in relief. Although primarily associated with a soft sedimentary substrate typically found on the outer West Antarctic continental shelf, MSGLs from the Gerlache-Boyd palaeo-ice stream emanate from bedrock highs, and upstream portions (~9 km) of the lineations are composed of bedrock (Canals et al. 2000; Clark et al. 2003; Heroy & Anderson, 2005). This link between MSGL initiation and bedrock highs has similarly been observed in Biscoe Trough (Amblas et al. 2006) and also on the northern Norwegian shelf (Ottesen et al. 2008).

3.3.2 *Grooved, gouged and streamlined bedrock*

Where the inner and mid shelf is composed of rugged crystalline bedrock, palaeo-ice streams often preferentially erode the underlying strata and create a gouged, grooved and streamlined submarine landscape (Fig. 4b) (Anderson et al. 2001; Wellner et al. 2001, 2006; Lowe & Anderson, 2002, 2003; Gilbert et al. 2003; Evans et al. 2004, 2005; Heroy & Anderson, 2005; Ó Cofaigh et al. 2005a,b; Amblas et al. 2006; Domack et al. 2006; Anderson & Oakes-Fretwell, 2008; Graham et al. 2009; Larter et al. 2009). Grooves and gouges tend to be concentrated along the axis of glacial troughs and reach lengths of >40 km with spacing of less than 10 m to over 1 km and amplitudes of a few metres up to >100 m (Wellner et al. 2006). Grooves with similar dimensions have been infrequently observed in terrestrial settings in the northern hemisphere (e.g. Jansson et al. 2003; Bradwell, et al. 2007, 2008) and these have typically been linked to the onset of streaming flow (cf. Bradwell et al. 2008). A genetic distinction must be applied between MSGLs formed in soft sediment and bedrock grooves, which can also exhibit elongation ratios >10:1. Grooved, gouged and streamlined bedrock are erosional landforms probably controlled, at least in part, by the underlying structural geology, as illustrated by the way the grooves often follow bedrock structures. Their association with palaeo-ice streams implies fast, wet-based ice flow (promoting high erosion rates) as a prerequisite for their genesis. Graham et al (2009) also suggest that, given typical erosion rates cited in the literature (e.g. Hallet et al. 1996; Koppes & Hallet, 2006; Laberg et al. 2009), the larger bedrock features would require high erosion rates over a sustained period of time. Thus, heavily eroded bedrock exhibiting grooving and streamlining could potentially indicate a legacy of repeated ice streaming over several glacial cycles or persistent ice-streaming during a single glacial cycle.

3.3.3 *Drumlinoid bedforms*

‘Drumlinoid’ bedforms, which in this paper encompass both bedrock (including roches moutonées and whalebacks) and sediment cored structures (though see Stokes et al., 2011), are commonly found clustered on the inner shelf in crystalline bedrock and at the transition between this bedrock and unconsolidated sediment further out on the shelf (Fig. 4c and Table 3) (Anderson et al. 2001; Camerlenghi et al. 2001; Wellner et al. 2001, 2006; Canals et al. 2002; Ó Cofaigh et al. 2002, 2005a,b; Lowe & Anderson, 2002; Gilbert et al. 2003; Evans et al. 2004, 2005; Heroy & Anderson, 2005; Domack et al. 2006; Mosola & Anderson, 2006; Graham et al. 2009; Larter et al. 2009). Drumlins observed on the inner Antarctic continental shelf are principally formed in bedrock, although not ubiquitously as demonstrated by sediment cored drumlins in Eltanin Bay (upstream section of Belgica Trough) and the central Ross Sea troughs (Wellner et al. 2001, 2006). At the bedrock-sediment transition, crag-and-tail bedforms are also prevalent (e.g. directly offshore from the Getz Ice Shelf and in Marguerite Bay) with their stoss ends formed in bedrock and their attenuated tails grading into sediment (Wellner et al. 2001, 2006; Ó Cofaigh et al. 2002; Heroy & Anderson, 2005; Graham et al. 2009).

The formation of drumlins and crag-and-tail landforms at this substrate transition has been related to accelerating/extensional flow at the onset of an ice stream (Wellner et al. 2001, 2006; Mosola & Anderson, 2006). Many drumlins are associated with crescentric overdeepenings around their stoss sides (e.g. Wellner et al. 2001, 2006; Ó Cofaigh et al. 2002, 2005a,b; Lowe & Anderson, 2002; Gilbert et al. 2003; Heroy & Anderson, 2005; Graham et al. 2009) and these are generally thought to result from localised meltwater production, possibly due to pressure melting (cf. Wellner et al. 2001; Ó Cofaigh et al. 2002, 2005a,b, 2010b). The role of meltwater in the formation of drumlins is, however, contentious (e.g. see Shaw et al. 2008; Ó Cofaigh et al. 2010b), with the geomorphic evidence failing to reconcile whether crescentric overdeepenings formed synchronously with, or subsequent to, drumlin genesis (cf. Ó Cofaigh et al. 2010b).

3.3.4 *Grounding Zone Wedges*

Grounding zone wedges (‘till deltas’), characterised by a steep distal sea-floor ramp and shallow back-slope are common features of Antarctic palaeo-ice stream troughs (Table 3 and Fig. 4d) (Larter & Vanneste, 1995; Vanneste & Larter, 1995; Anderson, 1997, 1999; Bart & Anderson, 1997; Domack et al. 1999; O’Brien et al. 1999; Shipp et al. 1999, 2002; Lowe & Anderson, 2002; Canals et al. 2003; Howat & Domack, 2003; Evans et al. 2005; Heroy & Anderson, 2005; Ó Cofaigh et al. 2005a,b, 2007; McMullen et al. 2006; Mosola & Anderson, 2006; Graham et al. 2009, 2010). They are composed of diamicton and are typically tens of kilometres long and tens of meters high (with GZWs in the Ross Sea up to 100 m high) with an acoustic signature which often includes inclined structures truncated by a gently dipping overlying reflector and capped by an acoustically transparent sedimentary unit, similar in geometry to Gilbert-style deltas (e.g. Alley et al. 1989; Domack et al. 1999; Shipp et al. 1999; Ó Cofaigh et al. 2005a). These features are interpreted as grounding zone wedges (GZWs) that are generally thought to have formed by the subglacial transport and then deposition of deformation till at the grounding-line during ice stream still-stands (Alley et al. 1989; O’Brien et al. 1999; Anandakrishnan et al. 2007). Whilst the capping unit reflects the direct emplacement of basal till at the ice stream bed, the inclined structures may relate to glacimarine sedimentation proximal to the grounding line, in particular deposition from sediment gravity flows.

An alternative interpretation presented by Christoffersen et al. (2010) highlights the role of basal freezing in the entrainment of sediment into basal ice layers (cf. Christoffersen & Tulaczyk, 2003). During stagnant phases of ice stream cycles, sediment is accreted in the ice via basal freezing, while during subsequent phases of fast ice-stream flow the sediment is transported to the grounding line and GZWs formed by melt-out of this basal debris (Christoffersen et al. 2010). MSGLs are commonly formed on the GZW surface, thereby demonstrating the persistence of streaming flow during the last phase of their formation (e.g. Ó Cofaigh et al. 2005a; Graham et al. 2010). Where MSGLs terminate at the wedge crest, the GZW is interpreted to have formed during episodic retreat of the ice stream (e.g. Ó Cofaigh et al. 2008). In contrast, GZWs completely overridden by MSGL, such as in outer Marguerite Trough, obviously document an advance of the ice stream over the wedge (Ó Cofaigh et al. 2005b). A paucity of MSGLs across the surface of a prominent GZW in Trough 5 of the Ross Sea combined with the presence of intervening morainal ridges gives evidence for a slow phase of ice-stream retreat (Mosola & Anderson, 2006).

The formation and size of GZWs is likely to be a function of sediment supply vs. duration of time the ice was grounded at the same position (Alley et al. 2007). Calculated subglacial sediment fluxes from modelled, palaeo- and contemporary ice streams generally range between 100 and 1000 m^3 yr^{-1} per meter wide (Alley et al. 1987, 1989; Hooke & Elverhøi, 1996; Tulaczyk et al. 2001; Shipp et al. 2002; Bougamont & Tulaczyk, 2003; Dowdeswell et al. 2004a; Anandakrishnan et al. 2007; Laberg et al. 2009; Christoffersen et al. 2010), although fluxes as high as 8000 m^3 yr^{-1} per meter wide have been estimated for the Norwegian Channel Ice Stream (Nygård et al. 2007) and several 10s of thousands m^3 yr^{-1} per meter width for the M'Clintock Channel Ice Stream in the Canadian Arctic (Clark and Stokes, 2001). In order to generate and sustain this magnitude of sediment flux, subglacial transport must be dominated by till deformation distributed over a considerable thickness (>10 cm) (e.g. Alley et al. 1989; Bougamont & Tulaczyk, 2003; Anandakrishnan et al. 2007) or via debris-rich basal ice layers (Christoffersen et al. 2010). Given these likely range of estimates, GZWs tens of km's long and tens of metres thick would typically take ~100-10,000 years to form (e.g. Alley et al. 1987, 1989; Anandakrishnan et al. 2007; Larter & Vanneste 1995; Graham et al. 2010).

Significantly, it has also been proposed that sedimentation at the grounding line and resultant GZW formation could cause temporary ice stream stabilization against small sea-level rises (Alley et al. 2007) similar to processes observed at tidewater glacier termini (Powell et al. 1991). This could act as a positive feedback mechanism with the proto-formation of a GZW stabilising the grounding line and therefore promoting further sediment deposition.

3.3.5 *Moraines*

In contrast to large GZWs, relatively small transverse ridges composed of soft till, with amplitudes of 1-10 m, spacings of a few tens to hundreds of metres, and overprinted by MSGLs have been identified in the JOIDES-Central Basin and troughs of the eastern Ross Sea (Table 3) (Shipp et al. 2002; Mosola & Anderson, 2006; Ó Cofaigh et al. 2008; Dowdeswell et al. 2008). In JOIDES-Central Basin, the ridges are found everywhere except the inner-most shelf and are typically 1-2 m high, symmetrical, closely spaced and straight crested (Shipp et al. 2002). In the eastern Ross Sea, the ridges are larger, straight to sinuous in plan form, and with orientations that are oblique or transverse to ice flow (Fig. 4e) (Mosola & Anderson, 2006). These smaller transverse ridges presumably reflect lower volumes of sediment transported to the grounding line and/or, possibly, lower ice velocities (i.e. slow retreat recessional moraines).

These ridges are interpreted as time-transgressive features formed at the grounding line by deposition and/or sediment pushing during minor grounding-line re-advances and stillstands, possibly on an annual cycle (Shipp et al. 2002). Their formation is thus consistent with De Geer moraine formation as identified in other marine-ice sheet settings (e.g. Ottesen & Dowdeswell, 2006; Todd et al. 2007). If the ridges were annually deposited, the retreat rate of the ice stream in JOIDES-Central Basin would have been ca. 40-100 m yr-1, which is consistent with independent dating controls (Domack et al. 1999; Shipp et al. 2002). In Lambert Deep, transverse ridges with scalloped edges have also been identified as push moraines formed during minor re-advances (O'Brien et al. 1999). Similar ridges have been identified in Prydz Channel (Table 3), although these features wedge out against the sides of flutes, are parallel between flutes, and display a convex geometry landward (O'Brien et al. 1999). These features are interpreted as sediment waves formed by ocean circulation in a sub-ice shelf cavity that had formed immediately after the floating of the formerly grounded ice sheet (O'Brien et al. 1999). In Pine Island Trough, transverse ridges are observed along the entire width of the trough, with amplitudes of 1-2 m and wavelengths of 60-200 m (Jakobsson et al. 2011). These bedforms are referred to as 'fishbone moraine' and are interpreted to have formed during the distintegration of an ice shelf. Each ridge is thought to represent one tidal cycle in which the remnant ice shelf lifted, moved seaward and then subsided onto the sea floor, squeezing sediment out to form a ridge (Jakobsson et al. 2011). A similar tidal mechanism has been proposed for ridges formed within iceberg scours in the Ross Sea (Wellner et al. 2006).

3.3.6 *Subglacial meltwater drainage networks*

The distribution and flow of water beneath an ice sheet is an important control on ice dynamics. This is demonstrated by observations that the water pressure beneath Whillans Ice Stream, West Antarctica, is almost at flotation point (Engelhardt & Kamb, 1997; Kamb, 2001). Basal lubrication can promote fast ice streaming by lowering the effective pressure, either within a soft, dilatant till layer, thus permitting deformation and/or sliding along the surface (e.g. Alley et al. 1986, 1987, 1989b; Engelhardt et al. 1990; Engelhardt & Kamb, 1997; Tulaczyk et al. 2000); or in association with a spatially extensive subglacial hydraulic network, which can develop on both hard and soft beds. The form of this drainage network has been variously described as a thin film (Weertman, 1972), a linked-cavity system (Kamb,

1987) and a channelized system of conduits incised either into the ice, sediment or bedrock (Rothlisberger, 1972; Nye, 1976; Alley, 1989; Hooke, 1989; Clark & Walder, 1994; Walder & Fowler, 1994; Fountain & Walder, 1998; Ng, 2000; Domack et al. 2006). In Antarctica, the drainage network is likely to be further modulated by the routing of water from active subglacial lakes beneath the ice sheet (Fricker et al. 2007; Stearns, 2008; Carter et al. 2009; Smith et al. 2009). There has been a growing recognition that the subglacial hydrodynamics of the ice sheet system exhibits considerable spatial and temporal variability (Kamb, 2001; Wingham et al. 2006; Fricker et al. 2007), which is also consistent with recent observations from beneath Rutford Ice Stream (Smith et al. 2007; Murray et al. 2008; Smith & Murray, 2008). Palaeo-ice stream beds provide a useful opportunity to describe the form, type and size of subglacial meltwater networks and to examine their evolution both downflow, and over both soft and hard substrates.

Extensive networks of relict subglacial meltwater channels and basins incised into crystalline bedrock have been identified on the inner shelf sections of Anvers-Hugo Island Trough, Marguerite Trough, Pine Island Trough and Dotson-Getz Trough (Anderson & Shipp, 2001; Anderson et al. 2001; Ó Cofaigh et al. 2002, 2005b; Lowe & Anderson, 2003, 2003; Domack et al. 2006; Anderson & Oakes-Fretwell, 2008; Graham et al. 2009; Nitsche & Jacobs 2010). There is also evidence of localised water flow from crescentric overdeepenings around the stoss ends of drumlins (section 3.3.3). Relict hydrological networks associated with palaeo-ice streams are shown to exhibit variable forms, with the rugged bedrock of the innermost shelf characterised by large isolated basins (e.g. Marguerite Bay, Anderson & Oakes-Fretwell, 2008), some of which have been interpreted as former subglacial lakes (e.g. Palmer Deep, Anvers-Hugo Island Trough, Domack et al. 2006).

Also present on the inner shelf of Marguerite and Pine Island troughs are tunnel valleys and anastomosing channel-cavity systems (Fig. 4f), which tend to follow the deepest portions of the bed, possibly along structural weaknesses and indicate a well organised subglacial drainage network (Lowe & Anderson, 2002, 2003; Domack et al. 2006; Anderson & Oakes-Fretwell, 2008; Graham et al. 2009). The largest tunnel valleys are up to 25 km long, 4.5 km wide and incise up to 450 m into the underlying substrate (Graham et al. 2009). Lowe & Anderson (2002, 2003) show that the anastomosing network in Pine Island Bay breaks down seaward into a dendritic channel system more aligned to the inferred former ice-flow direction. This progressive evolution and organisation in subglacial meltwater flow seems to be analogous to channel networks along palaeo-ice stream beds elsewhere in West Antarctica (Domack et al. 2006; Anderson & Oakes-Fretwell, 2008). Isolated straight and radial channels also occur across bedrock highs and along the flanks of basins (e.g. Anderson & Oakes-Fretwell, 2008). It has been suggested that, similar to the erosional bedforms on the inner shelf (section 3.3.2), the subglacial meltwater channel networks may have formed over multiple glacial cycles, possibly since the Mid-Miocene (Lowe & Anderson 2003; Smith et al. 2009).

In contrast to the discrete subglacial meltwater channel networks incised into the crystalline bedrock of the inner shelf, evidence of meltwater flow across the sedimentary substrate of the middle to outer shelf is largely absent or undetectable. Possible exceptions include a meltwater channel and small braided channels at the mouth of Belgica Trough (Noormets et al. 2009) and a large tunnel valley in Pennell Trough, western Ross Sea (Wellner et al. 2006). Gullies and channels on the continental slope in-front of the glacial troughs (Table 3) were interpreted to have formed by the drainage of sediment-laden meltwater from ice grounded at the shelf break (Wellner et al. 2001, 2006; Canals et al. 2002; Dowdeswell, et al. 2004b, 2006, 2008b; Evans et al. 2005; Heroy & Anderson, 2005; Amblas et al. 2006; Noormets et al. 2009) and therefore may demonstrate the evacuation of meltwater from the substrate. However, erosion of these gullies and channels solely by turbidity current activity and/or the down-slope cascading of dense shelf water masses has also been proposed (e.g., Michels et al. 2002; Dowdeswell et al. 2006, 2008b; Hillenbrand et al. 2009; Muench et al. 2009), whilst the role of groundwater outflow at the continental slope may also be significant (Uemura et al. 2011).

3.3.7 *Trough Mouth Fans*

Trough Mouth Fans (TMFs) are large sedimentary depo-centres on the continental slope and rise, located directly offshore of the mouth of palaeo-ice stream troughs (Vorren & Laberg et al. 1997). They form over repeated glacial cycles due to the delivery of large volumes of glacigenic sediment from the termini of fast flowing ice streams grounded at the shelf-break. TMFs on the Antarctic continental margin have been identified from seaward bulging bathymetric contours, large glacigenic debris-flow deposits, and pronounced shelf progradation observed in seismic profiles (e.g. Bart et al. 1999; Ó Cofaigh et al. 2003; Dowdeswell et al. 2008b; O'Brien et al. 2007).

Compared to the northern hemisphere (e.g. Dowdeswell et al. 1996; Vorren et al. 1989, 1998; Vorren & Laberg, 1997), TMFs are relatively rare around the continental margin of Antarctica and, to date, have only been recognized at four localities (Table 3): Northern Basin TMF in the western Ross Sea (Bart et al. 2000), Belgica TMF in the southern Bellingshausen Sea (Ó Cofaigh et al. 2005a; Dowdeswell et al. 2008b), Crary TMF in the southern Weddell Sea (Kuvaas & Kristoffersen, 1991; Moons et al. 1992; Bart et al. 1999) and Prydz Channel Fan (Kuvaas & Leitchenkov, 1992; O'Brien 1994, 2007. In contrast, most sections of the Antarctic margin are dominated by gullies and channels eroded either by meltwater and/or dense shelf water flowing down-slope (see section 3.3.6), or by turbidity currents originating in debris flows (e.g. Dowdeswell et al. 2004b, 2006; Hillenbrand et al. 2009, Noormets et al. 2009), with debris-flow frequency depending on glacigenic sediment supply, shelf width and, crucially, the gradient of the continental slope (Ó Cofaigh et al. 2003). One explanation proposed to explain the absence of TMFs along many areas of the Antarctic margin is that the relatively steep slopes promote rapid down-slope sediment transfer by turbidity currents resulting in sediment bypass of the upper slope, thereby precluding formation of debris flow dominated TMFs by facilitating development of a gully/channel system (Ó Cofaigh et al. 2003). However, there is a 'chicken and egg' problem to this interpretation with TMFs typically creating shallow slopes, whereas the surrounding continental margin may have much steeper slopes.

3.3.8 *Impact of contrasting retreat rates on ice stream geomorphology*

The genetic association between subglacial bedforms and processes (see Section 3.3) has allowed the rate of palaeo-ice stream retreat to be inferred from their geomorphic imprint. Three distinctive suites of landform assemblage, each of which represents a characteristic retreat style (rapid, episodic and slow) have been proposed, see Fig. 5 (Dowdeswell et al. 2008a; Ó Cofaigh et al. 2008).

Palaeo-ice streams characterised by the preservation of unmodified MSGLs that have not been overprinted by other glacial features and with a relatively thin deglacial sedimentary unit (Fig. 5) (e.g. Marguerite Trough) are consistent with rapid deglaciation. In contrast, a series of transverse recessional moraines and GZWs on the palaeo-ice stream bed indicate slow and episodic retreat, respectively (Fig. 5). In particular, De Geer-style moraines are diagnostic of slow retreat, with each ridge possibly representing an annual stillstand, such as in the JOIDES-Central Basin (Shipp et al. 2002; Dowdeswell et al. 2008a; Ó Cofaigh et al. 2008). When grounding-line retreat was slow, it is likely that a thick deglacial sequence, including sub-ice shelf sediments, would have been deposited, although this is dependent on sedimentation rates (e.g. Domack et al. 1999; Willmott et al. 2003; Ó Cofaigh et al. 2008).

3.3.9 Marine palaeo-ice stream landsystem model

The general distribution of glacial landforms associated with a typical palaeo-ice stream on the Antarctic continental shelf (discussed throughout Section 3.3) is summarised in Fig. 6. This landsystem model (cf. Graham et al. 2009) illustrates the different glacial bedforms associated with crystalline bedrock and unconsolidated sediments and the seaward transition of glacial features and their inferred relative velocities (see also models presented by Canals et al. 2002; Wellner et al. 2001, 2006). Models initially highlighted the general down-flow evolution of bedforms associated with a corresponding increase in velocity (Ó Cofaigh et al. 2002), especially marked at the boundary between crystalline bedrock and sedimentary substrate (e.g. Wellner et al. 2001). More recent attempts to produce a conceptual model of the palaeo-ice stream landsystem have acknowledged the role of substrate in landform genesis. Graham et al. (2009) try to distinguish between a sedimentary substrate on the outer shelf where landforms are dominated by MSGL and record the final imprint of ice streaming, and the rugged bedrock-dominated inner-shelf where landforms, such as meltwater channels, bedrock-cored drumlins, and streamlined, gouged and grooved bedrock could have formed time-transgressively over multiple glaciations and therefore represent an inherited signal (Fig. 6). These authors also highlight the bedform complexity, especially on the inner shelf, with rough, bare rock zones (i.e. potential sticky spots) interspersed with patches of lineations composed of unconsolidated sediment (i.e. enhanced sliding/deformation), which collectively indicates a complicated mosaic of palaeo-flow processes.

Antarctic palaeo-ice stream beds also exhibit less variation in landform type and distribution when compared to northern hemisphere palaeo-ice sheet beds. For example, eskers have not been reported anywhere on the Antarctic shelf and ice stagnation features, such as kames, also appear to be absent, presumably because of the general lack of surface melting in Antarctica. Drumlin fields are also uncommon on Antarctic palaeo-ice stream beds, apart from bedrock influenced features at the transition from hard to soft substrate (e.g. Wellner et al. 2001, 2006). Drumlinoid forms cut into bedrock are also apparent over the inner Antarctic shelf (e.g. Ó Cofaigh et al. 2002). In comparison, on terrestrial ice stream beds in the northern hemisphere, drumlins have been mapped in the onset zone and towards the termini (e.g. Dyke & Morris, 1988; Stokes & Clark, 2003). Finally, ribbed moraine, which have been found in some ice stream onset zones in terrestrial settings of the northern hemisphere (Dyke & Morris, 1988) and as sticky spots further downstream (Stokes et al. 2008), have not been observed to date on the Antarctic shelf. These differences may, in part, be due to the likely lesser knowledge of Antarctic palaeo-ice streams and the scale of observations and data acquisition. Indeed, higher resolution datasets are beginning to uncover new bedforms that have hitherto gone unrecognised (e.g. Jakobsson et al. 2011).

4 AGE CONSTRAINTS ON RATES OF ICE-STREAM RETREAT AND DEGLACIATION

Accurately constraining the timing and rate of ice-stream retreat in Antarctica is crucial for: (i) identifying external drivers, which could have triggered deglaciation; (ii) assessing the sensitivity of individual ice streams to different forcing mechanisms; (iii) identifying regional differences in retreat histories; and (iv) determining the phasing between northern and southern hemispheric retreat and their relative contributions to sea-level change. The following section discusses some of the difficulties encountered when attempting to date palaeo-ice stream retreat from Antarctic shelf sediments. We then present a compilation of radiocarbon ages constraining the minimum age and rate of retreat from the Antarctic continental shelf since the LGM in order to investigate regional and inter-ice stream trends in their behaviour during deglaciation.

4.1 Problems in determining the age of grounding-line retreat from the Antarctic shelf by dating marine sediment cores

Providing constraints on the timing and rate of ice sheet retreat on the Antarctic continental shelf from marine radiocarbon dates is notoriously difficult (cf. Andrews et al. 1999; Anderson et al. 2002; Heroy & Anderson, 2007; Hillenbrand et al. 2010b for detailed reviews). Because of the scarcity of calcareous (micro-)fossils in Antarctic shelf sediments, ^{14}C dates are usually obtained from the acid-insoluble fraction of the organic matter (AIO). The corresponding AIO ^{14}C dates are, however, often affected by contamination with reworked fossil organic carbon resulting in extremely old ^{14}C ages (e.g. up to 13,525±97 uncorrected ^{14}C yrs BP for modern seafloor sediments on the eastern Antarctic Peninsula shelf: see Pudsey et al. 2006). To try and counter this effect, downcore AIO ^{14}C dates in Antarctic shelf cores are usually corrected by subtracting the uncorrected AIO ^{14}C age of sediment at the seafloor. This approach assumes that both the degree of contamination with fossil organic carbon and the ^{14}C age of the contaminating carbon have remained constant through time. This assumption, however, is probably invalid for dating sediments from the base of the deglacial unit, which is required for obtaining an accurate age of grounding-line retreat. These sediments are dominated by terrigenous components and therefore contain only small amounts of organic matter, i.e. even a small contribution of fossil organic carbon can cause a large offset between the ^{14}C age obtained from the sediment horizon and the true time of its deposition. In addition, the supply of fossil organic matter was probably higher, and the ^{14}C age of the contaminating carbon different, from the modern contamination, because the grounding line of the ice sheet (and therefore the source of the contaminating carbon) was located closer to the core site. This problem results in a drastic down-core increase of ^{14}C ages in the deglacial unit (e.g. Pudsey et al. 2006) and is evident from a so-called 'dog leg' in age-depth plots for the sediment cores (e.g. Heroy & Anderson 2007).

Despite uncertainties regarding absolute deglaciation chronologies, the approach of AIO ^{14}C dating often produces meaningful results for dating ice-sheet retreat, particularly when AIO ^{14}C dates can be calibrated against more reliable ^{14}C ages derived from carbonate or diatom-rich sediments (Licht et al. 1996, 1998; Domack et al. 1998, 1999, 2005; Cunningham et al. 1999; Andrews et al. 1999; Heroy & Anderson, 2005, 2007; Ó Cofaigh et al. 2005b; Leventer et al. 2006; McKay et al. 2008; Hillenbrand et al. 2010a,b; Smith et al. 2011). Additionally, carbonate ^{14}C dates, although much more dependable than the AIO dates, still need to be corrected for the marine reservoir effect (MRE) (^{14}C offset between oceanic and atmospheric carbon reservoirs).

In this paper, for consistency and ease of comparison, we use a uniform MRE of 1,300 (±100) years (see Table 4), as suggested by Berkman & Forman (1996) for the Southern Ocean and in agreement with most other studies (Table 4), thereby assuming that the MRE has remained unchanged since the end of the LGM. Ideally, in order to obtain the best possible age on grounding-line retreat, calcareous (micro-)fossils from the transitional glaciomarine sediments lying directly above the till (i.e. the deglacial facies) should be radiocarbon-dated. Where carbonate ^{14}C dates cannot be obtained from this terrigenous sediment facies, the chronology for ice-sheet retreat is often constrained only from ^{14}C dates on calcareous (micro-)fossils in the overlying postglacial glaciomarine muds or AIO ^{14}C dates from diatom-rich sediments. These ages actually record the onset of open marine conditions and thus provide only minimum ages for grounding-line retreat (Anderson et al. 2002; Smith et al. 2011). Wherever available, we also used core chronologies based on palaeomagnetic intensity dating (Brachfeld et al. 2003; Willmott et al. 2007; Hillenbrand et al. 2010b) to constrain the age of grounding-line retreat (Table 4). All deglaciation dates are reported as calibrated ages (Table 4).

4.2 Database of (minimum) ages for post-LGM ice-stream retreat

Heroy and Anderson (2007) compiled a database of radiocarbon dates related to the retreat of grounded ice from the Antarctic Peninsula shelf following the LGM. Since then, a number of additional dates have been published for the Antarctic Peninsula shelf (e.g. Heroy et al. 2008; Michalchuk et al. 2009; Milliken et al. 2009; Kilfeather et al. 2010) and here we present an updated synthesis of deglacial ages that record the retreat of grounded ice from the entire Antarctic continental shelf (Table 4 and Fig. 7). This is the most complete compilation of published deglacial dates recording the retreat of the Antarctic ice sheets since the LGM. A number of dates, despite being sampled from transitional glacimarine sediments directly above the diamicton, give ^{14}C ages of >25,000 cal. yrs BP, such as those from JOIDES Basin, the eastern Ross Sea and the south-eastern Weddell Sea (cf. Anderson & Andrews, 1999; Licht & Andrews, 2002; Mosola & Anderson, 2006; Melis & Salvi, 2009). The anomalously old deglaciation ages probably indicated that, in these regions, grounded ice did not extend to the core sites since the LGM defined as the time interval from 23,000-19,000 cal yrs BP in the Southern Hemisphere (Gersonde et al. 2005).

4.3 Regional trends in ice-stream retreat

The deglaciation of Antarctica is generally thought to have begun around 18 ka BP, in response to atmospheric warming (Jouzel et al. 2001). However, dates from the continental shelf show that ice streams exhibited considerable variation in the timing of initial retreat (Table 4; Fig. 8a,b). This is also supported by peaks in ice-rafted debris (IRD) in the central Scotia Sea at 19.5, 16.5, 14.5 and 12 ka which appear to indicate independent evidence for multiple phases of ice-sheet retreat (Weber et al. 2010). Ages for the onset of deglaciation range from 31-8 cal. ka BP, with the majority of ages bracketed between 18 and 8 cal. ka BP (Fig. 8a,b). This pattern is broadly consistent with results by Heroy & Anderson (2007) for the Antarctic Peninsula, who constrained the onset of deglaciation from the shelf edge from ~18-14 cal. ka BP.

The chronology of ice sheet retreat from the shelf in East Antarctica is less well constrained, although the general consensus was of a much earlier deglaciation than in West Antarctica and the Antarctic Peninsula (cf. Anderson et al. 2002). Sparse dates from outside palaeo-ice stream troughs in the south-eastern Weddell Sea provide some support for ice recession from its maximum position prior to the LGM (Elverhøi, 1981; Bentley & Anderson, 1998; Anderson & Andrews, 1999; Anderson et al. 2002). However, our new compilation of deglacial ages (Table 4) favours a much later deglaciation of the East Antarctic palaeo-ice streams, with Fig. 8a and 8b illustrating that initial retreat occurred within a time frame coincident with elsewhere in Antarctica. In Mac. Robertson Land, for example, Nielsen Palaeo-Ice Stream started to retreat from the outer shelf at ~14 cal. ka BP (Mackintosh et al. 2011), with Iceberg Alley and Prydz Channel ice streams subsequently receding at ~12 cal. ka BP (Domack et al. 1998; Mackintosh et al. 2011).

By comparing ages of the initial phase of retreat with global climate records and local bathymetric conditions it is possible to investigate the triggers and drivers of ice stream retreat. Fig. 8b indicates a good match between the onset of circum-Antarctic deglaciation (~18 cal. ka BP) and atmospheric warming (Jouzel et al. 2001). This cluster of dates also occurred just after a period of rapid eustatic sea-level rise: the 19 cal. ka BP meltwater pulse (Yokoyama et al. 2000; Clark et al. 2004). Another cluster of palaeo-ice stream deglacial ages are coincident with meltwater pulse 1a, suggesting that sea-level rise may have been an important factor during that period, whilst palaeo-ice streams that underwent initial retreat between 12-10 cal. ka BP have, in contrast, been related to oceanic warming (e.g. Mackintosh et al. 2011).

Given the asynchronous retreat history (Figs. 8 & 9), internal factors are likely to have modulated the initial (and subsequent) response of palaeo-ice streams to external drivers. We have indicated on Fig. 8c the initial geometry and gradient of the troughs on the outer shelf; and also correlated the (isostatically adjusted) trough depth and trough width against the minimum age of deglaciation. Our results show poor correlations, with high scatter (low R^2) for both depth and width (Fig. 8c). There is, however, a weak positive trend between the minimum age of deglaciation and both trough width and trough depth; with wider/deeper troughs associated with earlier retreat (Fig. 8c). However, this weak trend is heavily influenced by the Belgica Trough outlier. Nonetheless, these general trends mirror what we might expect, with deeper troughs more sensitive to changes in sea-level, whilst wider troughs are less sensitive to the effects of lateral drag, which can modulate retreat. In contrast, trough geometry and bed gradient seem to have little influence on the initial timing of retreat (Fig. 8b).

The overall pattern of palaeo-ice stream retreat to their current grounding-line positions is also highly variable, reflected by the large scatter of deglaciation ages throughout all regions of the Antarctic shelf (Fig. 9). This scatter is attributed to the variable behaviour of individual ice streams as opposed to errors inherent within the data. However, in the Antarctic Peninsula, Heroy & Anderson (2007) identified two steps in the chronology at ~14 cal. ka BP and ~11 cal. ka BP that corresponded to meltwater pulses 1a and 1b, respectively (Fairbanks, 1989; Bard et al. 1990, 1996). In summary, it can be concluded that palaeo-ice stream retreat was markedly asynchronous, with a number of internal factors likely responsible for modulating the response of ice streams to external forcing.

4.4 Palaeo-ice stream retreat histories

Given the variability in palaeo-ice stream retreat histories (Figs. 8 & 9), it is useful to produce high-resolution reconstructions of individual palaeo-ice streams to better understand the controls driving and modulating grounding-line retreat. Of the palaeo-ice streams identified in this paper, only those from Anvers Trough, Marguerite Trough, Belgica Trough, Getz-Dotson Trough and Drygalski Basin have well-constrained retreat histories supported by glacial geomorphic data (Fig. 10). In addition, ice-stream retreat in JOIDES-Central Basin is well constrained by De Geer-style moraines which allow the calculation of annual retreat rates (Shipp et al. 2002). Note that the retreat rates calculated for these palaeo-ice streams are maximum retreat rates because they are based on minimum deglacial ages and because they are averaged, over shorter timescales they are likely to have undergone faster and slower phases of retreat.

Anvers palaeo-ice stream (Antarctic Peninsula) retreated at a mean rate of 24 m yr^{-1} (based on carbonate and AIO ^{14}C dates) (Table 5), with retreat from the outer shelf commencing at ~16.0 cal. ka BP (Fig. 10a) (Pudsey et al. 1994; Heroy & Anderson, 2007) and Gerlache Strait on the innermost shelf becoming ice free by ~8.4 cal. ka BP (Harden et al. 1992) (Fig. 10a). Deglaciation of the inner fjords is corroborated by cosmogenic exposure ages from the surrounding terrestrial areas suggesting that final retreat occurred between 10.1 and 6.5 cal. ka BP (Bentley et al. in press). According to the most reliable deglacial ages (^{14}C dates on carbonate material), retreat accelerated towards the deep inner shelf of Palmer Deep (Fig. 10a), in accordance with an increase in the reverse slope gradient (Fig. 2d) (Heroy & Anderson, 2007). This retreat pattern is supported by the identification of GZWs on the outer shelf (Table 1) that are indicative of a punctuated retreat (Larter & Vanneste, 1995).

Getz-Dotson Trough palaeo-ice stream was characterised by a two-step pattern of ice stream retreat back to the current ice-shelf front. Initial retreat was underway by ~22.4 cal. ka BP (Fig. 10b) and was slow (average retreat rate: 18 m yr^{-1}), with ice finally reaching the mid-shelf by ~13.8 cal. ka BP (Smith et al. 2011). Conversely, grounding-line retreat accelerated towards the inner shelf (retreat rates ca. 30-70 m yr^{-1}). The three inner-shelf basins directly north of the Dotson and Getz ice shelves deglaciated rapidly, with ice free conditions commencing between 10.2 and 12.5 cal. ka BP (Fig. 10b) (Hillenbrand et al. 2010a; Smith et al. 2011). The increase in the rate of retreat through the three tributary troughs is characterised by a corresponding steepening in sea-floor gradient into the deep basins (up to 1600 m) on the inner shelf (Graham et al. 2009; Smith et al. 2011). Contrary to the retreat chronology of Getz-Dotson palaeo-ice stream, the associated geomorphology displays uninterrupted MSGL on the outer shelf, in the zone of slow retreat, with a number of GZWs on the inner shelf, where the fastest rates of retreat are observed (Table 1) (Graham et al. 2009). The position of the GZWs in this zone of rapid retreat implies episodic, yet rapid retreat, with the GZWs formed over relatively short (sub-millennial) time scales (Smith et al. 2011).

Drygalski Basin palaeo-ice stream (western Ross Sea) is thought to have receded from its maximum position, just north of Coulman Island on the mid-outer shelf, by ~14.0 cal. ka BP (Fig. 10c) (Frignani et al. 1998; Domack et al. 1999; Brambati et al. 2002). An additional carbonate-based deglaciation date of ~16.8 cal. ka BP from the outer-shelf of the neighbouring Pennell Trough (Licht & Andrews, 2002) provides an additional constraint on deglaciation in the western Ross Sea (Fig. 10c). However, Mosola & Anderson (2006) suggest that the core may have sampled an iceberg turbate and therefore could be less reliable than initially reported.

Development of open marine conditions in the vicinity of Drygalski Ice Tongue was complete by ~10.5 cal. ka BP (Finocchiaro et al. 2007), with grounded-ice reaching south of Ross Island by 11.6 cal. ka BP (hot water drill core taken through Ross Ice Shelf [HWD03-2] – McKay et al. 2008) and open marine conditions established by 10.1 cal. ka BP (Fig. 10c) (McKay et al. 2008). This new date for the development of ice-free conditions in the vicinity of Ross Island is earlier than previously reported (7.4 cal. ka BP) (Licht et al. 1996; Cunningham et al. 1999; Domack et al. 1999; Conway et al. 1999). Mean rates of retreat towards Ross Island were calculated to be ~50 m yr^{-1} (Shipp et al. 1999), with retreat towards its present grounding-line position thought to have been much faster (~140 m yr^{-1}) (Shipp et al. 1999). This chronology suggests that ice retreated rapidly from its maximum position (mean retreat rate: 76 m yr^{-1}) with retreat accelerating south of Dryglaski Ice Tongue (average: 317 m yr^{-1}) (Fig. 10c) (McKay et al. 2008). Further recession to the current grounding line position proceeded at ~90 m yr^{-1}, with the Ross Ice Shelf becoming pinned against Ross Island during this period (McKay et al. 2008). Although geomorphological data of the sub-ice shelf section of Drygalski palaeo-ice stream is not available, the bedform evidence north of Ross Island is dominated by MSGLs, with a large GZW marking its LGM limit (Shipp et al. 1999; Anderson et al. 2002).

The neighbouring JOIDES-Central Basin palaeo-ice stream, which also reached a maximum position on the mid-outer shelf during the LGM (Licht et al. 1996; Domack et al. 1999; Shipp et al. 1999, 2002), is floored by a series of transverse ridges that overprint all other landforms (Shipp et al. 2002). These features are interpreted as annually deposited De Geer moraines, formed at the grounding line during ice stream recession (see section 3.3.5) and have been used to estimate a retreat rate of 40-100 m yr^{-1} (Table 5) (Shipp et al. 2002). Open marine conditions in outer JOIDES Basin were established by 13.0 cal. ka BP (Domack et al. 1999). Thus, the rates of recession calculated from both the transverse moraines and the timing of retreat inferred from radiocarbon ages in JOIDES-Central Basin are broadly consistent with those from Drygalski Basin (Domack et al. 1999), even though the geomorphic signatures are different.

The Marguerite Trough palaeo-ice stream (western Antarctic Peninsula) underwent a stepped pattern of retreat, with rapid retreat across the outer 140 km of the shelf at ~14.0 cal. ka BP (Fig. 10d) (Kilfeather et al. 2010). This rapid phase of retreat is consistent with well-preserved and uninterrupted MSGLs on the very outer shelf of the trough (Ó Cofaigh et al. 2008), whilst a number of GZWs further inland suggest that retreat became increasingly punctuated (Livingstone et al. 2010). However, as in Getz-Dotson Trough, the rapid retreat rates (i.e. within the error of the dates) suggest that GZW formation must have occurred over a relatively short (sub-millennial) time-scale (Livingstone et al. 2010). This was followed by a slower phase of retreat on the mid-shelf, which was also associated with the break-up of an ice-shelf. Thereafter, the ice stream rapidly retreated to the inner shelf at ~9.0 cal. ka BP (Fig. 10d) (Kilfeather et al. 2010). This latter phase of rapid retreat is supported by cosmogenic exposure ages at Pourquoi-Pas Island that indicate rapid thinning (350 m) at 9.6 cal. ka BP (Bentley et al., in press). George VI Sound is thought to have become ice free between ~6.6-9.6 cal. ka BP (Fig. 10d), based on ages from foraminifera and shells (Sugden and Clapperton, 1981; Hjort et al. 2001; Smith et al. 2007). The drivers for these two phases of rapid grounding-line retreat at 14.0 cal. ka BP and 9.0 cal. ka BP have been suggested as meltwater pulse 1a and the advection of relatively warm Circumpolar Deep Water (CDW) onto the continental shelf (Kilfeather et al. 2010). The mean retreat rate of the palaeo-ice stream along the whole trough was ~80 m yr^{-1} (Table 5) (and ranged between 36-150 m yr^{-1}: Figure 10d) which is noticeably faster than Anvers palaeo-ice stream. Crucially, the two phases of rapid retreat (outer-mid shelf and mid-inner shelf (see above)) are associated with even greater rates of recession and actually within the error of the radiocarbon dates.

The deglacial chronology of Belgica Trough (southern Bellinghausen Sea) is significantly different to that experienced by other West Antarctic palaeo-ice streams because the ice stream in Belgica Trough had receded from its maximum position on the shelf edge as early as ~30.0 cal. ka BP (Fig. 10e) (Hillenbrand et al. 2010a). Grounding-line retreat towards the mid-shelf proceeded slowly, and the middle shelf eventually became free of grounded ice by ~24.0 cal. ka BP (Fig. 10e). The inner shelf had deglaciated by ~14.0 cal. ka BP in Eltanin

Bay and by ~4.5 cal. ka BP in Ronne Entrance (Fig. 10e) (Hillenbrand et al. 2010a). Mean retreat rates varied between 7-55 m yr^{-1} (Table 5), with deglaciation thought to be prolonged and continuous (Hillenbrand et al. 2010a). However, a series of GZWs on the inner shelf of Belgica Trough suggests that retreat was characterised by episodic still-stands (Ó Cofaigh et al. 2005a).

5 DISCUSSION

Given the advantages of conducting research on palaeo-ice streams (see Section 1), and the information that has been obtained from their beds, it is important to contextualise observations within the broader themes of (palaeo)glaciology to inform discussions on how Antarctic ice streams may respond to future external forcings. The aim of this section, therefore, is to critically discuss the geological evidence for palaeo-ice streaming in terms of its implications for understanding ice-stream processes and its relevance to predictions of future Antarctic Ice Sheet behaviour.

5.1 Examples of contrasting Antarctic ice-stream retreat styles

Where detailed glacial geomorphic data exists along the length of palaeo-ice stream flow path and/or a deglacial chronology can be used to constrain the retreat rate (see Table 4), we have categorised Antarctic palaeo-ice streams into discrete retreat styles (Table 6). To discriminate between different asynchronous, multi-modal retreat patterns, Table 6 differentiates between palaeo-ice streams that exhibit slow/episodic retreat from the outer and middle shelf followed by rapid retreat from the inner shelf, and *vice-versa*. A good example of a palaeo-ice stream that underwent accelerated retreat from the inner shelf is Pine Island Trough. Five GZWs on the mid-outer shelf demarcate still-stand positions (Graham et al. 2010), whilst the deep, rugged inner shelf is characterised by a thin carapace of deglacial sediment with no evidence of morainal features (Lowe & Anderson 2002). In contrast, the Robertson Trough palaeo-ice stream has deposited no morainal features on the outer shelf (lineations which exhibit localised cross-cutting), whereas the mid-shelf is interrupted by a series of GZWs up to 20 m high (Gilbert et al. 2003; Evans et al. 2005). This is consistent with a switch from continuous and rapid retreat across the outer shelf to episodic retreat on the middle shelf (cf. Evans et al. 2005).

The range of retreat rates and variability in the timing of deglaciation around the Antarctic shelf highlights that local factors such as drainage-basin size, bathymetry, bed roughness and ice-stream geometry are important in modulating grounding-line retreat. Even neighbouring palaeo-ice streams can exhibit strikingly different retreat styles, as exemplified by the different frequency, localities and sizes of GZWs in adjacent troughs in the eastern Ross Sea (Mosola & Anderson, 2006). A paucity of deglacial sediment within this region (<1 m) has been used to infer rapid collapse of the palaeo-ice streams (Mosola & Anderson, 2006), although the large and multiple GZWs point towards repeated still-stands and thus episodic retreat (Table 6). This apparent contradiction between the formation of large GZWs (up to 180 m thick) and an apparent lack of deglacial sediment highlights current deficiencies in the understanding of: (i) rates of sub-, marginal- and pro-glacial sediment supply and deposition; (ii) speed of GZW formation; and (iii) depositional processes at the grounding line and beneath ice shelves.

5.2 Controls on the retreat rate of palaeo-ice streams

A number of general observations can be made regarding characteristics which tend to be symptomatic of distinctive retreat styles (Table 6). Firstly, there appears to be a correlation between bathymetric gradient of the trough floor and the rate of grounding-line retreat, as predicted by theoretical modelling (e.g. Schoof, 2007). This is demonstrated by the acceleration in grounding-line retreat on the reverse-slope, inner-shelves of the Anvers-Hugo Island and Getz-Dotson palaeo-ice streams (Figs. 2d; 10a,b; Table 6) (also see Smith et al. 2011). On a local scale, a link between lower gradients (average slope: 0.015°) and GZW development for Pine Island Trough has also been demonstrated (Graham et al. 2010). However, it is apparent that bed slope is not the only factor controlling grounding-line retreat and, indeed, can be modulated or in some circumstances suppressed by other factors. For example, GZWs, which have been linked with stable grounding-line positions, are commonly observed along reverse gradients, such as in Marguerite Trough, Belgica Trough, Pine Island Trough, Getz-Dotson Trough and all of the Ross-Sea palaeo-ice stream beds (Tables 2 & 6). Significantly, slow retreat rates have also been described within some troughs with reverse bed slopes (Shipp et al. 2002; Table 6). Belgica Trough is characterised by multiple GZWs on the inner shelf, where the trough dips steeply into Eltanin Bay (Ó Cofaigh et al. 2005a). Moreover, despite being characterised by a relative acceleration in grounding-line retreat towards Palmer Deep, the absolute retreat rates within Anvers-Hugo Island Trough are low (Fig. 10a; Table 6).

The retreat of palaeo-ice streams over rugged bedrock-dominated inner shelves, which exhibit large variations in relief and well-defined banks (e.g. Gerlache Strait, Marguerite Bay, Anvers-Hugo Island and Pine Island Bay), was not universally slow. This is again contrary to theoretical studies, which propose slower rates of grounding-line retreat where lateral and basal drag is greatest (Echelmeyer et al. 1991, 1994; Alley, 1993a; MacAyeal et al. 1995; Whillans & van der Veen, 1997; Joughin et al. 2004; Siegert et al. 2004; Rippin et al. 2006; Stokes et al. 2007). Possible reasons for this discrepancy include the preferential flow of relatively warm oceanic waters to the grounding line due to the large changes in relief (high roughness) via pre-existing meltwater drainage routes and deep basins (Jenkins et al. 2010), or increased lubrication generated as water starts to penetrate, and fill, deep 'hollows' in the rough bed (e.g. Bindschadler & Choi, 2007) as the ice reaches flotation.

The palaeo-ice streams characterised by the highest retreat rates tend to be the smallest glacial systems, whilst those that underwent episodic/slow retreat are typically associated with large drainage basins and/or broad troughs (Tables 2 & 6) (also see Ó Cofaigh et al. 2008). This relationship is what you might expect, with the response times of large drainage systems less sensitive to perturbations than a small drainage basin that can quickly re-adjust to a new state of equilibrium (e.g. thickness divided by mass balance rate).

Our review of the retreat styles of Antarctic palaeo-ice streams highlights the potential for using glacial landform signatures to investigate grounding-line retreat and reinforces the notion that local factors, such as trough width, drainage basin size, bed gradient, bed roughness, and substrate, play a critical role in modulating ice-stream retreat. It is also likely that subglacial meltwater (e.g. Bell, 2008) and the thermo-mechanical coupling between the ice and the underlying sediments (e.g. Tulaczyk & Hossainzadeh, 2011) also plays a major role in modulating ice-stream speed and retreat. However, these processes are harder to quantify from the palaeo-record. Our attempt to categorise palaeo-ice streams into discrete retreat styles has revealed the importance of drainage basin area and reverse slope gradient as potentially key controls governing the sensitivity of ice streams to grounding-line retreat.

5.3 Influence of underlying bedrock characteristics on ice-stream dynamics

The role of substrate (i.e. underlying bedrock geology and roughness) in controlling ice-stream dynamics is hard to quantify due to the lack of process understanding regarding how and over what time-period glacial landforms actually form in bedrock and, to an extent, in sediments as well (see Sections 3.2 & 3.3.8). Thus, although the generally 'higher' roughness and hardness of bedrock will almost certainly impact upon flow velocities, it is hard to unequivocally determine this relationship. Determining the role of substrate on ice velocity is therefore problematic for areas of bedrock, such as the inner Antarctic shelf. Indeed, the morphological signature of ice streaming over bedrock remains largely unresolved despite recent attempts to relate mega-grooves, roché moutonees and whalebacks to palaeo-ice streams (Roberts & Long, 2005; Bradwell et al. 2008). This is best exemplified by the downflow evolution of bedforms across the bedrock-sedimentary substrate transition in the Antarctic palaeo-ice stream troughs, which is mirrored by an increase in their elongation ratio (Fig. 6), and generally attributed to acceleration at the onset of streaming flow (Shipp et al. 1999; Wellner, et al. 2001, 2006; Canals et al. 2002; Ó Cofaigh et al. 2005a; Evans et al. 2006). However, it is unclear, whether this elongation change is caused by a genuine transition in ice velocity (i.e. zone of acceleration) or whether it simply reflects a change in underlying geological substrate and its potential for subglacial landform formation (Graham et al., 2009).

The strong substrate control on ice streaming implied by Antarctic geophysical observations (e.g. Anandakrishnan et al. 1991; Bell et al. 1998; Bamber et al. 2008) does support a genuine velocity transition. However, there are contemporary examples where streaming is thought to have occurred over a predominantly hard bedrock, e.g. Thwaites Glacier, West Antarctica (Joughin et al. 2009). In addition, the 'bundle structures' on the inner shelf of the Gerlache-Boyd palaeo-ice stream and the highly elongate grooves in Smith Trough and Sulzberger Bay Trough, which are eroded into bedrock and perhaps overconsolidated glacial till, may result from fast ice-stream flow. These examples suggest that thermo-mechanical feedbacks can cause fast flow in deep troughs irrespective of roughness or substrate.

The role of rougher bedrock areas in the transition zone are also likely to be important in generating (and retaining) meltwater through strain heating to lubricate the bed and initiate and maintain streaming flow (Bell, et al. 2007; Bindschadler & Choi, 2007). This is supported by the widespread presence of large meltwater channels, basins and even subglacial lakes on the rugged inner shelf (see Section 3.3.6). Bed roughness evolves as substrate is eroded or buried by sediment deposition. This is especially relevant in the bedrock dominated onset zone regions, where we hypothesise that changes in roughness could influence ice stream dynamics and potentially lead to upstream migration of the onset zone. Given typical erosion rates cited for temperate valley glaciers (Bogen et al. 1996; Hallet et al. 1996) it is feasible that large-scale (potentially 100s meters amplitude) bedrock landforms can be smoothed over sub-Quaternary time scales (e.g. Jamieson et al. 2008). Furthermore, it is likely that as bed roughness evolves in response to glacial erosion and deposition, the behaviour of the ice stream will also evolve. However, the spatial and temporal scale and significance of such a potential feedback mechanism has not been systematically investigated in the context of ice streams.

5.4 Atmospheric circulation and precipitation patterns

The interaction between the cryosphere and atmosphere is fundamental in controlling ice sheet and glacier dynamics. Indeed shifting ice-dispersal centres and complex ice-flow in response to changing patterns of accumulation has been widely reported in former ice sheets of the northern hemisphere (e.g. Kleman et al. 2006).

The first order control on Antarctic precipitation is topography, which differs significantly between East and West Antarctica. East Antarctica has a steep coastal escarpment, a relatively small area of ice shelves and a high, large inland plateau, while West Antarctica has extensive ice shelves and gentler slopes (van de Berg et al. 2006). In the steep coastal margins precipitation is dominated by orographic lifting of relatively moist, warm air associated with transient cyclones that encircle the continent. The steep ice-topography acts as a barrier to the inland propagation of storm tracks and thus inland accumulation rates are very low, especially on the interior plateau of East Antarctica, which is effectively a polar desert (Vaughan et al. 1999; Arthern et al. 2006; van de Berg et al. 2006; Monaghan et al. 2006a,b). During times of ice expansion the zone of orographic precipitation will also move seawards, leading to further starvation of the interior of the ice sheet and consequently little thickening. Thus, given the general mass balance distribution over Antarctica, it is perhaps, not surprising that the most prominent and largest concentration of palaeo-ice stream troughs occur where precipitation is highest, in the West Antarctic and Antarctic Peninsula ice sheets. Indeed, reduced precipitation in the interior of East Antarctica may have affected the ability of ice streams to reach the shelf edge in the late Pleistocene (O'Brien et al. 2007). For example, during the LGM, Prydz Channel Ice Stream became dominated by ice-flow out of Ingrid Christensen

Coast rather than along the axis of the Amery Ice Shelf, and this has been attributed to the topography setting of the Amery drainage basin relative to the circum-polar trough and associated storm tracks (O'Brien et al. 2007).

A further control on the pattern of accumulation is the Southern Annular Mode (SAM), whereby the synoptic-scale circum-polar vortex of cyclones oscillates between the Antarctic Coast and the mid-latitudes on week to millennial timescales. Typically, when the cyclones track across the Antarctic coastal slopes higher snow accumulation rates are observed (Goodwin et al. 2003, 2004). Precipitation is also affected by the regional variability in sea-ice extent around Antarctica. For example, according to Gersonde et al. (2005), LGM precipitation over the EAIS sector, between 90°E and 120°E, would have been much higher than over the EAIS sector, between 10°E and 30°W, because in the former sector the LGM summer sea-ice edge (and thus open water) was much closer to the continent than in the latter sector.

5.5 Landform-process interactions and subglacial sediment transport

The central tenet behind reconstructing palaeo-ice sheets from geological evidence is the causal link between subglacial processes and landform genesis. Success in using landforms and subglacial sediments to extract information on bed properties, and therefore in reconstructing the evolution of the ice sheet, relies upon a thorough comprehension of the genesis of landforms and sediments used in the 'inversion model' (cf. Kleman & Borgström, 1996; Kleman et al. 2006). This section investigates these linkages in reconstructing palaeo-ice streams by discussing the characteristics of palaeo-ice streams and, in particular, the two-tiered till structure and the formation of MSGL and GZWs in soft sediment. A further issue is that ice dynamics during ice sheet build up is not well constrained and, as such, some of the tills/landforms observed may be wholly or partially inherited from earlier ice flow events.

5.5.1 *Origin of the upper soft and lower stiff tills*

Three hypotheses were proposed by Ó Cofaigh et al. (2007) to account for the upward transition from stiff to soft till exhibited by palaeo-ice stream beds (Section 3.2): (1) till deposition during separate glacial advances, with the soft till associated with the development of an ice stream during the most recent phase; (2) a process transition from lodgement to deformation, with the deformation till associated with the onset of streaming flow; and (3) an upwards increase in dilatancy related to A/B horizons in a deformation till, a characteristic that has been previously observed beneath and in front of contemporary Icelandic glaciers (cf. Sharp, 1984; Boulton & Hindmarsh, 1987). Ó Cofaigh et al. (2007) view these hypotheses as 'end members', with aspects of each mechanism exhibiting some compatibility with the field evidence. They concluded that the soft till was a 'hybrid', formed by a combination of subglacial sediment deformation and lodgement. Reinardy et al. (2011a) identify significant (micro-scale) differences between the two till-types, which they also attribute to a deforming bed continuum. Initial deposition of till as ice advanced across the shelf produced ductile structures, with brittle structures produced subsequently following compaction and dewatering. The soft till was produced by a switch to streaming flow that resulted in deformation of the upper part of the stiff till (Ó Cofaigh et al. 2007; Reinardy et al. 2011a). The origin of the lower stiff and upper soft tills has important implications for understanding the behaviour of ice streams over the last glacial cycle. Whereas hypotheses (1) and (2) imply that ice streams switched on at a late stage in the glacial cycle, subsequent to ice expansion onto the outer continental shelf, and possibly associated with the onset of deglaciation, hypothesis (3) could imply continuous streaming during both ice sheet advance and retreat. A lack of buried MSGL on top of the stiff till support the interpretation that it is not associated with streaming conditions.

5.5.2 *MSGL formation*

Despite the use of MSGLs as a diagnostic landform for identifying palaeo-ice streams in the geologic record (Section 3.3.1), understanding of the genesis of this landform remains incomplete. There are four main hypotheses for their genesis: (1) as a product of subglacial erosion by high-discharge, turbulent meltwater floods (Shaw et al. 2000, 2008; Munro-Stasiuk & Shaw, 2002); (2) groove-ploughing of soft-sediment by ice keels formed at the base of an ice stream (Tulaczyk et al. 2001; Clark et al. 2003); (3) subglacial deformation of soft sediment from a point source, such as a bedrock obstacle or zone of stiff till beneath an ice-stream (Clark, 1993; Hindmarsh, 1998); and (4) the instability theory, which can be extended to include lineation genesis when a local subglacial drainage system is included in the calculations (Fowler, 2010). Hypothesis (3) has significant overlap with the groove-ploughing mechanism (hypothesis 2) reinforcing the idea that MSGL were formed by deformed sediment, in this case as the ice keels plough through the sediment (Clark et al. 2003).

The mega-flood hypothesis is contentious (e.g. Clarke et al. 2005; Ó Cofaigh et al. 2010), especially given recent observations of actively forming MSGL beneath Rutford Ice Stream, West Antarctica, in the absence of large discharges of meltwater (King et al. 2009). In Marguerite Trough, individual lineations show evidence for bifurcation or merging along their lengths, gradual increases in width and amplitude downflow, and also subtle ‘seeding points’ comprising flat areas devoid of lineations at their point of initiation (Ó Cofaigh et al. 2005b). These observations do not fit the expected landform outcome for the groove-ploughing mechanism (cf. Clark et al. 2003). Indeed, there does not appear to be a consistent correlation between bedrock roughness elements upstream and MSGL distribution downstream, but there are locations where the formation of MSGL is obviously linked to bedrock roughness. This suggests that groove-ploughing was not the only mechanism for MSGL formation on the Antarctic continental shelf despite supporting evidence. For example, the influence of bedrock roughness in Biscoe Trough and Gerlache-Boyd palaeo-ice stream coupled with some observations of an undulating subbottom reflector (marking the boundary between the stiff lodgement till and the soft deformation till) indicates localised ploughing (Ó Cofaigh et al. 2007).

The palaeo-ice stream landsystem model associates MSGLs with rapid ice-flow (Clark & Stokes, 2003) and they are often thought to record the final imprint of streaming (Graham et al. 2009). However, the preceding discussion highlights our lack of understanding regarding their formative mechanisms (cf. Ó Cofaigh et al. 2007), their relation to flow velocity (i.e. the potential interplay between velocity, ice-flow duration and sediment supply) and their implication for sediment transport and deposition at the ice-sheet bed. For example, is the evolution in elongation ratios along-flow related to their synchronous formation, with increased velocities towards the terminus, or a time-integrated signature related to changes in velocity as a function of grounding-line retreat? Are lineations transient features constantly (and rapidly?) being created and destroyed (depending on the prevalent bed conditions) or stable features capable of withstanding changes in basal processes? Resolving these genetic problems has implications for how ice flows, the bed properties and subglacial processes. For example, if we understand how MSGLs form, we will better understand ice stream flow mechanisms and how to parameterize flow laws in numerical models.

5.5.3 *GZW formation*

There are two main hypotheses to explain GZW formation that are not mutually exclusive: (1) subglacial transport and then deposition of till at the grounding-line during still stands (e.g. Alley et al. 1989); and (2) the melt-out of basal debris at the grounding-line, with the debris entrained by basal freeze-on (e.g. Christoffersen & Tulaczyk, 2003). Each of these mechanisms has important implications for our understanding of bed properties and the mode and rate of sediment transport to the grounding line. The few available sediment supply calculations (and assuming hypothesis 1) suggest that GZWs can form between 1,000 to 10,000 years with typical sediment fluxes ranging from 100 m^3 yr^{-1} per meter width to 1,000 m^3 yr^{-1} per meter width (Section 3.3.4). Despite these gross calculations, little is known about sediment transport beneath ice streams.

Recently, four 10-40 m thick, 5-10 km long and up to 8 km wide GZWs have been observed on the mid-outer shelf of Marguerite Trough (Livingstone et al. 2010). What is interesting is that these GZWs are situated within a zone of the trough where radiocarbon dates indicate ice stream retreat was rapid (i.e. within the error of the dates: Kilfeather et al. 2010). Thus, their occurrence hints at the potential for high sediment fluxes in GZW formation, with the quoted range of sediment fluxes suggesting that the GZWs would have taken between 500-5,000 years to form. High sediment fluxes of up to 8,000 m^3 yr^{-1} per meter width have also been estimated for the Norwegian Channel (Nygård et al. 2007). Indeed, deglacial ages on the inner shelf of the Getz-Dotson Trough constrain GZW genesis to a ca. 1,650 year period (Smith et al. 2011). However, fluxes are dependent upon the large-scale mobilization of sediment, which is limited by the rate of subglacial erosion and the depth of deformation below the ice-stream base. For example, Anandakrishnan et al. (2007) estimated that the calculated sediment flux at the grounding line of Whillans Ice Stream (~150 m^3 yr^{-1} per meter width) would require distributed upstream deformation of the subglacial sediment (hypothesis 1) over a considerable thickness (several tens of cm's).

A viscous till rheology could theoretically account for these large fluxes, but doubt has been cast on this particular assumption from both *in situ* borehole measurements (e.g. Engelhardt & Kamb, 1998; Fischer & Clarke, 1994; Hooke et al. 1997; Kavanaugh & Clarke, 2006) and laboratory experiments (e.g. Kamb, 1991; Iverson et al. 1998; Tulaczyk, 2000; Larsen et al. 2006). The plastic bed model is also potentially problematic as deformation can collapse to a single shear plane and thus limit sediment fluxes (Christoffersen et al. 2010). However, it has been shown that, over large scales, multiple failures can integrate to transport large volumes of sediment subglacially (e.g. Hindmarsh et al. 1997, 1998; Ó Cofaigh et al. 2007). Moreover, a change in basal thermal conditions from melting to freezing can cause a short-term movement of the shear plane into the till layer (Bougamont & Tulaczyk, 2003; Bougamont et al. 2003; Christoffersen & Tulaczyk, 2003a,b; Rempel, 2008). This change in thermal conditions is associated with basal freeze-on (hypothesis 2) and has resulted in the entrainment of large volumes of sediment within Kamb Ice Stream (Christoffersen et al. 2010). Subglacial sediment advection caused by changes in basal thermomechanical conditions also implies asynchronous erosion and transport and punctuated sediment delivery to the grounding line (cf. Christoffersen et al. 2010). Till ploughing by ice/clast keels offers an additional mechanism for mobilizing and transporting sediment (Tulaczyk et al. 2001).

A further potential mechanism for delivering pulses of increased sediment delivery is subglacial meltwater transport, especially given observations of major drainage events associated with active subglacial lakes beneath modern day ice masses (Fricker et al. 2007; Stearns et al. 2008; Carter et al. 2009; Smith et al. 2009). Jökulhlaups in Iceland, which may transport on the order of 10^7-10^8 tons of sediment, provides some support for this mechanism (Björnsson, 2002) and it is entirely plausible that meltwater may contribute to sediment being deposited at ice stream grounding lines.

The locally high rates of subglacial erosion (1 m yr^{-1}) monitored beneath Rutford Ice Stream (Smith et al. 2007) are up to four orders of magnitude greater than measured and interpreted values for subglacial environments (0.1-100 mm yr^{-1}) (e.g. Hallet et al. 1996; Alley et al. 2003) and indicate that sediment can be mobilized rapidly for subsequent redistribution. Indeed, these locally high erosion rates are much greater than published sediment fluxes for the formation of GZWs. This spatial variability is a common attribute from beneath active ice streams and reveals a dynamic sedimentary system that is characterised by spatial and temporal evolution in bed properties and the ability to undergo significant changes in erosion and deposition on decadal timescales (Smith, 1997; Smith et al. 2007; King et al. 2009). This conclusion is supported by the geological evidence, because many GZWs occupy discrete locations on ice stream beds, rather than spanning entire trough widths (e.g. Ó Cofaigh et al. 2005b; Graham et al. 2010), implying that sediment advected at the ice-stream bed can vary spatially at the macro-scale (tens of kms).

5.6 Sub-ice stream hydrological system

Catastrophic meltwater drainage (hypothesis 1) and/or sequential meltwater erosion over multiple glacial cycles (hypothesis 2) have been suggested to account for the large, deeply eroded subglacial meltwater channels and tunnel valleys incised into the bedrock-floored, inner continental shelf (Section 3.3.6). Tunnel valleys have been used as evidence for catastrophic meltwater discharge of water stored beneath ice sheets, with drainage occurring under bankfull conditions (Shaw et al. 2008). Recent observations show that active subglacial lake systems, intimately associated with outlet glaciers and ice streams, are characterised by periodic drainage along discrete flow-paths and thus provide some support for this hypothesis (Fricker et al. 2007; Stearns et al. 2008; Carter et al. 2009; Smith et al. 2009). However, it is equally plausible that the meltwater channels reflect an inherited signal of sequential erosion over multiple glaciations (hypothesis 2) (Lowe & Anderson 2003; Smith et al. 2009), with meltwater drainage pathways routing and re-routing along channel networks (Ó Cofaigh et al. 2010). Hypothesis 2 implies that meltwater streams across bedrock form stable well-organised drainage systems that become progressively more ‘fixed’ over time as the geometry of the channel becomes increasingly important relative to the geometry of the overlying ice mass.

The scarcity of meltwater channels on the outer shelf (apart from the shelf break) likely results from a soft, mobile bed, which precludes formation of a stable meltwater system and instead adjusts transiently to fluctuations in subglacial water pressure (cf. Noormets et al. 2009). Meltwater transfer by Darcian flow through the uppermost sediment layer is generally considered the primary mode of drainage under ice streams underlain by sedimentary substrate (e.g. Tulaczyk et al. 1998) and shallow “canals” may form temporarily, where excess water occurs (e.g. Walder & Fowler, 1994). These shallow canals have been both predicted from theoretical studies (Walder & Fowler, 1994; Ng, 2000) and also observed on geophysical records from beneath the modern Rutford Ice Stream (King et al. 2004). Hence, it cannot be ruled out that these networks are present on the outer shelf parts of the Antarctic palaeo-ice stream troughs but that their dimensions lie below the spatial resolution of the swath bathymetry data and have therefore not been detected.

The paucity of eskers on the Antarctic continental shelf can be attributed to the polar climate, with cold temperatures preventing supra-glacial meltwater production. This additional input of meltwater, penetrating from the surface to the ice sheet bed, is thought to be critical in forming an esker (Hooke & Fastook, 2007). Furthermore, Clark & Walder (1994) have theorised that eskers should be rare in regions with subglacial deformation, because drainage should be dominated by many wide, shallow canals as opposed to relatively few, stable channels (see Section 3.3.6). The unconsolidated sedimentary substrate that floors the mid-outer continental shelf of Antarctic palaeo-ice stream troughs and the lack of meltwater channels within this soft bed supports this theory. However, it might also be that case that the lack of eskers reflects limits in the observable resolution of our instruments and is therefore merely a scale problem.

6. FUTURE WORK

Numerical ice-sheet and ice-stream models, developed for the prediction of the future contribution of Antarctic Ice Sheet dynamics to sea-level change, are only as good as our current level of understanding regarding the mechanics of ice drainage, the processes and feedbacks operating within the system and also the scale and spatial extent at which we make observations and collect data. This review has highlighted recent advances, but now also considers some key areas for further work, which we summarise as follows:

- The need for a better understanding of subglacial sediment erosion, transport and deposition. For example, what do GZWs actually tell us about grounding-line stability, and how quickly do they form?
- An improved understanding of the timescales over which basal roughness is changed by glacial erosion and deposition and the consequent feedbacks with ice dynamics; i.e. if rough onset zones can prevent headward migration of ice streams, then over what timescales may this be overcome and is there a threshold roughness scale beyond which streaming is not possible?
- Further work examining how ice stream flow influences erosion of hard bedrock. This includes distinguishing between the relative influences of ice-stream substrate, changes in slope gradient/steepness and ice-flow velocity on bedform genesis.
- Resolving genetic links between landforms and subglacial processes, thereby allowing better parameterisation of the bed properties in numerical ice-stream and ice-sheet models.
- Further work on subglacial meltwater flow and its role in ice-stream dynamics. For example, how does water flow over or through unconsolidated sediment, and can 'outburst floods' deliver large pulses of sediment to the grounding line?
- Identifying the sensitivity of grounding-line retreat to external triggers and quantifying the influence of internal characteristics in either accelerating or reducing retreat rates.

The following paragraphs briefly detail two approaches that may provide some scope for investigating these problems.

Glacial-geomorphological mapping has been widely applied to palaeo-ice sheets in the northern hemisphere in order to reconstruct complex ice-flow dynamics and bed properties. So far, however, this detailed mapping has only been replicated in Antarctica for the Getz-Dotson Trough (see Graham et al. 2009). There is an urgent need for a more comprehensive analysis of the bed properties and their spatial and temporal variations for Antarctic palaeo-ice streams, especially given the fine-scale at which we can now identify glacial bedforms (e.g. Jakobssen et al. 2011). Detailed mapping must be augmented by further age constraints on the deglacial history. These are crucial to building up a database of individual palaeo-ice stream retreat styles, rates and timings, directly comparing against modern observations beneath contemporary ice streams (e.g. King et al. 2009) and investigating external and internal controls on grounding-line retreat.

Reconstructing the behaviour of palaeo-ice streams has typically involved examination of empirical data from individual ice stream beds or numerical/theoretical treatments. However, there has been little attempt to integrate, compare and validate numerical modelling experiments against the observational record of ice-stream retreat (e.g. Stokes & Tarasov, 2010). A combined observational and modelling approach to investigate palaeo-ice streams would provide a powerful tool for identifying the controlling factors governing grounding-line retreat, as model simulations could be compared against distinctive retreat styles. This approach could also be used to compare relict meltwater channel networks to modelled drainage routeways (e.g. Wright et al. 2008; Le Brocq et al. 2009) and to investigate subglacial hydrological systems. Incorporation of basal sediment transport within ice stream simulations (e.g. Bougamont & Tulaczyk, 2003) could provide invaluable information into the subglacial mobilisation, transport and deposition of sediment.

7. CONCLUSIONS

The importance of ice streams is reflected in the fact that they act as regulators of ice sheet stability and thus the contribution of ice sheets to sea level. From recent changes we know that ice streams are characterised by significant variability over short (decadal) timescales. In order to extend the record of their behaviour back into the geological past and to glean important information on their bed properties, investigations have turned to palaeo-ice streams. In this paper we have compiled an inventory of all known circum Antarctica palaeo-ice streams, their basal characteristics, and their minimum ages for retreat from the LGM.

At the LGM, palaeo-ice streams in West Antarctica and the Antarctic Peninsula extended to the shelf edge, whereas in East Antarctica ice was typically (although not universally) restricted to the mid-outer shelf. All of the known palaeo-ice streams occupied cross-shelf bathymetric troughs of variable size, dimension and gradient, and were distinguished by a range of glacial bedforms (see Tables 3 & Fig. 5). Typically, the outer shelf zone of Antarctic palaeo-ice streams is characterised by unconsolidated sediment, which can often be further sub-divided into soft (upper) and stiff (lower) till units. Where unconsolidated sediment is present, and this is sometimes as patches of till on the inner shelf, MSGLs and GZWs are commonly observed. The inner shelf, by contrast, is generally characterised by crystalline bedrock and higher bed roughness. It is on this more rigid substrate that drumlins, gouged and grooved bedrock and meltwater channels are commonly observed.

The retreat history of the Antarctic Ice Sheet since the LGM has been characterised by significant variability, with palaeo-ice stream systems responding asynchronously to both external and internal forcings (cf. Fig. 10). This includes both the response of palaeo-ice streams to initial triggers of atmospheric warming, oceanic warming and sea-level rise, and the subsequent pattern of retreat back to their current grounding-line positions. Thus, the recent spatial and temporal variability exhibited by ice streams over short (decadal) timescales (e.g. Truffer & Fahnestock, 2007) can actually be placed within a much longer record of asynchronous retreat. Whilst grounding line retreat may be triggered, and to some extent paced, by external factors, the individual characteristics of each ice stream will, nonetheless, modulate this retreat. Consequently, some ice streams will retreat rapidly, whereas others will retreat more slowly, even under the same climate forcing. It is therefore imperative that ice stream behaviour and grounding-line retreat is treated as unique to each ice stream and this highlights the importance of obtaining knowledge of their subglacial bed properties and bed geometry for constraining future ice stream behaviour.

The inherent association linking subglacial bedform genesis with subglacial processes permits the geomorphological signature to be used as a proxy for reconstructing ice stream retreat behaviour. Based on preliminary research into the retreat styles and characteristics of individual ice streams (Section 5.2), it seems that ice streams with small drainage basins and steep reverse slopes are most sensitive to rapid deglaciation. In contrast, palaeo-ice streams with large drainage basins were generally the slowest to deglaciate.

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Figures:

Fig. 1: Locations of the main palaeo-ice streams known on the Antarctic continental shelf. Approximate locations of ice streams are depicted by a black arrow and the numbers refer to the corresponding citation and evidence outlined in Table 1.

Fig. 2: Typical bathymetric long profiles taken along the axial length of: A: Pine Island (Eastern Trough) palaeo-ice stream (from modern ice-front); B: Marguerite Bay palaeo-ice stream (from ice-shelf front); C: Belgica Trough (Eltanin Bay palaeo-ice stream (from ice-shelf front); and D: Anvers-Hugo Island palaeo-ice stream (from modern ice front). Vertical exaggeration 90:1.

Fig. 3: TOPAS sub-bottom profiler records showing acoustic sedimentary units from outer Marguerite Bay: A: cross-line showing MSGL formed in acoustically transparent sediment; B: cross-line showing MSGL formed in acoustically transparent sediment, with a locally grooved sub-bottom reflector and sediment drape; and C: line parallel to trough long axis showing MSGL formed in acoustically transparent sediment (Reprinted from Ó Cofaigh et al. 2005b, with permission from Elsevier).

Fig. 4: Examples of glacial geomorphic landforms identified at the beds of marine-based Antarctic palaeo-ice streams: A: MSGLs from the outer shelf of Marguerite Bay (Reprinted from Ó Cofaigh et al. 2008, with permission from Wiley); B: Grooved and gouged bedrock on the mid-shelf of Marguerite Bay (light direction from the NE; x8 vertical exaggeration); C: Drumlins from Belgica Trough showing cresentric overdeepenings (light direction from the E; x8 vertical exaggeration); D: Grounding zone wedges in Marguerite Trough (note the subtle change in lineation direction across the GZW) (light direction from the N; x8 vertical exaggeration); E: morainal ridges in the Eastern Basin of the Ross Sea, orientated oblique to ice-flow direction (modified from Mosola & Anderson, 2006); F: Channel network cut into bedrock of the Palmer Deep Outlet Sill (Reprinted from Domack et al. 2006, with permission from Elsevier).

Fig. 5: A landsystem model of palaeo-ice stream retreat, Antarctica. The three panels represent different retreat styles: rapid (A), episodic (B) and slow (C) (Reprinted from Ó Cofaigh et al. 2008, with permission from Wiley).

Fig. 6: A landsystem model (based on Canals et al. 2002; Wellner et al. 2001, 2006) showing the general distribution of glacial landforms associated with a typical (Getz-Dotson) marine palaeo-ice stream on the continental shelf of Antarctica (Reprinted from Graham et al. 2009, with permission from Elsevier).

Fig. 7: Sediment core locations of (minimum) ages for post-LGM ice sheet retreat (see Table 4 for corresponding details): A: Antarctica; B: Antarctic Peninsula; and C: Ross Sea.

Fig. 8: Chronology of initial deglaciation for Antarctic palaeo-ice streams and comparison with climate proxy records and bathymetric conditions for the period 32-5 ka BP: A: Colour coded map illustrating initial timing of retreat from the Antarctic continental shelf since the LGM (dates in black refer to the dots and represent initial retreat of the palaeo-ice streams; dates in grey refer to initial retreat of the ice sheet). The black line shows the reconstructed position of grounded ice at the LGM. The dotted line indicates approximate grounding-line positions due to a paucity of data. B: Timing of deglaciation for marine palaeo-ice streams around the Antarctic shelf are distinguished by region with one sigma error. The grey line is the EPICA Dome C δD ice core record which is used as a proxy for temperature, whilst the grey bands refer to periods of rapid eustatic sea-level rise. Bed gradient of the outer shelf (– gradient = negative gradient; + gradient = positive gradient); Width geometry of the outer shelf (= geometry = a constant trough width; > geometry = narrowing trough; < = widening trough). C: Plot of trough width and trough depth against timing of initial deglaciation for circum-Antarctic palaeo-ice streams. Trough depth has been adjusted to account for the effect of isostasy at the LGM (Whitehouse et al. in prep).

Fig. 9: Deglacial history of Antarctic palaeo-ice streams by sediment facies and carbon source (Hollow circles: AIO; Filled-in circles: Carbonate; red: Antarctic Peninsula; blue: East Antarctica; and green: West Antarctica; TGM = transitional glaciomarine).

Fig. 10: Retreat chronologies of: A: Anvers palaeo-ice stream (the core at 0 km [DF86-83] indicates when Gerlache Strait was ice free); B: Getz-Dotson Trough; C: Drygalski Basin palaeo-ice stream. The oldest date is a carbonate age from the outer shelf of Pennell Trough; HWD03-2 is a hot-water drill core taken from Ross Ice Shelf [Mckay et al. 2008]; The terrestrial date is from algal remains at Hatherton Glacier [Bockheim et al. 1989]; D: Marguerite Bay palaeo-ice stream (terrestrial dates are from shells and foraminifera along the margin of George VI Ice Shelf [Sugden & Clapperton, 1981; Hjort et al. 2001; Smith et al. 2007]); and E: Belgica Trough palaeo-ice stream. Hollow circles refer to AIO carbon; filled-in circles indicate a carbonate source; and squares are additional terrestrial dating constraints. Red = Transitional Glaciomarine; Green = Iceberg Turbate; Blue = Diatomaceous Ooze; Diamict = Black; and Purple = glaciomarine. The dotted line is the predicted retreat history as based on the most reliable dates. Reliable ages are determined by the carbon source and the sediment sampled.

Fig. 1a

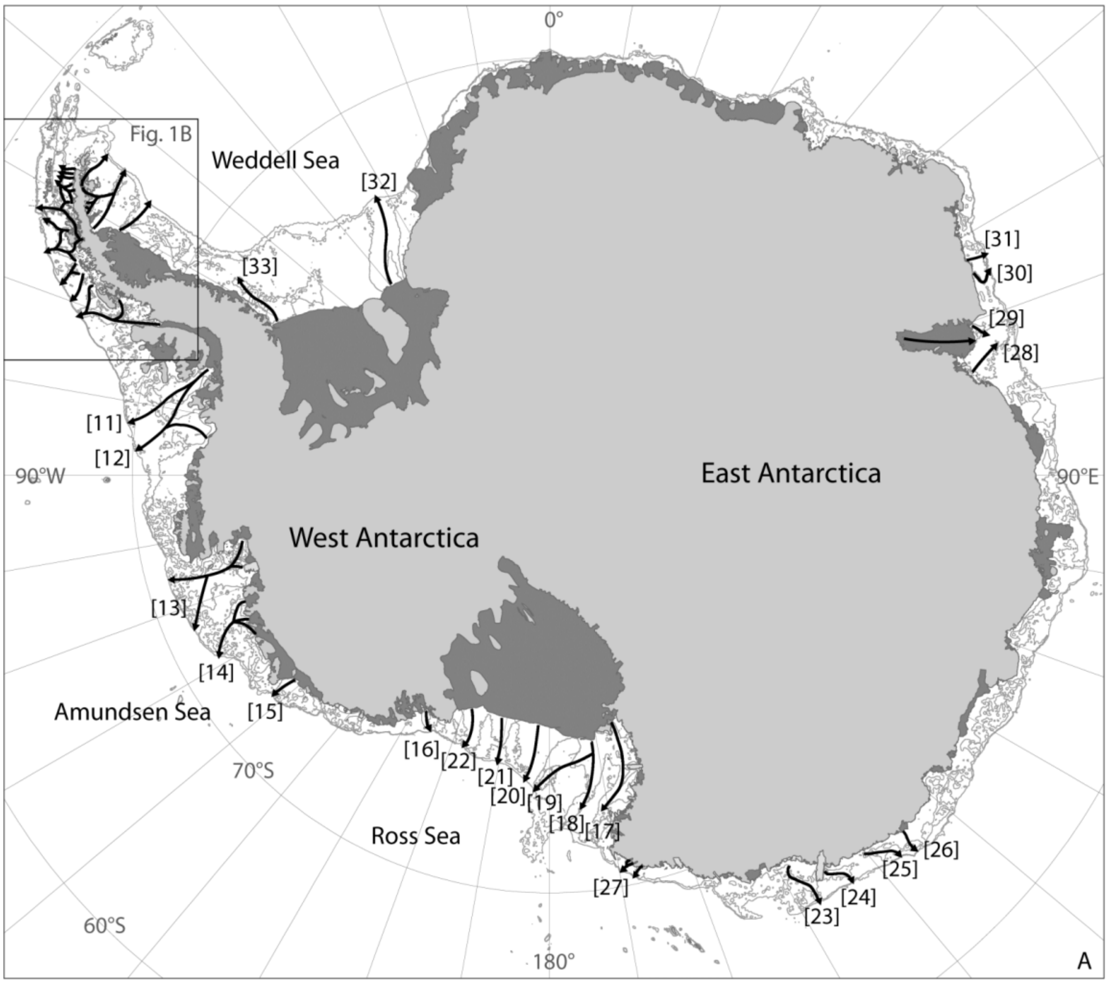
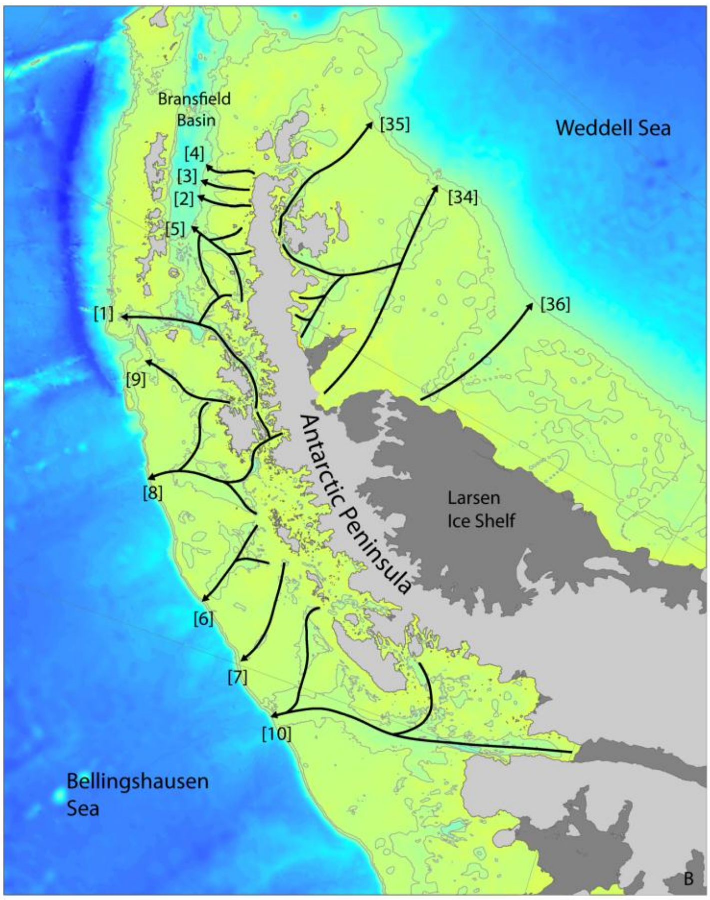

0°
Fig. 1B
Weddell Sea
[32]
[31]
[30]
[33]
[29]
[28]
[11]
[12]
90°W
90°E
East Antarctica
West Antarctica
[13]
[14]
Amundsen Sea
[15]
[16]
[22]
[21]
[20]
[19]
[18]
[17]
70°S
Ross Sea
[26]
[25]
[24]
[27]
[23]
60°S
180°
A

Fig. 1b


Bransfield Basin
[35]
Weddell Sea
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Antarctic Peninsula
Larsen Ice Shelf
[8]
[6]
[7]
[10]
Bellingshausen Sea
B

Fig. 2

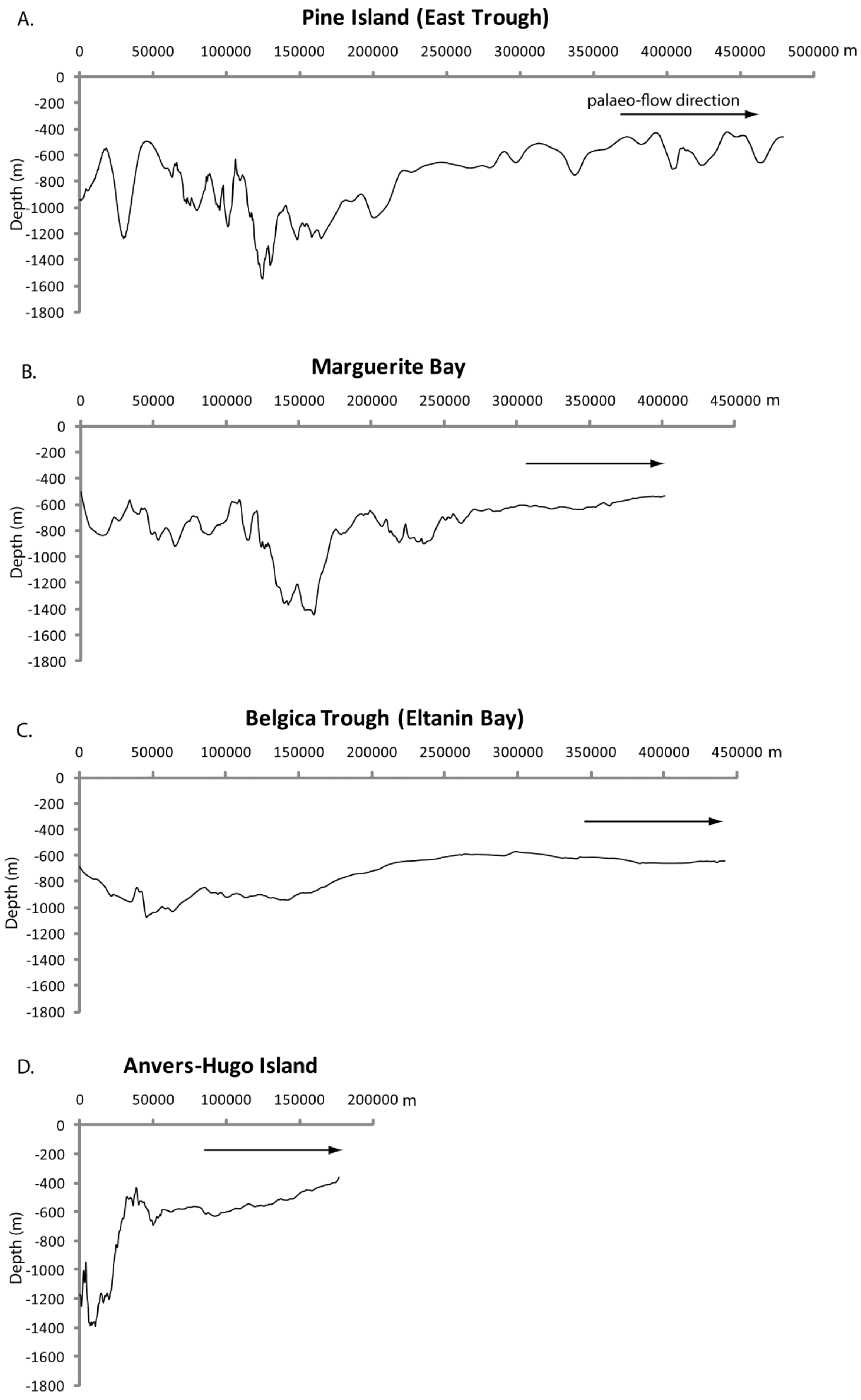

A.
Pine Island (East Trough)
0 50000 100000 150000 200000 250000 300000 350000 400000 450000 500000 m
0 -200 -400 -600 -800 -1000 -1200 -1400 -1600 -1800
Depth (m)
palaeo-flow direction
B.
Marguerite Bay
0 50000 100000 150000 200000 250000 300000 350000 400000 450000 m
0 -200 -400 -600 -800 -1000 -1200 -1400 -1600 -1800
Depth (m)
C.
Belgica Trough (Eltanin Bay)
0 50000 100000 150000 200000 250000 300000 350000 400000 450000 m
0 -200 -400 -600 -800 -1000 -1200 -1400 -1600 -1800
Depth (m)
D.
Anvers-Hugo Island
0 50000 100000 150000 200000 m
0 -200 -400 -600 -800 -1000 -1200 -1400 -1600 -1800
Depth (m)

Fig. 3

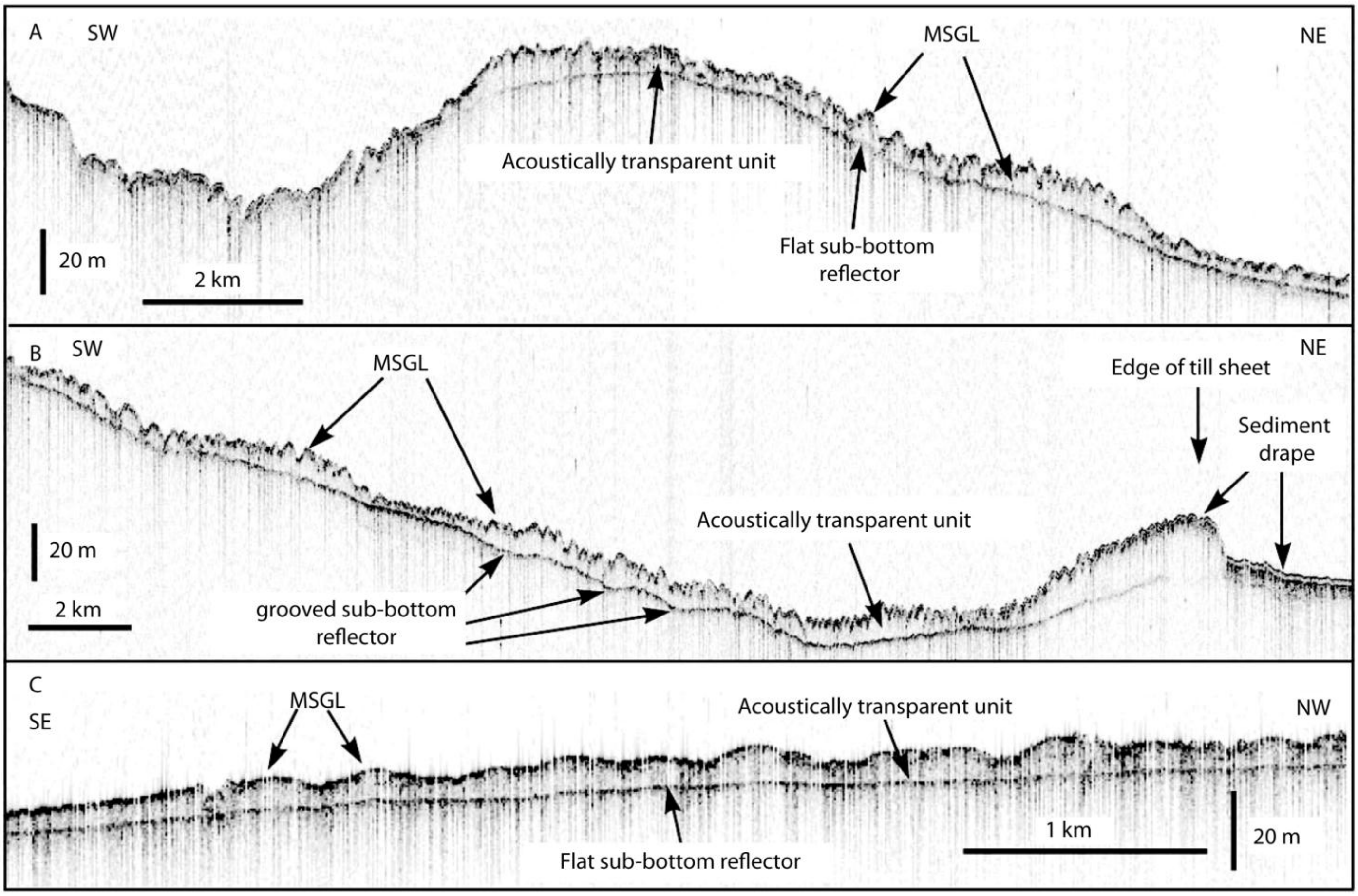
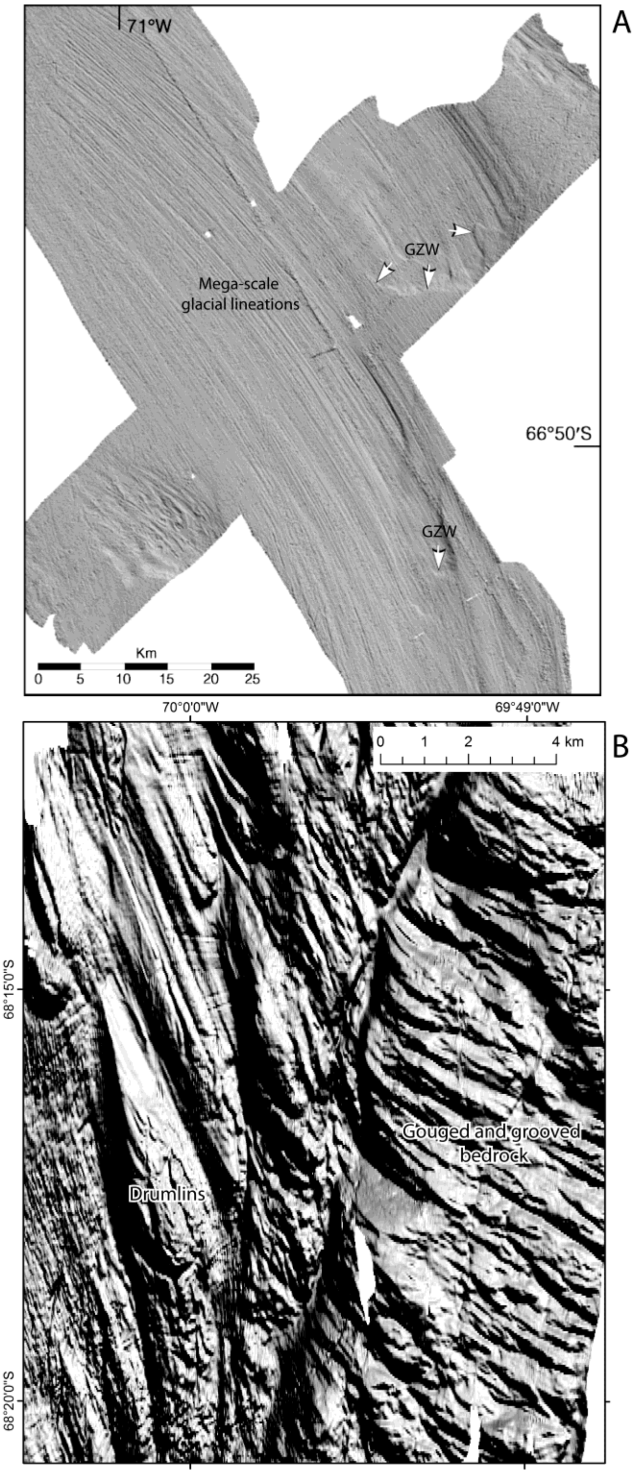
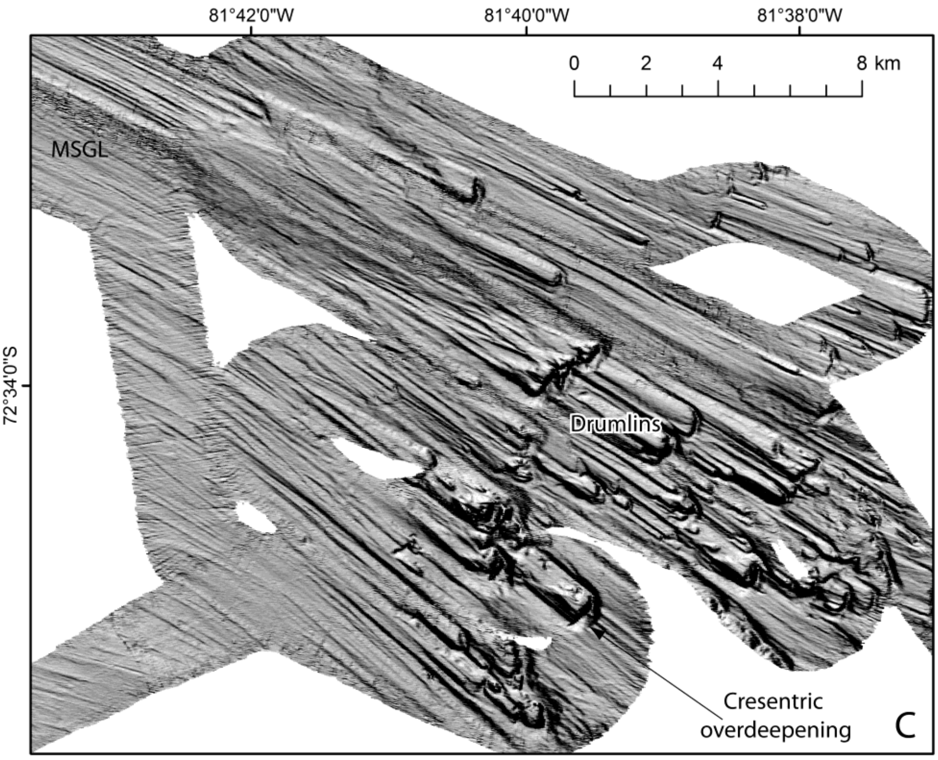

A
SW
NE
MSGL
Acoustically transparent unit
Flat sub-bottom reflector
20 m
2 km
B
SW
NE
MSGL
Edge of till sheet
Sediment drape
Acoustically transparent unit
20 m
2 km
grooved sub-bottom reflector
C
SE
NW
MSGL
Acoustically transparent unit
Flat sub-bottom reflector
1 km
20 m

Fig. 4a-b



71°W
A
GZW
Mega-scale
glacial lineations
66°50′S
GZW
Km
0 5 10 15 20 25
70°0'0"W
69°49'0"W
B
0 1 2 4 km
68°15'0"S
Gouged and grooved
bedrock
Drumlins
68°20'0"S

Fig. 4c-d


81°42'0"W
81°40'0"W
81°38'0"W
0
2
4
8 km
MSGL
72°34'0"S
Drumlins
Cresentric overdeepening
C

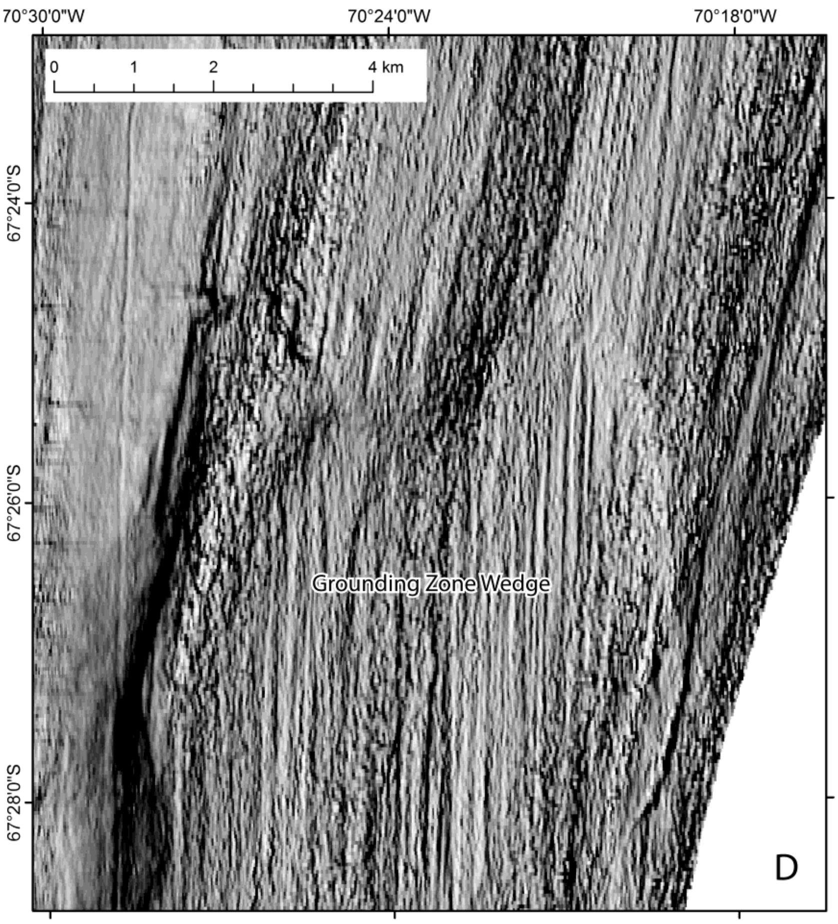

70°30'0"W
70°24'0"W
70°18'0"W
0
1
2
4 km
67°24'0"S
67°26'0"S
67°28'0"S
Grounding Zone Wedge
D

Fig. 4e-f

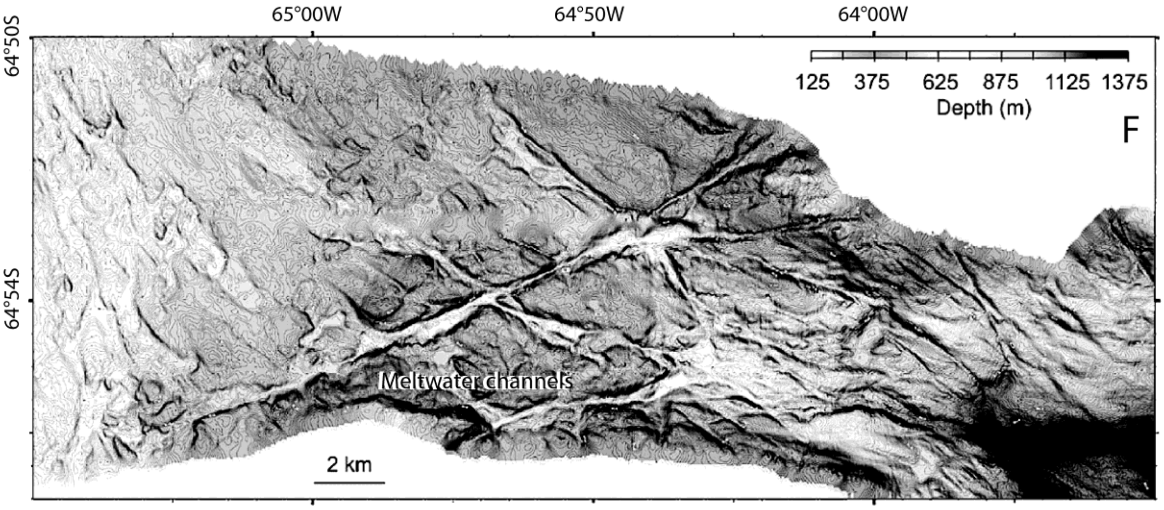

65°00W
64°50W
64°00W
64°50S
64°54S
125 375 625 875 1125 1375
Depth (m)
F
Meltwater channels
2 km

Fig. 5

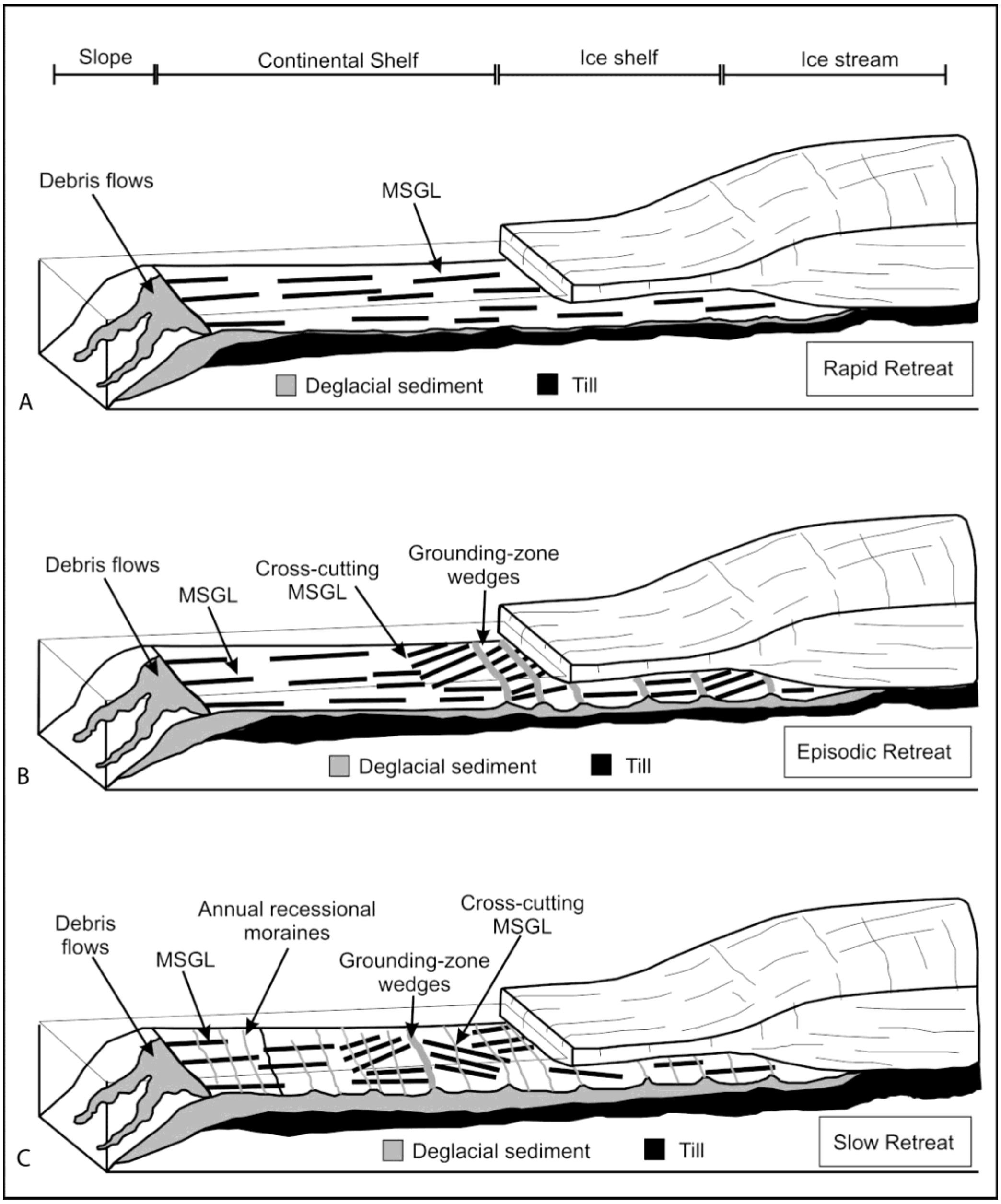
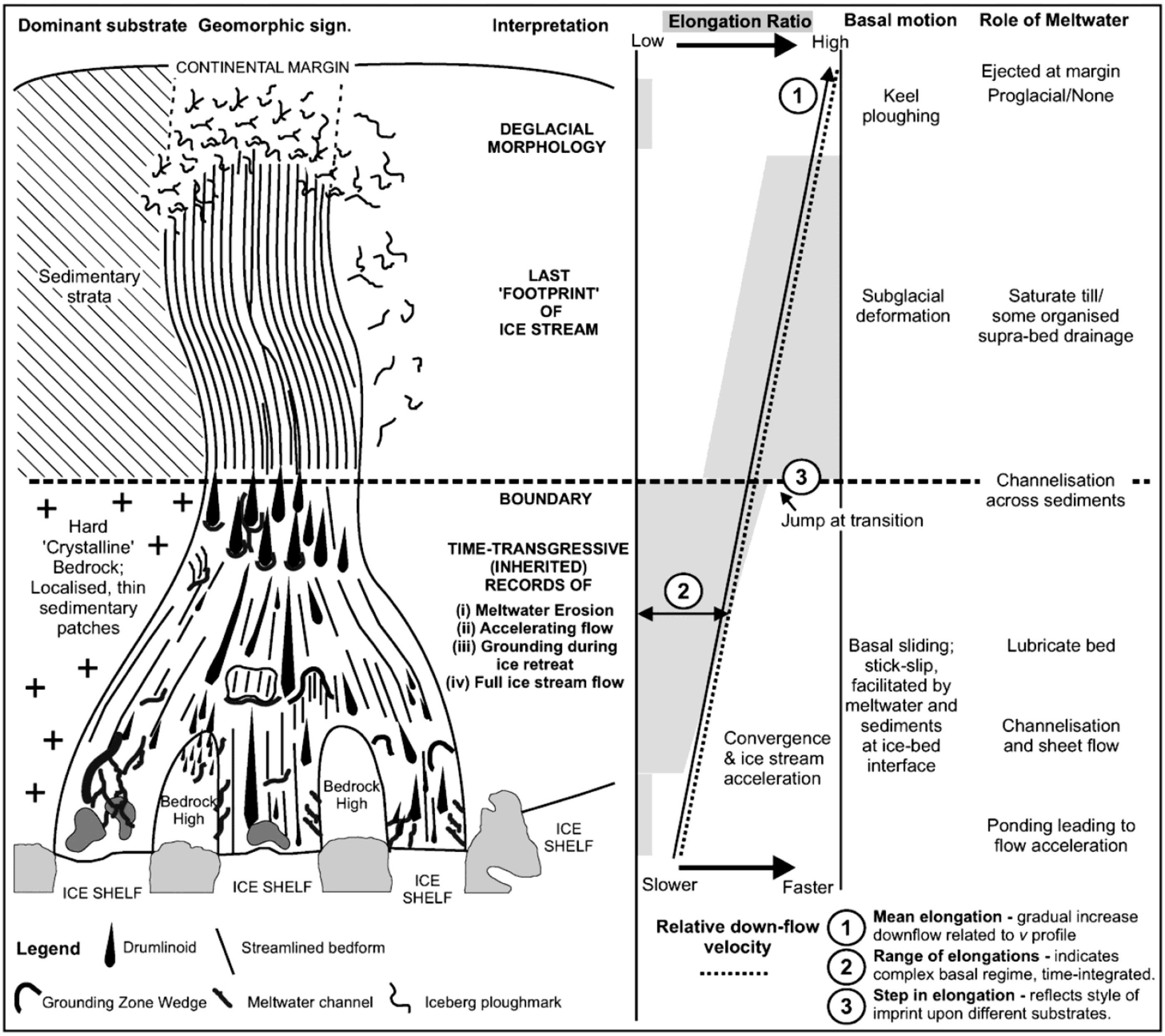
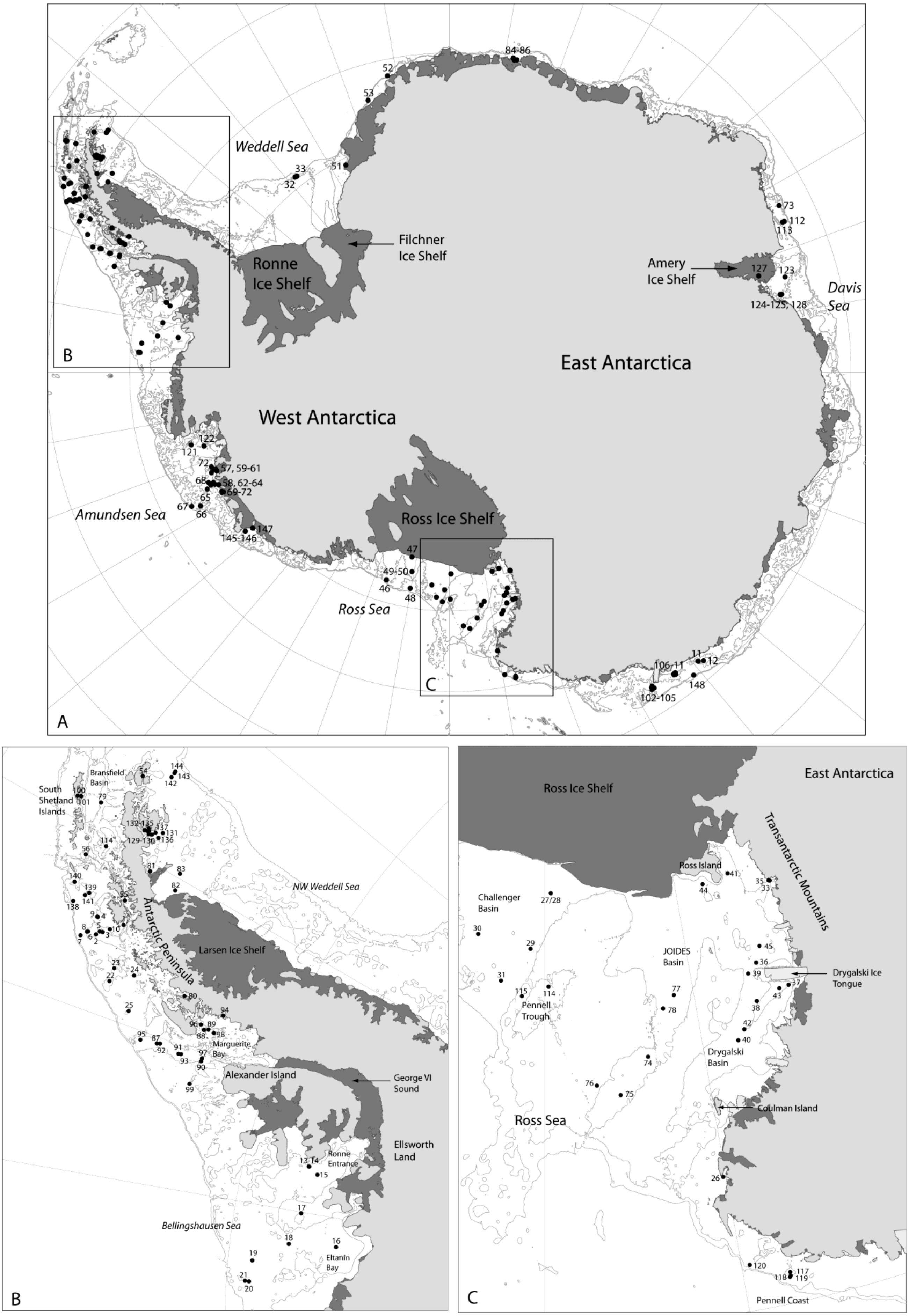

Slope
Continental Shelf
Ice shelf
Ice stream
Debris flows
MSGL
Deglacial sediment
Till
Rapid Retreat
A
Debris flows
MSGL
Cross-cutting MSGL
Grounding-zone wedges
Deglacial sediment
Till
Episodic Retreat
B
Debris flows
MSGL
Annual recessional moraines
Grounding-zone wedges
Cross-cutting MSGL
Deglacial sediment
Till
Slow Retreat
C

Fig. 6



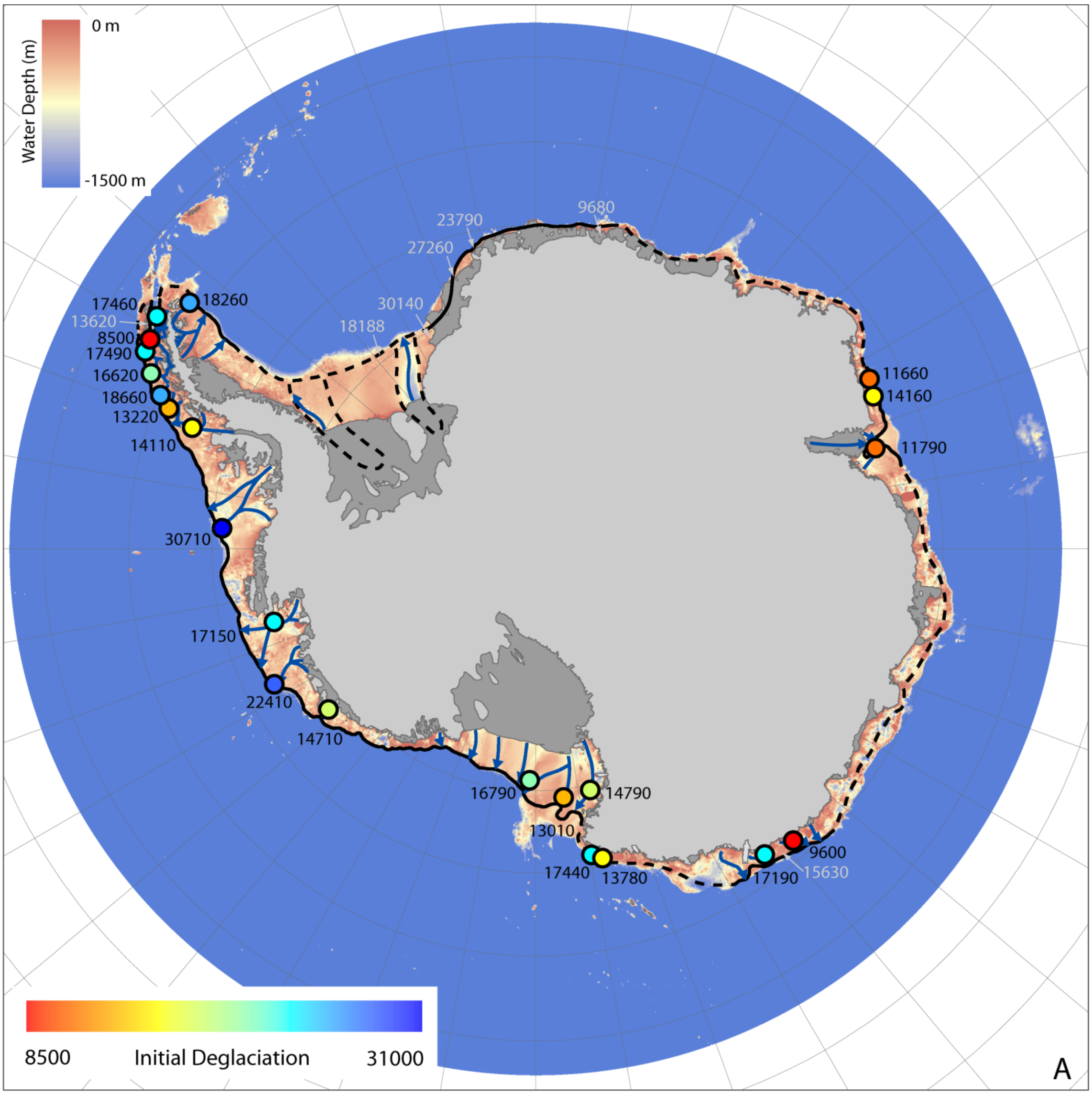
Dominant substrate
Geomorphic sign.
Interpretation
Elongation Ratio
Low
High
Basal motion
Role of Meltwater
CONTINENTAL MARGIN
Sedimentary strata
DEGLACIAL MORPHOLOGY
LAST 'FOOTPRINT' OF ICE STREAM
Keel ploughing
Ejected at margin
Proglacial/None
Subglacial deformation
Saturate till/ some organised supra-bed drainage
BOUNDARY
Channelisation across sediments
Jump at transition
Hard 'Crystalline' Bedrock; Localised, thin sedimentary patches
TIME-TRANSGRESSIVE (INHERITED) RECORDS OF
(i) Meltwater Erosion
(ii) Accelerating flow
(iii) Grounding during ice retreat
(iv) Full ice stream flow
Basal sliding; stick-slip, facilitated by meltwater and sediments at ice-bed interface
Lubricate bed
Convergence & ice stream acceleration
Channelisation and sheet flow
Ponding leading to flow acceleration
Bedrock High
Bedrock High
ICE SHELF
ICE SHELF
ICE SHELF
ICE SHELF
ICE SHELF
Slower
Faster
Legend
Drumlinoid
Streamlined bedform
Grounding Zone Wedge
Meltwater channel
Iceberg ploughmark
Relative down-flow velocity
1 Mean elongation - gradual increase downflow related to v profile
2 Range of elongations - indicates complex basal regime, time-integrated.
3 Step in elongation - reflects style of imprint upon different substrates.

Fig. 7



Weddell Sea
Filchner Ice Shelf
Ronne Ice Shelf
Amery Ice Shelf
Davis Sea
East Antarctica
West Antarctica
Amundsen Sea
Ross Ice Shelf
Ross Sea
A
B
C
Bransfield Basin
South Shetland Islands
NW Weddell Sea
Antarctic Peninsula
Larsen Ice Shelf
Marguerite Bay
Alexander Island
George VI Sound
Ellsworth Land
Ronne Entrance
Bellingshausen Sea
Eltanin Bay
Ross Ice Shelf
East Antarctica
Transantarctic Mountains
Ross Island
Challenger Basin
JOIDES Basin
Drygalski Ice Tongue
Pennell Trough
Drygalski Basin
Coulman Island
Ross Sea
Pennell Coast

Fig. 8



Water Depth (m)
0 m
-1500 m
9680
23790
27260
30140
18188
17460
13620
8500
17490
16620
18660
13220
14110
18260
11660
14160
11790
30710
17150
22410
14710
16790
14790
13010
17440
13780
17190
9600
15630
8500
Initial Deglaciation
31000
A

Fig. 8b-c

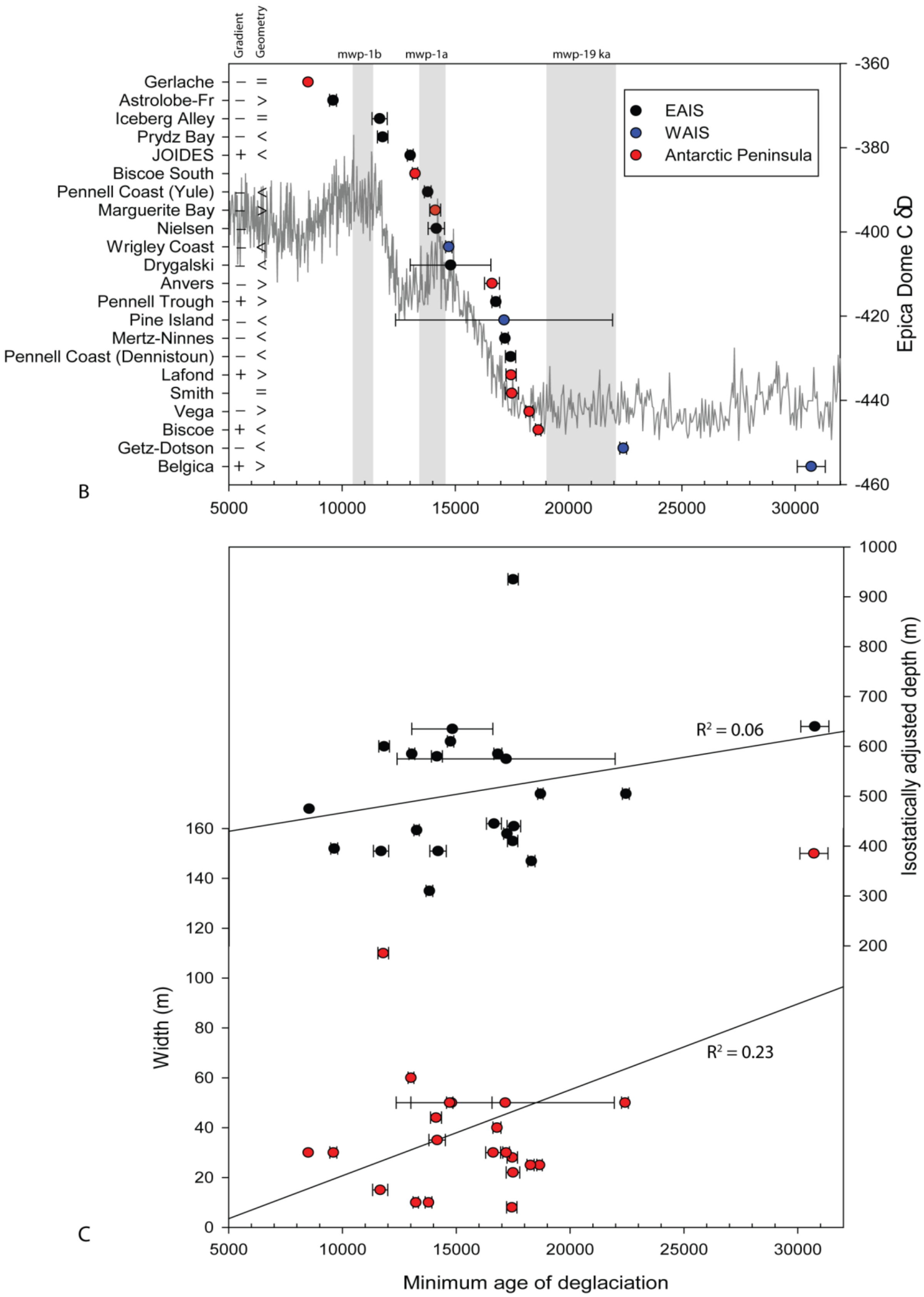
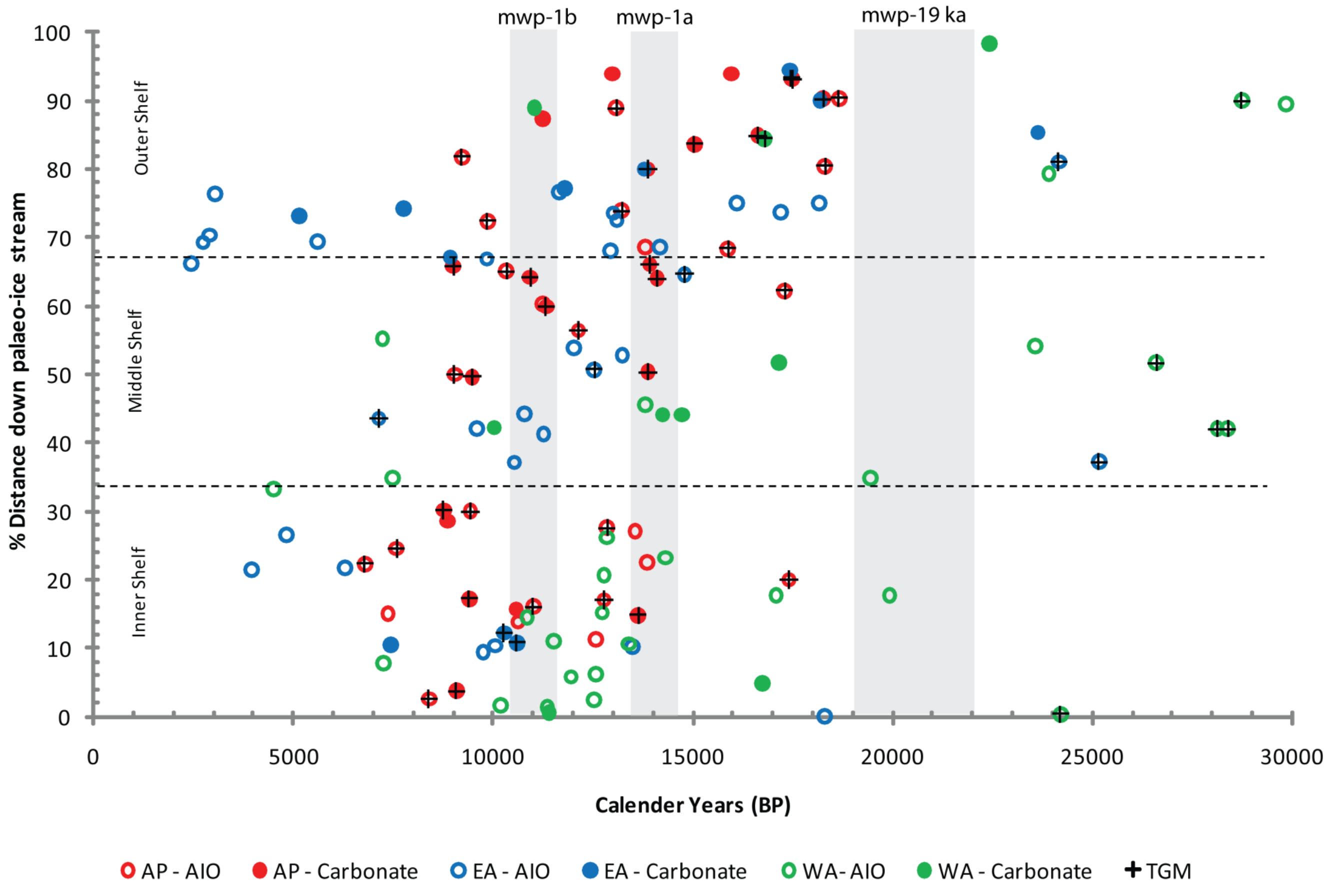
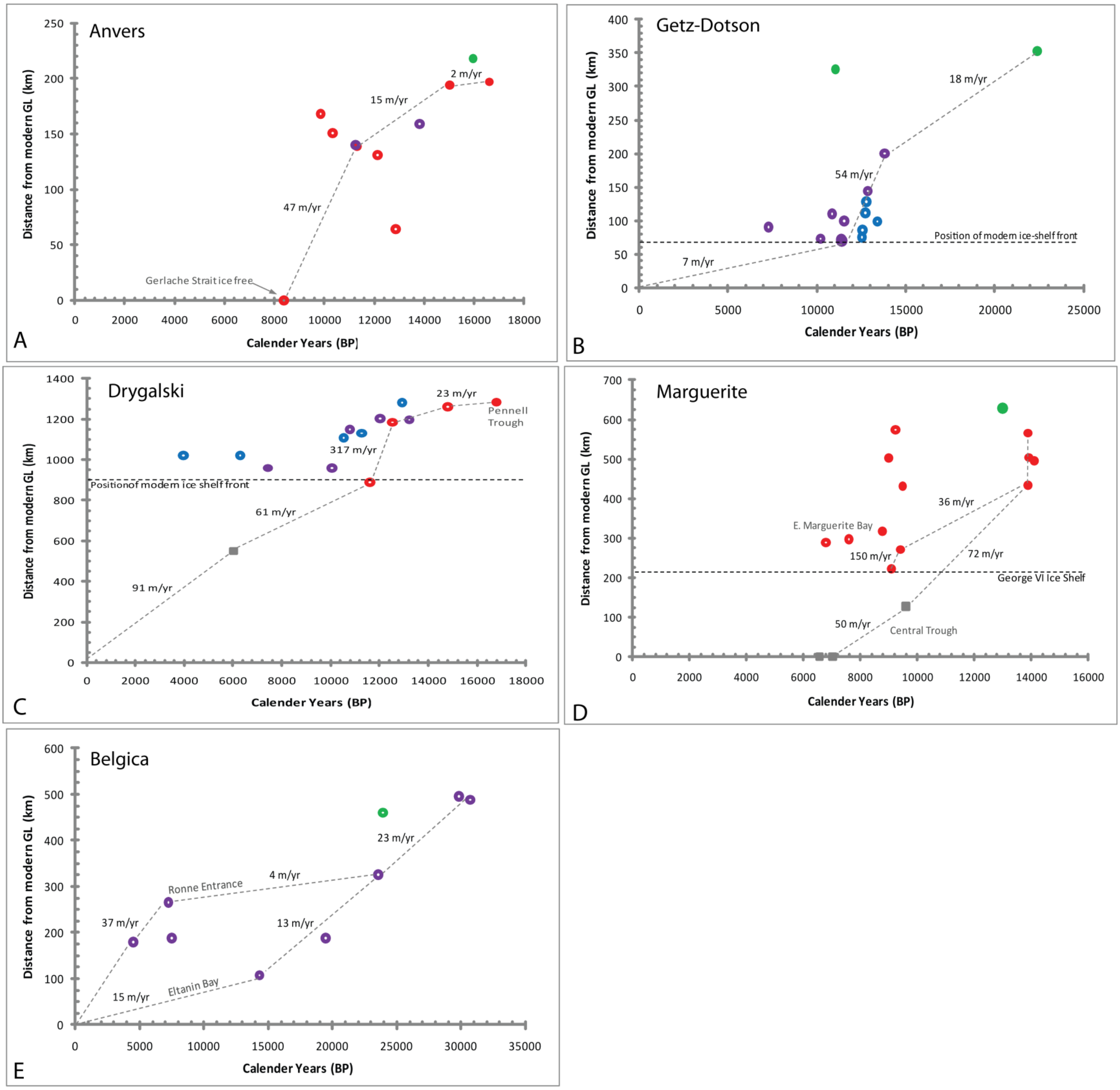

Gradient
Geometry
mwp-1b
mwp-1a
mwp-19 ka
Gerlache – =
Astrolobe-Fr – >
Iceberg Alley – =
Prydz Bay – <
JOIDES + <
Biscoe South
Pennell Coast (Yule) – <
Marguerite Bay – >
Nielsen –
Wrigley Coast – <
Drygalski – <
Anvers – >
Pennell Trough + >
Pine Island – <
Mertz-Ninnes – <
Pennell Coast (Dennistoun) – <
Lafond + >
Smith =
Vega – >
Biscoe + <
Getz-Dotson – <
Belgica + >
EAIS
WAIS
Antarctic Peninsula
Epica Dome C δD
-360
-380
-400
-420
-440
-460
B
5000
10000
15000
20000
25000
30000
R² = 0.06
R² = 0.23
Width (m)
Isostatically adjusted depth (m)
Minimum age of deglaciation
C

Fig. 9



mwp-1b
mwp-1a
mwp-19 ka
% Distance down palaeo-ice stream
Outer Shelf
Middle Shelf
Inner Shelf
100
90
80
70
60
50
40
30
20
10
0
0
5000
10000
15000
20000
25000
30000
Calender Years (BP)
AP - AIO
AP - Carbonate
EA - AIO
EA - Carbonate
WA- AIO
WA - Carbonate
TGM

Fig. 10



Anvers
Distance from modern GL (km)
Calender Years (BP)
2 m/yr
15 m/yr
47 m/yr
Gerlache Strait ice free
A
Getz-Dotson
18 m/yr
54 m/yr
7 m/yr
Position of modern ice-shelf front
B
Drygalski
23 m/yr
Pennell Trough
317 m/yr
Position of modern ice shelf front
61 m/yr
91 m/yr
C
Marguerite
36 m/yr
E. Marguerite Bay
150 m/yr
72 m/yr
George VI Ice Shelf
50 m/yr
Central Trough
D
Belgica
23 m/yr
4 m/yr
Ronne Entrance
37 m/yr
13 m/yr
Eltanin Bay
15 m/yr
E

Tables:

Table 1: Proposed palaeo-ice streams of the Antarctic Ice Sheet during the last glacial and main lines of evidence used in their identification (numbers in square brackets refer to their location in Fig. 1).

Table 2: Physiography of the identified Antarctic palaeo-ice streams (collated from the literature: Table 1).

Table 3: Geomorphic features observed at the former beds of Antarctic marine ice streams.

Table 4: Compiled uncorrected and calibrated marine radiocarbon ages representing minimum estimates of glacial retreat.

Table 5: Retreat rates of palaeo-ice streams

Table 6: Palaeo-ice streams, their inferred style of retreat and the evidence for this.

Table 1: Proposed palaeo-ice streams of the Antarctic Ice Sheet during the last glacial period and the main lines of evidence used in their identification (numbers in square brackets refer to their location in Fig. 1, whilst those palaeo-ice streams with a question mark are less certain).

REFERENCES	ICE STREAM	DRAINAGE BASIN	EXTENT AT LGM	PRINCIPLE EVIDENCE FOR ICE STREAM ACTIVITY
Canals et al. (2000, 2003); Willmott et al. (2003) Evans et al. (2004); Heroy & Anderson (2005).	[1] Gerlache-Boyd	Western Bransfield Basin	Shelf-break	A convex-up elongate sediment body comprising parallel to sub parallel ridges and grooves (bundles) up to 100 km long; a convergent ice-flow pattern exhibiting a progressive increase in elongation into the main trough; and an outer-shelf sediment lobe seaward of the main trough.
Banfield & Anderson (1995); Canals et al. (2002); Heroy & Anderson (2005).	[2] Lafond	Central Bransfield Basin	Shelf-break	Deeply incised U-shaped trough with a drumlinised bed on the inner shelf and elongate grooves and ridges on the outer shelf; and a well developed shelf-edge lobe and slope debris apron.
Banfield & Anderson (1995); Canals et al. (2002); Heroy & Anderson (2005).	[3] Laclavere		Shelf-break	Deeply incised U-shaped trough with a drumlinised bed on the inner shelf and elongate grooves and ridges on the outer shelf; and a well developed shelf-edge lobe and slope debris apron.
Banfield & Anderson (1995); Canals et al. (2002); Heroy & Anderson (2005).	[4] Mott Snowfield		Shelf-break	Deeply incised U-shaped trough with a drumlinised bed on the inner shelf and elongate grooves and ridges on the outer shelf; and a well developed shelf-edge lobe and slope debris apron.
Bentley & Anderson (1998); Heroy et al. (2008)	[5] Orleans Trough		?	A large cross-shelf trough. Streamlined bedforms including drumlins and scalloped features.
Canals et al. (2003); Amblas et al. (2006).	[6] Biscoe Trough	Antarctic Peninsula	Shelf-break	Biscoe Trough exhibits a convergent flow pattern at the head of the ice stream, with well-developed MSGL observed throughout. These bedforms show a progressive elongation towards the shelf edge, with the less elongate landforms on the inner shelf formed in bedrock and interpreted as roche moutonées.
Heroy & Anderson (2005); Wellner et al. (2006).	[7] Biscoe South Trough (= Adelaide Trough)		?	A distinctive cross-shelf bathymetric trough characterised by rock cored drumlins on the inner shelf and MSGL on the outer shelf.
Pudsey et al. (1994); Larter & Vannester (1995); Vanneste & Larter (1995); Domack et al. (2006).	[8] Anvers-Hugo Island Trough		Shelf-break	Comprised of three tributaries which converge on a central trough. The inner-shelf is characterised by streamlined bedrock, and meltwater channels which cut across the mid-shelf high. The outer trough is floored by sediment and dominated by MSGL, with grounding zone wedges also identified in this zone.
Heroy & Anderson (2005).	[9] Smith Trough		?	A cross shelf bathymetric trough containing streamlined bedrock features such as grooves and drumlins, with elongations ratios of up to 20:1.
Kennedy & Anderson (1989); Anderson et al. (2001); Wellner et al. (2001);Oakes & Andeson (2002); Ó Cofaigh et al. (2002, 2005b, 2007, 2008); Dowdeswell et al. (2004a, b); Anderson & Oakes-Fretwell (2008); Noormets et al. (2009); Kilfeather et al. (2010).	[10] Marguerite Trough	Marguerite Bay	Shelf-break	Streamlined subglacial bedforms occur in a cross-shelf bathymetric trough; the bedforms show a progressive down-flow evolution from bedrock drumlins and ice-moulded bedrock on the inner shelf to MSGL on the outer shelf in soft sediment; and the MSGL are formed in subglacial deformation till which is not present on the adjacent banks.

Ó Cofaigh et al. (2005a); Graham et al. (2010).	[11] Latady Trough	Ronne Entrance	?	MSGL located in a cross-shelf bathymetric trough.
Ó Cofaigh et al. (2005a); Dowdeswell et al. (2008b); Noormets et al. (2009); Hillenbrand et al. (2009, 2010a); Graham et al. (2010b).	[12] Belgica Trough	Eltanin Bay & Ronne Entrance	Shelf-break	Elongate bedforms are located in a cross-shelf bathymetric trough; the head of the ice stream is characterised by a strongly convergent flow pattern; the trough exhibits a down-flow transition from drumlins to MSGL, with the MSGL formed in subglacial deformation till; and large sediment accumulations have been observed including a TMF in front of Belgica Trough and a series of GZWs on the mid and inner shelf.
Anderson et al. (2001); Wellner et al. (2001); Lowe & Anderson (2002, 2003); Dowdeswell et al. (2006); Evans et al. (2006); Ó Cofaigh et al. (2007); Noormets et al. (2009); Graham et al. (2010a).	[13] Pine Island Trough	Pine Island Trough	Shelf-break	MSGL with elongation ratios of >10:1 in the middle/outer shelf composed of soft till formed by subglacial deformation; the bedforms are concentrated in a cross-shelf bathymetric trough; and a bulge in the bathymetric contours in-front of the trough indicates progradation of the continental slope.
Anderson et al. (2001); Wellner et al. (2001); Larter et al. (2009); Graham et al. (2009); Hillenbrand et al. (2010b); Smith et al. (in press).	[14] Getz-Dotson	Bakutis Coast	Shelf-break	Bedforms converge into a central trough from three main tributaries; the bedforms which occupy the trough have elongation ratios up to 40:1 and comprise drumlins, crag-and-tails and MSGL; MSGL on the outer shelf are formed in soft till; and the inner and mid-shelf contain a series of GZWs.
Wellner et al. (2001, 2006); Anderson et al. (2002).	[15] Wrigley Gulf	Wrigley Gulf	Outer-shelf/shelf break	A cross-shelf bathymetric trough containing drumlins and grooves on the inner shelf and MSGL on the outer shelf.
Wellner et al. (2001, 2006).	[16] Sulzberger	Sulzberger Bay	?	A prominent trough aligned with the structural grain of the coast. The bedrock floored trough is characterised by roche moutonées and erosional grooves which are concentrated along the axis of the trough; MSGL occupy the trough on the outer shelf.
Anderson (1999); Domack et al. (1999); Shipp et al. (1999); Bart et al. (2000); Licht et al. (2005); Melis & Salvi (2009).	[17] Drygalski Basin (Trough 1)	Western Ross Sea	Outer shelf	A narrow trough containing MSGL formed in deformation till, a distinctive dispersal train and a GZW on the outer shelf.
Anderson (1999); Domack et al. (1999); Shipp et al. (1999, 2002); Bart et al. (2000); Anderson et al. (2001); Howat & Domack (2003); Licht et al. (2005); Farmer et al. (2006); Melis & Salvi (2009).	[18] JOIDES-Central Basin (Trough 2)		Outer shelf	JOIDES-Central Basin forms a narrow trough characterised by MSGL along its axial length. The trough also exhibits a distinctive dispersal train, and has a GZW on the outer shelf.
Domack et al. (1999); Shipp et al. (1999); Howat & Domack (2003); Licht et al. (2005); Mosola & Anderson (2006); Salvi et al. (2006).	[19] Pennell Trough (Trough 3)	Central Ross Sea	Shelf-break	A narrow cross-shelf trough characterised by a drumlinised inner shelf, with MSGL extending across the mid and outer shelf. The MSGL are formed in deformation till, with gullies on the continental shelf in-front of the trough. The trough also exhibits a distinctive dispersal train.
Domack et al. (1999); Shipp et al. (1999); Licht et al. (2005); Mosola & Anderson (2006).	[20] Eastern Basin (Trough 4)		Shelf-break	Cross-cutting MSGL, formed in deformation till, extend along the entire axis of the trough, with gullies on the continental slope in-front of the broad trough. GZWs have also been identified along the length of the trough. The trough is also characterised by a distinctive dispersal train.
Licht et al. (2005); Mosola & Anderson (2006).	[21] Eastern Basin (Trough 5)	Eastern Ross Sea	Shelf-break	The broad trough that hosted this ice stream is characterised by MSGL and multiple GZWs and terminates in gullies on the continental slope. The MSGL are formed in deformation till. The trough is also characterised by a distinctive dispersal train.

Licht et al. (2005); Mosola & Anderson (2006).	[22] Eastern Basin (Trough 6)		Shelf-break	Trough 6 is a broad depression comprising MSGL and 3 GZWs. The MSGL are formed in a deformation till and there is sharp lateral boundary into non-deformed till. The continental slope is dominated by gullies. The trough is also characterised by a distinctive dispersal train.
Barnes (1987); Domack (1987); Eittreim et al. (1995); Escutia et al. (2003); McCullen et al. (2006); Crosta et al. (2007).	[23] Mertz Trough	Wilkes Land Coast	?	A broad cross-shelf trough floored by deformation till and characterised by MSGL and GZWs. The outer shelf is characterised by prograding wedges, with steep foresets composed of diamict.
Barnes (1987); Domack (1987); Eittreim et al. (1995); Beaman & Harris (2003, 2005); Escutia et al. (2003); Presti et al. (2005); Leventer et al. (2006); McCullen et al. (2006); Crosta et al. (2007); Denis et al. (2009).	[24] Mertz-Ninnes Trough		?	A broad cross-shelf trough floored by deformation till and characterised by MSGL and GZWs. The outer shelf is characterised by a number of GZWs, with steep foresets composed of diamict. Further evidence is provided by the presence of lateral moraines on the adjacent banks.
Eittreim et al. (1995); Escutia et al. (2003); Crosta et al. (2007); Denis et al. (2009).	[25] Astrolabe-Français		?	A broad cross-shelf trough. Steeply prograded GZW on the outer shelf and MSGL on the inner and mid shelf.
Eittreim et al. (1995); Escutia et al. (2003); Crosta et al. (2007).	[26] Dibble Trough		?	A broad cross-shelf trough. The outer shelf is characterised by GZWs, with steep foresets composed of diamict.
Wellner et al. (2006).	[27] Pennell Coast (?)	Pennell Coast	?	The shelf offshore of Pennell Coast has two prominent troughs which merge on the inner shelf; the troughs are characterised by erosional grooves (amplitudes of over 100 m and wavelengths of 10 m to 1 km).
O'Brien (1994); O'Brien et al. (1999, 2007); O'Brien & Harris (1996); Domack et al. (1998); Taylor & McMinn (2002); Leventer et al. (2006).	[28] Prydz Channel	Prydz Bay	Inner-shelf/mid-shelf	The shelf is bisected by a large trough with elongate bedforms (flutes) along its floor. The flutes are overprinted by a series of transverse moraines; in front of the trough there is a bulge in the bathymetric contours, typical of a TMF; and GZWs have been observed on the inner shelf.
O'Brien (1994); O'Brien et al. (1999, 2007); O'Brien & Harris (1996); Domack et al. (1998); Taylor & McMinn (2002); Leventer et al. (2006).	[29] Amery		Inner-shelf/mid-shelf	The shelf is bisected by a large trough with elongate bedforms (flutes) along its floor. The flutes are overprinted by a series of transverse moraines; in front of the trough there is a bulge in the bathymetric contours, typical of a TMF; and GZWs have been observed on the inner shelf.
Harris & O'Brien (1996, 1998); Leventer et al. (2006); Mackintosh et al. (2011).	[30] Nielsen	Mac.Robertson Land	Outer shelf/shelf break	A deep trough that strikes across the shelf. The trough contains GZWs, MSGL and parallel grooves.
O'Brien et al. (1994); Harris & O'Brien (1996); Stickley et al. (2005); Leventer et al. (2006); Mackintosh et al. (2011).	[31] Iceberg Alley		Outer shelf/shelf break	A narrow trough containing a GZW and MSGL.
Melles et al. (1994); Bentley & Anderson (1998); Bart et al. (1999); Anderson & Andrews (1999); Anderson et al. (2002); Bentley et al. (2010).	[32] Crary Trough (?)	Southern Weddell Sea	?	A broad trough on the continental shelf of SE Weddell Sea and an oceanward-convex bulge in the bathymetric contours in front of the trough (TMF).
Haase (1986); Bentley & Anderson (1998).	[33] Ronne Trough (?)		?	Shallow trough on the inner shelf

Gilbert et al. (2003); Evans et al. (2005); Brachfeld et al. (2003); Domack et al. (2005); Pudsey et al. (2006); Curry & Pudsey (2007); Ó Cofaigh et al. (2007); Reinardy et al. (2009, 2011a,b).	[34] Robertson Trough (Prince Gustav channel; Larsen-A, -B, BDE, Greenpeace)	NW Weddell Sea	Shelf-break	A strong convergence of multiple tributaries (and bedforms) into a large central trough on the outer shelf; bedforms which range from short bedrock drumlins, grooves and lineations on the inner shelf to MSGL on the outer shelf are confined to the cross-shelf troughs; bedforms show a progressive elongation down-flow and where formed in sediment are associated with soft till; prominent GZWs on the inner shelf document still-stand positions.
Anderson et al. (1992); Bentley & Anderson, (1998); Carmelenghi et al. (2001); Heroy & Anderson (2005).	[34] North Prince Gustav channel-Vega Trough		Outer shelf/shelf-break	Subglacial bedforms, including mega-flutes, drumlins, crag-and-tails have been identified within a deep trough which broadens towards the shelf edge. On the outer shelf MSGL are common and there is a prominent GZW.
Bentley & Anderson (1998).	[35] Jason Trough (?)		?	A large cross-shelf trough.

Table 2: Physiography of the palaeo-ice stream troughs collated from the literature (see Table 1). * Derived from GEBCO; [1] the gradient (degrees) is averaged from a long profile extracted along the axial length of the trough (from the shelf edge to the modern ice-front) using the GEBCO data. [2] From Graham et al. (2010). [3] Grounding of ice in Ronne Trough and Crary Trough at the LGM is disputed. [4] Crary Trough =Thiel Trough =Filchner Trough. EB = Eltanin Bay; RE = Ronne Entrance; PG = Prince Gustav channel; R = Robertson Trough.

Palaeo-ice stream trough	Length (km)	Width (km)	Major tributaries	Drainage Basin (km^2)	Water depth (m)				Gradient[1] (main trough)
					Shelf-break	Mid-outer shelf	Inner shelf	banks	
[1] Gerlache-Boyd	340	5-40	2	23,000	400-500	500-800	1200	300-400	-0.0001
[2] Lafond	75*	10-28	1		650-900	700	200-610	100-200*	0.0048
[3] Laclavere	70*	10-28	1		650-900	700	200-610	100-200*	0.0067
[4] Mott Snowfield	70*	10-28	1		650-900	700	200-610	100-200*	0.0028
[5] Orleans*	150	10-35	3		750-800	500-800	550-800	100-300	0.0019
[6] Biscoe	170	23-70	1		450-500	300-450	600-800	200-350*	-0.0007
[7] Biscoe South (Adelaide)*	180	15-30	1		450-500	450-600	450-550	300-350	0.00005
[8] Anvers-Hugo Island	240	15-30	3		400-430	300-800	500-1400*	200-350	-0.0036 (outer: -0.0018)
[9] Smith*	190*	5-22	1		800	400-900	200-800	300-400	0.0003
[10] Marguerite	445	6-80	2	10,000-100,000	500-600	500-600	1000-1600	400-500	-0.0007 (outer: -0.0009)
[11] Latady	510	up to 80	1		400	600-800	600-1000	400-500	0.0001
[12] Belgica	490(EB) 540 (RE)	75-150	2	217,000-256,000	600-680	560-700	500-1200	400-500	-0.0009 (EB) -0.0003(RE)
[13] Pine Island	450	50-95	2	330,000	480-540	490-640	1000-1700	400-500	-0.0012 (west) -0.0008 (east) (outer: 0.015)[2]
[14] Getz-Dotson	290	17-65	3		500	600	1100-1600	350-450*	-0.0015
[15] Wrigley Gulf*	145	50-70	1		500-600	600-800	600-1000	200-450	-0.0015
[16] Sulzberger*	130	25	1		500	500-1300	600-900	200-400	-0.0033 (outer: -0.0089)
[17] Drygalski (1)	560*	45-65	1		500	600	800-1000	250	-0.0003
[18] JOIDES-Central (2)	470*	45-65	1	1.6 million & 265,000	450-550	500-620	800-1000	250	-0.0001
[19] Pennell (3)	400	100	1		500-600	600-700	600-800	500	-0.0006
[20] Eastern Basin (4)	300*	150-240	1		500-600	600-700	600-800	500	-0.0003
[21] Eastern Basin (5)	240*	100-200	1		500-600	600-700	600-800	500	-0.0001
[22] Eastern Basin (6)	200*	125	1		500-600	600-700	600-800	500	-0.0008
[23] Mertz	≤280	50-100	1		450-500	450-500	450-1000	<400	-0.0004
[24] Mertz-Ninnes	≤160	50	1		450-500	450-500	450-1000	<400	-0.0027
[25] Astrolabe-Français	230	40-80	1		300	600-850	600-1100	200-350	-0.0014
[26] Dibble	130	50-80	1		450-550	400-1000	300-500	200-350	0.0016
[27] Pennell Coast*	≤70	10-15	2		350	400-1200	600-1100	200-300	-0.0082
[28] Prydz Channel	220-350	150	1		500-600	600-800	700-800	100-400	-0.0015

[29] Amery	>450	150	2	1.48 million (present)	500-600	600-800	800-2200	100-400	-0.003
[30] Nielsen	≤140	30-40	1		250-350*	550-800*	600-1200	<200	-0.0053
[31] Iceberg Alley	103*	15*	1		300*	450-550	450-500	<150	0.0003
[32] Crary[4]	≤460	120-170*	1		630	550-800	650-1140	350-400*	-0.0017
[33] Ronne	≤300	50-140*	1		400-500*	400-600	500-600	350-400*	-0.0006[3]
[34] Robertson	310	25-100	5		450	400-550	500-1200	300-400	-0.001 (PG) 0.0004 (R)
[35] North Prince Gustav-Vega	≤300	5-25*	2		300-400*	400-500	350-1240	300	-0.0026
[36] Jason*	≤220	20-120	1		750-800	550-900	450-600	300-400	0.0013

Table 3: Geomorphic features observed at the beds of Antarctic marine palaeo-ice streams.

Landform	Defining characteristics	Palaeo-ice streams
Mega-scale glacial lineations (MSGL)/'bundle structures'	>10:1 elongation, parallel bedform sets formed in the acoustically transparent seismic unit.	All palaeo-ice streams except Smith Trough and Sulzberger Bay Trough
Drumlinoid bedforms	Lobate/teardrop/ovoid-shaped bedforms formed either wholly or partially in bedrock and occasionally with overdeepenings around their upstream heads.	Anver-Hugo Island Trough, Bakutis Coast, Belgica Trough, Biscoe South Trough, Bransfield Basin, Central Ross Sea, Gerlache-Boyd Strait, Getz-Dotson Trough, Marguerite Trough, Pine Island Trough, Robertson Trough, Sulzberger Bay Trough, Vega Trough
Crudely streamlined and grooved bedrock	Elongate grooves/ridges formed in bedrock.	Anver-Hugo Island Trough, Belgica Trough, Biscoe Trough, Getz Ice Shelf, Getz-Dotson, Marguerite Trough, Pennell Coast, Pine Island Trough, Robertson Trough, Smith Trough, Sulzberger Bay Trough
Crag-and-tails	Large bedrock heads with tails aligned in a downflow direction.	Getz-Dotson Trough, Marguerite Bay
Subglacial meltwater channel systems	Straight to sinuous channels with undulating long-axis thalwegs and abrupt initiation and termination points.	Anvers-Hugo Island Trough, Central Ross Sea, Getz-Dotson, Marguerite Trough, Pine Island Bay,
GZW (Grounding Zone Wedges)	Steep sea-floor ramps with shallow backslopes and wedge-like profiles. Formed within till and often associated with lineations, which frequently terminate at the wedge crests.	Anvers-Hugo Island Trough, Belgica Trough, Gerlache-Boyd Strait, Getz-Dotson, Iceberg Alley, Laclavere Trough, Lafond Trough, Marguerite Trough, Mertz Trough, Nielsen Trough, Pine Island Trough, Prydz Channel; Robertson Trough, Ross Sea troughs, Vega Trough
Transverse moraines	Transverse ridges, 1-10 m high with spacings of a few tens to hundreds of metres. Straight to sinuous in plan-form.	Eastern Ross Sea, JOIDES-Central Basin, Prydz Channel
TMF (Trough Mouth Fan)	Seaward bulging bathymetric contours, large glacigenic debris-flow deposits and prominent shelf progradation (prograding sequences in seismic profiles).	Belgica Trough, Crary Trough, Prydz Channel, western Ross Sea troughs
Gully/channel systems	Straight or slightly sinuous erosional features on the continental slope, which occasionally incise back into the shelf edge. The gully networks on the upper slope show a progressive organisation into larger and fewer channels down-slope.	Anvers-Hugo Island Trough, Belgica Trough, Biscoe Trough, Biscoe Trough South, Bransfield Basin, Gerlache-Boyd Strait, Marguerite Trough, Pine Island Trough, Robertson Trough, Smith Trough, Weddell Sea trough, western Ross Sea troughs
Iceberg scours	Straight to sinuous furrows, uniform scour depths, cross-cutting and seemingly random orientation.	All palaeo-ice streams

Table 4: Compiled uncorrected and calibrated marine radiocarbon ages representing minimum estimates of glacial retreat.
Astrolobe-Fr = Astrolabe-Français Trough

[a]For core locations see Fig. 7.

[b]Dist. (%) = (Distance of core along palaeo-ice stream flow line/Total length of ice stream) x 100.

[c]TGM = transitional glaciomarine ; IT = iceberg turbate; DO = diatomaceous ooze; GM = glaciomarine.

[d]AIO = acid insoluble organic carbon; F = foraminifera; (m) = mixed benthic and planktic; (b) = benthics; Geomag. = geomagnetic palaeointensity.

[e]R = reservoir correction (ΔR = 400 – R for CALIB program). For all carbonate samples, a marine reservoir correction of 1300 (±100) years was applied (Berkman & Forman, 1996). For AIO samples we used the reported core-top ages. DO samples in the Getz-Dotson Trough were corrected by 1300 (±100) years (Berkman & Forman, 1996) as discussed in Hillenbrand et al. 2010b.

[f]Calibrated using the CALIB program v 6.0 (Stuiver et al. 2005), reported in calendar years before present (cal. yr BP). Ages rounded to the nearest ten years.

Dates in bold are inferred to be the most reliable minimum ages constraining initial palaeo-ice stream retreat.

Reference	**Core**	**Location**	**[a]Map No.**	**[b]Dist. (%)**	**[c]Sediment Facies**	**[d]Carbon Source**	**Conventional ^{14}C age**	**Error (± years)**	**[e]R (yrs)**	**Corrected age (yrs)**	**[f]Median cal. age (yrs)**	**1 σ error**	**2 σ error**
Domack et al. (2001)	ODP-1098C	Anvers	1	27.6	TGM	AIO	12250	60	1260	10990	12850	120	207
Pudsey et al. (1994)	GC51	Anvers	2	72.4	TGM	AIO	12730	130	4020	8710	9860	219	371
Pudsey et al. (1994)	GC49	Anvers	3	65.1	TGM	AIO	13110	120	4020	9090	10340	153	365
Yoon et al. (2002)	GC-02	Anvers	4	60.3	GM (above till)	AIO	12840	85	3000	9840	11250	115	320
Nishimura et al. (1999)	GC1702	Anvers	5	68.5	GM	AIO	14320	50	2340	11980	13810	78	179
Heroy & Anderson (2007)	PC-24	Anvers	6	83.6	TGM	F(m)	14020	110	1300	12720	15030	334	775
Heroy & Anderson (2007)	PC-25	Anvers	7	94.0	IT	F(m)	14450	120	1300	13150	15960	449	732
Heroy & Anderson (2007)	KC-26	Anvers	8	84.9	TGM	F(m)	14880	200	1300	13580	**16620**	330	803
Heroy & Anderson (2007)	PC-23	Anvers	9	59.9	TGM	Shell	11168	81	1300	9868	11320	146	429
Pudsey et al. (1994)	GC47	Anvers	10	56.5	TGM	AIO	12280	150	1870	10410	12140	373	730
Domack et al. (1991)	302	Astrolabe-Fr	11	26.5	DO	AIO	5515	132	1300	4215	4840	234	434
Crosta et al. (2007)	MD03-2601	Astrolabe-Fr	12	42.2	DO	AIO	10855	45	2350	8505	9600	154	331
Hillenbrand et al. (2010a)	JR104-GC358	Belgica	13	34.8	GM	AIO	21433	168	5131	16302	19450	163	488
Hillenbrand et al. (2010a)	JR104-GC359	Belgica	14	34.8	GM	AIO	11736	120	5131	6605	7500	114	241
Hillenbrand et al. (2010a)	JR104-GC360	Belgica	15	33.3	GM	AIO	8415	95	4450	3965	4520	167	309
Hillenbrand et al. (2010a)	JR104-GC366	Belgica	16	23.3	GM	AIO	16193	196	3914	12279	14320	384	659
Hillenbrand et al. (2010a)	JR104-GC357	Belgica	17	55.2	GM	AIO	12140	191	5810	6330	7230	210	419
Hillenbrand et al. (2010a)	JR104-GC368	Belgica	18	54.1	GM	AIO	25240	565	5484	19756	23560	668	1342
Hillenbrand et al. (2010a)	JR104-GC371	Belgica	19	79.3	IT	AIO	22507	436	2464	20043	23910	536	1033
Hillenbrand et al. (2010a)	JR104-GC372	Belgica	20	87.2	GM	AIO	27900	797	1731	26169	**30710**	618	1451
Hillenbrand et al. (2010a)	JR104-GC374	Belgica	21	89.6	GM	AIO	27512	721	2464	25048	29830	714	1316
Heroy & Anderson (2007)	PC-55	Biscoe	22	90.4	TGM	AIO	18420	130	2999	15421	**18660**	120	224
Heroy & Anderson (2007)	PC-57	Biscoe	23	62.2	TGM	AIO	19132	87	4913	14219	17300	203	371
Pope & Anderson (1992)	PD88-42	Biscoe	24	14.8	TGM	F(m)	13120	100	1300	11820	13630	150	289
Heroy & Anderson (2007)	PC-30	Biscoe S	25	73.9	TGM	AIO	17660	110	6300	11360	**13220**	118	283
Finocchiaro et al. (2005)	ANTA02-CH41	Cape Hallett	26	N/A	DO (varved)	AIO	10920	50	1790	9130	10380	110	194
Licht & Andrews (2002)	NBP9501-18tc	Central Ross Sea	27	17.7	GM	AIO	20490	260	3735	16755	19920	297	565
Licht & Andrews (2002)	NBP9501-18pc	Central Ross Sea	28	17.7	GM	AIO	17760	115	3735	14025	17090	170	337
Licht & Andrews (2002)	NBP9501-11	Central Ross Sea	29	51.7	TGM	AIO	25870	245	3735	22135	26600	406	830
Licht & Andrews (2002)	NBP9501-24	Central Ross Sea	30	79.0	TGM	AIO	30635	445	3735	26900	31280	270	768
Licht & Andrews (2002)	NBP9401-36	Central Ross Sea	31	56.0	TGM	AIO	30220	420	3735	26485	31010	274	562

Anderson & Andrews (1999)	IWSOE70_2-19-1	Crary Trough	32	90.0	IT	F(b)	16190	70	1300	14890	**18188**	114	381
Elverhøi (1981)	212	Crary Trough	33	N/A	IT?	Shell	31290	1700	1300	29990	34460	1831	3461
Domack et al. (1999)	NBP95-01_PC26	Drygalski	34	21.8	DO	AIO	7690	65	2210	5480	6300	87	191
Domack et al. (1999)	NBP95-01_PC29*	Drygalski	35	21.4	DO	AIO	5770	75	2210	3560	3960	128	263
Domack et al. (1999)	NBP95-01_KC31	Drygalski	36	41.3	DO	AIO	12280	95	2430	9850	11270	136	339
Licht et al. (1996)	DF80-102	Drygalski	37	52.9	GM	AIO	12640	80	1270	11370	13230	81	163
Licht et al. (1996)	DF80-108	Drygalski	38	53.9	GM	AIO	11545	95	1270	10275	12020	201	394
Licht et al. (1996)	DF80-132	Drygalski	39	44.3	GM	AIO	10730	80	1270	9460	10790	154	259
Domack et al. (1999)	NBP95-01_KC37	Dryglaski	40	68.0	DO	AIO	13840	95	2780	11060	12930	138	239
Licht et al. (1996)	DF80-57	Dryglaski	41	10.5	GM	Bivalve	7830	60	1300	6530	7440	105	209
Frignani et al. (1998)	ANTA91-28	Dryglaski	42	64.6	TGM	AIO	17490	930	5090	12400	**14790**	1780	3430
Frignani et al. (1998)	ANTA91-29	Dryglaski	43	50.7	TGM	AIO	17370	60	6710	10660	12530	173	395
McKay et al. (2008)	DF80-189	Dryglaski	44	10.4	GM	AIO	11331	45	2470	8861	10060	126	268
Finocchiaro et al. (2007)	ANTA99-CD38	Dryglaski	45	37.1	DO	AIO	12270	40	3000	9270	10530	50	127
Mosola & Anderson (2006)	NBP99-02_PC15	Eastern Ross Sea	46	91.7	TGM	AIO	30620	400	4590	26030	30730	297	565
Mosola & Anderson (2006)	NBP99-02_PC04	Eastern Ross Sea	47	0.4	TGM	AIO	23950	230	3663	20287	24200	287	640
Mosola & Anderson (2006)	NBP99-02_PC13	Eastern Ross Sea	48	90.0	TGM	AIO	28520	300	4613	23907	28740	390	699
Mosola & Anderson (2006)	NBP9902_PC06	Eastern Ross Sea	49	42.1	TGM	AIO	27330	290	3704	23626	28380	349	705
Mosola & Anderson (2006)	NBP99-02_TC05	Eastern Ross Sea	50	42.1	TGM	AIO	27000	260	3663	23337	28120	266	580
Anderson & Andrews (1999)	IWSOE70_3-7-1	SE Weddell Sea	51	N/A	TGM	F	26660	490	1300	25360	**30140**	482	898
Anderson & Andrews (1999)	IWSOE70_3-17-1	SE Weddell Sea	52	N/A	TGM	F	23870	160	1300	22570	**27260**	338	624
Elverhøi (1981)	234	SE Weddell Sea	53	N/A	TGM	Bryozoan	21240	760	1300	19940	**23790**	834	1915
Michalchuk et al. (2009)	NBP0602-8B	Firth of Tay	54		TGM	Shell	8700	40	1300	7400	8260	107	228
Harden et al. (1992)	DF86-83	Gerlache	55	2.6	TGM	AIO	10240	250	2760	7480	8390	383	751
Willmott et al. (2007)	JPC-33	Gerlache	56	73.8	GM	N/A					**8500**		
Smith et al. (in press)	VC408	Getz-Dotson	57	14.5	GM	AIO	14646	63	5135	9511	10850	126	253
Smith et al. (in press)	VC415	Getz-Dotson	58	1.7	GM	AIO	13677	57	4723	8954	10200	96	234
Smith et al. (in press)	VC417	Getz-Dotson	59	1.4	GM	AIO	16307	76	6405	9902	11350	124	319
Smith et al. (in press)	VC418	Getz-Dotson	60	7.9	GM	AIO	11469	47	5135	6334	7270	72	132
Smith et al. (in press)	VC419	Getz-Dotson	61	0.7	Gravity flow	Benthics	11237	40	1300	9937	11410	113	307
Hillenbrand et al. (2010b)	VC424	Getz-Dotson	62	20.7	DO	AIO	12183	51	1300	10883	12770	119	201
Hillenbrand et al. (2010b)	VC425	Getz-Dotson	63	10.7	DO	AIO	12868	54	1300	11568	13400	106	246
Smith et al. (in press)	VC427	Getz-Dotson	64	15.2	DO	AIO	12139	55	1300	10839	12730	114	210
Smith et al. (in press)	VC428	Getz-Dotson	65	45.5	GM	AIO	15841	72	3865	11976	13810	112	223
Smith et al. (in press)	VC430	Getz-Dotson	66	89.0	IT	Benthics	10979	40	1300	9679	11040	143	280
Smith et al. (in press)	VC436	Getz-Dotson	67	98.3	IT	Benthics	20115	71	1300	18815	**22410**	150	307
Smith et al. (in press)	PS69/267-2	Getz-Dotson	68	26.2	GM	AIO	15108	66	4124	10984	12850	128	216
Hillenbrand et al. (2010b)	PS69/273-2	Getz-Dotson	69	2.4	DO	AIO	11945	38	1300	10645	12540	71	265
Hillenbrand et al. (2010b)	PS69/274-1	Getz-Dotson	70	6.2	DO	AIO	11967	49	1300	10667	12570	80	284
Hillenbrand et al. (2010b)	PS69/275-1	Getz-Dotson	71	5.9	DO	AIO	11543	47	1300	10243	11950	233	478
Smith et al. (in press)	PS69/280-1	Getz-Dotson	72	11.0	GM	AIO	17021	80	7019	10002	11520	206	360
Leventer et al. (2006)	JPC43B	Iceberg Alley	73	76.7		AIO	11770	45	1700	10070	**11660**	335	623
Domack et al. (1999)	NBP95-01_KC39	JOIDES	74	73.6	DO	AIO	14290	95	3140	11150	**13010**	131	254
Frignani et al. (1998)	ANTA91-14	JOIDES	75	81.1	TGM	AIO	24000	620	3800	20200	24160	1607	3217
Melis & Salvi (2009)	ANTA91-13	JOIDES	76	85.3	TGM	F	21100	75	1300	19800	23640	189	415
Finocchiaro et al. (2000)	ANTA99-8	JOIDES	77	37.2	TGM	AIO	24830	110	3800	21030	25160	794	1634
Finocchiaro et al. (2000)	ANTA96-9	JOIDES	78	43.6	TGM	AIO	10100	60	3800	6300	7150	585	1178
Banfield & Anderson (1995)	DF82-48	Lafond	79	93.3	TGM	F(m)	15665	95	1300	14365	**17460**	230	414
Shevenell et al. (1996)	GC-01	Lallemand	80	N/A	TGM	Shell	9358	70	1300	8058	9080	169	343
Brachfeld et al. (2003)	KC-23	Larsen-A	81	3.5	TGM	Geomag.	10700	500	n/a	n/a	10700	500	
Domack et al. (2005)	KC-02	Larsen-B	82	15.7	GM	F	10600	55	1300	9300	10571	147	340
Domack et al. (2005)	KC-05	Larsen-B	83	28.6	GM	F	9210	45	1300	7910	8857	158	311
Gingele et al. (1997)	PS2028-4	Lazarev Sea shelf	84	N/A	GM	Bryozoan	9850	130	1300	8550	**9680**	211	416
Gingele et al. (1997)	PS2226-3	Lazarev Sea shelf	85	N/A	GM	Bryozoan	8430	95	1300	7130	7800	144	290
Gingele et al. (1997)	PS2058-1	Lazarev Sea shelf	86	N/A	GM	Bryozoan	6530	95	1300	5230	6030	150	300
Ó Cofaigh et al. (2005b)	VC304	Marguerite Bay	87	81.3	TGM	AIO	11670	250	3473	8197	9220	311	664

Harden et al. (1992)	DF86-111	Marguerite Bay	88	22.3	TGM	AIO	10180	170	4260	5920	6790	272	501
Harden et al. (1992)	DF86-112	Marguerite Bay	89	24.6	TGM	AIO	10800	160	4100	6700	7590	204	418
Kilfeather et al. (2010)	GC002	Marguerite Bay	90	50.0	TGM	F	13340	57	1300	12040	13870	117	271
Kilfeather et al. (2010)	GC005	Marguerite Bay	91	65.6	TGM	F	13390	56	1300	12090	13920	119	314
Pope & Anderson (1992)	PD88-85	Marguerite Bay	92	79.5	TGM	F(m)	13335	105	1300	12035	13870	156	405
Pope & Anderson (1992)	PD88-99	Marguerite Bay	93	63.6	TGM	F(m)	13490	140	1300	12190	**14110**	240	631
Allen et al. (2010)	JPC-43	Marguerite Bay	94	3.8	TGM	F(m)	9360	50	1300	8060	9080	152	321
Pope & Anderson (1992)	PD88-76	Marguerite Bay	95	93.3	IT	F(m)	12425	110	1300	11125	12990	166	290
Heroy & Anderson (2007)	PC-48	Marguerite Bay	96	17.2	TGM	Shell	9640	60	1300	8340	9400	123	287
Heroy & Anderson (2007)	KC-51	Marguerite Bay	97	49.3	TGM	Shell	9640	n/a	1300	8340	9480		
Heroy & Anderson (2007)	PC-49	Marguerite Bay	98	30.2	TGM	Shell	9126	95	1300	7826	8760	183	351
Heroy & Anderson (2007)	PC-52	Marguerite Bay	99	65.4	TGM	Shell	9320	220	1300	8020	9000	310	558
Kim et al. (1999)	A10-01	Marian Cove	100	N/A	TGM	AIO	13461	98	5200	8261	**9310**	259	317
Milliken et al. (2009)	NBP0502-1B	Maxwell Bay	101	N/A	TGM	Shell	13100	65	1300	11800	**13620**	132	248
McMullen et al. (2006)	KC-1	Mertz	102	73.2	DO	Molluscs	5755	35	1300	4455	5140	145	280
McMullen et al. (2006)	KC-2	Mertz	103	76.4	DO	AIO	6240	50	3410	2830	3050	140	260
McMullen et al. (2006)	KC-13	Mertz	104	69.3	DO	AIO	5980	40	3410	2570	2750	106	243
McMullen et al. (2006)	KC-12	Mertz	105	70.4	DO	AIO	6110	40	3410	2700	2890	109	211
Domack et al. (1991)	DF79-12	Mertz-Ninnes	106	66.3	DO	AIO	7350	80	5020	2330	2450	327	668
Maddison et al. (2006)	JPC10	Mertz-Ninnes	107	72.5	DO	AIO	13550	50	2340	11210	13100	71	192
Harris et al. (2001)	26PC12	Mertz-Ninnes	108	75.0	GM	AIO	15469	70	2241	13228	16100	366	731
Harris et al. (2001)	17PC02	Mertz-Ninnes	109	69.4	GM	AIO	7294	60	2431	4863	5620	298	618
Harris et al. (2001)	24PC10	Mertz-Ninnes	110	73.8	GM	AIO	16807	80	2700	14107	**17190**	149	387
Harris et al. (2001)	27PC13	Mertz-Ninnes	111	66.9	GM	AIO	11148	60	2453	8695	9840	210	342
Harris & O'Brien (1998)	149/12/GC12	Nielsen	112	75.0	GM	AIO	17150	280	2170	14980	18150	472	961
Mackintosh et al. (2011)	JPC40	Nielsen	113	68.6	GM	AIO	13895	40	1700	12195	**14160**	363	694
Heroy et al. (2008)	PC-61	Orleans	114	50.0	TGM	AIO	10859	53	2833	8026	9040	139	287
Salvi et al. (2006)	ANTA96-5BIS	Pennell Trough	115	72.8	TGM	AIO	37000	1400	3820	33180	37940	2037	4087
Licht & Andrews (2002)	NBP9501-7	Pennell Trough	116	84.5	TGM or IT	F	14970	135	1300	13670	**16790**	175	516
Anderson et al. (2002)	NBP9801-22	Pennell Coast	117	67.1	GM	Algae	9260	70	1300	7960	8930	185	343
Anderson et al. (2002)	NBP9801-17	Pennell Coast	118	74.3	GM (above till)	Algae	8200	90	1300	6900	7770	129	254
Anderson et al. (2002)	NBP9801-19	Pennell Coast	119	80.0	GM (above till)	Bryozoan	13260	80	1300	11960	**13780**	147	291
Anderson et al. (2002)	NBP9801-26	Pennell Coast	120	94.4	GM	F	15645	95	1300	14345	**17440**	227	415
Lowe & Anderson (2002)	NBP9902_PC39	Pine Island	121	51.8	GM	F	15800	3900	1300	14500	**17150**	4789	9422
Lowe & Anderson (2002)	NBP9902_PC41	Pine Island	122	42.2	GM	F	10150	370	1300	8850	10030	460	962
Domack et al. (1998)	KROCK 24	Prydz Channel	123	77.3	GM	AIO	12680	110	2510	10170	**11790**	233	495
Leventer et al. (2006)	JPC25	Prydz Channel	124	10.9	TGM	Scaphopod	10625	35	1300	9325	10600	137	308
Barbara et al. (2010)	JPC24	Prydz Channel	125	12.3	TGM	Shell	10315	35	1300	9015	10280	117	254
Domack et al. (1991)	740A-3R1	Prydz Channel	126	9.4	GM	AIO	11140	75	2510	8630	9750	155	287
Hemer & Harris (2003)	AM02	Prydz Channel	127	0.0	DO	AIO	21680	160	6548	15132	18290	285	567
Taylor & McMinn (2002)	GC29	Prydz Channel	128	10.3	GM	AIO	14140	120	2493	11647	**13500**	209	426
Pudsey et al. (2006)	VC242	Robertson	129	20.0	TGM	AIO	20300	160	6000	14300	17410	312	562
Pudsey & Evans (2001)	VC244	Robertson	130	17.1	TGM	AIO	17450	60	6550	10900	12770	96	73
Pudsey & Evans (2001)	VC275	Robertson	131	30.0	TGM	AIO	14810	50	6450	8360	9430	60	126
Pudsey et al. (2006)	VC238	Robertson	132	11.3	GM	AIO	16700	120	6000	10700	12570	250	523
Pudsey & Evans (2001)	VC236	Robertson	133	16.1	TGM	AIO	15660	50	6030	9630	11010	114	214
Pudsey et al. (2006)	VC243	Robertson	134	15.2	GM	AIO	12450	40	6000	6450	7370	82	168
Pudsey et al. (2006)	VC237	Robertson	135	13.9	GM	AIO	15330	80	6000	9330	10630	197	393
Pudsey et al. (2006)	VC277	Robertson	136	27.1	GM	AIO	17714	96	6000	11714	13550	163	313
Pudsey et al. (2006)	VC276	Robertson	137	22.6	GM	AIO	18006	98	6000	12006	13840	196	432
Yoon et al.(2002)	GC-03	Smith	138	88.9	TGM	AIO	14210	90	3000	11210	13080	127	251
Heroy & Anderson (2007)	PC-20	Smith	139	64.2	TGM	F(b)	10870	270	1300	9570	10930	354	772
Heroy & Anderson (2007)	PC-22	Smith	140	93.2	TGM	F(m)	15680	200	1300	14380	**17490**	298	538
Nishimura et al. (1999)	GC1705	Smith	141	68.4	TGM	AIO	16110	60	3000	13110	15880	389	645
Heroy & Anderson (2007)	PC-04	Vega	142	80.5	TGM	AIO	21170	140	6000	15170	18300	104	314
Heroy & Anderson (2007)	PC-05	Vega	143	87.4	IT	F(b)	11121	67	1300	9821	11240	161	416

Heroy & Anderson (2007)	PC-06	Vega	144	90.2	TGM	F(m)	16340	120	1300	15040	**<u>18260</u>**	157	364
Anderson et al. (2002)	NBP9902-23	Wrigley Gulf	145	44.1	GM	Bryozoan	13873	86	1300	12573	**<u>14710</u>**	312	553
Anderson et al. (2002)	NBP9902-22	WrigleyGulf	146	44.1	GM	Shell	13576	74	1300	12276	14250	284	547
Anderson et al. (2002)	NBP9902-26	Wrigley Gulf	147	4.8	GM	Shell	14194	82	1300	12894	16750	141	365
Domack et al. (1989)	4	Adelie Bank	148	N/A	Sand	Benthics	14260	140	1300	12960	**<u>15630</u>**	416	800

Table 5: Mean retreat rates of palaeo-ice stream grounding lines.

Palaeo-ice stream	Mean retreat rate along the whole trough (Range in mean retreat rates)	Deglaciation chronology (and reference)	Reference for retreat rate
Anvers Trough	~24 m yr^{-1} (7-54 m yr^{-1})	Carbonate and AIO ^{14}C dates (Pudsey et al. 1994; Nishimura et al. 1999; Heroy &Anderson 2007)	This paper
Belgica Trough	~15 m yr^{-1} (7-55 m yr^{-1})	AIO ^{14}C dates (Hillenbrand et al., 2010a)	This paper (Hillenbrand et al., 2010a)
Drygalski Basin	~76 m yr^{-1} (23-317 m yr^{-1})	Carbonate and AIO ^{14}C dates (Licht et al. 1996; Frignani et al. 1998; Domack et al. 1999; Finocchiaro et al. 2007; McKay et al. 2008)	This paper
	~50 m yr^{-1} to Ross Island ~140 m yr^{-1} to current grounding line position from Ross Island		Shipp et al. (1999)
Getz-Dotson Trough	(18-70 m yr^{-1})	Carbonate and AIO ^{14}C dates, palaeomagntic intensity dating (Hillenbrand et al. 2010b; Smith et al. in press)	Smith et al. (in press)
JOIDES Basin	(40-100 m yr^{-1})	Annual De Geer moraine	Shipp et al. (2002)
Marguerite Trough	~80 m yr^{-1} (36-150 m yr^{-1})	Carbonate and AIO ^{14}C dates (Harden et al. 1992; Pope & Anderson 1992; Ó Cofaigh et al. 2005b; Heroy & Anderson 2007; Kilfeather et al. 2010)	This paper

Table 6: Inferred retreat styles of Antarctic marine palaeo-ice streams since the LGM based on available geomorphic and chronological evidence.

Mode of Retreat	Palaeo-ice stream	Evidence	Defining characteristics of palaeo-ice stream
Rapid	[3/4] Central Bransfield Basin (Laclavere & Mott Snowfield)	Lineations that are not overprinted by GZWs or moraines (Canals et al. 2002).	Small troughs and drainage basin area. Normal slope and well defined, deeply incised U-shaped troughs and shallow banks.
	[7] Biscoe Trough	Lineations that are not overprinted by GZWs or moraine (Amblas et al. 2006).	Small glacial system, shallow outer trough.
	[10] Marguerite Trough	Dates suggest a catastrophic retreat (over a distance of >140 km) from the outer shelf followed by a pause and then further rapid retreat (Ó Cofaigh et al. 2005b; Kilfeather et al. 2010). The outer shelf is characterised by pristine MSGL. Mean retreat rates are ~80 m yr^{-1}, although during rapid collapse of the outer and inner shelf they must have been significantly greater (i.e. within the error of the radiocarbon dates).	Deep, rugged inner shelf with well-developed meltwater network; drainage area: 10,000-100,000 km^2.
	[16] Sulzberger Bay Trough	Erosional grooves that are not overprinted. Thin deglacial sediment.	Small trough and drainage basin. Steep reverse slope.
Episodic (fast then slow)	[34] Robertson Trough	Inner and mid-shelf – lineations overprinted by GZWs (up to 20 m thick). 3-4 m of deglacial sediment and pelletized facies. Four generations of cross-cutting lineations on the outer shelf.	Large, shallow and wide outer trough which splits into a series of tributary troughs on the inner-shelf.
Episodic (slow then fast)	[1] Gerlache-Boyd Strait	Thick layer of deglacial sediment (7-60 m) on the outer shelf. Thick morainal wedge south of sill at end of Gerlache Strait. <2 m deglacial sediment in the bedrock scoured inner shelf.	Narrow inner shelf trough with large changes in relief. Small drainage basin.
	[13] Pine Island Trough	Five GZWs on mid and outer shelf associated with changes in subglacial bed gradient (Graham et al. 2010). The inner shelf is dominated by bedrock with a thin carapace (<2.5 m) of deglacial sediment.	Drainage area ~330,000 km^2. Rugged, bedrock dominated inner shelf with a major meltwater drainage network.
	[14] Getz-Dotson Trough	Deglacial dates suggest that initial retreat from the outer to the mid shelf was extremely slow (about 18 m yr^{-1}). Further retreat back into the three tributary troughs was characterised by faster rates of retreat (54 m yr^{-1} on average & up to 70 m yr^{-1}).	Small tributary troughs with very deep inner basins and a high reverse slope.
	[17] Drygalski Basin	Dates suggest a mean retreat rate of 76 m yr^{-1} (based on dates from McKay et al. (2008)). Similar calculations by Shipp et al. (1999) gave a retreat rate of ~50 m yr^{-1}. Further rates of retreat to the current grounding line position (900 km further inshore) may have been considerably faster 89-140 m yr^{-1}, whilst grounding line retreat from Drygalski Ice Tongue to Ross Island was also thought to be rapid (317 m yr^{-1}). Large GZW on the outer shelf and MSGL preserved along entire length of mapped trough (Shipp et al. 1999).	Long, narrow trough with shallow banks. Fed by ice from East Antarctica and floored predominantly by unconsolidated strata.
Episodic	[2] Lafond Trough	Three morainal ridges on the mid-shelf (Bentley & Anderson, 1998).	Small trough and drainage basin area. Normal slope and well defined, deeply incised U-shaped trough and shallow banks.
	[19] Central Ross Sea	One GZW on inner shelf (25 m) & one on outer shelf (50 m). Series of back-stepping ridges on outer shelf (Shipp et al. 1999).	Narrow trough.

	[20] Central Ross Sea	One GZW on inner shelf (c. 50 m thick), one on mid-shelf (40 m) & two on outer shelf (50 m & 70 m).	Large trough with deep banks.
	[21] Eastern Ross Sea	Three GZWs on inner (50 m), mid & outer shelves (180 m). Some moraine ridges.	Large trough with deep banks. Predominantly unconsolidated strata.
	[22] Eastern Ross Sea	Two GZWs on inner shelf (both 100 m); one on outer shelf (50-100 m).	Large trough with deep banks. Predominantly unconsolidated strata.
	[23] Mertz Trough	Up to 7 m of deglacial sediment and two prominent GZWs on the outer shelf (up to 80 m high).	Broad (50-100 km) trough.
	[28] Prydz Channel	Multiple GZWs and small transverse ridges on the mid and inner shelf (O'Brien et al. 1999). >3 m of deglacial and sub-ice shelf sediments (Domack et al. 1998).	Convergent flow with Amery palaeo-ice stream, which has a large drainage basin (currently drains ~20% of ice from East Antarctica) and deep inner shelf (>1,000 m). Ice only reached mid-shelf at LGM.
Slow	[8] Anvers-Hugo Island Trough	Very slow retreat, with mean retreat rates on the outer and mid-shelf of 2-15 m yr^{-1}. However, the inner-most shelf was subject to slightly faster rates (~47 m yr^{-1}), with Gerlache Strait ice free by ~8.4 cal. ka BP. GZW and up to 12 m of deglacial sediment on the outer shelf is consistent with a slow overall rate of retreat (Larter & Vanneste, 1995; Vanneste & Larter, 1995).	Small trough with a very deep inner basin (Palmers Deep: >1400 m). Mid shelf high at ~300 m. Three tributaries.
	[12] Belgica Trough	Mean retreat rate of 7-55 m yr^{-1}, with the outer shelf deglaciating slightly faster (~23 m yr^{-1}) than the inner shelf (Hillenbrand et al. 2010). Multiple small GZWs on the inner shelf. The outer shelf is heavily iceberg scoured so the geomorphic evidence is limited.	Large glacial trough, with drainage area of >200,000 km^2. Seaward dipping middle-outer shelf profile (angle: ~0.08°). Primarily composed of unconsolidated strata.
	[18] JOIDES-Central Basin	GZWs (3-80 m high) & corrugation moraine (De Geer?). De Geer moraine suggests a retreat rate of 40-100 m yr^{-1} (Shipp et al. 2002).	Long, quite narrow trough with shallow banks. Nourished by ice from East Antarctica (drainage area: >1.8 million km^2. Predominantly unconsolidated strata.