



How dynamic are ice-stream beds?

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Abstract

Projections of sea-level rise contributions from West Antarctica's dynamically thinning ice streams contain high uncertainty because some of the key processes involved are extremely challenging to observe. An especially poorly observed parameter is sub-decadal stability of ice-stream beds. Only two previous studies have made repeated geophysical measurements of ice-stream beds at the same locations in different years, but both studies were limited in spatial extent. Here, we present the results from repeat radar measurements of the bed of Pine Island Glacier, West Antarctica, conducted 3-6 years apart, along a cumulative ~60 km of profiles. Analysis of the correlation of bed picks between repeat surveys show that 90% of the ice-stream bed displays no significant change despite the glacier increasing in speed by up to 40% over the last decade. We attribute the negligible detection of morphological change at the bed of Pine Island Glacier to the ubiquitous presence of a deforming till layer, wherein sediment transport is in steady state, such that sediment is transported along the basal interface without inducing morphological change to the radar-sounded bed. Significant change was only detected in one 500 m section of the bed where a change in bed morphology occurs with a difference in vertical amplitude of 3-5 m. Given the precision of our measurements, the maximum possible erosion rate that could go undetected along our profiles is 500 mm a⁻¹, far exceeding erosion rates reported for glacial settings from proglacial sediment yields, but substantially below subglacial erosion rates of 1000 mm a⁻¹ previously reported from repeat geophysical surveys in West Antarctica.

1. Introduction

Glaciological studies over the past three decades have revealed that the West Antarctic Ice Sheet (WAIS) is losing mass at an accelerating rate, raising concerns over its potential future contribution to global sea-level (Shepherd et al., 2012, DeConto and Pollard, 2016). Between 2010 and 2013, around 33% of the ice sheet's net mass loss came from Pine Island Glacier (hereafter PIG), a major ice stream draining to the Amundsen Sea Embayment (McMillan et al., 2014). There, from satellite-altimetry observations, mass loss was first expressed at the grounding zone in the mid-1990s, and has now propagated all the way up to the divides (McMillan et al., 2014) by a set of processes broadly termed dynamic thinning (Shepherd et al., 2001; Pritchard et al., 2009). At PIG the dynamic thinning has incorporated an upstream expansion of regions both of accelerating ice flow (e.g. Scott et al., 2009; Mouginot et al., 2014), ice-surface lowering (e.g., Konrad et al., 2017), and upstream migration, by several 10s of km, of the grounding line (e.g. Park et al., 2013; Rignot et al., 2014), prompting suggestions that PIG is in runaway retreat (Joughin et al., 2014). However, the detailed processes by which dynamic thinning works, especially upstream from the grounding zone, are not well constrained. Particularly unclear is how the coupling between basal ice and the bed may evolve or vary over time and whether this needs to be accounted for in models projecting future ice response.

Monitoring the beds of ice streams is also important for understanding processes of erosion and sediment transport that can provide information on landscape evolution (e.g., Jamieson et al., 2010, Herman et al., 2011), basal processes (Alley et al.,



1997) and the supply of nutrients to polar oceans (Raiswell et al., 2006). A review of previously published erosion rates (Hallet et al., 1996) indicates low erosion rates in polar settings (0.01 mm a^{-1}) contrasting with relatively high rates beneath temperate alpine glaciers ($10\text{--}100 \text{ mm a}^{-1}$). However, more recent studies have identified far more rapid erosion rates of 4.8 mm a^{-1} in Greenland (Cowton et al., 2012) and as much as 1000 mm a^{-1} or more in Alaska (Motyka et al., 2006) and Antarctica (Smith et al., 2007; 2012). Critically, measurements of erosion rates in Antarctica are limited both temporally and spatially, making it difficult to assess whether such high rates of erosion are typical or exceptional.

A significant difficulty in assessing temporal changes to the beds of WAIS ice streams is lack of access to the bed. Theoretical and geophysical constraints have shown that ice streams typically achieve fast flow by overriding dilated till that provides low basal drag (Alley et al., 1986; Bentley et al., 1998). It has been inferred that even relatively small fluctuations to hydrological conditions in these locations can induce significant changes to basal drag, in the most extreme circumstances causing ice streams to switch on or off (e.g. Anandakrishnan and Alley, 1997; Conway et al., 2002; Vaughan et al., 2008). However, to our knowledge, only two studies have attempted directly to capture temporal changes to ice-stream bed conditions over decadal to sub-decadal timescales commensurate with available satellite records of surface elevation and velocity change. The first was a single-location repeat measurement of ice thickness and surface elevation on PIG (location, Fig. 1) made firstly in 1960 and re-measured in 2009 (Smith et al., 2012). That study found surface lowering with no significant change in ice thickness, suggesting that mean erosion of the ice-stream bed of up to 1 m a^{-1} took place during the study period (Smith et al., 2012). This rate is well above the range reported elsewhere (Hallet et al., 1996). The second, more detailed study, consisted of three repeat-seismic surveys of the bed of Rutford Ice Stream (hereafter RIS; Fig. 1c) obtained in 1991, 1997 and 2004 (Smith et al., 2007). There, across a $\sim 0.5 \text{ km}$ width of the bed, 6 m of sediment were removed from the ice bed between 1991 and 1997 followed by the appearance of a drumlin 10 m high and 100 m wide between 1997 and 2004. Both of the studies from West Antarctica therefore implied that active sediment erosion or deposition at rates $\sim 1\text{--}1.4 \text{ m a}^{-1}$, and landform evolution, are possible beneath ice streams on decadal to sub-decadal timescales.

In this paper, we present and analyse results from three repeat surveys of the bed of PIG, whereby we geophysically surveyed the same profiles of PIG's bed in different years along a cumulative $\sim 60 \text{ km}$ of traverses. Our principal aim was to ascertain whether erosion or deposition, and any morphological changes, were detectable at the bed of PIG over intervals of 3–7 years, a period over which PIG (from satellite monitoring) has undergone considerable ice-surface lowering and ice acceleration (McMillan et al., 2014; Mouginot et al., 2014).

2. Methods

Our data consist of three repeat radar surveys of PIG's bed acquired with a low-frequency ice-penetrating radar system in austral seasons 2007/08 and 2013/14 (two profiles of 18 km and 16 km long separated by ~ 6 years) and 2010/11 and 2013/14 (one 25 km profile separated by ~ 3 years) (Figs. 1a,b). The 18 km profile, R1, was acquired 3 km downstream and parallel to an 18 km seismic survey acquired in austral season 2007/08 (Smith et al., 2013), 5 km of which was resurveyed in austral season 2014/15 (Brisbourne et al., 2017) (profiles S1₂₀₀₇ and S1₂₀₁₄ on Fig. 1b respectively). Each of the repeat-survey locations on PIG experienced significant ice-surface lowering and acceleration over the encompassing period (Table 1). In our analysis we consider our findings from PIG against the results from Smith et al.'s (2007) repeat-seismic surveys of the bed of RIS (profile C1 in Fig. 1c), where ice flow has remained relatively stable over decadal timescales and negligible surface lowering has been observed (Table 1).

2.1 Data acquisition



All radar profiles were acquired with the British Antarctic Survey's "DELORES" (Deep-Look Radio-Echo Sounder) system, a skidoo-towed monopulse array (see King et al., 2016, for general specifications). During the 2007/08 field season, the system operated using 40 m half-dipole antennae, resulting in a centre frequency of 1.2 MHz. In 2010/11 and 2013/14, 20 m half-dipole antennae were used, giving a centre frequency of 3 MHz. Along-track traces were sampled at <1 m intervals, stacked
 5 for noise reduction to produce datapoints at ~5 m spacing, and georeferenced with dual-frequency differential GPS mounted on the radar system.

The repeat-radar surveys on PIG were each acquired approximately orthogonal to ice flow at sites 150, 120 and 115 km upstream from the 2011 grounding line (Park et al., 2013) (R1, R2 and R3 respectively in Fig. 1b), in each case close to, or
 10 forming part of, more extensive radar surveys of patches of the ice bed conducted in the three field seasons (Bingham et al., 2017). Repeat-radar profile R1 comprises a traverse ~18 km in length, first surveyed in January 2008 and then again in December 2013. The January 2008 profile also represents the most downstream traverse of a more extensive set of radar profiles used to image a 108 km² patch of PIG's bed in January 2008 surrounding the 18 km seismic profile acquired in the same season (patch 2007t1 and seismic profile S1₂₀₀₇ in Fig. 1b). Profile R2 is derived from a 16 km traverse first driven in
 15 December 2007 and then again in December 2013; this profile lies ~2 km downstream of an extensive survey of 150 km² of PIG's bed also surveyed in December 2013 (patch iSTART1 in Fig. 1b). Profile R3, ~25 km in length, was first surveyed in January 2011 as the upstream profile of multiple transects used to image a 425 km² patch of PIG's bed (patch 2010tr in Fig. 1b), and then again in December 2013 to yield a survey gap of ~3 years.

20 **Table 1: Mean velocities for each repeat radar survey profile 2007-2017.**

Survey	Velocity (m a ⁻¹) 2007-2009 (MEaSURES)	Velocity (m a ⁻¹) 2017*	Velocity change (m a ⁻¹)
C1 (RIS)	370	407	37
R1 (PIG)	287	361	74
R2 (PIG)	384	523	139
R3 (PIG)	435	609	174

*Velocities derived from Sentinel-1 image pairs obtained in April 2017. Details of processing methodology are provided in Hogg et al., 2017).

2.2 Data processing

25 Radar data were processed using ReflexW processing software (Sandmeier Scientific Software). A data processing workflow was applied which included a gain function to improve the strength of reflections at greater depth, and bandpass and 2D median filters to reduce data noise. Finite-difference (FD) migration was used to contract diffraction hyperbolae and to recover the correct locations of individual reflectors. The onset time of the bed reflector was determined at 5 m intervals at the peak in the amplitude of the bed reflector using a semi-automated picking procedure. Bed picks were then converted to depth using a radar
 30 wave speed of 0.168 m ns⁻¹. We smoothed bed picks by applying a moving average over a 50 m window to all bed picks to remove high-frequency noise.

To assess changes at the bed, we focus on comparing the morphological character of the picked bed along repeat profiles rather than deriving changes in absolute bed elevation between surveys. This is for two reasons. Firstly, we do not have the data to
 35 assess whether firn properties, that impact upon radar wave speed, changed over the periods between repeat surveys. For this reason we make no firn correction to our derived ice thicknesses. Secondly, differences in the triggering mechanism of the radar system between survey years meant that we could not match directly the onset waveforms between repeat surveys. In



both cases, the effects preclude recovery of absolute ice thickness or bed elevation, but allow capture of relative bed profile morphology per survey.

The seismic surveys S1₂₀₀₇ and S1₂₀₁₄ in the vicinity of R1 were processed and analysed by Smith et al. (2013; 2007/08 profile only) and Brisbourne et al. (2017; both profiles). Brisbourne et al. (2017) primarily report on the calculation of acoustic impedance at the bed from both profiles, which we will consider in our discussion below. For this paper, we also investigated the possibility of directly picking the bed for the repeated 5 km section in an analogous manner to the radar picking described above. However, low signal-noise ratios along large parts of the bed, resulting from the similarity of ice and bed acoustic impedance (Brisbourne et al., 2017), precluded the recovery of results with sufficiently low picking errors to have confidence in identifying any change or lack thereof.

2.3 Errors

For each profile R1-R3, the ability to detect changes in bed morphology is largely determined by the precision with which the bed reflector can be picked, and the degree to which the second radar profile of a repeat survey follows or diverges slightly from the path driven by the first profile.

The signal-to-noise ratio in all our radar profiles is high and the strength of the basal reflector produces a clearly discernible, high-amplitude wavelet (e.g. radargrams in Figs. 2-5) that requires little user interaction during the semi-automatic picking procedure. With such clear data, the uncertainty with which the bed reflection can be picked is determined primarily by the system rise-time of 250 V ns⁻¹, the recording-system bandwidth of 50 MHz, and the digitisation interval of 10 ns. Considering also the uncertainty in our GPS-derived elevations, we estimate that our radar data have a vertical range precision of ±3 m.

Survey lines were repeated by following a route programmed into a dashboard GPS unit mounted on the radar skidoo. Due to the higher accuracy and precision of the dual-frequency GPS compared to the dashboard units, and the challenges of navigating in featureless terrain, navigational divergences were registered. These divergences were mostly < 50 m (Table 2), however, where bed topography is rough, even small navigational divergences may lead to incorrect interpretation of bed change. We assessed the variability in bed topography by analysing nine additional intersecting radar lines driven orthogonal to repeat survey lines (three per repeat survey line). We calculated a mean of the standard deviation of bed elevation for available intersecting lines within 50 m east and west of repeat survey lines (Table 2). From this analysis we find that the maximum variability in bed elevation within the range of navigational divergence between repeat surveys is just over a metre (Table 2).

Table 2. Analysis of navigational divergence and associated variability in bed elevation.

Repeat survey line	R1	R2	R3
Repeat survey divergence (m)			
Maximum	45.1	54.0	35.0
Mean	23.7	13.7	20.0
Standard Deviation	±10.3	±12.0	±8.0
Bed variability (m)*			
	±1.1	±1.0	±0.2

*Mean of the standard deviation of bed elevation



3. Results

Bed picks for the repeat-geophysical surveys from PIG are shown in Figures 2 and 3, alongside associated geophysical images. A visual inspection of the radar/seismic images and bed picks (Figs. 2 and 3) shows that there is remarkable consistency in the morphology of the bed at all three of the repeat-survey sites. This qualitative impression is confirmed by calculating the Pearson correlation coefficient (r) of each repeat profile's bed picks across 500 m moving windows (Figs. 2 and 3). For 90% of the repeat radar tracks $r > 0.9$, underscoring that there has been negligible morphological change for much of the surveys. For context, we performed the same correlation routine for the bed picks of repeat-seismic survey data acquired in 1991, 1997 and 2004 along the 3.5 km profile C1 of RIS previously reported by Smith et al. (2007): these results are shown in Fig. 4e. At C1, the ~ 0.5 km section of track for which Smith et al. (2007) reported basal erosion between 1991 and 1997 yields $r \sim 0.5$, and the ~ 100 m length of profile interpreted as hosting the growth of a drumlin between 1997 and 2004 returns $r \sim -0.2$.

There is only one location on PIG where r is considerably lower than 0.9; this occurs between 8.5 and 9 km along profile R1 where r spikes around a value of -0.2 (Fig 2). Closer inspection of this location reveals a subtle change in the morphology of the bed picks between 2007/08 and 2013/14 (Fig. 5). At this site two bumps of ~ 1 -2 m height in 2007/08 are replaced in 2013/14 by a central ridge with two troughs ~ 3 m in depth.

4. Discussion

The prevailing picture that emerges from our 58 km of repeat surveys of PIG's bed is one of little measurable change to the ice stream's basal topography and morphology over the 3-7 year timescale. This is in spite of significant changes occurring to ice-flow speeds and ice-surface elevations over the same periods (Table 1), and the observations of active erosion and deposition made over similar timescales at RIS (Fig. 4), where ice has experienced little to no dynamic thinning and ice flow is essentially stable (Table 1). In the following discussion we firstly consider how these apparently different behaviours between the ice beds of PIG and RIS can be reconciled (Section 4.1). We then turn to the implications of our results for the understanding of processes of sediment erosion, transport and deposition beneath ice streams and implications for future monitoring of the bed (Section 4.2).

4.1 Are ice-stream beds dynamic?

The only precedent for the measurements obtained here from PIG in terms of repeat geophysical survey over comparable timescales is that from RIS where erosion and deposition on the order of $\sim 1 \text{ m a}^{-1}$ were observed in association with morphological changes at the bed (Smith et al, 2007). Here we assess whether the lack of comparable changes observed at PIG can be explained by contrasting glaciological, hydrological and basal characteristics to the RIS survey site, and whether either is likely to be more representative of wider changes occurring at the beds of Antarctic ice-streams.

The first notable difference between PIG and RIS is the broad subglacial topography. Each of the repeat surveys on PIG was conducted across the 30-km-wide, ~ 2000 m deep, main ice-stream trunk where, at the multi-km wavelength, the bed is largely flat along and across flow (Vaughan et al., 2006). By contrast, the repeat-survey location on RIS, though also traversing the ice-stream trunk, overlies a notable topographic ridge ("Central low ridge") that abuts ~ 350 m vertically upwards into the central ~ 5 km width of the ~ 30 km-wide, ~ 2000 m deep ice-stream trunk (Fig. 4a). There is, therefore, a clear contrast in the gross topographic shape of the cross-sectional bed profile between the PIG and RIS repeat-survey sites; and we note that the flat-bedded trough of PIG is more characteristic of the majority of ice streams in West Antarctica (Fretwell et al., 2013). Turning to the finer, sub-km morphological character of each site, detailed ground-based radar surveys of PIG's trunk (Figs 2



and 3, and see further imagery in Bingham et al., 2017) have depicted ubiquitous mega-scale glacial lineations (MSGL) across all regions of PIG's trunk, indicative of the widespread presence of deforming sediment (Clark, 1993, Stokes and Clark 2001; Spagnolo et al., 2016). MSGL with mean amplitudes ~10 m and a mode spacing of 300-400 m are also pervasive surrounding the RIS repeat-survey site (Fig. 4a) but, unlike PIG, within the MSGL themselves linear features with much higher amplitudes (up to 70 m) are observed; King et al. (2016) termed these features "tapering drumlins".

The physical properties of the bed between PIG and RIS also differ. Seismic reflection surveying of several sites along PIG's trunk, including upstream and downstream of our repeat-survey sites, has confirmed that the bed immediately below the ice pervasively consists of deforming sediments, which are at least several metres thick (Smith et al., 2013; Brisbourne et al., 2017). Along seismic profile S1₂₀₀₇ potential-field data indicate that a transition from sedimentary to crystalline bedrock lies beneath the cap of deforming sediments (at position 9 km on Fig. 2d) (Smith et al., 2013), but the repeat survey of S1 in 2014/15 exhibited no change to acoustic impedance anywhere along this profile (Brisbourne et al., 2017), reinforcing the notion of a relatively stable basal environment despite the transition in geology below the deforming till layer. By contrast, the bed around RIS profile site C1 is characterised by a "patchwork" of soft, deforming sediments contrasting with regions of basal sliding indicative of more consolidated sediments (Smith and Murray, 2009; Smith et al., 2015). Notably, the areas of deforming sediment coincide with topographic highs such as the feature known as "The Bump" (Smith, 1997) (Fig. 4c). Changes in acoustic impedance were detected for parts of the RIS repeat-profile between 1997 and 2004, and were interpreted as changes in hydrological conditions within subglacial sediments affecting till porosity (Smith et al., 2007).

A third key difference between the PIG and RIS repeat-survey sites is the degree to which the survey sites are affected by tidal influences, and the potential effects this can have on subglacial hydrology. High-resolution GPS monitoring of ice motion at RIS site C1, 10 km upstream of the grounding line, has shown that horizontal ice velocity varies by ~20% on fortnightly timescales in response to tidally-modulated vertical displacement of Filchner-Ronne Ice Shelf (Gudmundsson, 2006, 2007; Murray et al., 2007; Minchew et al., 2017). In addition passive, seismic monitoring has demonstrated increased seismic activity during tidal cycles (Adalgeirsdóttir et al., 2008). Numerical modelling studies have suggested that the velocity oscillations are transmitted by tidally-forced fluctuations in basal water pressure, that alter the pore pressure of basal sediments and thus effective pressure at the bed (Thompson et al., 2014; Rosier et al., 2015). These observations combine to suggest that the hydrological system beneath the RIS survey site is prone to significant dynamism and reorganisation over short timescales. By contrast, high-resolution GPS monitoring at several sites along PIG's main trunk showed no tidal signal in ice motion even 55 km upstream from the grounding line (Scott et al., 2009), which lies well downstream of our repeat radar sites R1-R3. The ubiquitous dilated till layer that overlies a relatively flat bed at each of the PIG survey sites provides suitable conditions for a stable, distributed drainage system (Weertman, 1972; Alley, 1989; Engelhardt et al., 1990, Engelhardt and Kamb, 1997) potentially in the form of a canal network as suggested in the upstream catchment of neighbouring Thwaites Glacier (Schroeder et al., 2013), upstream RIS (King et al., 2004) and Whillans Ice Stream (Engelhardt and Kamb, 1997). In the absence of a dynamic hydrological system, sediment mobility may be restricted, thereby limiting the rate of erosion and sediment transport detectable within the precision of repeat geophysical measurements.

We therefore consider the possibility that rapid erosion and bed reorganisation on the scale observed beneath RIS is an exception rather than the rule. Geophysical surveys of other West Antarctic ice streams (Alley et al., 1986, Blankenship et al., 1986; Peters et al., 2006) have revealed shallow bed gradients and widespread deforming till similar to the surveyed sites on PIG. These characteristics are also evident in offshore records of palaeo-ice stream beds on the outer continental shelf of West Antarctica, where ice streams occupied shallow troughs in sedimentary basement (Lowe and Anderson, 2002; Wellner et al., 2006; Larer et al., 2009; Graham et al., 2010). These offshore regions are characterised by ubiquitous MSGL (Spagnolo et al.,



2014) that are also observed in the more extensive grid surveys surrounding our repeat surveys (Fig. 2a,3a,f) (Bingham et al., 2017). Assuming MSGL do evolve through time (Spagnolo et al., 2017), the presence of these landforms associated with a bed that shows little morphological change in up to 6 years, suggests that MSGL may not evolve on sub-decadal timescales. Alternatively, it is possible that the MSGL imaged here have reached a mature configuration, as reflected by the apparent
 5 stable bed.

4.2 Implications for subglacial sediment transport and future surveys

The absence of detectable morphological change to the bed over the majority of the ~60 km of bed profiles on PIG could be interpreted in three ways: (1) that no sediment erosion/transport/deposition is occurring at the measured sites; (2) that erosion
 10 /deposition is occurring but at rates too low to be detected within the vertical range resolution of the radar; or (3) that the subglacial till flux is in a steady state wherein sediment transport is active but is not altering the shape of the bed.

Prior to this study the few repeat geophysical surveys of the ice-bed interface in Alaska and Antarctica yielded subglacial erosion rates of 100-0 - 3900 m a⁻¹ (Motyka et al., 2006; Smith et al., 2007; 2012), far exceeding the 0.01 - 100 mm a⁻¹ range
 15 traditionally reported as characteristic of glacial settings using proglacial sediment yields (Hallet et al., 1996; Koppes and Hallet, 2006; Koppes & Montgomery, 2009; Cowton et al., 2012; Herman et al., 2015) (Fig. 6). Consequently, there has been growing consensus that subglacial erosion and transport is likely to be high, i.e. of the order of m a⁻¹, beneath thick, warm-based ice as manifested by polar ice streams. Observations of substantial till deposition at the grounding lines of contemporary and palaeo West Antarctic ice streams support this view (Anandakrishnan et al., 2007; Batchelor and Dowdeswell, 2015). The
 20 precision of our measurements, essentially defined by the vertical range resolution of the radar, means that the maximum possible erosion rate that could go undetected along our profiles is 500 mm a⁻¹.

Aside from the repeat geophysical studies conducted in Antarctica by Smith et al. (2007, 2012), the only other location where this method has been used is southeast Alaska where Motyka et al., (2006) found exceptionally high erosion rates of up to
 25 3900 ± 800 mm a⁻¹ over a period of 14 years (Fig. 6). This erosion took place during an advance of Taku Glacier over glaciomarine and outwash sediments. Erosion rates in southern Alaska are the highest reported for any region. These high erosion rates are associated with high precipitation rates, active tectonic uplift and glaciofluvial evacuation of unlithified sediments (Hallet et al., 1996; Motyka et al., 2006). The equally rapid erosion observed by Smith et al. (2007, 2012) therefore evokes a glaciofluvial mechanism. Emerging evidence of subglacial hydrology from West Antarctica shows that meltwater
 30 can be stored and released over short-timescales through interconnected subglacial lakes and channels (Wingham et al., 2006; Fricker et al., 2007; Fricker and Scambos, 2009; Smith et al., 2017). In our survey area there is no evidence from radar (Bingham et al., 2017) or seismic data (Brisbourne et al., 2017) that indicates the presence of subglacial lakes or concentrated meltwater channels. However, in Smith et al.'s (2012) repeat seismic measurement over ~50 years, rapid erosion of 1 m a⁻¹ was detected ~25 km downstream of survey R3 (EHT31 in Fig. 1b). At this location the bed is also smooth and there is no
 35 indication from satellite or geophysical data of rapid meltwater drainage events or channelised meltwater flow (Smith et al., 2012). A possible explanation for the discrepancy between erosion observed at this location and the results of this study is a timescale bias (Ganti et al., 2016). Smith et al.'s (2012) study may have detected intermittent episodes of erosion not captured by our 3-6 year survey intervals.

40 In Section 3, we showed that while the majority of the repeat-radar profiles evince negligible morphological change, there is one ~0.5 km section of profile R1 where a possible reorganisation of the bed is expressed as a morphological change in the basal reflector (Fig. 5). This region of the bed broadly coincides with a transition from thin (≤ 10 m) sediment overlying a crystalline basement to a deep sedimentary basin imaged in seismic and potential field surveys (Smith et al., 2013). A change



in basal drag and ice velocity is also associated with this boundary (Smith et al., 2013).. It is likely that such boundaries influence subglacial hydrological pathways and, hence, basal ice motion. This exemplifies the importance of the sampling location and periods of measurement to any study attempting to capture subglacial sedimentary processes. For example, the location where Smith et al. (2007) monitored deposition (drumlin formation) on RIS between 1997-2004 is positioned at the end of an MSGL subsequently imaged in a wider radar survey of the region (King et al., 2009; 2016) (Fig. 4b). These repeat measurements may have been fortuitously timed to capture the extension of this bedform. The anisotropic morphology of tapering drumlins and MSGL present beneath RIS (King et al., 2016) may not lend itself well to the detection of change from 2-D surveys driven orthogonal to ice flow. Dynamic change in these bedforms may be expressed as downstream migration of a propagating sediment front that would be difficult to detect without more extensive and/or higher-density coverage of repeat surveys over longer timescales.

5.0 Conclusions

We have analysed ~60 km of repeat radar surveys acquired from Pine Island Glacier, West Antarctica, between austral seasons 2007/08 and 2013/14. The results showed that little morphological change occurred at the bed over this period. The absence of large signals of erosion or deposition contrasts with erosion/deposition measured on nearby Rutford Ice Stream between 1991 and 2004 (Smith et al., 2007), and with inferred erosion of $\sim 1 \text{ m a}^{-1}$ at Pine Island Glacier over a half-century (1960-2009; Smith et al., 2012). We attribute the negligible detection of morphological change at the bed of Pine Island Glacier on the sub-decadal timescale to the ubiquitous presence of a deforming till layer, wherein sediment transport is in steady state such that sediment is transported along the basal interface without inducing measurable vertical displacement to the radar-sounded basal interface. Moreover, none of the survey sites on Pine Island Glacier have experienced short-term variations in ice velocity (Scott et al., 2009) that may indicate active hydrological systems capable of rapidly altering basal conditions (Thompson et al., 2014; Rosier et al., 2015). By comparison, the high subglacial erosion and deposition rates reported from Rutford Ice Stream occurred where the ice is overriding a topographic rise 350 m high, there are sharp contrasts in subglacial sediment properties, and the subglacial system likely experiences short-term variability influenced by tidal oscillations (Aðalgeirsdóttir et al., 2008; Thompson et al., 2014; Rosier et al., 2015).

The surveys presented in this study have increased the length, by an order or magnitude, of the available records of repeat measurements of Antarctic ice-stream beds. However, these environments remain poorly sampled. The surveys of Pine Island Glacier and Rutford Ice Stream have shown that improved understanding of the dynamism of the ice-sheet bed may best be gained from a multi-method survey approach, involving: (1) high-density radar grid surveying to image the subglacial landscape and provide spatial context, (2) seismic survey to discern the physical properties of the basal interface, and (3) repeated seismic and/or radar profiles at the same site undertaken over subdecadal to decadal timescales. Our findings also show the importance of the physical properties of the bed at each site to the sediment erosion and deposition that could be detected, underscoring the requirement to sample sites that, as a whole, capture a representative range of basal conditions. Models of basal drag inverted from satellite-imaged surface properties (e.g. Joughin et al., 2009; Arthern et al., 2015) offer the best opportunities to guide site selection.

Author contributions

Radar and seismic data from 2007/08 were acquired by RGB; radar data from 2010/11 by ECK; radar data from 2013/14 by DD, RGB, AMS and DGV; and seismic data from 2014/15 by AMB, AMS and DD. Radar data were processed by DD, RGB and ECK and seismic data by AMB and AMS. Sentinel-1 velocity estimates for each geophysical site were produced by AEH. DD analysed the data and wrote the paper. All authors contributed edits to the final manuscript.



Competing interests

The authors declare that they have no conflict of interest.

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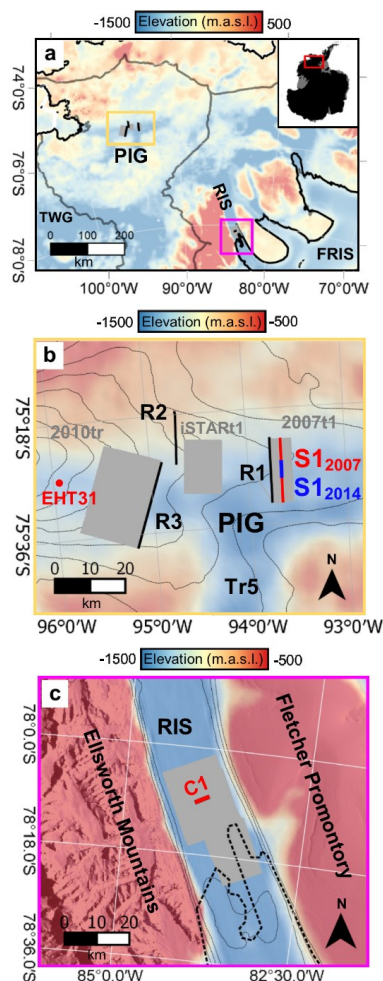


Figure 1. Location maps of the study area. Colourscale shows BEDMAP2 bed elevation (Fretwell et al., 2013). (a) Map of the Pine Island Glacier (PIG) and Rutford Ice Stream (RIS) catchments. Grey lines show drainage boundaries (Zwally et al., 2012). (b) Location of PIG surveys. Dashed lines are 100 m a^{-1} ice surface velocity contours (Rignot et al., 2011). Grey shaded boxes show the locations of radar grid DEMs in Fig 2a, 3a,f, black lines are repeat radar surveys. S1₂₀₀₇ is a seismic survey presented in Smith et al., (2013) and S1₂₀₁₄ is a section of this profile resurveyed in 2014 (presented in Brisbourne et al., 2017). EHT31 is a single point repeat survey presented in Smith et al., (2012). (c) Location of RIS surveys. Thick dashed line represents the MODIS grounding line. Thin dashed lines are 100 m a^{-1} velocity contours (Rignot et al., 2011). Grey shaded box is the area of the radar grid DEM in Fig. 5a.

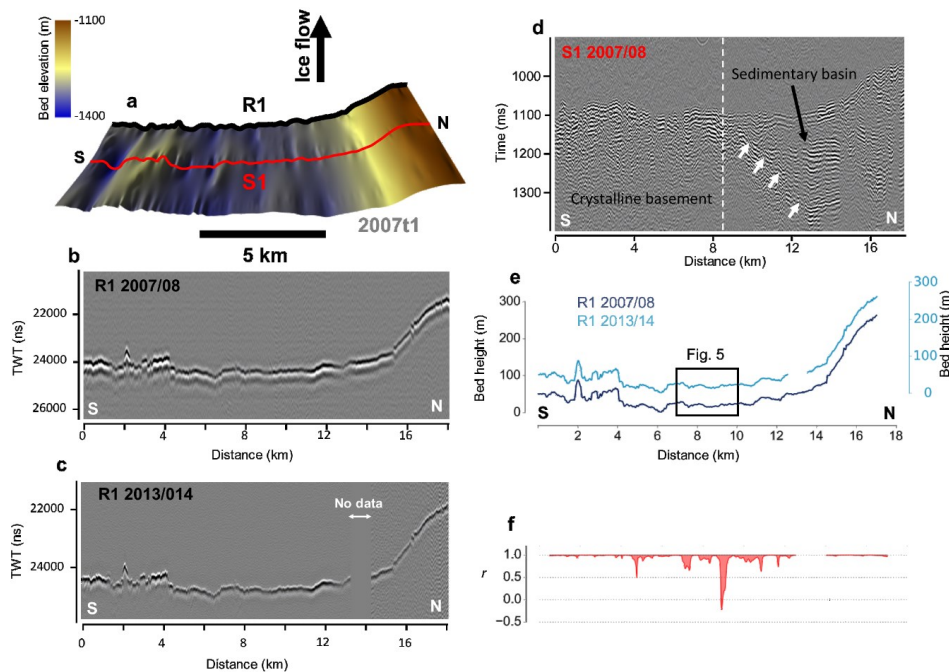


Figure 2: Geophysical data and bed picks from PIG. (a) Radar derived DEM from PIG showing the location of radar (black line) and seismic survey (red line). (b, c), processed radargrams of repeat radar surveys. (d) Processed seismic section of S1 (Smith et al., 2013). White arrows denote a dipping reflector demarcating a sedimentary basin. (e) Bed picks from repeat radar survey R1. (f) Correlation coefficient of repeat bed picks between 2008/09 and 2013/14 across a 500 m moving window.

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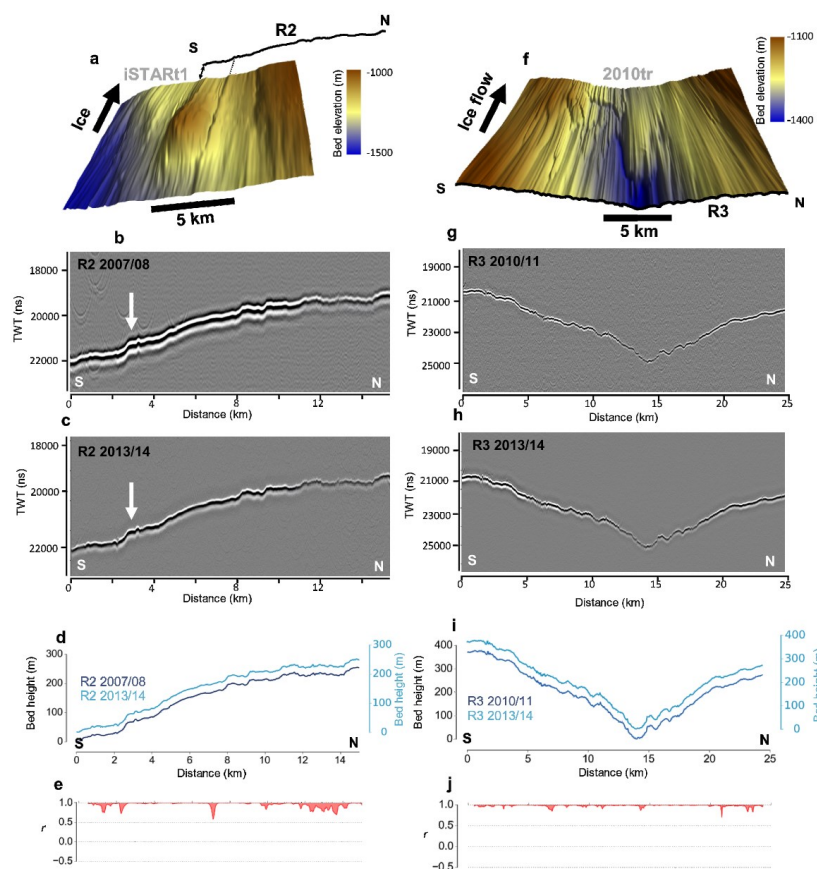


Figure 3: Geophysical data and bed picks from PIG. (a) radar derived bed DEM (Bingham et al., 2017) showing the location of survey R2 2km downstream. (b,c) processed radargrams of repeat radar surveys. (d) bed picks from repeat radar survey R2. (e) Correlation coefficient of repeat bed picks between 2008/09 and 2013/14 across a 500 m moving window. (f) radar derived bed DEM (Bingham et al., 2017) showing the location of survey R3 at the upstream limit of the radar grid. (g,h) processed radargrams of repeat radar surveys. (i) bed picks from repeat radar survey R3. (j) Correlation coefficient of repeat bed picks between 2010/11 and 2013/14 across a 500 m moving window.

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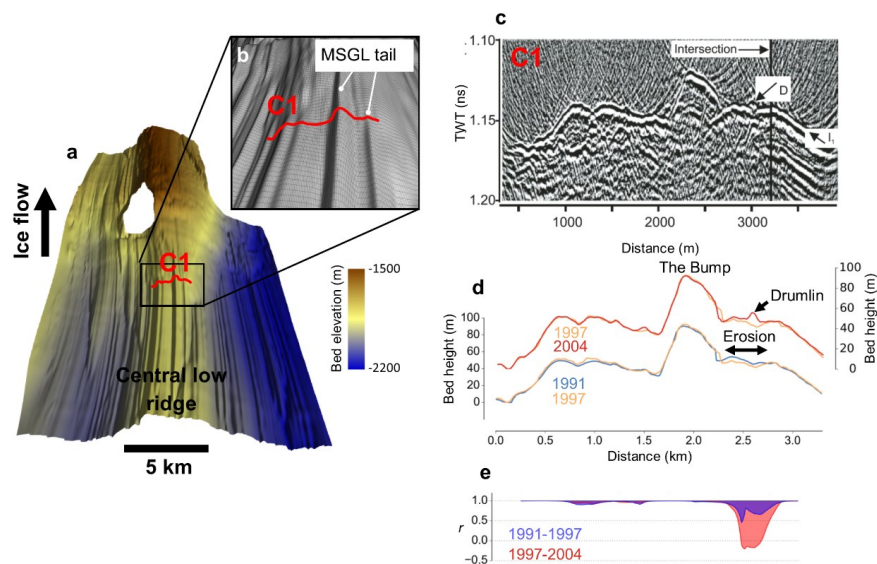


Figure 4: . Geophysical data and bed picks from RIS (a) radar derived bed DEM for RIS (King et al., 2016) showing geomorphology of the bed and the location of survey C1. (b) close-up of the location of survey C1 showing the 3-dimensional character of the bed. (c) Processed seismic section of survey C1. (d) bed picks from repeat seismic surveys showing erosion and drumlin formation between 1991 and 2004 (Smith et al., 2007). (e) Correlation coefficient of repeat bed picks between 1991 and 2004 across a 500 m moving window

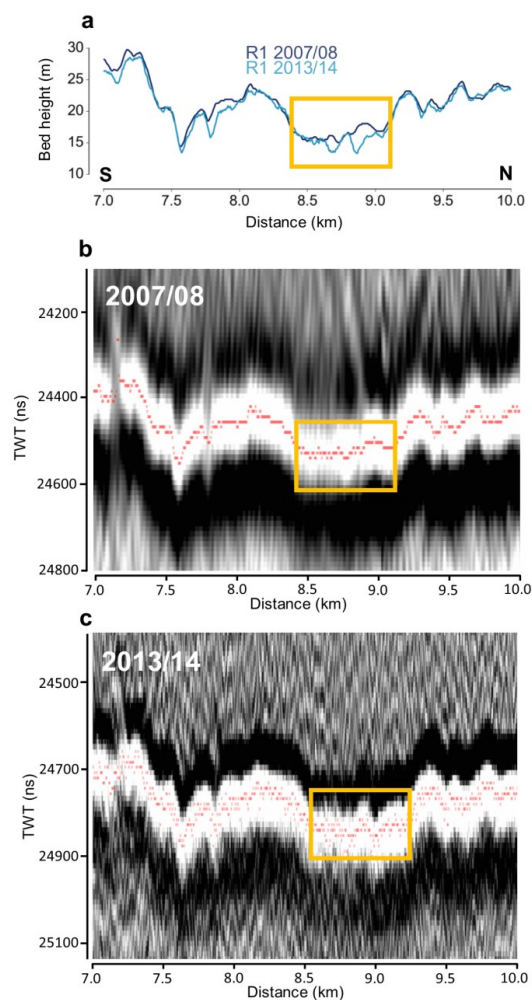
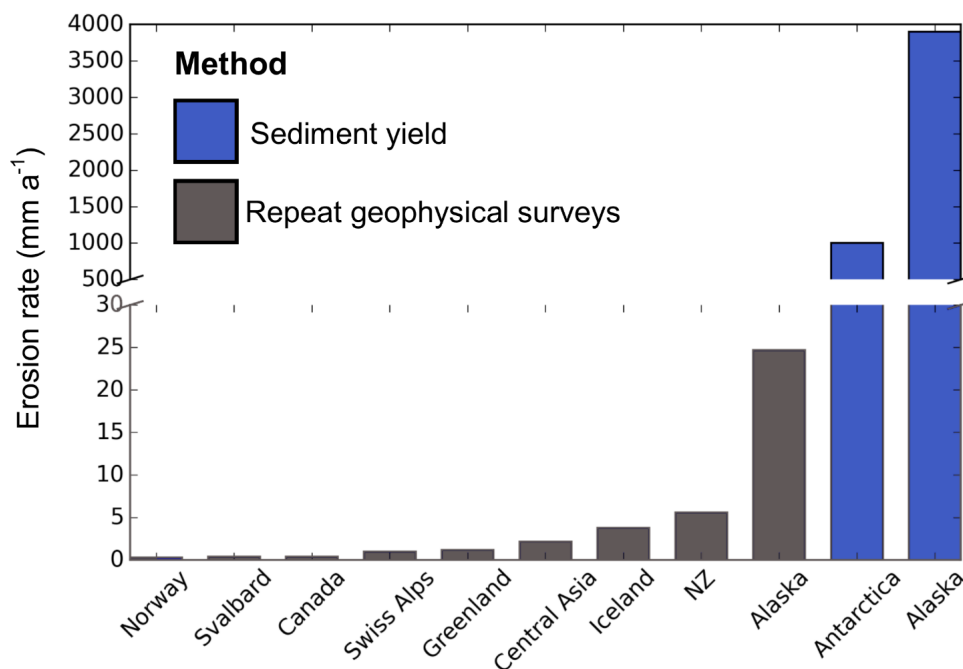


Figure 5: Close-up view of change in bed morphology between 2007/08 and 2013/14 identified in survey R1. Yellow boxes outline the area of change identified as having a low correlation ($r = -0.2$) (a) smoothed bed picks from radar data. (b, c) raw bed picks (red dots) of the basal reflector in 2007/08 and 2013/14.

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Figure 6: Mean erosion rates from previously published literature in a range of locations. Data taken from Hallet et al., (1996 and references therein), Koppes and Hallet (2006), Motyka et al., (2006), Cowton et al., (2012) and Smith et al., (2007,2012).