

Ancient pre-glacial erosion surfaces preserved beneath the West Antarctic Ice Sheet

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This discussion paper is/has been under review for the journal Earth Surface Dynamics (ESurFD). Please refer to the corresponding final paper in ESurf if available.

Ancient pre-glacial erosion surfaces preserved beneath the West Antarctic Ice Sheet

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Received: 16 June 2014 – Accepted: 29 June 2014 – Published: 15 July 2014

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Published by Copernicus Publications on behalf of the European Geosciences Union.

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be used to make inferences about ice-dynamic regimes (Siegert et al., 2005; Bingham et al., 2007; Bingham and Siegert, 2007, 2009). Most recently, Rippin et al. (2014) used a two-parameter Fast Fourier Transform to determine total roughness and the standard deviation of along-track bed topography to assess roughness directionality across the IMIS (see detailed methods therein). Here, we further develop the study by Rippin et al. (2014) by comparing regional-scale roughness characteristics and landscape geometry.

4 Results

4.1 Subglacial topography

4.1.1 Radar echograms

Figure 2 displays a series of radar echograms representing two cross-profiles and a single long-profile across the central IMIS survey area. The cross-profile radar echograms illustrate a smooth, gently sloping bed in the region between the Robin Subglacial Basin and the Pirrit and Martin–Nash Hills (Fig. 2a and b). Proximal to the Robin Subglacial Basin, the lateral extent of this smooth bed is truncated on either side by the Ellsworth Trough and Transitional Basins (Fig. 2a). Further inland, the smooth bed is still evident but is less obviously continuous (Fig. 2b, red dashed lines), being dissected by a number of typically U-shaped valleys. This radar echogram also shows evidence of similar (but more dissected) near-level surfaces in the region of the Marginal Basins, east of the Transitional Basins (Fig. 2b, green dashed lines). The long-profile radar echogram shows that there is a pronounced break in slope and change in elevation along this profile, approximately 80 km inland from the edge of the Robin Subglacial Basin (Fig. 2c). This marks the presence of two discrete topographic elements, which form a lower and an upper surface. The lower surface appears to dip towards the Robin Subglacial Basin, whilst the upper surface shows a distinct tilt inland.

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4.1.2 DEM

The DEM shows the full extent of the surfaces identified in the radar echograms (Fig. 2), revealing a smooth, gently sloping, roughly rectangular topographic block, located in the region between the Robin Subglacial Basin and the Pirrit and Martin–Nash Hills (Figs. 1b, dashed black box, and 3a, black dot-dash line). The block is ~ 200 km × 150 km in size and is bounded laterally by the Ellsworth Trough and the Transitional Basins. Slope analysis highlights a number of geomorphic features in the topography (Fig. 3c). Steep slopes ($> 7^\circ$) are associated with the Pirrit and Martin–Nash Hills. They are also found at the lateral boundaries of the block, along the margins of the Ellsworth Trough and Transitional Basins. These areas are characterised by steep-sided valley sides and flat bottomed floors, indicative of U-shaped cross-profiles. The block itself has generally low slope gradients ($< 4^\circ$), reflecting a gently dipping surface profile. The middle of the block is dissected by a linear zone of higher slope values ($6\text{--}8^\circ$), transverse to ice flow, which divide it into two sections (Fig. 3c). This break in slope is in keeping with observations from radar echograms at the ice surface and bed (Fig. 2c), and we note it is also visible in DEMs of the ice-sheet surface (e.g. Bamber et al., 2009; Fretwell et al., 2013). Under present-day ice cover, the elevation of the topographic block is predominantly below sea level (within a range of 200 m to -1500 m; mean of -656 m). When an isostatic correction is applied (Fig. 3b, contours), the elevation of the block rises (within a range of 600 to -1100 m), but mean elevation remains below sea level (mean of -272 m). It is also noted that the estimated sea level contour (0 m) corresponds with the break in slope identified across the centre of the block (Fig. 3b and c).

4.1.3 Hypsometry

The hypsometry of the IMIS drainage basin displays two area-elevation maxima that are skewed downward (H_{\max} values of -900 and -400 m), demonstrating that the majority of the basin lies below sea level (within a range of 1591 to -1971 m; mean of

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–643 m; Fig. 4a and b). The style of hypsometry is more consistent with glacial, rather than fluvial, environments (Egholm et al., 2009; Fig. 4b, inset). The area-elevation distribution of the gently sloping topographic block (Fig. 3a, black dot-dash line) was also determined, in order to assess its setting within the context of the IMIS basin hypsometry. Because the radar data (Fig. 2c), in conjunction with slope analyses (Fig. 3c), reveal that the block is divided into two surfaces, its hypsometry was also sub-divided into a lower surface proximal to Robin Subglacial Basin and an upper surface located further inland (Fig. 4a, red and blue dashed lines).

The hypsometric distributions of the two surfaces fall within the central elevation range of the IMIS drainage basin bed (within a range of –100 to –1100 m). Both the lower and upper surfaces show hypsometric maxima that are in line with the lower and upper H_{\max} peaks recorded in the IMIS basin (lower $H_{\max} = -800$ m, upper $H_{\max} = -500$ m, respectively; Fig. 4b). This highlights that the majority (almost 60 %) of the drainage basin corresponds with the gently sloping block. The two surfaces that comprise the block lie almost completely below sea level (< 1 % area above 0 m), in keeping with their broad sea embayment setting. For each surface, a large proportion of the area lies within an elevation range of 200 m, highlighting the low relief of this block. Specifically, 50 % of the lower surface area lies between elevations of –800 to –1000 m and 40 % of the upper surface area lies between elevations of –400 to –600 m.

The glacio-isostatically corrected bed topography (Fig. 3b, contours) allows comparison between present-day and pre-glacial hypsometric distributions (Fig. 4b and c, respectively). In the absence of ice cover to depress the lithosphere, the mean elevation of the IMIS drainage basin rises to –240 m (Fig. 4c). A similar pattern is observed where two hypsometric maxima are recorded in the IMIS basin and these correspond with the individual maxima for the lower and upper block surfaces (lower $H_{\max} = -400$ m, upper $H_{\max} = 0$ m). In particular, H_{\max} for the upper surface is at sea level (0 m) under ice free conditions and 58 % of the landscape area lies within 200 m of glacial-isostatic sea level (between 100 and –100 m).

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4.1.4 Roughness

A thorough analysis of spectral basal roughness across the IMIS was carried out by Rippin et al. (2014). Here, we discuss the total basal roughness and roughness directionality derived from that work (Fig. 5), and its relation to the morphological analyses presented above. Total basal roughness shows an area of low roughness (≤ 0.1) across the block (Fig. 5b). A band of slightly higher total roughness (0.1–0.2) is also evident in the centre of this region, corresponding with the break in slope in the topography (Figs. 2c and 3c). Low roughness values are also found across the sediment-filled Robin Subglacial Basin and the Bungenstock Ice Rise. Higher roughness values (> 0.2) are typically associated with the surrounding Ellsworth Mountains and Pirrit and Martin–Nash Hills. Again, when examining the directionality of roughness relative to present-day ice flow, the lowest values are associated with the Robin Subglacial Basin and Bungenstock Ice Rise, whilst the highest values generally correlate with the mountains. In other regions, such as the Transitional Basins, we find that roughness parallel to flow is typically lower than that orthogonal to flow. In contrast, however, the block shows a different pattern, whereby roughness is lower (0.24) orthogonal to ice flow (Fig. 5c), but higher (0.35) parallel to flow (Fig. 5d). These patterns reveal that the block has generally low basal roughness, dominated by shorter wavelength variations (typically parallel to ice flow), which gives it a distinct roughness character within the IMIS basin (Rippin et al., 2014). Similarly, the Robin Subglacial Basin is also defined by its own distinct (extremely smooth) roughness character that is particularly evident in the patterns of roughness directionality (Fig. 5c and d).

4.2 Ice sheet surface imagery

MODIS MOA imagery highlights a number of key morphological features in the ice sheet surface that reflect distinct changes in the nature (e.g. roughness, elevation) of the underlying topography (Fig. 3e). The isolated granite intrusions of the Pirrit and Martin–Nash Hills are characterised by a rough surface texture (Fig. 3f, blue lines),

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continuous wave-cut platform. Given their similarity and likely shared origin, it is interesting to note that both the Ross and Weddell Sea embayments have been affected by the same marine geomorphic processes, despite the difference in their geological setting. The wave-cut platforms in the Ross Sea embayment are part of the West Antarctic Rift System, whilst those in the Weddell Sea embayment are near the much older Weddell Sea Rift and the tectonic boundary between west and east Antarctica.

This previous work demonstrates that wave-cut platforms often extend for many 100 s km around coastlines. We have identified one particular area in the Weddell Sea embayment where a large component of the wave-cut platform has been well-preserved. However, we also find additional evidence indicating that this platform may have been more extensive in the past. Radar echograms, for example, hint to the existence of a more laterally continuous erosion surface extending across to the Transitional and Marginal Basins (Fig. 2b, green dashed lines). This is indicated by the presence of a few, isolated, flat topped hills (or inselbergs) between the basins (Figs. 1b and 2b). These basins are tectonically controlled (Jordan et al., 2013) and it is likely that they developed due to glacial erosion exploiting pre-existing tectonic structures and down-cutting through the original coastal surface, retaining only remnants of this surface as inselbergs (Burbank and Anderson, 2012). Given that glacial erosion is likely to have been extensive across these basins, greater modification, and a less continuous surface, would be expected. This is particularly true of the Transitional Basins, which underlie one of the primary IIS tributaries and where low roughness values parallel to ice flow highlight that glacial erosion has been particularly effective (Fig. 5d). Indeed, the preservation of the large area of the topographic block highlights its unique setting in this region. Elsewhere, glacial erosion has had a much more significant impact on the landscape. This is further demonstrated by the fact that 60 % of area-elevations in the IMIS drainage basin correspond with the block, but the block only accounts for 15 % of the area of the drainage basin. From this we may infer that the ancient wave-cut platform once occupied a significant proportion of the embayment and it has since been dissected by glacial erosion. Therefore, with the exception of the well-preserved

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of material. Third, to demonstrate this further, we can use radar echograms to compare the appearance of the erosion surfaces with the Robin Subglacial Basin. The latter has a very smooth and flat appearance, and records the lowest roughness values for the survey area (Fig. 5), giving it a distinct roughness character (Rippin et al., 2014) indicative of being sediment covered (Siegert et al., 2004; Ross et al., 2012). In contrast, the gently sloping erosion surfaces display small-scale undulations and intermittent valleys (Fig. 2a), giving this region a different (short wavelength) roughness (Fig. 5a), that distinguishes it from the likely sediment filled Robin Subglacial Basin (Rippin et al., 2014). Although we cannot rule out that there may be pockets of sediments or even a thin sedimentary drape in places, there is no evidence from radar or gravity data for a saturated or thick sedimentary deposit in this region. The surfaces are, therefore, more in keeping with an erosional bedrock landscape setting.

We also favour marine erosion, rather than fluvial or glacial erosion, as the dominant process responsible for the formation of the surfaces. Fluvial erosion processes erode towards a base level, typically at sea level. However, given the broad extent of the surfaces and their setting in a marine embayment, we consider destructional marine terrace formation (Burbank and Anderson, 2012) to be the dominant erosion process in this case. This is particularly applicable given that the erosion surfaces may have been much more extensive around the Weddell Sea embayment, as inferred from the regional-scale topography (Fig. 1a). We consider that a glacial origin for the erosion surfaces is also unlikely. Under present-day ice sheet conditions, the comparatively thin ice and low ice flow velocities ($< 50 \text{ m a}^{-1}$), coupled with low roughness values for this region (Figs. 3d and 5), are consistent with low rates of erosion. This region was also unlikely to have been a fast flow zone at earlier times in the Quaternary, as there would be slow, interior ice sheet-type flow across this region when ice was at the shelf edge. Furthermore, glacial erosion is more commonly associated with widening and deepening existing valley features (e.g. Siegert and Dowdeswell, 1996; Kessler et al., 2008), rather than shearing off flat bedrock surfaces (Embleton and King, 1975). In North America, for example, glacial erosion was responsible for the removal of sedimentary

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deposits and the development of a “knock and lochan” style landscape, but the underlying Canadian Shield geology was the dominant control on the gently undulating bedrock surface retained (Embleton and King, 1975). Therefore, whilst processes of areal scour may have modified the landscape on the micro-scale, at a macro-scale the dominant mechanism generating the gently-sloping surfaces identified is likely to be marine-erosional processes.

In addition, glacial erosion by ice is largely independent of sea level, with glacial overdeepenings, for example, frequently forming 100s to 1000s m below sea level. The deep troughs that flank the erosion surfaces, such as the Ellsworth Trough, indicate where this has occurred. Similarly, further inland, the hills and subglacial highlands show evidence of well-developed, alpine-style glacial erosion (Ross et al., 2014). Roughness values are also much higher, reflecting the mountainous relief and extent of warm-based glacial erosion. In contrast, the erosion surfaces do not show alpine-glacial features, such as hanging valleys and truncated spurs (Fig. 2). Instead they display their own distinct roughness signal, with comparatively high values parallel to ice flow vs. those orthogonal to flow (Fig. 5c and d). This indicates that the landscape has not experienced significant streamlining by ice, but rather retains a signal of pre-glacial geomorphic processes (Rippin et al., 2014).

A marine origin for the erosion surfaces is applicable for a number of reasons. The lateral continuity of the surfaces and the fact that they average the same elevation over large distances are both highly indicative of wave-cut platforms (Wilson and Luyendyk, 2006a, b). This is further demonstrated when, following correction for glacial isostatic adjustment, bed elevations approximate sea level (Figs. 3b and 4c), whereas glacial erosion functions independent of sea level. Marine erosion processes are concentrated at the interface between land and sea through the constant action of waves impacting a shoreline, often during a period of tectonic quiescence (Burbank and Anderson, 2012). The setting of the erosion surfaces, in the centre of the embayment, would be particularly susceptible to marine erosion, receiving maximum exposure to wave action. Given that the topography is located around where one would expect sea

in their lee, thereby maximising the potential for landscape preservation. These mountains significantly modulate regional ice flow, reducing downstream ice velocities and focusing flow along the adjacent troughs and basins that now host the fast flowing ($> 50 \text{ m a}^{-1}$) tributaries of the IIS (Fig. 3d).

Despite the present-day setting, it is evident that the surfaces have been subject to some degree of glacial erosion following formation. A few larger (often U-shaped) valleys are visible in cross-profile A, particularly in proximity to the Ellsworth Trough (Fig. 2a); whilst further inland cross-profile B has been more significantly dissected by broader U-shaped valleys (Fig. 2b). These intermittent, U-shaped valleys are suggestive of selective linear erosion by small- to regional-scale, warm-based, ice masses (Sugden and John, 1976; Hirano and Aniya, 1988). The location of the valleys may reflect pre-existing fluvial networks that have been exploited (e.g. Baroni et al., 2005; Rose et al., 2013; Ross et al., 2014). The scale and style of this glacial erosional overprinting is characteristic of warm-based, outlet glaciers, prior to the onset of extensive West Antarctic glaciation. These ice masses would be subject to topographic steering and could therefore flow around the Pirrit and Martin–Nash Hills, enabling glacial incision of the erosion surfaces to occur. Ross et al. (2012) use the location of troughs and elevated bars to infer the position of a former grounding line upstream of the Robin Subglacial Basin. Their findings indicate that, even under a smaller-scale, restricted ice sheet configuration, the majority of ice drainage would likely follow the linear topographic valleys that flank this block. This would minimise the degree of glacial incision, helping to retain pre-glacial landscape signals across this region. Furthermore, once the ice sheet expanded to a continental scale, buttressing of ice sheet flow by the Pirrit and Martin–Nash Hills would also restrict extensive landscape modification by glacial erosion downstream of these major mountain massifs. The existence of an extensive, although incised, wave-cut platform demonstrates that it is possible to preserve ancient surfaces beneath an ice sheet, even at low elevations and not just at high elevations associated with thin, cold-based ice (Rose et al., 2013; Ross et al., 2014). Furthermore, it is interesting to note that whilst regions of smooth and flat topography are typically

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sea level elevations we infer (Zachos et al., 2001). Following the onset of Antarctic glaciation (~ 34 Ma), we may estimate the most recent period in the past when the surfaces experienced significant marine erosion. In the Ross Sea sector, the erosion surfaces identified by Wilson and Luyendyk (2006a, b) are interspersed with glacial marine sediments. These sediments can often be dated and their stratigraphy used to interpret the glacial history of the region (e.g. Naish et al., 2001, 2008). Using the location of the erosion surfaces in relation to these sediments, Wilson and Luyendyk (2006b) infer that the most recent period of sustained marine erosion was during the Mid-Miocene Climatic Optimum (17–15 Ma). At this time, temperatures (and therefore, sea level) were high enough to provide sufficiently ice free conditions at the coast in order for wave erosion to act over a prolonged period (in this case ~ 2 Myr) and thereby contribute to the development of the wave-cut platform (Harwood et al., 1989; Zachos et al., 2001; Miller et al., 2005).

It seems feasible, therefore, that the wave-cut platform identified in the Weddell Sea sector was subject to a similar erosion history. In order to form, ice free conditions must prevail at the coast during periods of ice retreat to allow wave action to occur. In addition, for elevations to locate at sea level, the topography could not have been subject to glacial isostatic depression. This makes it unlikely that a major ice mass was established in the region at that time. Such ice-free conditions would be met in the Middle Miocene. However, from ~ 14 Ma, proxy records indicate that a continental-scale ice sheet was established across Antarctica, in response to a gradual cooling of the climate (Zachos et al., 2001; Miller et al., 2005). It is likely, therefore, that ice expanded to fill the Weddell Sea embayment from this time. This would shift the marine margin towards the continental shelf and lead to the glacial isostatic depression of the landscape, as observed today.

Temperatures were also warmer than present in the Pliocene (~ 5–3 Myr ago) (Dowsett, 2007) and the marine glacial record indicates considerable oscillations in the WAIS during this time (Naish et al., 2009). Marine erosion during these intervals of periodic ice sheet collapse cannot be ruled out. However, these variations have been

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linked to obliquity paced cycles of ~ 40 kyr (Naish et al., 2009), suggesting that the duration of any marine erosion processes would be much shorter than the suggested 100 kyr needed for significant and persistent erosion (Wilson and Luyendyk, 2006a, b). It is also unclear whether sea ice was fully dispersed during interglacials and the retention of such ice within the embayment would hinder marine erosion at wave base (Burbank and Anderson, 2012). We, therefore, favour the Mid-Miocene Climatic Optimum as a more likely candidate for the last phase of significant marine erosion. We also envisage that the large fluctuations in ice extent recorded during the Pliocene (Naish et al., 2009) may be responsible for the signal of warm-based glacial erosion that is overprinted on the erosion surfaces.

7 Conclusions

A new DEM, built from an extensive and high resolution modern airborne radar survey, provides a detailed view of subglacial topography across the IMIS, where previously only sparse data were available (Bingham and Siegert, 2007). We have examined radar echograms and MODIS MOA imagery and applied morphometric analyses to the DEM, in order to characterise the landscape. In doing so, we have: (1) identified a smooth, laterally continuous, gently sloping topographic block in the region between the Robin Subglacial Basin and the Pirrit and Martin–Nash Hills, (2) characterised this block as two surfaces separated by a distinct break in slope, (3) determined that pre-glacial elevations for these surfaces approximate a glacial-isostatic sea level, (4) shown that erosion rates across the surfaces are currently low, precluding formation via present-day glacial erosion; and (5) interpreted these features as marine erosion surfaces, indicative of wave-cut platforms.

Our findings show that it is possible for ancient pre-glacial erosion surfaces to be preserved at low elevations beneath ice sheets. We have also identified the Pirrit and Martin–Nash Hills as having played a key role in the long-term landscape evolution of this region. By buttressing upstream ice and reducing downstream ice velocities they

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have limited glacial erosion rates and enabled long-term preservation of the wave-cut platforms. By modulating ice dynamics, these mountain massifs have facilitated a region where smooth, low geometry basal topography does not correspond with the fast ice flow typically associated with ice streams. We are not able to constrain the timing of formation for the wave-cut platforms. However, we propose that the dominant phase of marine erosion likely occurred during the Paleocene and Eocene (~ 66–34 Ma). Following the onset of widespread Antarctic glaciation, the most recent period when warm atmospheric and oceanic temperatures would allow prolonged exposure of this region to marine erosion processes is during the Mid-Miocene Climatic Optimum (17–15 Ma). This is in keeping with the interpretations of Wilson and Luyendyk (2006a, b) for wave-cut platforms in the Ross Sea embayment.

Acknowledgements. This project was funded by UK NERC AFI grant NE/G013071/1. Carl Robinson (Airborne Survey engineer), Ian Potten and Doug Cochrane (pilots), and Mark Oostlander (air mechanic) are thanked for their invaluable assistance in the field.

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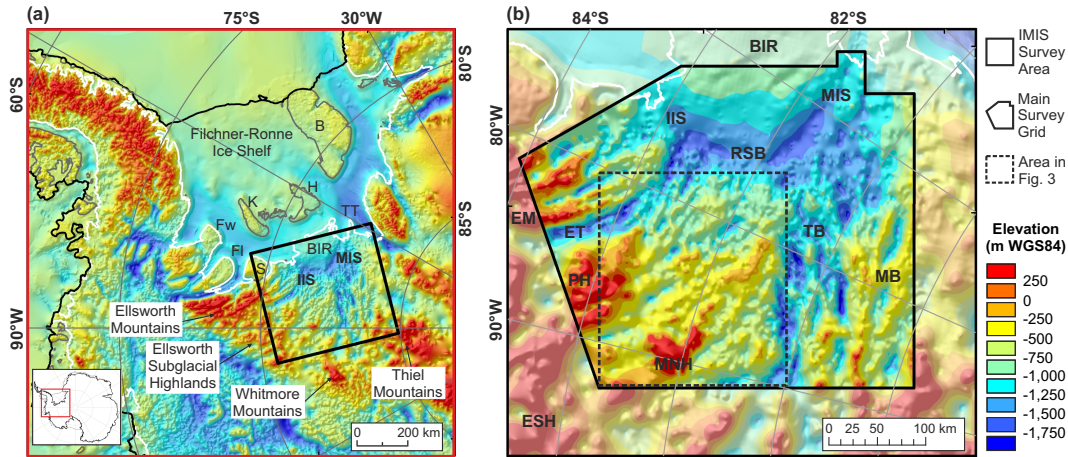


Figure 1. (a) Subglacial topography of the Weddell Sea sector, West Antarctica (Fretwell et al., 2013). White line marks the grounding line, black line marks the outer ice shelf edge, closed grey lines mark islands (derived from MODIS MOA imagery; Haran et al., 2005, updated 2013). Inset with red box shows the location of (a) in West Antarctica. (b) Subglacial topography of the Institute and Möller ice streams (IMIS), derived from a high resolution, modern airborne survey (Ross et al., 2012), and overlain on semi-transparent Bedmap2 topography. Annotations: B – Berkner Island; BIR – Bungenstock Ice Rise; EM – Ellsworth Mountains; ESH – Ellsworth Subglacial Highlands; ET – Ellsworth Trough; F – Fletcher Promontory; Fw – Fowler Peninsula; H – Henry Ice Rise; IIS – Institute Ice Stream; K – Korff Ice Rise; MB – Marginal Basins; MIS – Möller Ice Stream; MNH – Martin–Nash Hills; PH – Pirrit Hills; RSB – Robin Subglacial Basin; S – Skytrain Ice Rise; TB – Transitional Basins; TT – Thiel Trough.

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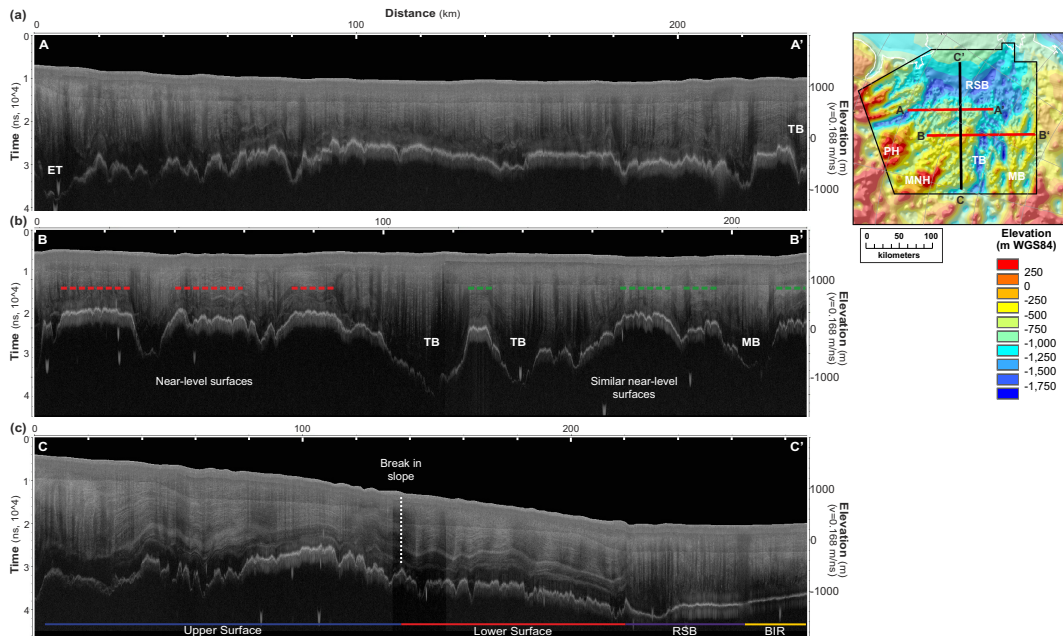


Figure 2. Radar echograms showing the gently sloping surfaces identified in subglacial topography. Inset shows the location of transects overlain on subglacial topography. **(a)** Cross-profile transect A-A', located proximal to the Robin Subglacial Basin (RSB), between the main tributaries of the Institute Ice Stream. **(b)** Cross-profile transect B-B', located further inland, extending between the Pirrit Hills (left) and the Marginal Basins (right), crossing the easternmost tributary of the Institute Ice Stream, which is underlain by the Transitional Basins. Cross-profiles show ice flow going in to the page. **(c)** Long-profile transect C-C', running from the inland edge of the survey grid (left) to southern edge of the Bungenstock Ice Rise (right). Ice flow is roughly from left to right, but changes orientation across the Robin Subglacial Basin. Elevations are relative, topography in all transects is located below sea level (see inset). Annotations: BIR – Bungenstock Ice Rise; ET – Ellsworth Trough; MB – Marginal Basin; MNH – Martin–Nash Hills; PH – Pirrit Hills; RSB – Robin Subglacial Basin; TB – Transitional Basins.

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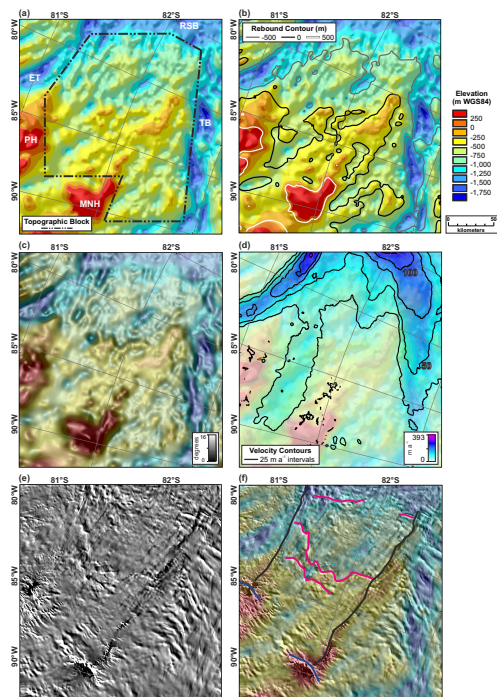


Figure 3. (a) Subglacial topography between the main Institute Ice Stream tributaries. Annotations: ET – Ellsworth Trough; MNH – Martin–Nash Hills; PH – Pirrit Hills; RSB – Robin Subglacial Basin; TB – Transitional Basins. (b) Present-day bed elevations, overlain with elevation contours derived from the isostatically adjusted topography, at 500 m intervals. (c) Surface slope (semi-transparent) overlain on subglacial topography. (d) Ice sheet surface velocity (semi-transparent) (Rignot et al., 2011), overlain on subglacial topography. (e) MODIS MOA imagery of the ice sheet surface. (f) MODIS MOA imagery (semi-transparent) overlain on subglacial topography and annotated to show the dominant morphological features observed. Extent of panels shown by dashed black box in Fig. 1b.

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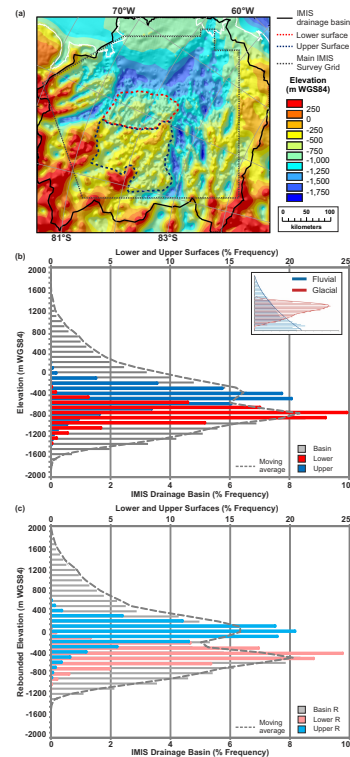


Figure 4. (a) Drainage basin determined for the subglacial topography in the region of the Institute and Möller ice streams (IMIS). (b) Hypsometry (area-elevation distribution) determined for the IMIS drainage basin, in comparison with the lower and upper surfaces. Inset shows characteristic hypsometric distributions for fluvial and glacial landscapes (Egholm et al., 2009). (c) Hypsometry determined for pre-glacial (isostatically corrected) elevations across the IMIS basin and the lower and upper surfaces.

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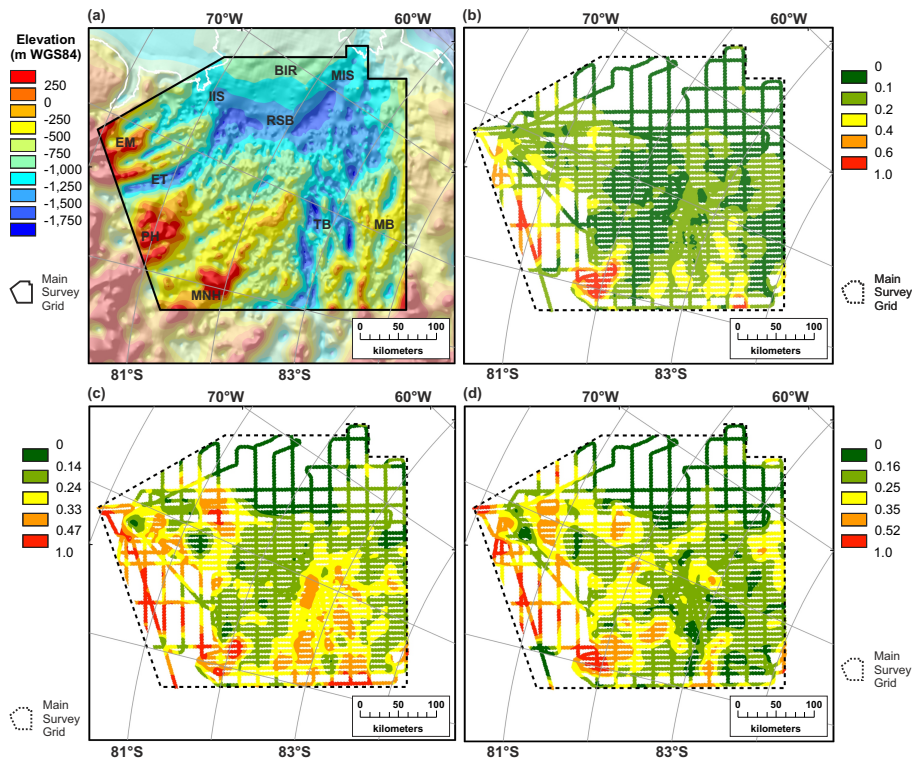


Figure 5. (a) Subglacial topography of the Institute and Möller ice streams overlain on semi-transparent Bedmap2 topography. Annotations: BIR – Bungenstock Ice Rise; EM – Ellsworth Mountains; ET – Ellsworth Trough; IIS – Institute Ice Stream; MB – Marginal Basins; MIS – Möller Ice Stream; MNH – Martin–Nash Hills; PH – Pirrit Hills; RSB – Robin Subglacial Basin; TB – Transitional Basins. (b) Total basal roughness determined from bed elevation data along IMIS survey flight lines. (c) Basal roughness determined using the standard deviation of along-track bed topography orthogonal to ice flow. (d) Basal roughness determined using the standard deviation of along-track bed topography parallel to ice to flow. Adapted from Rippin et al. (2014).

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