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

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Abstract

We present ice-penetrating radar evidence for $\sim 150 \text{ km}$ wide planation surfaces beneath the upstream Institute and Möller Ice Streams, West Antarctica. Accounting for isostatic rebound under ice-free conditions, the surfaces would be around sea level.

- ⁵ We, thus, interpreted the surfaces as ancient, marine erosion (wave-cut) platforms. The scale and geometry of the platforms are comparable to erosion surfaces identified in the Ross Sea embayment, on the opposite side of West Antarctica. Their formation is likely to have begun after the development of the deep ocean basin of the Weddell Sea (~ 160 Myr ago). In order to form wave-cut platforms, the sea must be relatively free of
- sea ice for a sustained period to allow wave erosion at wave base. As a consequence, the most recent period of sustained marine erosion is likely to be the Mid-Miocene Climatic Optimum (17–15 Ma), when warm atmospheric and oceanic temperatures would have prevented ice from blanketing the coast during periods of ice-sheet retreat. The erosion surfaces are preserved in this location due to the collective action of the Pirrit
- and Martin–Nash Hills on ice-sheet flow, which results in a region of slow flowing, coldbased ice downstream of this major topographic barrier. This investigation shows that smooth, flat subglacial topography does not always correspond with regions of either present or former fast ice flow, as has previously been assumed.

1 Introduction

The Institute and Möller Ice Streams (IMIS) drain around 20% of the West Antarctic Ice Sheet (WAIS) (Fig. 1). Despite their significance as fast flowing outlet glaciers, within an ice sheet that is regarded as potentially unstable (Joughin et al., 2014), relatively little was known about the glacial history of this region (Bingham and Siegert, 2007). In order to address this issue, an aerogeophysical survey was undertaken across the ice streams and surrounding locations (Ross et al., 2012). Mapping ice-sheet bound-





ary conditions in these little explored regions is of great importance, as subglacial

topography can exert a strong control on ice dynamics (Joughin et al., 2009) and may retain a long-term record of surface processes and ice-sheet evolution (Young et al., 2011). It is also possible to use bed topography to make inferences about the nature and evolution of palaeo-landscapes in Antarctica (Rose et al., 2013).

- ⁵ The IMIS survey has provided a wealth of new information, elucidating the tectonic, topographic and hydrological settings of this region (Ross et al., 2012; Jordan et al., 2013; LeBrocq et al., 2013; Siegert et al., 2014). These, in turn, have contributed to our understanding of early ice inception (Ross et al., 2014) and ice sheet sensitivity in the Weddell Sea sector (Siegert et al., 2013; Wright et al., 2013, TCD). However,
- the macro-scale geomorphology of the IMIS sector has yet to be considered. In particular, the region between the Robin Subglacial Basin and the mountain ranges of the Ellsworth–Whitmore block has received little attention (Fig. 1). Here, we focus on this striking region, where a zone of apparently flat and smooth topography, bounded laterally by the deeper elongate troughs and basins that underlie the fast flowing tributaries of the Institute Ice Stream (IIS), is imaged and mapped. We inspect and analyse the
- ¹⁵ of the institute fice Stream (IIS), is imaged and mapped. We inspect and analyse the morphology of this sector, in order to determine the origin and likely evolution of the landscape in relation to long-term glacial history.

2 Regional setting

The Weddell Sea sector of the WAIS (Fig. 1), in which the IMIS are contained, is characterised by a broad, marine (below sea level) embayment (Ross et al., 2012). The ice streams drain into the Filchner–Ronne Ice Shelf and are separated by the Bungenstock Ice Rise at the grounding line (Fig. 1b). The area immediately inland of the grounding line is dominated by the ~ 1.5 km deep, (likely) sediment-filled, Robin Subglacial Basin. Two primary tributaries feed the fast ice flow of the IIS (> 50 m a⁻¹). These are underlain by the Ellsworth Trough and a series of right-stepping, en-echelon, Transitional Basins (Jordan et al., 2013). In contrast, the onset of fast flow of the Möller Ice Stream





and the northern margin of the Transitional Basins. Topography on the eastern (MIS) side of the survey area lies at the boundary of West and East Antarctica and is strongly influenced by the tectonic history of this region (Jordan et al., 2013). Here, there are six elongate Marginal Basins flanked by a series of magnetic lineaments, interpreted as faults (Jordan et al., 2013). The IMIS survey area is bordered by numerous uplands that include the Ellsworth, Thiel and Whitmore Mountains, and the Ellsworth Subglacial Highlands (Fig. 1a). A number of Jurassic granite intrusions, emplaced between ~ 175–165 Ma (Storey et al., 1988; Lee et al., 2012), form isolated peaks across the study area, including the Pirrit Hills and Martin–Nash Hills. Aerogeophysical data suggest
the granites at the Pirrit and Martin–Nash Hills are restricted to the local subglacial highlands (Jordan et al., 2013). The main focus of this paper is the inter-tributary area of the IIS, between the Robin Subglacial Basin and the Pirrit and Martin–Nash Hills further inland.

3 Approach

15 **3.1 Data collection**

In the austral summer of 2010/2011 an airborne geophysical survey was carried out across the IMIS. A central survey grid, with flightline spacing of 7.5 km and tie lines of 25 km, was established over the ice streams, covering an area of \sim 350 km × 400 km. Additional exploratory lines, with 50 km spacing, were also flown to link with previous re-

- gional surveys (Ross et al., 2012). Approximately 25 000 line-km of radio-echo sounding (RES), gravity and magnetic data were collected (Jordan et al., 2013). Here, we discuss only the RES data. These were collected using a coherent system with a 12 MHz bandwidth and 150 MHz carrier frequency (Corr et al., 2007), providing an approximate 10 m along-track sampling interval. Differential GPS, with a horizontal accuracy
- ²⁵ of ~ 5 cm, was used for positioning. Doppler processing was applied to migrate radarscattering hyperbolae in the along-track direction (Hélière et al., 2007). The seismic





processing software PROMAX was then used to carry out a semi-automated picking sequence that identifies the onset of the received bed echo.

Ice thickness was determined from the two-way traveltime of the bed pick, using a velocity of $0.168 \,\mathrm{m \, ns^{-1}}$, coupled with a firn layer correction of 10 m (Ross et al.,

⁵ 2012). Bed elevations were then calculated by subtracting ice thickness measurements from ice surface elevations. The latter were determined from measurements of terrain clearance derived from radar/laser altimeter measurements, relative to the WGS84 ellipsoid. Cross-over RMS errors are ~ 18 m (Ross et al., 2012). These data have since contributed to the Bedmap2 depiction of Antarctic subglacial bed elevation (Fretwell et al., 2013).

3.2 DEM, radar echograms and satellite imagery

A DEM of subglacial topography was produced by rendering RES-derived bed elevations onto a 1 km grid mesh, using the "Topo to Raster" iterative finite difference interpolation function in ArcGIS. This algorithm employs a nested grid strategy to calculate suscessively finer gride until the user enceified resolution is obtained (Hutabingen

- ¹⁵ late successively finer grids until the user specified resolution is obtained (Hutchinson, 1988, 1989); and it has been shown to be particularly effective in rendering glacial terrain (Fretwell et al., 2013). In addition, an isostatic correction was applied to the DEM to account for the removal of the modern ice-sheet load. This comprised a simple Airy-type compensation, with an ice density of 915 kg m⁻³ and a mantle density
- of 3330 kg m⁻³. We appreciate that this does not take into account the full complexity associated with glacio-tectonic interactions. However, it does provide an indication of pre-glacial elevations across the region, offering insight into the landscape setting prior to glaciation. Radar echograms were also studied in order to assess detailed (along-track) basal topography and englacial structure. These images offer an immedi-
- ate regional-scale perspective on the nature of the sub-ice landscape and add specific context to the DEM. Furthermore, the recently updated Moderate-Resolution Imaging Spectroradiometer (MODIS) Mosaic of Antarctica (MOA) (Haran et al., 2005, updated 2013) was used to understand the morphology of the ice sheet surface (Scambos et al.,





g roughı ,

2007), as this can provide detailed insight into the nature of the underlying subglacial terrain (Ross et al., 2014).

3.3 Geomorphometry

The DEM was analysed in order to identify, map and quantify subglacial morphology and the spatial variability of geomorphic features. In particular, bed slope and hypsometry (area-elevation distribution) were quantified, as outlined below. Geomorphic features were also compared with maps of spectral bed roughness, derived from the IMIS survey radar transects (from Rippin et al., 2014). From the DEM, bed slope represents the gradient (rate of maximum change in *z* value) from each grid cell to its neighbours and is measured in degrees (Burrough and McDonell, 1998). Abrupt changes in slope can reveal distinct changes in landscapes and specific topographic features, as well as provide an overview of general surface texture.

Hypsometric analysis quantifies the distribution of land surface area with altitude (Strahler, 1952). It is commonly used to understand the relationship between local and
regional tectonics and the spatial variability in fluvial and glacial surface processes (Montgomery et al., 2001; Pedersen et al., 2010). Hypsometry identifies the dominant signal of landscape erosion, but is generally scale-dependent (Brocklehurst and Whipple, 2004). In order to assess hypsometry, therefore, the topographic drainage basin encompassing the IMIS was delineated. While this represents an artificial drainage divide boundary, as topography is predominantly below sea level within the survey area, it is a well-defined region that reflects the upslope area that contributes flow (normally water) to a common outlet (topographic low point along the basin boundary). Although it does not necessarily relate to a viable hydrological system (because rivers grade to sea level), it does provide a discrete scale at which to calculate hyp-

Basal roughness is a measure of vertical variation with horizontal distance (e.g. Taylor et al., 2004; Li et al., 2010; Rippin et al., 2011). As the nature of basal topography exerts a control on ice flow, quantifying roughness at the ice-bed interface can





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
 ⁵ 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 land-scape geometry.

4 Results

4.1 Subglacial topography

10 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 (Fig. 2c).
- ²⁵ Basin, whilst the upper surface shows a distinct tilt inland.





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
⁵ (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°) 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°), reflecting a gently dipping surface profile. The middle of the block is dissected by a linear zone of higher slope values (6–8°), 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

²⁵ 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 hypsom-

- ⁵ etry. 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.
- ²⁰ 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 ²⁵ correspond with the individual maxima for the lower and upper block surfaces (lower $H_{\text{max}} = -400 \text{ m}$, upper $H_{\text{max}} = 0 \text{ m}$). 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).





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 analy⁵ ses 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 10 (> 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
¹⁵ 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

4.2 Ice sheet surface imagery

in the patterns of roughness directionality (Fig. 5c and d).

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),





forming clustered "corrugated" features, orientated transverse to ice flow. Downstream of each hill is a linear feature, aligned with ice flow in the direction of the grounding line (Fig. 3f, black lines). Located between these are a series of linear to U-shaped features that lie transverse to ice flow (Fig. 3f, pink lines). The most distinct U-shaped surface feature runs across the middle of this region. These features, respectively, correspond with the lateral boundaries of the block and the break in slope identified in radar echograms and the DEM (Figs. 2c and 3c).

5 Interpretation – wave-cut platforms

Due to the gently sloping, low roughness geometry of the block, and given that "icefree" rebounded elevations tally with sea level, we interpret the region as two pre-glacial erosion surfaces that likely represent ancient, extensive wave-cut platforms. Such topography is commonly found globally and is often interlinked with glacial environments. In the Northern Hemisphere, wave-cut platforms, known as strandflats, are found along almost the whole west coast of Norway (Klemsdal, 1982), and in Greenland, Alaska and Scotland (Dawson et al., 2013). They are thought to result primarily from marine

- abrasion, with contributions from subsequent weathering and glacial erosion processes (Fredin et al., 2013). In Norway, strandflats are found both above and below sea-level. Typically, they lie in front of higher land or coastal mountains and form either level or gently dipping bedrock surfaces from the coast (Klemsdal, 1982; Fredin et al., 2013).
- We find a similar setting in the IMIS drainage basin, where the Ellsworth Subglacial Highlands and other hills are found inland of the erosion surfaces (Fig. 1a), and a gentle coastal dip is evident in the lower surface (Fig. 2c). In West Antarctica, a sequence of erosion surfaces has been identified in the Ross Sea sector, located between 100 and 350 m below sea level (Wilson and Luyendyk, 2006a, b). They are 100 s km wide and discontinuous, separated by troughs that are occupied by the ice streams of the Siple
- Coast. Using additional evidence from gravity and marine seismic data sets, Wilson and Luyendyk (2006a, b) also interpret these surfaces as the remnants of a former,





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 Bift System, whilst those in the Weddell Sea embayment are near the much

⁵ 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).
- ¹⁵ 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
- ²⁰ 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 wavecut 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





topographic block, only much smaller, localised, remnants of this surface remain, which are harder to identify from the DEM.

If we examine the landscape at a regional-scale, it is easy to see how a largerscale, more continuous wave-cut platform may have existed across the Weddell Sea ⁵ embayment in the past. In particular, Berkner Island and the Henry and Korff Ice Rises stand out as regions of gently dipping, smooth topography now encompassed by the Filchner–Ronne Ice Shelf (Fig. 1a). These regions also rebound close (±300 m) to sea level following isostatic correction. Closer to the present-day grounding line, the Fowler Peninsula, Fletcher Promontory and Skytrain Ice Rise also provide evidence for low gradient, low relief surfaces. The geometry of these features is in keeping with the topographic block we identify between the primary IIS tributaries. If we assume that they once formed part of a more continuous erosion surface, coupled with the block, then marine erosion was clearly a significant regional-scale process across the Weddell Sea embayment prior to glaciation, rather than a local one.

15 6 Discussion

6.1 Formation of the flat surfaces

We interpret that the flat surfaces discussed above reflect erosional rather than depositional features for three reasons. First, they are located inland of the grounding line where the movement of grounded ice and subglacial water is more likely to result in
²⁰ erosion. In particular, in the recent past when the ice sheet margin was located at the continental shelf, the erosion surfaces would have been located even further in the ice sheet interior, demonstrating that this is not an obvious position for major sedimentary deposition. Second, at both a regional scale and within the detailed IMIS survey area, it is clear from the relief of the landscape that this region has been predominantly
²⁵ subject to processes of erosion (Fig. 1). For example, the presence of troughs and basins below sea level, particularly the Ellsworth Trough, indicates significant removal



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

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- ²⁰ 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 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
- ²⁵ 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





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 1000sm 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 high-lands 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-

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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 preglacial 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 concen-

trated 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





level to be prior to glaciation, the break in slope may correspond to a former sea level (Fig. 3b and c). It may, therefore, represent the inner edge of the wave-cut platform, which closely approximates sea level at the time of formation. Wave-cut platforms typically have a gentle seaward slope (< 1°), which helps to transfer eroded material from

- the inner platform edge towards the deep ocean. Erosion does not typically extend below a wave base of ~ 10 m. However, here the extent of the erosion surface (largely concentrated within a 200 m elevation band; Fig. 4) may be the result of multiple occupations of similar relative sea level positions, the action of subsequent erosion processes and/or a gradual rise in sea level (Burbank and Anderson, 2012). This may also
 point to a more complex tectonic history in this region that cannot solely be explained
 - by glacial isostatic processes (Jordan et al., 2013).

In order to form these surfaces, the coastal region must have been relatively free of sea ice, to allow wave erosion to occur at wave base over a period of several 100 kyr (Wilson and Luyendyk, 2006a, b). In turn, this dictates that the wave-cut platform must

- ¹⁵ pre-date the occurrence of grounded ice and stable floating ice in this region. Since formation, however, the land has been depressed below sea level due to crustal loading, resulting from an expansion to present-day continental-scale ice cover. Indeed, the inland dip of the upper surface (Fig. 2c) suggests it may have been subject to regional tilt since formation (Burbank and Anderson, 2012). This may be the result of ice experience and/or due to testen within the West Anteretia Diff. Sustem between the
- ²⁰ cover and/or due to tectonic motion within the West Antarctic Rift System between the Cretaceous and Cenozoic (Jordan et al., 2010).

6.2 Preservation

The identification of ancient pre-glacial erosion surfaces between the IIS tributaries highlights that there has been long-term landscape preservation in this sector. MODIS MOA imagery reveals two linear surface features in the lee of the Pirrit and Martin–Nash Hills, which appear to delineate the "protected" area, corresponding with the lateral boundaries of the block (Fig. 3e and f). The Pirrit and Martin–Nash Hills clearly act as an ice-sheet pinning point, buttressing the upstream ice and minimising erosion



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 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 over-printing is characteristic of warm-based, outlet glaciers, prior to the onset of extensive
- ¹⁵ 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





associated with relatively fast ice flow and often inferred marine sediments (Joughin et al., 2006; Peters et al., 2006; Bingham and Siegert, 2007), we do not find that relationship here. Instead, the fastest ice flow exploits the deeply incised troughs and basins that flank the erosion surfaces.

5 6.3 Timing

Although long-term topographic elevations may be influenced by a range of tectonic processes, such as thermal subsidence, deep mantle flow, erosion and isostatic responses, it is interesting to debate the most likely age of formation for the wave-cut platforms. The Jurassic granites that comprise the Pirrit and Martin–Nash Hills were emplaced between ~ 175–165 Ma (Storey et al., 1988; Lee et al., 2012) and have since been exposed by erosion. Whilst this provides a maximum age for the onset of erosion across this region, the relief today between the mountain peaks and the upper erosion surface demonstrates that a significant amount of material (at least 2 km) has been removed since the Jurassic. It is unfeasible that all of this erosion occurred via ma-

- rine processes. Instead, the formation of the wave-cut platforms will reflect a significant stage of erosion with a continuum of landscape development. It is likely that fluvial erosion played an important role in the early evolution of the landscape, as rivers work to reduce surface elevations to a base level through the erosion, transportation and deposition of material (Burbank and Anderson, 2012). This region was most likely first
- exposed to marine processes after the deep ocean basin of the Weddell Sea began to form ~ 160 Myr ago (Jokat et al., 2003). As we lack dating controls it is not possible to provide a firm constraint on the timing of initial formation of the surfaces. Clearly, initial formation must have occurred before the onset of widespread Antarctic glaciation, prior to 34 Ma. It is likely that a combination of fluvial and marine erosion processes were
- operating under the sub-tropical to warm temperate climate conditions experienced in the Late Mesozoic to Early Cenozoic eras (Francis et al., 2008). We favour that the majority of formation occurred in the latter half of this period during the Palaeocene and Eocene (~ 66–34 Ma) when sea level was more likely to correspond with the pre-glacial





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
⁵ 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 continentalscale 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 ($\sim 5-3$ 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



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





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).
- ¹⁰ This is in keeping with the interpretations of Wilson and Luyendyk (2006a, b) for wavecut platforms in the Ross Sea embayment.

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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; FI – 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.



Elevation

-250 -500

-750

-1,000 -1,250 -1,500

-1.750





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.







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.







Figure 5. (a) Subglacial topography of the Institute and Möller ice streams overlain on semitransparent 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 alongtrack 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).

