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# Quantifying subglacial bed roughness in Antarctica: implications for ice-sheet dynamics and history

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ABSTRACT

Glaciated landscapes consist of complex assemblages of landforms resulting from ice flow dynamic regimes and ice-sheet history, superimposed over, and in turn modifying, preglacial topography, lithology and geological structure. Insights into the formation of glaciated landscapes can, in principle, be obtained by analysing modern ice-sheet beds, but terrain analyses beneath modern ice sheets are restricted by the inaccessibility of the bed. It is, however, possible to quantify roughness, the vertical variation of the subglacial interface with horizontal distance, along two-dimensional images of the bed obtained from radio-echo sounding (RES). Here we collate several case studies from Antarctica, where roughness calculations have been used as a glaciological tool to infer basal processes and ice-sheet history over large (>500 km$^2$) areas. We present two examples from West Antarctica, which demonstrate the utility of bed roughness in determining the presence and extent of subglacial sediments, glacial dynamics and former ice-sheet size. We also present two examples from East Antarctica, which illustrate how roughness provides knowledge of ice-sheet dynamics in the interior and pre-Quaternary ice-sheet histories. In modern ice-sheet settings, characterising bed roughness along RES tracks has the twin advantages of being relatively simple to calculate while producing informative subsurface data, and is especially powerful at furthering understanding when coupled with knowledge of ice flow from field, satellite and modelling investigations. The technique also offers significant potential for the comparison of modern and former ice-sheet terrains, contributing to an improved understanding of the formation and evolution of glaciated landscapes.



## 1. Introduction

Glaciated landscapes may be considered palimpsests of landforms (Kleman, 1992) that reflect changing ice dynamic and thermal regimes with ice-sheet history superimposed over, and in turn modifying, preglacial topography and the underlying lithology and geological structure. These landscapes can be used to reconstruct former ice dynamics and thermal regimes in the Quaternary ice sheets that covered much of northern North America and Eurasia at the Last Glacial Maximum (~21,000 B.P.) (Siegert, 2001; Evans, 2003). This goal has been aided particularly in recent years with the advent of remote sensing techniques able to visualise landscapes on the continental scale (>500 km$^2$) from space (e.g., Clark et al., 2004; Napieralski et al., 2007; Ó Cofaigh and Stokes, 2008). In principle, insights into the formation of glaciated landscapes can be obtained by analysing modern ice-sheet beds as analogues, but continental-scale terrain analyses of subglacial topography beneath modern ice sheets are restricted by the inaccessibility of the bed. Through ice several 100s to 1000s of m thick, the only effective means of capturing bed topography over the continental-scale is through airborne radio-echo sounding (RES). Interpolation between bed elevations obtained largely by airborne RES, but also some seismic and oversnow radar measurements, has been used to compile a digital elevation model (DEM) of Antarctic subglacial topography (BEDMAP; Lythe et al., 2001). Unfortunately, the resolution of BEDMAP (5 km) limits its ability to capture most glacial landforms, and its reliance on interpolation between survey tracks, especially in data-sparse regions, precludes its direct use in meaningful DEM terrain analysis techniques. However, the subglacial interfaces imaged directly along the RES flight tracks themselves constitute high resolution bed profiles with the potential to supply critical insights concerning subglacial landscapes beneath modern ice sheets.

In this paper, we discuss a scheme that employs Fast Fourier Transform (FFT) analysis to assess bed roughness, the vertical variation of the subglacial interface with horizontal distance, at the

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multi-kilometre scale along RES-derived bed profiles collected across several regions of the Antarctic Ice Sheet. We review the findings from four previous studies in which we have extracted FFT bed roughness measurements from either the marine-based West Antarctic Ice Sheet (WAIS) or the continentally grounded East Antarctic Ice Sheet (EAIS), and discuss their applications in addressing fundamental issues of ice sheet evolution and stability. All of the studies we review have previously addressed different specific objectives. Here we build upon the previous studies in three ways. Firstly, we assemble for the first time all of our previous Antarctic subglacial roughness measurements in one map. Secondly, we present a unified framework under which bed roughness can be exploited systematically for the interpretation of glaciated landscapes, and discuss the findings from our previous studies through this framework. Finally, we consider the wider implications of bed roughness patterns from Antarctica for studies of Quaternary glaciated landscapes, and provide some suggestions for future research on bed roughness in modern (glacierised) and palaeoglacial (glaciated) settings.

## 2. Methods

### *2.1. RES data collection*

All of the RES data used to investigate bed roughness in this article were obtained during extensive airborne surveys flown across Antarctica between 1971 and 1979 by a consortium of the U.K. Scott Polar Research Institute, the U.S. National Science Foundation and the Technical University of Denmark (hereafter referred to as SPRI-NSF-TUD; see Drewry, 1983, and Siegert, 1999, for further details). Although numerous further RES surveys have been conducted in Antarctica since 1979 (e.g., Holt et al., 2006; Vaughan et al., 2006), the SPRI-NSF-TUD dataset remains the most extensive survey of the continent by area of coverage (Fig. 1). The data, gathered along >400,000 km of flight tracks, span about two-thirds of West Antarctica, principally the areas draining to the Ross and Ronne Ice Shelves, and about a third of East Antarctica, east of Dome A and encompassing Vostok, Dome C, and the coastal regions from Oates Land to Wilkes Land (Fig. 1). Collected with a 250 ns 60 MHz pulsed RES system, bed returns were recorded to digital files at 20 s intervals using a semi-automatic trace reader, and were corrected and tied to navigation records obtained by an aerogeophysical inertial system. The bed is successfully imaged in 86% of cases; and is typically absent only where crevassing scatters the signal or where ice is particularly thick (both because attenuation increases with ice depth, and with thicker ice the basal zone is more likely temperate, increasing signal scattering). The along-track sampling resolution ranges from $\sim$1.8 to 3 km. Errors in absolute position are generally <5 km, but more typically $\pm$1 km, and errors in the sampling interval are at least two orders of magnitude lower. Although these navigational errors are much greater than incurred by modern GPS-constrained surveys, within the context of identifying regional-scale patterns of multi-kilometre bed roughness, as discussed in this paper, the errors can be neglected. The vertical resolution of the data is on the order of $\pm$25 m. In summary, the SPRI-NSF-TUD data stand out as the most regionally extensive in Antarctica, making them ideal for investigating macro-scale bed roughness in a wide-ranging and consistent manner. It is also worth noting that in several regions, especially those particularly remote from the major scientific bases, the SPRI-NSF-TUD surveys still provide to this day the only bed returns available.

### *2.2. Definition and determination of bed roughness*

We define bed roughness qualitatively as the vertical variation of the subglacial interface with horizontal distance along a bed profile. To quantify bed roughness, we employ Fast Fourier

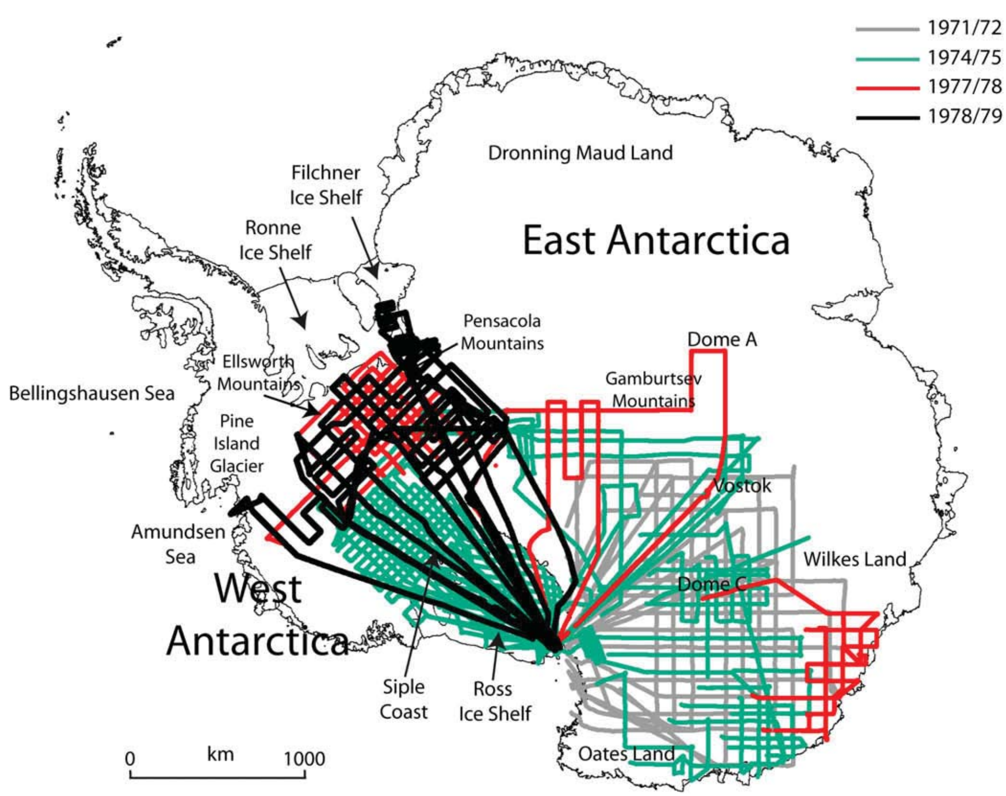


**Fig. 1.** RES flightlines completed by the SPRI-NSF-TUD consortium over the Antarctic Ice Sheet during four seasons in the 1970s. The data comprise the most extensive survey (by area) of the ice sheet and remain the only record of bed information in many parts of Antarctica.

Transform (FFT) analysis following Taylor et al. (2004). The method requires straight sections of 'quasi-continuous' bed returns; by 'quasi-continuous' we recognise that each profile actually consists of a series of datapoints (individual stacks of RES returns) which are distributed along-track at a very high resolution (the 1.8–3 km) compared with the spacing between parallel flight tracks (typically 10s of km; see Fig. 1). Each FFT requires a section of track consisting of $2^N$ datapoints, and a suggested value of $N = 5$ (after Brigham, 1988) gives sections that are 32-datapoints long. Where the sampling resolution is $\sim$1.85 km, as is the case for much of the data over West Antarctica, quasi-continuous bed data are therefore required over a minimum 'window' length of 60 km; similarly, where sample spacing is $\sim$3 km, as is the case over much of East Antarctica, bed data must span at least 100 km. Bed profiles which do not reach these lengths are removed from further consideration. Where data gaps are long (>10 km) bed roughness cannot be determined. However, there are many instances where flight tracks are quasi-continuous but for small gaps of a few datapoints, which, if not interpolated, would preclude FFT analysis over an otherwise quasi-continuous window of the bed. To address this, Taylor et al. (2004) proposed that one can justifiably interpolate the bed linearly over gaps of not more than five consecutive datapoints. Sensitivity testing has shown that the overall effect of interpolating over these small gaps does not influence the results, yet increases the number of bed returns that can be analysed by >50%. To meet the requirement that only straight sections of flight track are analysed, flight tracks are broken into separate components wherever they change direction (usually at 90° turns).

Bed roughness is determined point-by-point along RES bed returns using a discrete forward FFT analysis routine (available in several commercial software packages). Each FFT is applied to a 32-datapoint window which moves progressively along the RES-determined bed. The FFT produces power spectra that are detrended linearly using least-squares regression to remove the dominance of very long wavelength roughness (where 'very long wavelength' equates to at least half of the analysis window length), and the amplitude is normalised with respect to $N$ (after Hubbard et al., 2000). Total bed roughness (plotted at the centre-point of each 32-datapoint analysis window) is defined as the integral of the resulting FFT power spectral power density plot for each analysis window. Essentially, bed roughness values approaching 0 indicate a flat smooth bed, with few vertical undulations. Bed roughness values increasing from 0 represent an increasing number of vertical undulations in the bed. Further details of the method can be consulted in Taylor et al. (2004). We now introduce a framework for the geomorphological interpretation of these roughness variations across glaciated landscapes.

### 2.3. Geomorphological interpretation of bed roughness

Since 2004 we have used FFT analyses to quantify bed roughness from SPRI-NSF-TUD RES data in different regions of Antarctica, in each case addressing distinct regional scientific issues (Siegert et al., 2004a, 2005b; Bingham et al., 2007; Bingham and Siegert, 2007a). In Fig. 2 the bed roughness results from these studies are assembled, for the first time, in a single map. The map highlights clear regional variations in bed roughness across Antarctica, discussed further in the case studies below. The main point to focus on at this stage is that the existence of spatial patterns and regional variations in bed roughness demonstrates that analysing the basal reflector in this way has potential for characterising the subglacial landscape; and there is further potential to transfer this scheme to

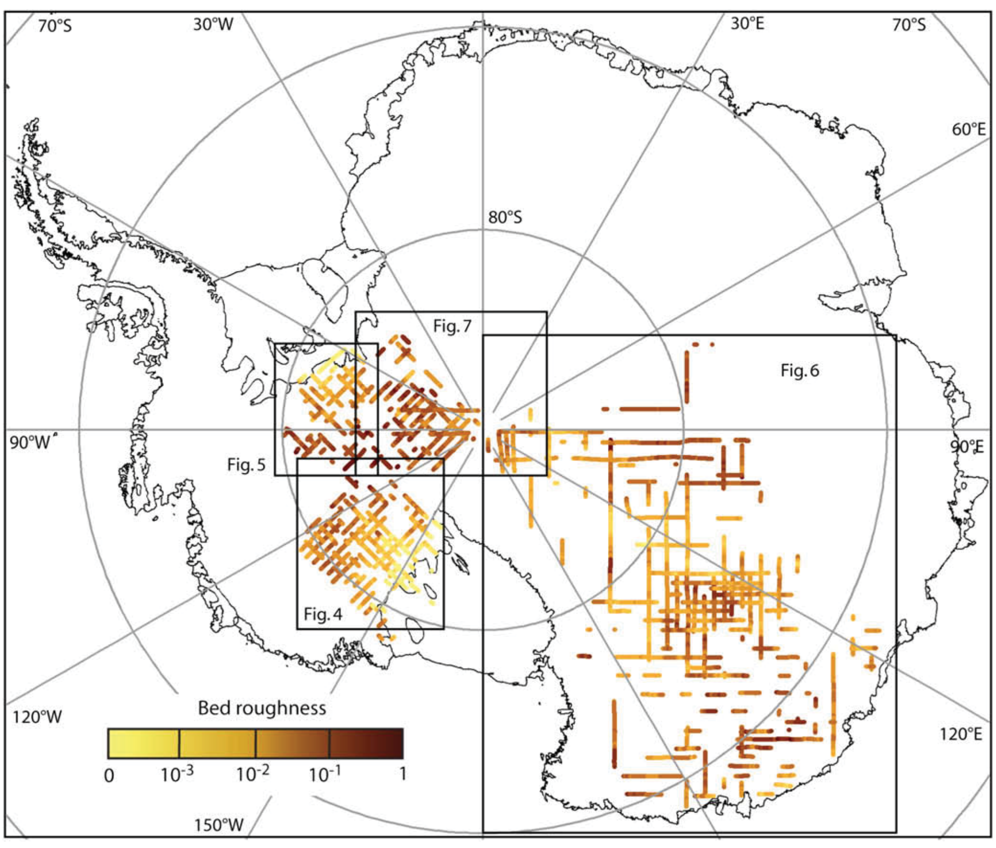


**Fig. 2.** Along-track roughness for the whole Antarctic ice sheet, using roughness data assembled from Siegert et al. (2004a, 2005b), Bingham et al. (2007) and Bingham and Siegert (2007a). Yellow shades denote the smoothest beds, dark brown denote the roughest beds.

Quaternary landscapes, forming the basis for direct quantitative comparisons between the sub-Antarctic and Quaternary glaciated landscapes. For such characterisation to be meaningful, however, it is necessary to consider how regional-scale bed roughness can be interpreted in the context of the geomorphological processes that shape the glaciated landscape.

In Fig. 3 we present schematically the principal factors that can affect bed roughness beneath the Antarctic Ice Sheet. It should be noted before further discussion that many of these factors are interlinked; for example, where an ice sheet overlies a marine setting, such as in the Siple Coast region of West Antarctica, marine sediments deposited during an Interglacial are likely to be conducive to subglacial deformation after glaciation, allowing ice streams (fast flow) to develop (Tulaczyk et al., 1998; Kamb, 2001), and in turn promoting further erosion and deposition (Bingham and Siegert, 2007a). Nevertheless, if bed roughness is to be used to aid the interpretation of glaciated landscapes, it is necessary to consider in turn the different factors that might shape the landscape in different regions. This requires, in each case, support from extra lines of evidence such as knowledge of regional geology, satellite-derived ice velocities, and ice temperatures.

Following Fig. 3, we suggest that the first factor to consider in interpreting bed roughness is whether the landscape being studied can be classified as a 'marine setting,' where much of the subglacial landscape lies below post-rebound sea-levels as exemplified by much of West Antarctica, or a 'continental setting,' where the subglacial landscape lies above sea level as exemplified by much of East Antarctica. The importance of making this distinction lies in the probability that marine settings are typically inundated by marine sedimentation during Interglacial periods. The nature of marine sedimentation, principally the slow settling of particles from an overlying water body, implies that preglacial topography – whether initially rough or smooth – becomes buried beneath a sediment layer whose upper boundary is likely smooth. In a marine setting, then, highly smooth subglacial topography is most likely a first-order indicator of the presence of marine sediments beneath the overlying ice; in a continental setting one cannot invoke sediment draping as such a likely cause, and must consider other factors.

In some regions preglacial topography, in particular orogenic centres where numerous alternations from highs to lows are reflected as rough subglacial topography, may play a prominent role in defining the roughness variations across a glaciated landscape. Beneath the Antarctic Ice Sheet, mountainous subglacial topography is most readily identified where nunataks betray its existence above the surface. There are, however, instances, for example the Gamburtsev Subglacial Mountain range (Drewry, 1983; location in Fig. 1), where our knowledge of subglacial orogenies is entirely based on RES. In such a location as this in central East Antarctica, marine sedimentation cannot have taken place, ice flow rates are low, and basal ice is likely cold (Siegert et al., 2005b). In consequence there has probably been little modification to the preglacial landscape and roughness variations primarily reflect preservation of preglacial topography.

After accounting for the geological setting and preglacial topography of a glaciated landscape, other factors which may induce variations in bed roughness are the basal thermal regime, ice dynamics, and subglacial erosion and deposition rates (Fig. 3). Again, these are all intimately linked: warm basal ice enables faster ice flow which promotes higher erosion and deposition rates, all leading to smoother subglacial topography; while cold basal ice restricts ice flow to slow internal deformation and limits subglacial erosion and deposition (Sugden, 1978; Stroeven et al., 2002), and the lower modification to the landscape that results enables rough topography to endure.

We now turn our attention to reviewing some regional applications of FFT bed roughness across areas of both West and East Antarctica.

## 3. Applications of bed roughness analysis across Antarctica

### 3.1. West Antarctica

The WAIS covers an area of $\sim$1.4 million km$^2$ and much of it is grounded several hundred metres below current sea level. Presently, $\sim$90% of the WAIS is drained by fast-flowing ice streams and glaciers into three major destinations: through the Siple Coast region into the Ross Ice Shelf, between the Ellsworth and Pensacola mountain ranges into the Ronne Ice Shelf, and into the Bellingshausen and Amundsen Seas (Fig. 1). Extensive glaciological studies of these ice streams over the last 30 years have revealed that they can be highly dynamic, some migrating laterally (Echelmeyer and

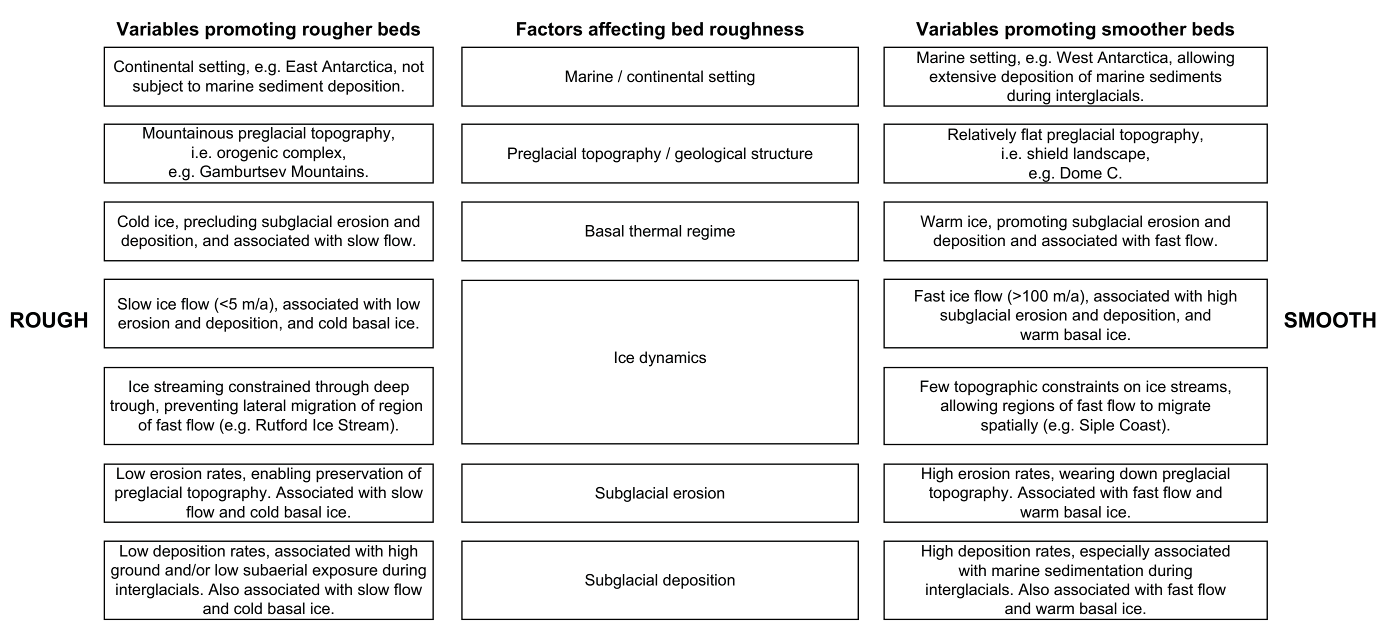


**Fig. 3.** Schematic framework for the geomorphological interpretation of bed roughness values from RES data, discussed in Section 2.3.

Harrison, 1999; Clarke et al., 2000; Raymond et al., 2006), and some oscillating between periods of activity and stagnation (Retzlaff and Bentley, 1993; Catania et al., 2006; Vaughan et al., 2008). It is generally agreed that the primary influence on most of these dynamic variations is provided by conditions at the ice–bed interface, but direct knowledge of the bed of the WAIS is limited to a handful of down-borehole samples collected in the Siple Coast region. There, the sediments sampled have a marine origin, indicating their emplacement when the ice sheet was smaller than today (Scherer et al., 1998). The WAIS therefore exemplifies the marine setting illustrated in Fig. 3, where variations in bed roughness can help to shed light primarily on which regions are underlain by marine sediments.

Aside from the use of bed roughness, discussed below, it should be noted that inferences about the extent and properties of sediments underlying various areas of the WAIS have been made from some other remote sensing techniques. For example, in parts of the Siple Coast (Blankenship et al., 1986; Atre and Bentley, 1993; Anandakrishnan et al., 1998; Peters et al., 2006) and Rutford Ice Stream (Smith, 1997; King et al., 2007; Smith et al., 2007), active seismics have been used to identify areas of basal sediments, to calculate their thickness, and to identify sediments that are dilatant and actively deforming. Another approach has been to identify areas of weak or stiff basal sediments by inverting from surface ice velocities derived from satellite observations (Joughin et al., 2004, 2006). While these techniques are providing powerful insights, both only cover limited areas: seismic data can typically be generated at a maximum of $\sim$5 km per day; while satellite-derived ice-velocity coverages are also limited across Antarctica. Using bed roughness to identify regions of basal sediments complements these techniques because it is able to characterise regions over the continental-scale ($>$500 km$^2$) and to investigate regions not examined locally with seismics or covered by satellite-derived velocities.

The SPRI-NSF-TUD data cover two of the three major regions draining the WAIS: the Siple Coast/Ross Ice Shelf catchment and the Ronne Ice Shelf catchment (Fig. 1). We now discuss FFT-derived bed roughness across both of these regions.

### *3.1.1. Bed roughness beneath the Siple Coast ice streams*

The Siple Coast region is an optimal sector of the WAIS over which to study bed roughness both because it is well covered by RES flight tracks and there is a wide range of ancillary glaciological information about the region available. Over 11,700 km of RES profiles were flown by the SPRI-NSF-TUD consortium over the region (in fact providing the data from which the major ice streams traversing it and feeding the Ross Ice Shelf were first identified; see Rose, 1979). Since then a coordinated series of RES, seismic and borehole sampling programmes has been conducted across the region (e.g., Alley and Bindschadler, 2001). The region also benefits, relative to many other parts of the WAIS, from comprehensive satellite observations of ice flow, used to delineate the spatial extent and velocity of the fast-flowing ice streams, the tributaries that feed them, and the intervening slow-flowing ridges that lie between them (e.g., Joughin et al., 2002). The orientation of the grid pattern flown during the RES surveys is also such that most of the bed reflectors analysed for bed roughness are oriented orthogonal or parallel to the main ice streams.

Bed roughness calculated over the Siple Coast region using the FFT method (after Siegert et al., 2004a) is shown in Fig. 4. The overall range in total bed roughness is large (0.0–0.877; mean 0.046) and the roughness variations are spatially organised in two distinct ways. Firstly, the beds beneath all of the ice streams are smooth, whereas the beds beneath the intervening slow-flowing ridge regions tend to be much rougher. This is exemplified by the contrast between the smooth beds of Bindschadler, Whillans and Mercer Ice Streams and the rough beds below Shabtaie and Conway Ice Ridges. A significant exception is Kamb Ice Stream, where the bed beneath both the ice stream itself *and* the bounding Engelhardt and Raymond Ice Ridges is smooth. This anomaly is discussed below. The second way in which spatial organisation in bed roughness manifests itself is in the clear progression from rough subglacial topography in upstream regions to smooth subglacial topography in downstream regions (Fig. 4). Beneath the area dominated by the current ice streams (including the slow-flowing Kamb Ice Stream) bed roughness is generally low ($<$0.025). Approaching the isostatically rebounded ice-free coastline higher in the catchments, however, bed roughness increases and there is an order of magnitude greater variation in the total roughness values. For three of the catchments (Bindschadler, Kamb and Whillans/Mercer) bed roughness generally increases with distance from the grounding line. This relationship is especially strong for the Kamb and Whillans/Mercer catchments, but less strong over Bindschadler Ice Stream.

Returning to our scheme for interpreting bed roughness as outlined in Fig. 3, we must first consider the likelihood of the region being underlain by marine sediments. Scherer et al. (1998) directly sampled Tertiary marine sediments beneath Whillans Ice Stream; and seismic studies conducted across the Siple Coast ice streams and their interstream ridges suggest that these sediments are widely distributed across the region (e.g., Atre and Bentley, 1993; Anandakrishnan et al., 1998; Peters et al., 2006). Control method inversions for basal shear stresses also indicate that the Siple Coast ice streams lie atop a bed composed of weak subglacial sediments (Joughin et al., 2004). Studinger et al. (2001) isostatically adjusted the subglacial topography across the Siple Coast region to demarcate a paleao-shoreline below which marine sedimentation would have been possible prior to the formation of the WAIS and/or during periods of ice-sheet collapse. Almost all of the SPRI-NSF-TUD datapoints analysed here for bed roughness lie downstream of this palaeo-shoreline, indicating that we need to interpret the roughness variations within the context of a subglacial landscape swathed by marine sediments.

The high degree of bed roughness variations found across the Siple Coast region (Fig. 4) suggests that the marine sediments are distributed nonuniformly. Siegert et al. (2004a) suggested such variations might result from: (i) variations in the thickness of sediment infills during marine inundation; (ii) incomplete burial of a rough preglacial surface leaving exposures of bedrock protruding above the sediments; and/or (iii) processes of erosion, transportation and deposition of sediments leaving some areas of the bed rougher than others. The high degree of correspondence between smooth areas of the bed and the fast-flowing ice streams, the converse coincidence of rougher beds with ice ridges, and the general downstream smoothening of the subglacial interface throughout the region (Fig. 4), all support the third hypothesis. In the context of Fig. 3, therefore, regional variations in bed roughness across the Siple Coast can be interpreted primarily in terms of variations in basal thermal regime, ice dynamics, and subglacial erosion and deposition of marine sediments. Thus, towards the coast and beneath the ice streams, very smooth subglacial reflectors most likely result from fast flowing, warm basal ice flowing over deforming marine sediments; while further inland and beneath the ice ridges rougher subglacial topography probably evinces a landscape less modified by glacial erosion beneath slow-flowing ice.

Kamb Ice Stream appears to represent a special case in the Siple Coast region in that its flow is currently slow, yet it is underlain by smooth subglacial topography, elsewhere indicative of fast flow/warm-based ice; and the subglacial topography beneath the flanking Engelhardt and Raymond Ice Ridges is also smooth, contrasting with the rough beds typically found beneath the other ice

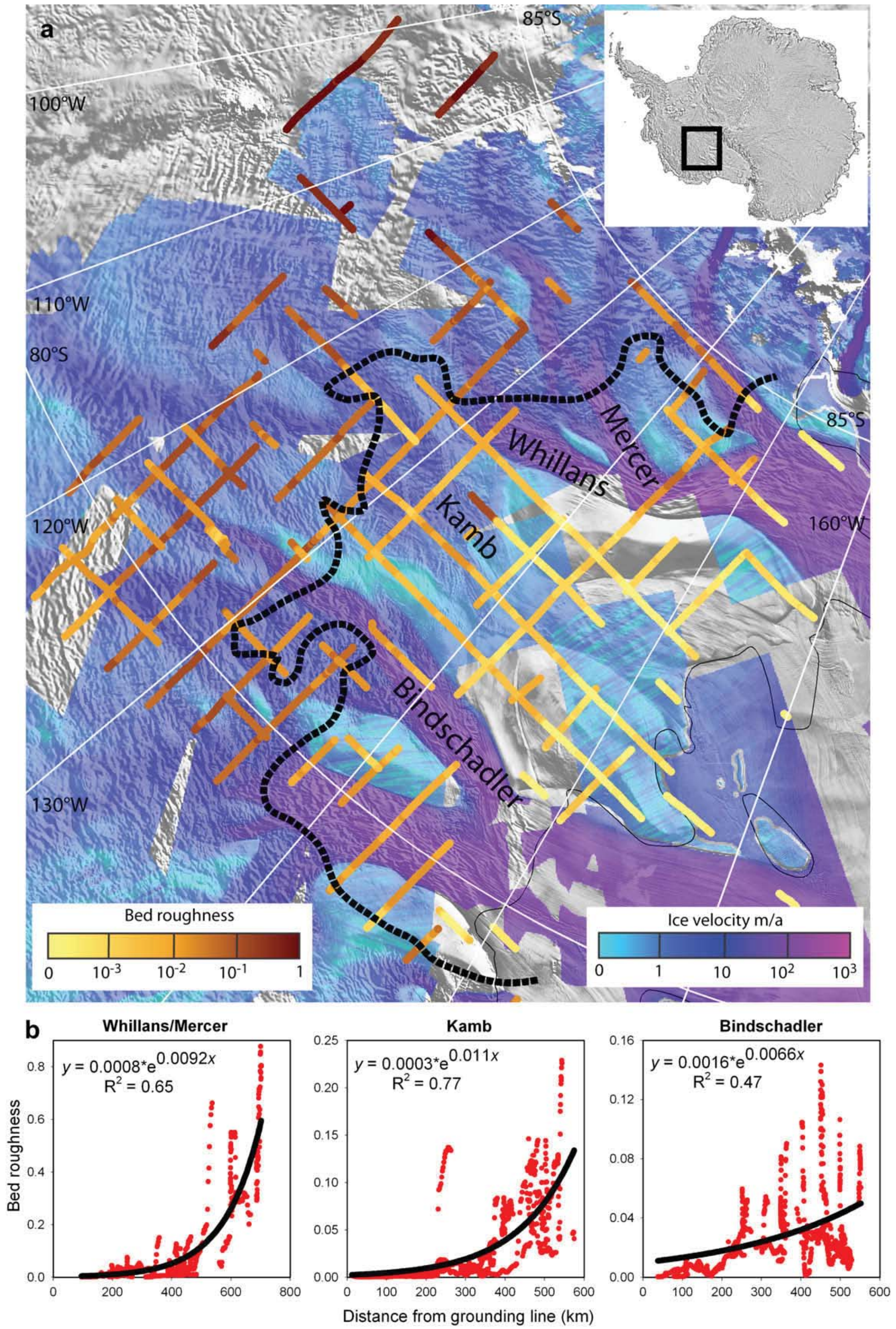


**Fig. 4.** (a) Bed roughness measurements across the Siple Coast region of West Antarctica, after Siegert et al. (2004a). Inset shows location; see also Fig. 2. In the background, ice velocities after Joughin and Tulaczyk (2002) are superimposed over MODIS visible imagery (MOA; Haran et al., 2006). Ice streams discussed in the text are labelled. The thin black line marks the approximate position of the grounding line. The dashed line, after Siegert et al. (2004a) separates regions of 'rough' and 'smooth' topography as defined in Section 3.1.3. (b) Plots of bed roughness versus distance from the grounding line for Whillans/Mercer, Kamb and Bindschadler Ice Stream catchments, after Bingham and Siegert (2007a).

ridges. The anomaly can be explained by considering the overall flow history of the ice stream and the lack of topographic constraints through which it flows. Although Kamb Ice Stream is currently 'stagnant,' flowlines on the Ross Ice Shelf, and folding and deformation of internal layers in the ice stream itself, have been used to show that previously it has been subject to fast flow like the other Siple Coast ice streams (Jacobel et al., 1993; Anandakrishnan et al., 2001; Ng and Conway, 2004; Catania et al., 2006; Hulbe and Fahnestock, 2007). Buried crevasses have been interpreted as evidence that the last period of fast flow may have ended as recently as 150 years ago (Retzlaff and Bentley, 1993). The smooth subglacial interface beneath Kamb Ice Stream is therefore likely to reflect sustained erosion/deposition of subglacial sediments during 'active' periods of fast flow in the ice stream's history; indeed, the smooth bed can be treated as additional evidence supporting the hypothesis that Kamb Ice Stream has not always been stagnant. There is also evidence that Kamb Ice Stream, because it does not flow through a topographically constrained deep trough, may have migrated laterally during its history (Catania et al., 2005). Such lateral migration promotes widespread smoothening of the subglacial landscape, as wider areas are subject to erosion/deposition under fast-flowing warm-based ice (Fig. 3), and may explain the unusually smooth beds extending beneath the ice ridges bounding Kamb Ice Stream today (Siegert et al., 2004a).

The Siple Coast findings demonstrate that bed roughness patterns in this part of the WAIS correspond with, and therefore reflect, variations in the distribution of marine sediments inferred from other forms of evidence. They suggest in particular that where bed roughness is low, there is a strong likelihood that marine sediments are present at the base, and have probably been modified (smoothened) further by erosion, transportation and deposition in a fast-flowing, warm-based subglacial environment. This suggests that bed roughness can be used elsewhere to improve our knowledge of the subglacial environment beneath the WAIS. Such an application might be of especial value where other forms of evidence (e.g., seismics, satellite observations) are lacking, as in the following case study.

### *3.1.2. Bed roughness beneath the Ronne ice streams*

Institute Ice Stream (IIS; 81.5°S, 75°W; Fig. 5) drains an area of 140,000 $km^2$ to the Ronne Ice Shelf at a rate of $22.7 \pm 2$ Gt $a^{-1}$ (Scambos et al., 2004; Joughin et al., 2006). The adjacent Möller

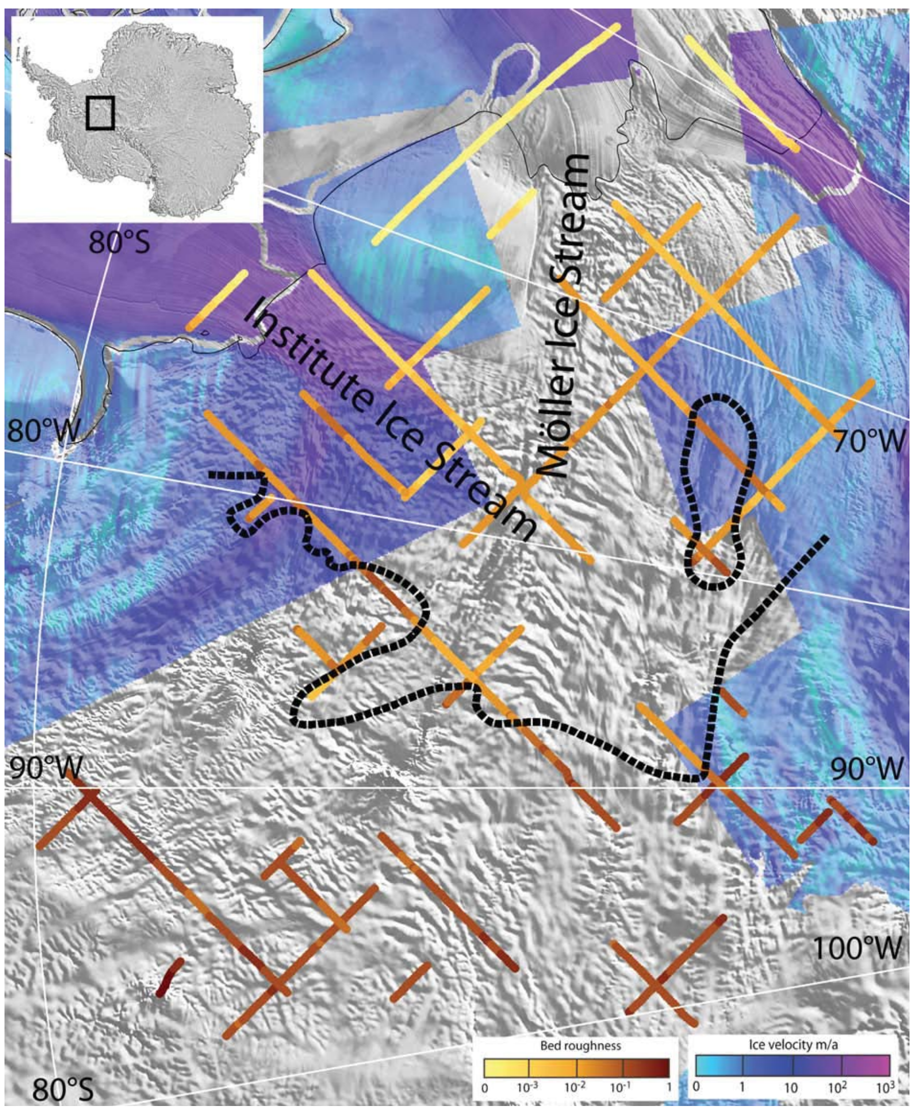


**Fig. 5.** Bed roughness measurements across Institute and Möller Ice Streams, after Bingham and Siegert (2007a). Inset shows location; see also Fig. 2. In the background, ice velocities after Joughin et al. (2006), available only in selected areas, are superimposed over MODIS visible imagery (MOA; Haran et al., 2006). The thin black line marks the approximate position of the grounding line. The dashed line separates regions of 'rough' and 'smooth' topography as defined in Section 3.1.3.

Ice Stream (MIS; 82.3°S, 65°W) drains an area of 66,000 km$^2$ but its ice flux across the grounding line is unknown. Together, these catchments cover ~10% of the WAIS and account for almost all of its ice flow to the Ronne Ice Shelf, yet their dynamic and basal characteristics rank amongst the most poorly known in Antarctica. This is because, in stark contrast to the Siple Coast region, the coverage of satellite-derived ice velocities is more limited, and few field studies have been undertaken across the area. However, 7000 km worth of RES profiles were collected across the IIS/MIS region during the SPRI-NSF-TUD surveys, and bed roughness has been extracted from the bed reflections (Fig. 5; see also Bingham and Siegert, 2007a).

Across IIS/MIS total bed roughness ranges between 0.0002 and 0.5040, averaging 0.0570; comparable to values measured across the Siple Coast region (see Section 3.1.1). The roughness variations exhibit clear spatial organisation (Fig. 5). Beneath the main trunk of IIS and the downstream portion of MIS the basal reflector is remarkably smooth. Away from the regions of active streaming, and as distance towards the ice divides decreases, the basal reflector becomes noticeably rougher. These observations are true regardless of whether the roughness windows analysed are quasi-parallel or quasi-orthogonal to ice flow (as depicted by satellite-derived surface ice velocities), and are qualitatively similar to the variations in bed roughness derived across the Siple Coast region above. Bingham and Siegert (2007a) further plotted bed roughness versus balance-flux modelled ice velocities over IIS/MIS, and found that very low bed roughness values are measured where ice speeds are high, while high bed roughness is only measured where ice speeds are low. Typically, away from the grounding line, where ice flows more slowly, bed roughness is high; whereas towards the grounding line, where ice velocities are more variable but can be very high, bed roughness is low. These patterns match those observed over the Siple Coast region (Siegert et al., 2004a).

That the subglacial properties beneath IIS and MIS are analogous to those found across the Siple Coast region implies that a similar geomorphological interpretation of the bed roughness patterns following Fig. 3 can be followed. With much of its subglacial topography well below post-rebound sea levels, and evidence for fast flow in combination with low driving stresses in the downstream sector where satellite ice velocities are available (Scambos et al., 2004; Joughin et al., 2006), the most likely explanation for the very smooth subglacial interface found beneath IIS is that the ice stream is underlain by marine sediments whose upper surface has been smoothened by processes of erosion, deposition and deformation associated with fast-flowing, warm-based ice flow above. There is no increase in bed roughness moving eastwards from the lower 200 km of IIS across the slow-flowing Bungenstock Ice Rise and into the downstream 200 km of MIS (Fig. 5), and this may reflect a thick basal sediment layer pervading most of the IIS/MIS region fringing the Ronne Ice Shelf (Bingham and Siegert, 2007a). As discussed for the case of the Kamb Ice Stream earlier, the presence of a thick covering of sediments over the subglacial landscape has implications for ice dynamics. Firstly, it provides a thick layer of material which may deform readily when wet, allowing fast flow to persist in spite of low driving stresses (e.g., Scambos et al., 2004) but stiffens when frozen, contributing to periods of ice stream stagnation (e.g., Anandakrishnan et al., 2001). Secondly, a large covering of thick sedimentation may 'iron out' subglacial protuberances, leaving few constraints to lateral migrations of the ice streams. That much of the subglacial topography across IIS/MIS, including beneath the slow-flowing Bungenstock Ice Rise, is smooth, perhaps evincing ubiquitous marine sedimentation over the region, therefore suggests that these ice streams feeding the Ronne Ice Shelf may be vulnerable to the types of dynamical instabilities observed over the Siple Coast region.

#### 3.1.3. WAIS extent during the Last Interglacial from bed roughness

The bed roughness variations determined across the WAIS here can be used to make a first-order estimate of the WAIS extent and its contribution to sea level during the Last Interglacial period (Eemian; 125,000 B.P.). Scherer et al. (1998) have shown that sediments sampled from the base of a borehole in Whillans Ice Stream have a marine origin and were likely to have been deposited during the Eemian period. We have shown above that across the Siple Coast region areas of smooth subglacial topography correspond with regions thought from other forms of evidence to be underlain by these marine sediments. Areas of smooth subglacial topography may therefore be considered indicators of areas of marine sedimentation beneath the WAIS and, hence, areas that were ice-free to allow marine sedimentation during the Last Interglacial. A difficulty, not easily answered definitively, is in deciding on a value of bed roughness indicative of the boundary between areas of sedimentation ('smooth') and areas not ubiquitously covered by sediments ('rough'), and so any value derived using this method can only be treated as an approximation. Here we use the median value of bed roughness over both the Siple Coast and IIS/MIS regions analysed above, 0.023, to draw a loose boundary between smooth 'sediment-covered' regions and rougher 'exposed' regions beneath the WAIS (drawn onto Figs. 4 and 5). Beneath IIS/MIS, the downstream region over which bed roughness is <0.023 covers ~89,000 km$^2$, 43% of the IIS/MIS catchments; beneath the four Siple Coast catchments analysed in Section 3.1.1 similarly-defined smooth beds underlie 218,000 km$^2$, 31% of their combined area.

We now follow Bahr et al. (1997) in using volume-area scaling, such that:

$$V_E = xA_E^{1.25}$$
$$x = V_M/A_M^{1.25}.$$

where 1.25 is the appropriate exponent for ice sheets (Bahr et al., 1997); $x$ is a scaling factor; $V_M$ and $A_M$ are the volume and area of the modern WAIS, respectively $3.6 \times 10^6$ km$^3$ (Lythe et al., 2001) and $1.9 \times 10^6$ km$^2$ (Bindschadler, 2006); and $V_E$ and $A_E$ are the volume and area of the Eemian WAIS. Here, $A_E$ is taken as $A_M$ minus the 89,000 km$^2$ of IIS/MIS and the 218,000 km$^2$ of the Siple catchments underlain by smooth subglacial topography, giving $A_E = 1.6 \times 10^6$ km$^2$ and $V_E = 2.9 \times 10^6$ km$^3$. Given that the modern WAIS may contribute 6 m to sea level (Bindschadler, 2006), a loss of $V_M - V_E = 0.5 \times 10^6$ km$^3$ accounts for a 1.2 m sea-level contribution during the Eemian Interglacial. This estimate of 1.2 m must be considered appropriately in the context that it is dependent on the value of bed roughness chosen to identify marine sedimentation, and that it is likely a minimum estimate taking into account the potential discovery of further regions of underlying marine sediments in areas of the WAIS away from IIS/MIS and the Siple Coast covered in this analysis.

### 3.2. East Antarctica

The EAIS covers an area of 10.35 million km$^2$ and has a volume of approximately 26 million km$^2$. Much of the ice-sheet bed is above sea level and, if the ice sheet decayed, postglacial uplift would result in sub-aerial exposure of the bulk of the continent. Hence, much of East Antarctica, though heavily glaciated, is a continental land mass, with significant regions of highlands and mountains, areas of deep lowlands and valleys, and relatively flat expanses of continental shield (Lythe et al., 2001). Referring to Fig. 3, variations in roughness beneath much of the EAIS cannot therefore be attributed, as beneath the WAIS, primarily to differences in marine sedimentation; instead, the preglacial topography, basal thermal regime and ice dynamics probably play the more prominent roles in imprinting variations in roughness over the subglacial landscape.

*3.2.1. Bed roughness and ice sheet evolution in East Antarctica*

Siegert et al. (2005b) undertook the first systematic analysis of bed roughness in East Antarctica, utilising >200,000 km of RES tracks collected by the SPRI-NSF-TUD consortium over the EAIS extending between Dome A and Wilkes Land (Fig. 6). The results yield distinct geographical areas that can be identified on the basis of their roughness characteristics: Dome A, Ridge B and the Porpoise Subglacial Highlands are all underlain by rough subglacial topography; while both the Wilkes and Aurora Subglacial Basins lie over smoother terrain (Fig. 6). An initial influence over these broad patterns might be provided directly by the preglacial topography: the areas of rougher topography overlie higher ground, in the case of the Porpoise region directly referred to as "Subglacial Highlands," in the case of Dome A at least partly overlying the Gamburtsev Subglacial Mountain Range (Bell et al., 2006); and if the higher ground can indeed be characterised as 'mountainous' then the numerous highs and lows constituting 'mountainous' terrain would yield high bed roughness values. Conversely, the terrain beneath the smoother-bedded regions is generally lower and occurs across two large subglacial 'basins.'

While preglacial topography may therefore be considered a first-order influence on bed roughness, it is necessary to consider the rates at which the preglacial topography has been modified since glaciation; partly a function of basal thermal regime which defines whether basal ice is cold-based (precluding erosion) or warm-based (allowing erosion). Siegert et al. (2005b) tackled this issue by applying an ice-sheet model to investigate likely thermal conditions at the base of the present-day ice sheet. Beneath Ridge B basal temperatures were calculated to be approximately $-10\,^{\circ}C$ while at Dome A they were as low as $-20\,^{\circ}C$ (Siegert et al., 2005b). Reconstructions of the EAIS (e.g., Huybrechts, 2002) show that ice thickness in central East Antarctica changes by only a few hundred metres over Glacial–Interglacial timescales; thus it is likely that the ice divides at Ridge B and Dome A have been cold-based, preserving the preglacial rough topography, for as long as the ice sheet has been at its present continental scale. West from the Porpoise Subglacial Highlands across Wilkes Land basal ice is also cold (around $-15\,^{\circ}C$), again promoting subglacial highland preservation. However, the subglacial lowlands across the Wilkes and Aurora Basins are characterised by warm modelled subglacial

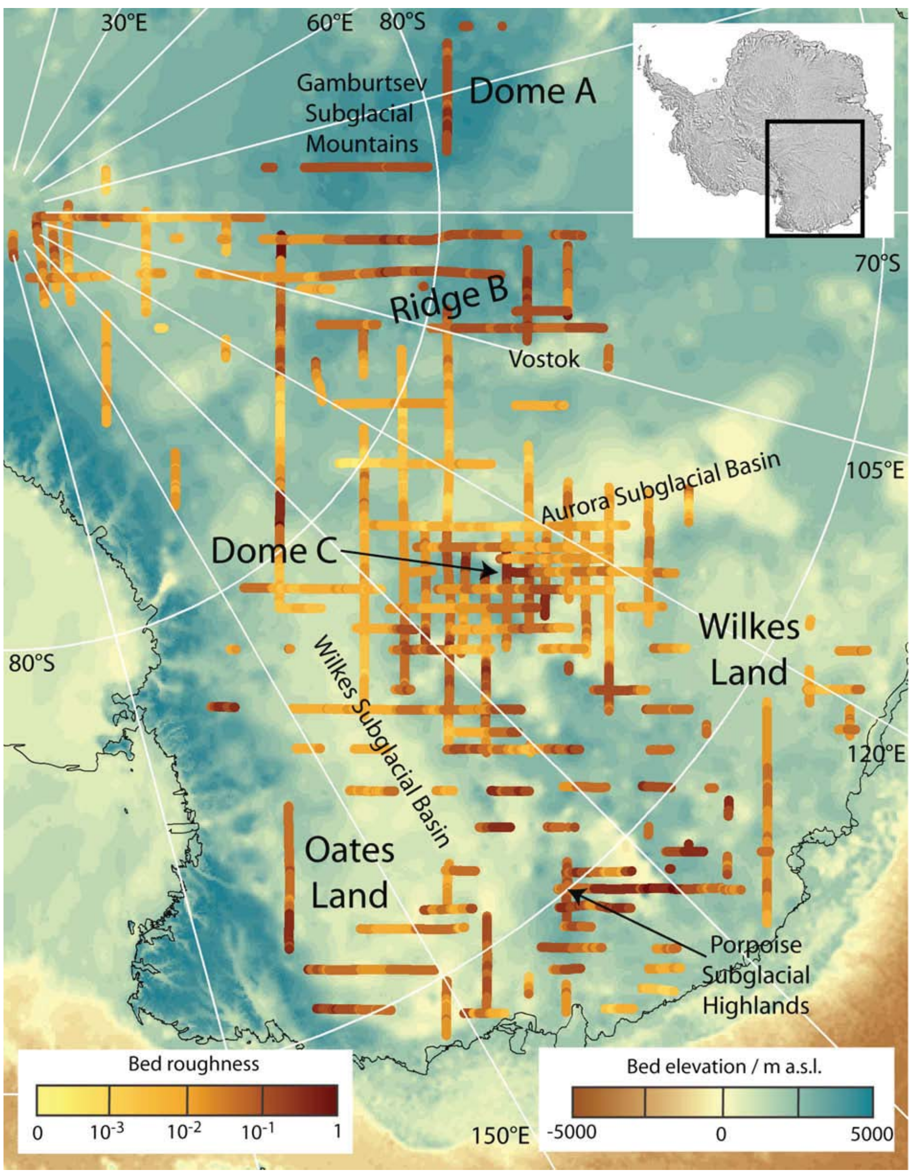


**Fig. 6.** Bed roughness measurements from SPRI-NSF-TUD surveys in East Antarctica, after Siegert et al. (2005b), superimposed over subglacial bed elevations (BEDMAP; Lythe et al., 2001). Inset shows location; see also Fig. 2.

temperatures (Siegert et al., 2005b), suggesting that basal ice is at the pressure melting point – this is also consistent with the identification of numerous subglacial lakes in both basins (Siegert et al., 2005a; Tabacco et al., 2006). Warm basal ice allows subglacial erosion, or deformation of subglacial sediments if present, and may allow erosional smoothening of the subglacial landscape (Fig. 3).

The bed roughness imprint beneath the Dome C region is more difficult to interpret in the context of the present thermal and dynamic ice regime. Over the central, highest part of Dome C, the bed is rough; but in the lowlands flanking it bed roughness is much lower than is found near the ice divides at Dome A and Ridge B (Fig. 6). Modelling suggests that basal ice is warm beneath Dome C (Siegert et al., 2005b), and the region is well populated by subglacial lakes (Siegert et al., 2005a; Tabacco et al., 2006); hence one might attribute the relatively smooth bed beneath Dome C to erosion allowed under the present-day warm basal conditions. However, the same model also calculates a subglacial slip coefficient, calculated as the difference between balance velocities and velocity due to the deformation of ice, divided by the gravitational driving stress; and across the Dome C region the subglacial slip coefficient is negligible (Siegert et al., 2005b; see also Rémy and Tabacco, 2000). Thus, taking into account *both* the modelled basal thermal *and* ice dynamic regime, and considering the factors outlined in Fig. 3, the *present* ice configuration across Dome C is unlikely to be conducive to basal erosion. An explanation proposed for the unusually smooth bed found across much of the region is that the subglacial morphology beneath Dome C likely predates the present ice configuration. In experiments attempting to replicate the growth and decay of the EAIS, ice-sheet modelling suggests that the main regions where ice decay is likely to happen first are the Wilkes and Aurora subglacial basins (Siegert et al., 2005b). Such decay would leave the Dome C region isolated, resulting in radial flow centred on the higher ground. This ice configuration is consistent with the interpretation of several large-scale valleys extending outwards from Dome C as overdeepened troughs (Siegert et al., 2005b). The climate required for such an ice configuration is broadly similar to that on Svalbard today, placing the timing of such an ice sheet before the Quaternary and probably before the Pliocene (Siegert et al., 2005b).

### *3.2.2. Bed roughness and ice flow from the South Pole to Filchner–Ronne Ice Shelf*

Like the IIS/MIS region discussed in Section 3.1.2, the region extending between the South Pole through the Support Force and Foundation catchments to the Filchner–Ronne Ice Shelf (Fig. 7) is poorly known in terms of its dynamic and basal characteristics. A particular problem here is that because most Earth observation satellites do not orbit at the poles surface altimeter data and satellite-derived ice velocities are either absent or not resolved to the level of other locations in Antarctica (Shepherd and Wingham, 2007; Rignot et al., 2008). Therefore RES data represent some of the only survey information available across much of the region, and roughness is an important tool for elucidating basal conditions and the likely ice dynamic history. The region differs from IIS/MIS in that it extends 100s of km inland into the EAIS and therefore mostly constitutes a 'continental,' rather than 'marine' setting (cf. Fig. 3).

Across the Support Force Ice Stream (SFIS) and Foundation Ice Stream (FIS) catchments there exist 11,500 km of RES data collected during the SPRI-NSF-TUD surveys. Bed roughness extracted from the subglacial interface along the flight tracks spanning SFIS/FIS is shown in Fig. 7. Bed roughness ranges between 0 and 0.902, averaging 0.102 (Bingham et al., 2007). In the lower parts of the

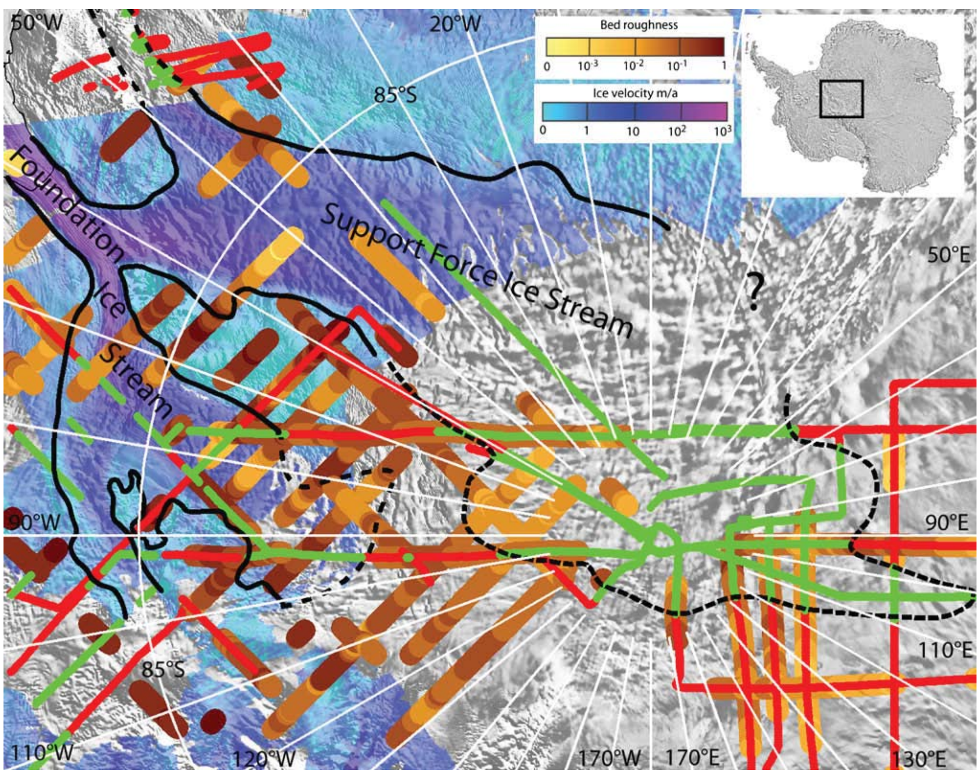


**Fig. 7.** Bed roughness and classification of internal layering from RES data between the South Pole and Filchner–Ronne Ice Shelf, after Bingham et al. (2007). Inset shows location; see also Fig. 2. Thin red lines denote RES tracks with undisrupted/continuous layering; thin green lines denote tracks with disrupted/buckled layering. In the background, ice velocities after Joughin et al. (2006), available only south of 85°S, are superimposed over MODIS visible imagery (MOA; Haran et al., 2006). The solid black line follows the 5 m a$^{-1}$ surface ice-velocity contour where InSAR data are available, which can be treated as the boundary between slow (interior) flow and enhanced (tributary) flow. The dashed black line marks the inland tributary extent inferred from boundaries between 'rough' and 'smooth' bed topography and disrupted/undisrupted internal layering, as described in Section 3.2.2. The question mark denotes a region with neither RES data nor satellite velocity coverage where the boundary cannot be drawn.

catchments, where satellite-derived ice velocities are available, bed roughness varies with the ice flow configuration, such that smoother beds are found in the regions of active streaming while rougher beds are found where ice flow is slower (Fig. 7). Bingham et al. (2007) further compared the bed roughness measurements across SFIS/FIS with the characteristics of internal layers imaged in the same RES profiles. Internal layers were classified as disrupted/buckled or undisrupted/continuous (following e.g., Rippin et al., 2003; Siegert et al., 2003), where disrupted/buckled layering is interpreted as diagnostic of current or previous fast or enhanced flow (e.g., Jacobel et al., 1996; Ng and Conway, 2004; Siegert et al., 2004b). Across SFIS/FIS disrupted layers are strongly associated with smooth beds, while undisrupted layers correspond with rougher subglacial topography (Bingham et al., 2007). The findings are significant for two reasons. Firstly, they suggest that the ice dynamic regime is a first-order influence on bed roughness variations between the South Pole and Filchner–Ronne Ice Shelf (cf., Fig. 3). Secondly, and following on from this, they suggest that the variations in bed roughness and internal layers observed in the RES data can be used to identify regions of fast or enhanced flow where satellite-derived ice velocities are not available.

In Fig. 7 we show the possible inland extent of enhanced flow around the South Pole region drawn from an analysis of bed roughness and internal layering across SFIS/FIS. The boundary between slow (interior) flow and enhanced (tributary) flow is inferred from the RES data by connecting transitions between disrupted and undisrupted internal layering and smooth and rough subglacial topography. The distinction between smooth and rough subglacial topography is made at a bed roughness value of 0.20, the average bed roughness value at all transitions between disrupted and undisrupted layering across SFIS/FIS. The head of the inferred ice drainage feature feeding SFIS covers much of the region surrounding the South Pole, where the existence of several mini ice domes (Fig. 7) evinces probable ice drawdown from the region. The existence of an enhanced flow tributary connecting the South Pole with Filchner–Ronne Ice Shelf is also supported by balance-flux modelling (Bamber et al., 2000; Le Brocq et al., 2006), but its area of inception, inferred from bed roughness and internal layering, covers a much wider area than the modelling predicts (Fig. 7).

## 4. Wider implications for Quaternary glaciological research

This paper has demonstrated, through a series of examples from Antarctica, that quantifying bed roughness over subglacial terrain can provide valuable insights on the dynamics and history of contemporary ice sheets. We have collated here, for the first time in one map, all of our bed roughness calculations from the entire SPRI-NSF-TUD RES dataset, collected between 1971 and 1979 (Fig. 2). We have concentrated on extracting bed roughness from this dataset to date because by using a single data type in all studies we could maintain methodological consistency; the dataset remains the most spatially extensive (by area) across Antarctica; and the surveys extend into a number of remote and little understood regions which have not been revisited since and which, in some cases (e.g., IIS, SFIS), have few satellite observations, so that bed roughness information from the SPRI-NSF-TUD dataset supplies a significant proportion of our regional knowledge. Yet there is potential to develop studies of bed roughness further, by: (i) applying the technique to other RES datasets collected across Antarctica, Greenland and other terrestrial ice caps; (ii) testing the utility of the FFT technique at different scales of RES data, i.e., with different along-track sampling resolutions; (iii) evaluating FFT determinations of bed roughness against other possible methods for calculating bed roughness; and (iv) applying the technique to formerly glacierised terrain such as offshore landscapes imaged by swath bathymetry and seismics, and the Quaternary glaciated landscapes of northern North America and Eurasia. In this section we make some recommendations for future bed roughness research based on these points.

Since the SPRI-NSF-TUD surveys were discontinued in 1979 further RES surveys have been conducted across many parts of the Antarctic Ice Sheet (Lythe et al., 2001; Bingham and Siegert, 2007b). Significant improvements in technology over 40 years have allowed digital recording of data at an ever higher sampling resolution: for example, an extensive survey of Pine Island Glacier in 2004–2005 collected bed echoes along ∼35,000 km of flight tracks with an along-track sampling resolution of ∼20 m (Vaughan et al., 2006; cf. 1.8–3 km in the SPRI-NSF-TUD dataset). Bed roughness measurements conducted on these newer datasets have the potential both to widen our regional understanding of the basal properties of the ice sheet and to improve our physical understanding of the bed roughness parameter itself. In applying FFT bed roughness analysis to newer RES datasets, a clear first stage would be to run the analysis over datasets subsampled to the same along-track resolution as that used in the earlier studies outlined here (i.e., between 1.8 and 3 km). Such an exercise, performed incrementally on RES datasets collected over different parts of the Antarctic Ice Sheet, would widen the consistent bed roughness coverage shown in Fig. 2, and provide a bed roughness dataset that could be interpreted region-by-region following the considerations outlined in Fig. 3. An obvious benefit to this would simply be expanding our knowledge of bed roughness to regions not covered by the SPRI-NSF-TUD data (e.g., Pine Island Glacier in West Antarctica, Dronning Maud Land); however, it would also be valuable to test whether bed roughness extracted from SPRI-NSF-TUD data is consistent with bed roughness extracted across the same region from a newer dataset following different survey tracks (e.g., the Siple Coast region, revisited with RES by e.g., Blankenship et al., 2001; Peters et al., 2005).

The order-of magnitude greater sampling resolution achieved by recent datasets also offers an opportunity to calculate FFT-derived bed roughness over much smaller sampling windows than has been achieved to date. In all of the studies reviewed here, each bed roughness datapoint reflects undulations of the bed along a track length of at least 57 km; now that there are RES datasets with sampling resolution as low as 20 m (Corr et al., 2007), bed roughness can be determined from windows as short as 6.4 km in length. In the studies discussed here, our interpretation of the results has been restricted to regional-scale physical considerations and processes that occur over scales of 60–100 km, an example being that smooth terrain extending along 100s of km in West Antarctica is a strong indicator of widespread marine sedimentation. The ability to extract bed roughness over smaller scales allows roughness to be used to investigate basal characteristics and basal physical processes over correspondingly smaller scales, approaching those of individual landform development. However, RES data required theoretically to extract bed roughness at the scales affecting ice flow through regelation and enhanced basal creep (20 cm to 1 m; Weertman, 1957; Hubbard and Hubbard, 1998) are not yet available.

So far in this discussion we have considered only the contemporary terrestrial ice sheet covering Antarctica as a candidate for further bed roughness studies. However, the Earth's only other continental ice sheet, the Greenland Ice Sheet, has also been extensively surveyed with RES (e.g., Gogineni et al., 2001). To date there has been no comprehensive bed roughness analysis conducted beneath the Greenland Ice Sheet. The same point is also true of many of the >10,000 km$^2$ ice caps distributed across Arctic Canada, Svalbard, Franz Josef Land and Severnaya Zemlya: RES data exist (Dowdeswell et al., 1984, 1999, 2002, 2004) but have not been queried for bed roughness characteristics. Determination of bed roughness from RES across these ice masses can add incrementally to our knowledge of landscapes glacierised at different scales and with varying dynamic and thermal histories.

One of our principal motivations in investigating bed roughness in contemporary continental-scale ice sheets has been to develop a scheme that can also be applied to the continental-scale glaciated landscapes left across northern North America and Eurasia by the Quaternary Laurentide and Eurasian ice sheets, and which can therefore be used to compare and contrast glacierised and glaciated terrains. Indeed, FFT methods have previously been applied to glaciated terrains to understand the interaction between roughness, glacial dynamics and the precipitation of calcite at the mm-scale (Hubbard and Hubbard, 1998; Hubbard et al., 2000). It was these novel methods that Taylor et al. (2004) upscaled and adapted for studying bed roughness at the continental scale beneath contemporary ice sheets – but an examination of bed roughness at the continental scale across a Quaternary glaciated landscape has not yet been conducted.

Studying bed roughness over the glaciated landscapes of North America and/or Eurasia would be of benefit both in testing methodologies for the determination of bed roughness, and in improving our knowledge of the physical manifestations of different roughness signatures. The fundamental advantage of a former, rather than current, ice-sheet bed, is the ability to 'see' the entire bed – the glaciated landscapes of northern North America and Eurasia have been imaged from space, both allowing widespread visualisation and classification of landforms associated with particular ice flow regimes (e.g., Stokes and Clark, 2001; Napieralski et al., 2007) and the construction of high-fidelity digital elevation models (DEMs; e.g., Siegert and Dowdeswell, 2004; Tarasov and Peltier, 2006). Beneath Antarctica, the bed is only captured along RES tracks, thus analyses of bed roughness are always limited by the preimposed orientation and sampling resolution of those tracks; however, no such analytical limitations are imposed on formerly glacierised terrains. Experiments are required over these 'glaciated landscape DEMs' to test: (i) whether tracks (2D profiles) distributed across the DEM and queried for bed roughness in the same manner as that done across Antarctica produce similar ranges of bed roughness values; (ii) the effects on bed roughness of varying the orientation of tracks with respect to former ice flow directions; (iii) the effects on bed roughness values of changing the along-track sampling resolution; and (iv) the correspondence of regional bed roughness signatures with different glacial landsystems. Having a fully 3D landscape on which to conduct methodological tests also affords the opportunity to compare and contrast FFT-derived bed roughness with a wide range of other 2D and 3D terrain analysis techniques. As well as the Quaternary landscapes discussed above, glaciated terrains on the continental shelves of Antarctica, Greenland and northern Europe are increasingly being imaged using swath bathymetry and offshore seismics (Lowe and Anderson, 2003; Ó Cofaigh et al., 2004, 2005; Evans et al., 2006; Dowdeswell et al., 2007). These ice-free terrains provide further opportunities for analysing bed roughness patterns over glaciated landscapes.

The main values of the bed roughness research outlined here are in being able to draw inferences about subglacial landscape variations across continental-scale ($>$500 km$^2$) regions; in being able to extrapolate across multi-kilometre distances over contemporary ice sheets site-specific basal information from, for example, seismic sounding and borehole drilling; and in developing a scheme that could transfer easily between continental-scale investigations of contemporary and former ice-sheet beds. At more local scales, recent strides have been made in imaging subglacial landforms at sub-km resolution beneath the Antarctic Ice Sheet using seismic and oversnow radar methods, providing some of the first images of modern subglacial landforms analogous to those observed across Quaternary glaciated landscapes (King et al., 2007; Smith et al., 2007). The bed roughness methods outlined here operate at a scale too large to capture these individual landforms, although it is possible that regional collections of similar landforms indicative of particular ice flow regimes (e.g., Stokes and Clark, 2001) may impart regional roughness signatures. Continental-scale analyses of bed roughness therefore have significant potential to complement local-scale imaging of subglacial/glaciated landforms.

## 5. Conclusions

Bed roughness values derived from ice-penetrating RES basal reflectors represent a rapid, non-invasive method for extrapolating first-order basal characteristics across large areas of modern ice sheets. The information, while simple to generate, is of high value as a means to compare regions, assess modern basal processes, and infer past changes. The method, first defined by Taylor et al. (2004), and employing a Fast Fourier Transform (FFT) methodology, has been used to calculate bed roughness across four regions of the Antarctic Ice Sheet. We have introduced a schematic framework outlining the principal factors that can affect roughness beneath the Antarctic Ice Sheet. The main influences include: (i) whether the ice sheet lies primarily in a marine setting (e.g., much of West Antarctica) and may be underlain by marine sediments; or a continental setting (e.g., most of East Antarctica) where marine sedimentation can be discounted; (ii) the nature of the preglacial topography and subglacial lithological structure; (iii) ice dynamics, both present and former, and whether the region is topographically constrained; (iv) basal thermal regime, both present and former; (v) and processes of subglacial erosion and deposition. Many of the above processes are interlinked.

The Antarctic roughness studies reviewed in this paper demonstrate that characterising bed roughness using FFT can draw out spatial patterns which can be used, in support of additional sources of information, to extrapolate local information on basal characteristics from, for example, seismic sounding and borehole drilling. In principle, the technique is readily transferable to characterising glaciated terrain that is now free of ice, such as now-offshore deglaciated regions imaged by swath bathymetry, and the Quaternary glaciated landscapes of northern North America and Eurasia. Therefore, bed roughness has strong potential to form part of a quantitative scheme by which subglacial landscapes beneath contemporary ice sheets can be used as analogues for Quaternary glaciated landscapes; and vice versa. Future investigations need to focus on the effects of varying the sampling resolution and orientation of RES tracks, and comparing different methods for calculating along-track bed roughness. New digital RES datasets from Antarctica, Greenland and the Arctic ice caps, and DEMs of Quaternary glaciated landscapes (onshore and offshore), provide the data on which these ideas can be tested.

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