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Boundary conditions of an active West Antarctic subglacial lake: implications for storage of water beneath the ice sheet

M. J. Siegert[1], N. Ross[2], H. Corr[3], B. Smith[4], T. Jordan[3], R. Bingham[5], F. Ferraccioli[3], D. Rippin[6], and A. Le Brocq[7]

[1]Bristol Glaciology Centre, School of Geographical Sciences, University of Bristol, BS8 1SS, UK
[2]School of Geography, Politics and Sociology, Newcastle University, NE1 7RU, UK
[3]British Antarctic Survey, Cambridge CB3 0ET, UK
[4]Applied Physics Lab Polar Science Center, University of Washington Seattle, WA 98105, USA
[5]School of Geosciences, University of Aberdeen, Aberdeen AB24 3UF, UK
[6]Environment Department, University of York, York YO10 5DD, UK
[7]School of Geography, University of Exeter, Exeter EX4 4RJ, UK

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Correspondence to: M. J. Siegert (m.j.siegert@bristol.ac.uk)

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Abstract

Repeat-pass IceSat altimetry has revealed 124 discrete surface height changes across the Antarctic Ice Sheet, interpreted to be caused by subglacial lake discharges (surface lowering) and inputs (surface uplift). Few of these active lakes have been confirmed by radio-echo sounding (RES) despite several attempts (notable exceptions are Lake Whillans and three in the Adventure Subglacial Trench). Here we present targeted RES and radar altimeter data from an "active lake" location within the upstream Institute Ice Stream, into which 0.12 km^3 of water is calculated to have flowed between October 2003 and February 2008. We use a series of transects to establish an accurate appreciation of the influences of bed topography and ice-surface elevation on water storage potential. The location of surface height change is over the downslope flank of a distinct topographic hollow, where RES reveals no obvious evidence for deep (> 10 m) water. The regional hydropotential reveals a sink coincident with the surface change, however. Governed by the location of the hydrological sink, basal water will likely "drape" over existing topography in a manner dissimilar to subglacial lakes where flat strong specular RES reflections are measured. The inability of RES to detect the active lake means that more of the Antarctic ice sheet bed may contain stored water than is currently appreciated. Variation in ice surface elevation datasets leads to significant alteration in calculations of the local flow of basal water indicating the value of, and need for, high resolution RES datasets in both space and time to establish and characterise subglacial hydrological processes.

1 Introduction

Over the last ten years, our appreciation of basal hydrology in Antarctica has changed significantly. In the most recent inventory, Wright and Siegert (2012) collated evidence for 379 Antarctic subglacial lakes. Most of these (~ 250) are evidenced solely by discrete and distinct radio wave reflections from flat ice-water interfaces, using ice penetrating radio-echo sounding (RES). A few have been interpreted from flat ice sheet surfaces derived from a large region of floating ice (Bell et al., 2006, 2007). The remainder (numbering 124) have been established from repeat-pass satellite altimetric measurements of ice-surface changes interpreted as evidence of water flowing into, and discharge from, "active" subglacial lakes (Smith et al., 2009). In some instances, RES data from a subglacial lake have been combined with repeat-pass altimetry, revealing compelling evidence for a substantial and active basal hydrological system in Antarctic ice streams (Christianson et al., 2012; Horgan et al., 2012). In one case, subglacial water discharge beneath Byrd Glacier, inferred from satellite altimetry, has been shown to coincide with an increase in measured ice-surface velocity (Stearns et al., 2008). While most surface elevation changes have been measured close to the ice-sheet margin, satellite altimetric evidence of basal water flow in the centre of East Antarctica (Wingham et al., 2006) points to a potentially highly-connected basal hydrological system linking large stores of water beneath the ice divide through fast flowing ice streams to the ice margin (Wright et al., 2012).

Several altimetrically-derived subglacial lakes have numerous repeat-pass transects, which define accurately both lake outline and the loss/gain of water (e.g. Lake Whillans, Fricker et al., 2007). However, around half of the "active lake" inventory (Smith et al., 2009) is comprised of evidence for losses/gains of subglacial water from fewer than five transects, often just two and occasionally only one. In such cases, interpretation of discrete ice-surface height changes as "active lakes" beneath the ice is plausible yet, given the paucity of data, inconclusive. Other explanations could include migration of "packets" of basal water (or sediment), as postulated by Gray et al. (2005) rather than volume loss/gain in distinct lake locations. Additionally, it is not known whether proposed "active lakes" drain completely before refilling, making them potentially ephemeral features. In both the Aurora and northern Wilkes subglacial basins, airborne geophysical data were acquired at the locations of proposed "active lakes" (Wright et al., 2012, 2013). Based on available data, no classic bright, flat and strong radio-wave reflections characteristic of deep-water lakes have been found to coincide with the locations of the "active lakes".

In a few cases, "active lakes" were found to coincide with radio reflections showing one or two of these traits. Explanations for this mismatch include both the inability of radar to detect subglacial lakes under certain glaciological conditions, and a lack of water during the time of RES data acquisition. Furthermore, for all "active lakes", surrounding physiographical information that might help to understand how water flows into and out of a region cannot be detected using satellite altimetry alone.

Here we present targeted RES and radar altimeter data from an "active lake" within the upstream Institute Ice Stream in West Antarctica, named "Institute E2". These data are coupled with IceSat ice-surface measurements to ascertain evidence for basal water, local and regional basal topography and hydrological potential. In doing so, we reveal how the subglacial system can store and route basal water without the build-up of deep-water lakes, and demonstrate the ability of (and necessity for) high-resolution surface altimetry and RES to detect and understand such a system.

2 Geophysical methods and data

Airborne geophysical data were acquired over Institute E2, centred on the coordinate 82.625° S, 79.013° E (Fig. 1). The subglacial lake was defined by two pairs of IceSat repeat-pass altimeter data (October 2003 and February 2008), between which the surface elevation rose by up to ∼ 5.5 m. According to calculations by Smith et al. (2009), the lake experienced a gain in water of 0.12 km^3 and, according to the most recent data, was still gaining water at the end of 2009 (Figs. 2 and 3). The lake was used to define the location of a deep field camp on the Institute Ice Stream where, in the austral summer of 2010/11, a RES campaign to measure basal topography and conditions of the Institute and Möller Ice Streams was undertaken (Ross et al., 2012). As the camp was positioned just a few km from the central lake coordinate, input/output survey flights were organised such that numerous RES transects were recovered from the lake at a variety of orientations. The survey geometry also led to RES data be-

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ing acquired from the immediately surrounding regions, at a flight spacing of 7.5 km (Fig. 1).

The RES equipment used is a coherent system with a carrier frequency of 150 MHz, a bandwidth of 12 MHz and acquires pulse-coded waveforms at a rate of 312.5 Hz. Aircraft position was obtained from differential GPS. Surface elevation of the ice sheet was derived from radar altimeter terrain-clearance measurements, with an accuracy of ±1 m. Doppler processing was used to migrate radar-scattering hyperbola in the along-track direction. The onset of the received bed echo was picked in a semi-automatic manner using PROMAX seismic processing software. The post-processed data rate was 6.5 Hz giving a spatial sampling interval of ~10 m. The travel time in the near-surface firn layer is taken as the sum of two components; solid ice and an air gap. When calculating ice thickness we use a nominal value of 10 m to correct for the firn layer. A spatial variation in density affects the equivalent air gap, however, and this could account for variations across the survey area in the order of ±3 m. This error is small relative to the overall error budget, which is dominated by the uncertainty in the overall ice thickness, estimated to be in the order of ±1 %.

Four transects (including one half the length of the others) were flown directly over the central coordinate of Institute E2 (Smith et al., 2009) in different orientations (Fig. 1). Measurement of ice thickness, subtracted from surface elevation, yields subglacial topography (Fig. 4a). Assuming water is driven by gravity and overburden pressure of ice according to Shreve (1972), the bed and ice-surface elevations can be used to calculate the hydrological potential (Fig. 4b), which in turn can deliver hydrological pathways (assuming the bed is wet everywhere) (Fig. 4). Finally, the basal radio-echo power returns, when normalised (i.e. corrected for englacial absorption and geometry spreading) to account for ice thickness, can provide information on basal properties (Fig. 5).

Radar altimeter data were gridded, and used to define the ice surface elevation around Institute E2. The ice surface elevations were compared with the most recent digital elevation model compilation of Antarctica (Bamber et al., 2009), and revealed significant (10–20 m) differences in elevation at a number of locations, most notably over the centre of Institute E2 (Fig. 6). We believe our surface elevation is more accurate over and around Institute E2 than Bamber et al.'s (2009) satellite-derived ice surface because of the greater density of airborne versus satellite tracks in this region and the fact that the data were acquired within a short survey window of no more than 3 weeks.

3 Analysis

3.1 Ice surface elevation, bed topography and lake sensitivity

Institute E2 lies beneath a hydrological potential minima, and on the downslope flank of a sizeable topographic hollow, ~2 km below the ice surface, at least 15 km in length and ~1–2 km wide (Fig. 4; Video in the Supplement). In many places, the influence of local topography on the flow of basal water is sufficiently large that subglacial basins are coincident with hydrological sinks and, hence, are topographically-controlled subglacial lakes (Bell et al., 2006; Siegert et al., 2007). While it is possible for the hydrological sink to be displaced from obvious topographic basins (Christianson et al., 2012), what makes Institute E2's topographic situation unusual (compared with RES-defined subglacial lakes) is that its bed slope is aligned along the approximate direction of ice flow. This makes Institute E2 controlled by surface slope rather than bed topography and, therefore, potentially susceptible to change if the ice-surface elevation is adjusted. Obviously, one way of adjusting the ice surface is to fill the lake, thus the stability of Institute E2 may be related to its own growth. Since Institute E2 is not confined by topography, any discharge will only cease once the surface-slope-driven hydrological sink is re-established. According to Wingham et al. (2006) this might not occur until the lake has completely discharged. Hence, Institute E2 may well be ephemeral.

The ice sheet in this region has only relatively recently relaxed from its Last Glacial Maximum form (perhaps within the last 500–1000 yr; Livingston et al., 2013; Siegert et al., 2013). Given Institute E2's sensitivity to ice surface elevation, it is unlikely to

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predate the most recent glaciological setting, making it probably relatively young compared with some topographically-controlled subglacial lakes (such as Lake Ellsworth, Woodward et al., 2010).

3.2 RES evidence of basal water

At the centre of Institute E2, there is little obvious qualitative evidence from RES for the presence of a subglacial lake (Figs. 5 and 7). Subglacial lakes can be characterised by strong reflections (10–20 dB greater than surrounding bed) that are specular and flat, provided they are sufficiently deep (> 10 m) (Oswald and Robin, 1973; Carter et al., 2009). RES from Institute E2 reveal bed reflections that are non-specular and non-flat, however. According to Gorman and Siegert (1999), VHF radio waves can penetrate through pure shallow water bodies and, in six examples, they demonstrated how RES reflections from the ceiling and floor of a subglacial lake can interfere, providing the water depth is < 10 m. This can lead to RES reflections that are non-characteristic of deep subglacial lakes, and provides one explanation for why a sharp, mirror like interface is not observed in Institute E2 (in accordance with observations elsewhere, e.g. Langley et al., 2011; Welch et al., 2009).

Relative RES power returns from Institute E2 have a noticeable spatial variability (Fig. 5). Over the centre coordinate of Institute E2 (Smith et al., 2009) they do not appear to be anomalously large. In both the centre of the topographic basin, and the centre of the hydrological potential sink (which is offset to one side of the proposed area of Institute E2), relative basal power returns are ∼ 10–20 dB greater than from surrounding regions, however. Similarly strong RES returns, and therefore basal water, are also recorded from two other regions in the survey area (Fig. 5). Conceptually, even if the water depths are very low (of the order of centimetres) enhanced RES reflections should still be observed compared with a dry bed. Hence, it is possible that the distribution of water in Institute E2 is discontinuous. This opens the possibility for bed roughness to further reduce basal power returns, which may also help to explain the spatial complexity observed (Fig. 5). Consequently, the water depth of Institute E2 may be less than 1 m in some places.

An alternative explanation for the distribution of basal power returns is that they portray defects in the RES system (either in data acquisition or processing). This is unlikely, given the system's ability to detect well-defined subglacial lake features in other regions of the wider survey, however.

3.3 Flow of basal water

According to Shreve (1972), based on our RES depiction of bed topography and radar measurement of ice-surface elevation, water will be driven from Institute E2 toward the main trunk of the Institute Ice Stream (Fig. 4), where a larger, deeper subglacial basin exists, which (considering the thickness of ice within it) is also very likely to contain subglacial water (Fig. 1). This larger basin is part of a significant, topographically-complex, valley which routes water northwards toward the trunk of the Institute Ice Stream and, from there, to the Filchner Ronne Ice Shelf (Le Brocq et al., 2013). Differences between our radar measurements of ice surface elevation and Bamber et al.'s (2009) DEM, which are of the order of 10–20 m in places, result in local changes to the calculation of the expected route of basal water flow (Fig. 4b), demonstrating a high level of sensitivity in basal hydrology to ice surface elevation in this region.

4 Basal processes and data quality

The measurement of a subglacial hydrological sink coincident with the IceSat-derived "active lake" is crucial to appreciating how water may both pond and discharge at the site. From RES alone, it is likely that the existence of the subglacial lake would have been missed due to its possible thickness (< 10 m) and/or its surface shape (i.e. not flat). Calculating the hydrological sink requires RES data, however, as does an appreciation of subglacial hydrological pathways. The example presented in this paper shows that RES alone may be both inadequate to fully comprehend the locations of "active lakes" yet crucial to appreciating the flow of water beneath the ice sheet. Currently, RES data are too sparse in Antarctica to accurately measure subglacial hydrological sinks other than those dominated by large scale topography. Sub-km resolution in RES data are needed, and yet although such data allow enhanced appreciation of the subglacial environment (e.g. Sun Bo et al., 2009; Woodward et al., 2010; King et al., 2009) the vast majority of the ice sheet base is sampled with transect separations normally > 5 km, and often far greater. Consequently, although the broad potential subglacial flow of water is understood (e.g. Siegert et al., 2007), information on the flow of basal water is currently spatially insufficient to fully comprehend how "active" subglacial lakes exchange water and, ultimately, influence ice sheet dynamics.

It is possible that Institute E2 discharged all of its water in the short period (~ 1 yr) between the end of the IceSat time series and when the RES data were acquired, thus explaining the lack of evidence for a lake in the RES data. It is also possible that the ice-surface elevation may have changed recently, thus affecting the flow of basal water and our potential ability to calculate it correctly. These are potentially serious data-related issues for the detection and understanding of basal water flow from active subglacial lakes, and their solution requires a time series of high-resolution data that is currently unavailable.

Institute E2 was chosen for investigation because of its location to a planned RES survey. Through this opportunity we have been able to reveal that the lake is unlikely to comprise deep (> 10 m) water, and that it exists as a consequence of the interrelation between basal topography, ice flow and hydrology. Our data provide no evidence for why the lake grew in its present location, and not within the topographic valley nearby. The lake is likely to be ephemeral but there is no certainty that once drained it will later reoccupy the same location. As it appears similar to many of the other 124 lakes identified using satellite altimetric changes, we should be confident that the mismatch between intuitive identification of subglacial lakes from RES and inferred subglacial hydrological changes from ice-surface altimetry may be repeated in some (potentially

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many) other regions of Antarctica. The implication of the inability of RES to independently detect the locations of "active" subglacial lakes (both at Institute E2 and in the Byrd Glacier catchment, Wright et al., 2013), leads to the possibility that much more of the Antarctic bed contains water than is currently appreciated from RES alone.

5 Summary

Between October 2003 and December 2009, IceSat surface altimetry revealed a discrete zone of ice-sheet uplift within the Institute Ice Stream, interpreted by Smith et al. (2009) to be due to the filling of a subglacial lake by at least 0.18 km^3 of water (named Institute E2). In the austral summer of 2010–2011, an airborne geophysical survey of the Weddell sector of West Antarctica targeted Institute E2 by flying a series of transects centred on the middle of the uplifted region. RES was used to measure subglacial topography and evidence for basal water. Radar altimetry was used to measure ice-sheet surface elevation. Institute E2 lies over the upstream flank of a subglacial hollow, revealing that the lake is not controlled solely by basal topography. Using altimetry information to calculate how water may flow beneath the ice sheet revealed a hydropotential minimum coincident with the IceSat-derived outline of Institute E2. RES also reveals relatively stronger (10–20 dB) basal reflections from beneath the hydropotential minimum than in adjacent regions, suggesting the presence of water. No classic RES reflections from a "deep-water" (> 10 m) subglacial lake were observed, however. Based on these data, we consider Institute E2 is controlled far more by ice-sheet surface slopes than basal topography, making it potentially susceptible to small ice-surface changes. Such changes may occur as a consequence of ponding by basal water, making the lake's stability potentially influenced by its own growth. As such, Institute E2 is likely to be ephemeral and shallow; we believe it unlikely that the water depth is greater than 10 m, and may be less than 1 m in places. As RES alone was incapable of detecting Institute E2 (among others), considerably more basal water may be present in Antarctica than currently appreciated from RES.

Supplementary material related to this article is available online at: http://www.the-cryosphere-discuss.net/7/2979/2013/tcd-7-2979-2013-supplement.zip.

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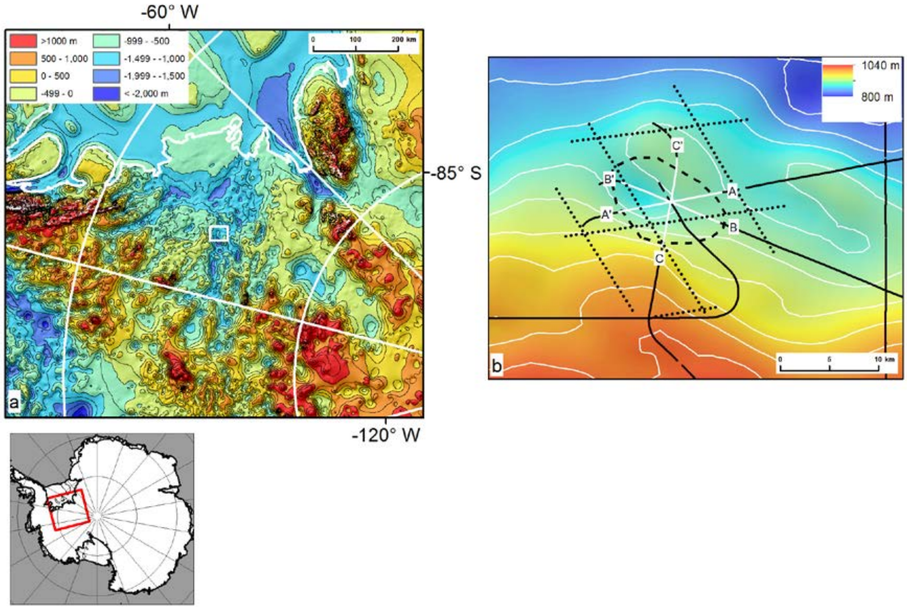

Fig. 1. (a) Subglacial topography of the Institute ice stream region (with insert for study region in West Antarctica). RES data used to compile the map are from Ross et al. (2012) and are included in BEDMAP II (Fretwell et al., 2013). **(b)** RES (black line) and IceSat (black dashed line) transects over and around Institute E2, as delineated by Smith et al. (2009) (white dashed line). Labels A–A′, B–B′ and C–C′ refer to RES transects provided in Fig. 7.

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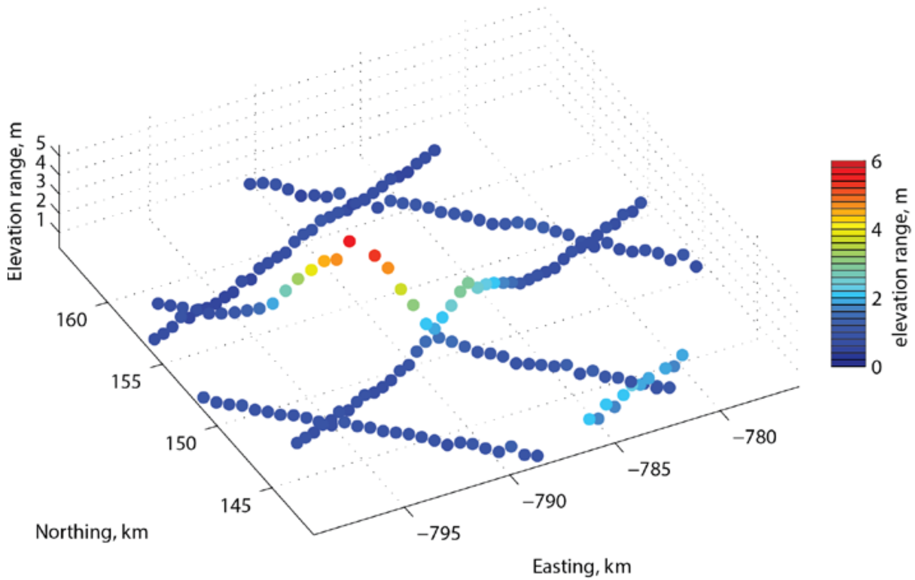

Fig. 2. Three-dimensional view of IceSat tracks and the uplift detected between October 2003 and February 2008 (Smith et al., 2009).

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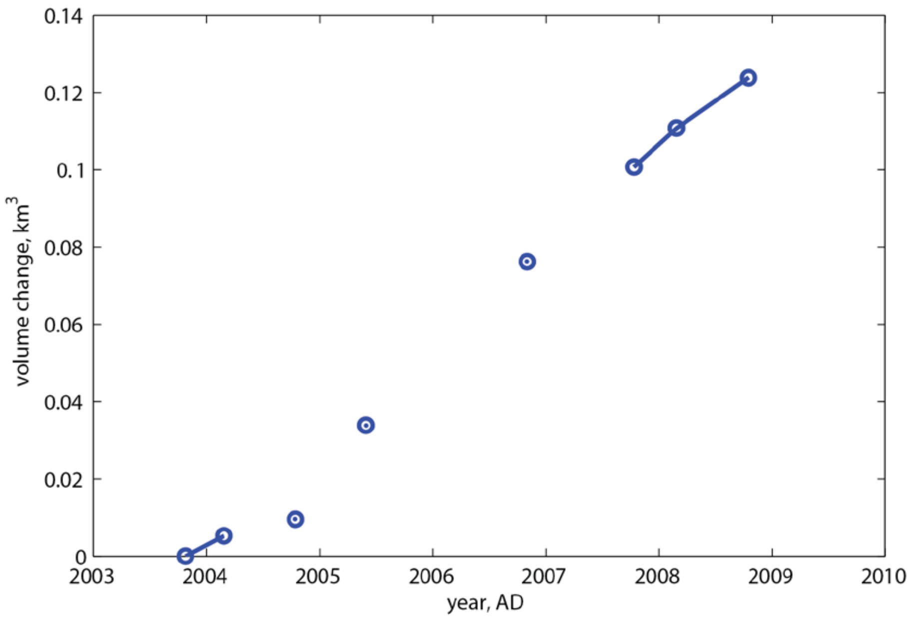

Fig. 3. Time-dependent change in the volume of Institute E2, using the method detailed in Smith et al. (2009) and IceSat surface elevation data between October 2003 and December 2009.

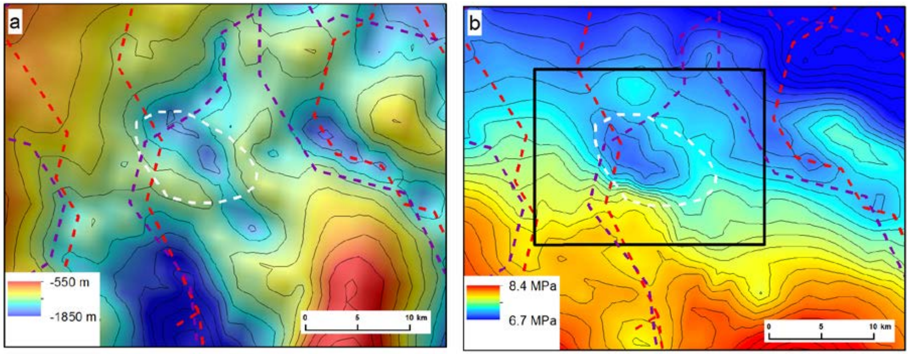

Fig. 4. Hydrological pathways (red dashed line) using basal topography (Fig. 1) and aircraft altimeter data (Ross et al., 2012) superimposed on **(a)** bed topography and **(b)** hydrological potential of the region surrounding Institute E2 (location as in Fig. 1b). The black box denotes the region depicted in Fig. 5. Dashed purple lines denote hydrological pathways calculated using surface elevations from Bamber et al. (2009). White dashed line locates edge of Institute E2 according to Smith et al. (2009). Contour lines are in 100 m **(a)** and 100 kPa intervals **(b)**.

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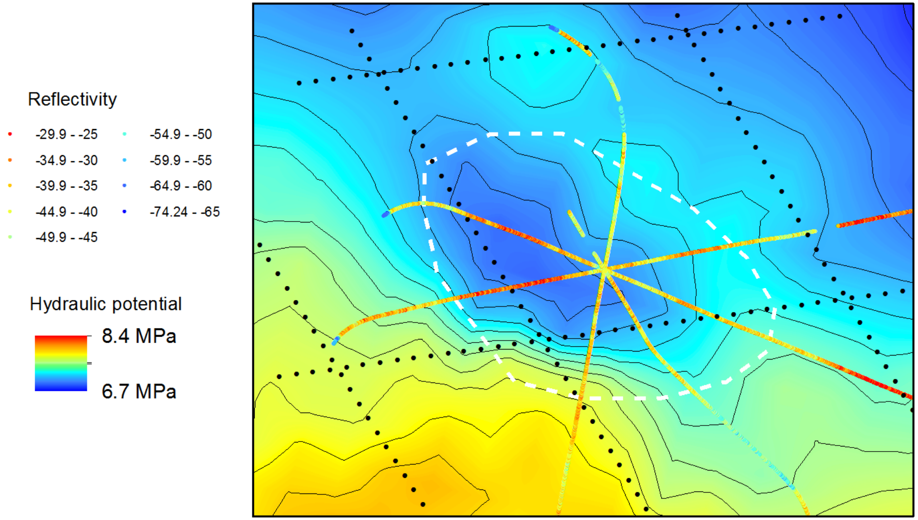

Fig. 5. Received RES strength (in dB) from basal reflections along lines defined in Fig. 1 over and around Institute E2 (see Fig. 4b for location). The reflection strengths have been corrected for geometric spreading with depth. The lines are superimposed onto a map of the ice sheet hydro-potential (as in Fig. 2b). Also shown is the outline of Institute E2, as proposed by Smith et al. (2009) (white dashed line). To calculate basal power returns, the RES data were first processed by unfocused SAR; a coherent integration over the returns from a section of the first Fresnel zone. The window length was chosen to be ~50 m within which the peak value of the basal reflector power was picked. The signal power was then compensated for differences in path length, transmitted output, and dielectric absorption (a nominal average absorption value of 3.5 dB (100 m)$^{-1}$ was applied). Reducing the rate of dielectric absorption increases the relative basal powers in shallower ice, but does not adversely affect the general distribution of received RES strength. Finally an estimated system performance figure of 200 dB was subtracted from the result.

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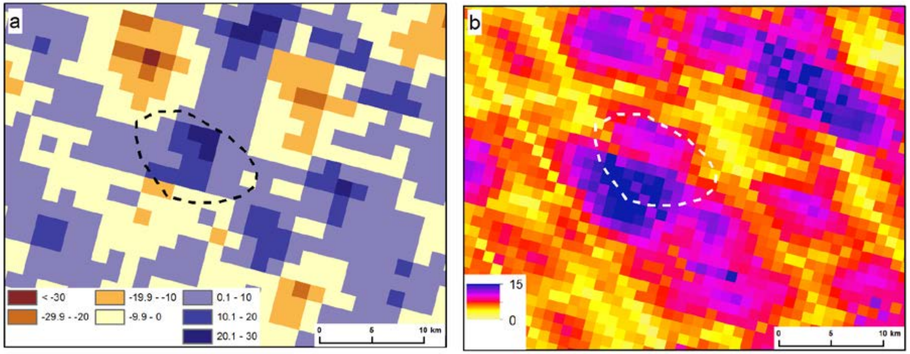

Fig. 6. Quantification (in meters) of errors in ice-sheet surface elevation. **(a)** Difference between Griggs and Bamber (2009) and this paper's (Fig. 1b) surface elevations. **(b)** Map of the distribution of RMS error in the DEM calculated using a multiple regression based on airborne validation data (adapted from Griggs and Bamber, 2009).

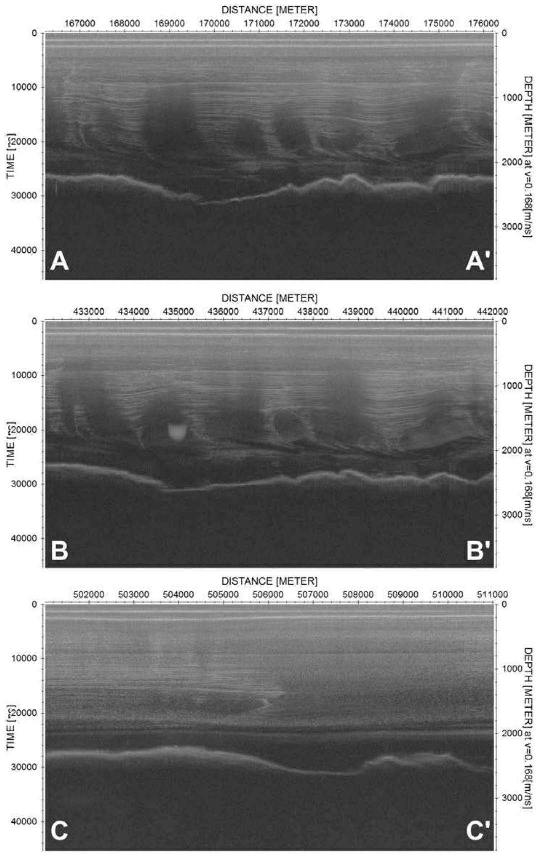

Fig. 7. RES transects centred on the "active" subglacial lake Institute E2. The locations of the transects are provided in Fig. 1b.

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