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Steep reverse bed slope at the grounding line of the Weddell Sea sector in West Antarctica

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The bed of the West Antarctic Ice Sheet is, in places, more than 1.5 km below sea level[1,2]. It has been suggested that a positive ice-loss feedback may occur when an ice sheet's grounding line retreats across a deepening bed[1–3]. Applied to the West Antarctic Ice Sheet, this process could potentially raise global sea level[4] by more than 3 m. Hitherto, attention has focussed on changes at the Siple Coast[5–7] and Amundsen Sea embayment[8–10] sectors of West Antarctica. Here, we present radio-echo sounding information from the ice sheet's third sector, the Weddell Sea embayment, that reveals a large subglacial basin immediately upstream of the grounding line. The reverse bed slope is steep, with about 400 m of decline over 40 km. The basin floor is smooth and flat, with little small-scale topography that would delay retreat, indicating that it has been covered with marine sediment[5,11] and was previously deglaciated. Upstream of the basin, well-defined glacially carved fjords with bars at their mouths testify to the position of a former ice margin about 200 km inland from the present margin. Evidence so far suggests that the Weddell Sea sector of the West Antarctic Ice Sheet has been stable, but in the light of our data we propose that the region could be near a physical threshold of substantial change.

Marine ice sheet instability (MISI) is thought possible if the bed is both below sea level and deepens upstream of the grounding line separating floating and grounded ice (a reverse ice-sheet bed slope)[1–3]. Although the existence of a large confined ice shelf could act to stabilize the grounded ice sheet, the extent of this influence is uncertain at present[3]. Understanding the risk of West Antarctic Ice Sheet (WAIS) change owing to MISI therefore requires knowledge of the subglacial bed and an assessment of where reverse bed slopes exist. Knowledge of past WAIS behaviour is also critical for evaluating present risk as regions susceptible to change can be identified. Radio-echo sounding (RES) provides information on each of these areas, by directly measuring subglacial elevations and, when collected across a region, quantifying glacial landscapes that may have been influenced by former ice-sheet configurations[12,13].

In the austral summer 2010/11, a UK team acquired >25,000 line kilometres of RES data across the Institute and Möller ice streams, inland of the Filchner Ronne Ice Shelf (FRIS) (Fig. 1a,b). From these data a new bed topography was established (Fig. 1c), filling one of the main gaps in our knowledge of subglacial West Antarctica[14] (Supplementary Fig. S1). The new survey reveals a significant subglacial basin—∼20,000 km^2, about the size of Wales—beneath the WAIS. The basin is divided into two components, which have depths of >1,600 m and >1,700 m below sea level (b.s.l.; regions A and B in Fig. 1c). The new survey shows the present Institute and Möller grounding lines to be perched at the inflexion of the reverse marine slope that leads into the basin (Fig. 2a,b).

The gradients of the reverse slopes of the Institute and Möller ice streams are greater than in the Siple Coast and Pine Island Glacier (PIG) and are comparable to that of upstream Thwaites Glacier (Fig. 2). Although basin A (upstream of the Institute Ice Sheet grounding line) is shallower than the Byrd Subglacial Basin (>2,000 m b.s.l.), there are two other important differences that make the Institute Ice Stream situation one of concern. First, unlike Thwaites Glacier, there is little small-scale topography on the Institute/Möller reverse slopes that could act as pinning points to delay retreat of the grounding line. Second, the basal shear of the Institute Ice Stream over basin A is very low[15], whereas upstream of Thwaites Glacier's grounding line the basal shear stress is significant and affects ice dynamics considerably[16]. Basal shear stress (that is, the resistance of the ice sheet bed to ice flow) is critical in determining the form of the overlying ice. Low basal shear stresses support low driving stresses and low-gradient ice-sheet surfaces, whereas high basal shear stresses support high driving stresses and steep ice-sheet surfaces. Therefore, although Thwaites Glacier is undergoing dynamic thinning at present[9], basal boundary conditions in its coastal margin reduce immediate susceptibility to MISI (Supplementary Fig. S1).

Between the Institute and Möller ice stream grounding lines and the basin the ice sheet is close to floatation (Figs 1d and 2a,b)[14,15]. Recent precise (∼2 cm) satellite laser measurements of the ice surface between the Institute and Foundation ice streams reveal evidence of tidal influence at the grounding zone[17], where large portions of the grounded ice plain (∼10 km) are lifted to floatation at high tide (which at 8 m (ref. 18) is one of the highest in the world). A ∼10-km-wide zone of the ice plain between the Möller and Foundation ice streams may have, within the past few decades, transitioned permanently from grounded ice to fully floating ice shelf[17].

One chief driver of grounding-line retreat, and a potential trigger for MISI, is thought to be thinning of the floating ice shelf by basal melting[19]. For the Institute and Möller ice streams this would reduce the buttressing effect of the ice shelf on the grounded ice, encouraging ice-flow acceleration and, as a consequence, ice-sheet mass loss. Although the extreme case would be the complete decay of the FRIS, resulting in zero buttressing, it is unlikely to occur imminently as the ice shelf is thought to be stable at present[20] and is, in fact, now thickening[21]. Migration of the grounding line is not necessarily solely dependent on ice-shelf decay, however; the

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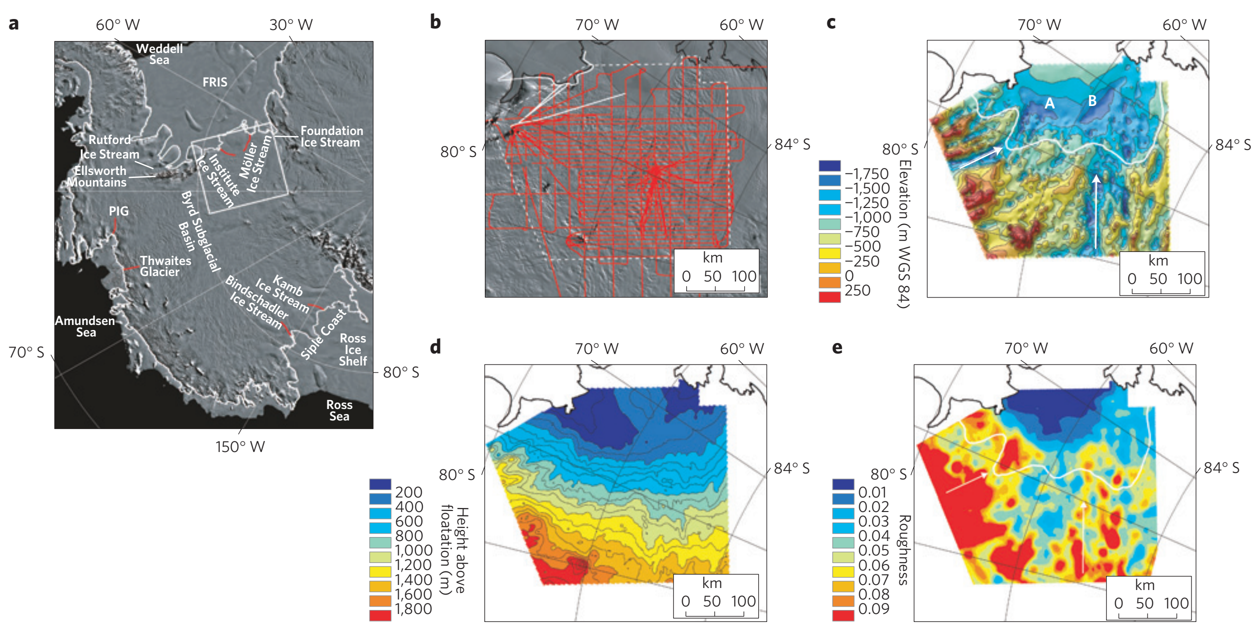

Figure 1 | Location, bed elevation, surface roughness and height above floatation of Institute and Möller ice streams. a, Location of the Institute and Möller ice streams and other features noted in the text. Red lines are locations of profiles shown in Fig. 2. White box is extent of **b**–**e**. Background image is the Moderate Resolution Imaging Spectroradiometer Mosaic of Antarctica imagery. White line is the Antarctic Surface Accumulation and Ice Discharge project's grounding line (shown as a black line in **b**–**e**). **b**, Aerogeophysical survey flights used to produce the bed-elevation grid. Red lines are data acquired in this survey. White lines represent BAS data from 2006–2007. White dashed line shows the extent of gridded data in **c**–**e**. **c**, Subglacial bed elevation with the proposed ice-sheet configuration after grounded ice decay within the subglacial basin. The white lines represent the former grounding line and the arrows depict likely ice-drainage routes. The basin is divided into two components; A, >1,600 m b.s.l. and B, >1,700 m b.s.l. WGS, World Geodetic System. **d**, Height above floatation of the Institute and Möller ice streams. **e**, Subglacial roughness of the Institute and Möller ice streams. Blue colours represent low roughness, reds show regions of high roughness.

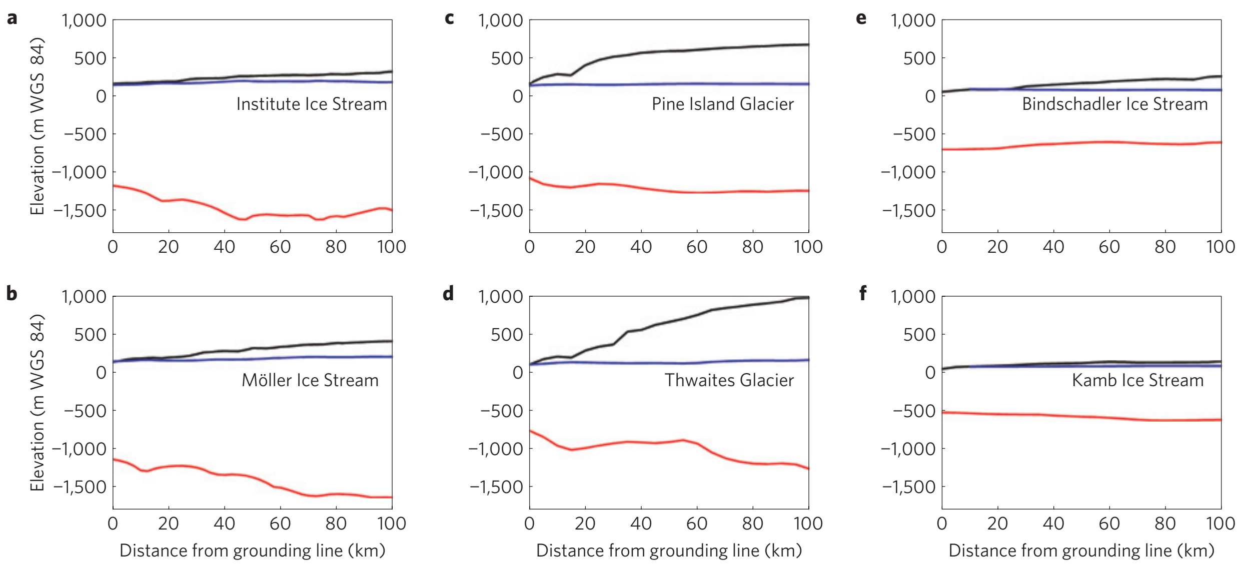

Figure 2 | Ice surface (black), bed elevations (red) and height above floatation (blue) over a distance of 100 km up ice flow of the grounding lines of six West Antarctic ice streams. a, Institute Ice Stream; **b**, Möller Ice Stream; **c**, Pine Island Glacier; **d**, Thwaites Glacier; **e**, Bindschadler Ice Stream (ice stream D); **f**, Kamb Ice Stream (ice stream C).

grounding line in the Ross Sea embayment has been shown to have retreated ~1,000 km since the start of the Holocene epoch, without loss to the ice shelf[22].

For enhanced ice-shelf thinning to occur, and/or basal mass loss from the grounding zone itself, warmth from the ocean is required[19]. The water mass thought to interact with the grounding line of ice streams feeding the FRIS is a derivative of high-salinity shelf water[20]. Although a colder water mass than the Circumpolar Deep Water, which is thought to be responsible for driving dynamic thinning in the Amundsen Sea region[2,19], modified high-salinity shelf water does generate melt at the grounding lines of the Foundation and Rutford ice streams[20,23]. Melt rates exceed $2.51 \pm 0.1\,\mathrm{m\,yr^{-1}}$ at the Rutford Ice Stream[23]. Basal melting at such deep grounding lines occurs because of the depression of the freezing point with increasing pressure[20]. In a scenario of MISI-driven retreat into the deep basins beneath the Institute and

Möller ice streams, grounding-line melt rates would increase as the grounding line deepened. Our ability to fully evaluate the risk associated with ocean-circulation-driven grounding-line melt is restricted, however, because cavity geometry and water circulation beneath the FRIS are constrained by few, sparse observations. Satellite methods indicate a low, or negligible, rate of grounding-line melt ($2 \pm 3\,\mathrm{m\,yr^{-1}}$) for the Institute and Möller ice streams[24]. Owing to their basal geometry and proximity to floatation (Figs 1d and 2a,b), however, even small changes in the rates of grounding-line melt may drive significant glaciological change.

Analysis of the basin's bed shows it to have a very low level of roughness (Fig. 1e; Supplementary Fig. S2), similar to the bed beneath the Siple Coast region of the WAIS (ref. 11) and smoother than much of the bed over Thwaites Glacier's catchment (Supplementary Fig. S2). The transition between the rough bed and the smooth basal zone is sharply defined at the edge of the subglacial basin. These features, and others noted on satellite data[25], are consistent with the presence of weak unconsolidated sediments deposited within the basin. Given that ice stream beds are primarily erosional environments, rather than depositional, these sediments are most likely to have been deposited in a marine setting when the basin was free of grounded ice[11]. Thus, the roughness of the basin floor indicates former deglaciation in this region. Although this may hint at former MISI, it does not prove it. The presence of soft basal sediments does, however, open the possibility of subglacial processes affecting ice-sheet dynamics, as observed in the Siple Coast[2,5–7], which could be associated with alterations to the position of the grounding line.

Upstream of the basin the RES data reveal a topography heavily influenced by the action of a former ice sheet terminating upstream of the trough, although the landscape may have developed initially by tectonic and then fluvial activity before glaciation. Several deep linearly incised channels are evident (Fig. 1c), sculptured through selective, linear basal erosion by fast-flowing ice streams or outlet glaciers[12,26]. At the mouths of the troughs, elevated bars are observed, indicative of a former grounded ice margin. Bedrock-trough mouth bars develop because of a downstream reduction in basal erosion caused by ice-flow divergence[26]. Hence, the subglacial landscape provides evidence of a grounded ice-sheet margin immediately upstream of the basin and, therefore, substantial former ice-sheet change in the Institute and Möller region.

A recent depiction of Antarctic Ice Sheet surface velocities revealed the position of major ice streams and their tributaries[27]. At present, the Möller Ice Stream is relatively restricted in extent and catchment. Conversely, the Institute Ice Stream is a major outlet for the WAIS, with tributaries that flow within the upstream linear topographic valleys. The former glacial landscape, therefore, has an important influence on modern ice dynamics. If the ice sheet were to retreat across the basin, it is likely that the tributaries would transform into discrete ice-stream outlets, implying enhanced erosion within the restricted ice mass compared with today (Fig. 1c). An ice-sheet reconfiguration of this nature would generate enhanced ice flow from major WAIS divides, increasing the potential for more widespread ice-sheet drawdown.

Numerical ice-sheet modelling, which included ice-grounding-line migration (albeit run using an ice-sheet bed less detailed than now available), supports the notion of the former WAIS being restricted to the highlands upstream of the subglacial basin[28]. Furthermore, cosmogenic exposure ages on erratics from the southern Ellsworth Mountains indicate continued progressive ice-sheet thinning in this region as it relaxes to a stable interglacial position[29]. Although we have no information on when the ice sheet last decayed across the Institute and Möller ice streams, modelling indicates that it may have occurred numerous times in the past[28], possibly as recently as 125,000 years ago (the last interglacial). Additionally, geochronological evidence from Bermuda of sea level 400,000 years ago up to 13 m higher than now points to major WAIS change at this time[30].

Bed information beneath the Institute and Möller ice streams depicts a glacial landscape formed by an ice sheet restricted to the region upstream of a large subglacial basin. Given that the present-day grounding line is now very close to (if not directly on) a significant reverse marine slope, and that satellite altimetry indicates changes to the grounding-line position, it is important to consider the Institute and Möller ice streams region of WAIS at risk of decay through grounding-line retreat.

Methods

Details of the RES acquisition and data processing. The British Antarctic Survey (BAS) airborne radar system was installed on a ski-equipped Twin Otter aircraft. This survey aircraft operated from two remote field camps located to maximize survey efficiency. Between 23 December 2010 and 9 January 2011 26 survey and positioning flights collected ~25,000 km of aerogeophysical data. Flights were flown in a stepped pattern. This design aimed to maintain a surface clearance of <500 m, optimizing radar data collection while achieving level flight required for simultaneous collection of airborne gravity data. The ice-sounding radar 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 global positioning systems and the surface elevation of the ice sheet was derived from radar/laser altimeter terrain-clearance measurements. 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 13 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 of the order of ±3 m. This error is small relative to the overall error budget, however, which is dominated by the uncertainty in the overall ice thickness, estimated to be of the order of ±1%. Crossover analysis (independent measurements at the same position) of the entire survey (including parts of the flights beyond the extent of our grid) yielded RMS differences of 18.29 m in ice thickness and 1.44 m in ice surface elevation ($n = 15{,}971$). Spatial variation in the RMS differences of ice thickness were apparent, however; an RMS error of 20.58 m ($n = 12{,}194$) was obtained for the upper parts of the gridded survey area, dropping to 6.00 m ($n = 3{,}531$) for the subglacial basin. This variation reflects the roughness of the underlying topography. Bed picks from the aerogeophysical survey were gridded, using a natural neighbour interpolation, along with data from a single flight line acquired during the 2006–2007 BAS IMAGE–GRADES survey, to produce a map of bed elevation. In Fig. 1, ice thickness above floatation (H^*) was calculated using: $H^* = H + (\rho_w/\rho_i) \times h$, where H = ice thickness, ρ_w = density of seawater ($1{,}028\,\mathrm{kg\,m^{-3}}$), ρ_i = density of ice ($917\,\mathrm{kg\,m^{-3}}$), and h = bed elevation.

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References

1. Mercer, J. H. West Antarctic Ice Sheet and CO_2 greenhouse effect: A threat of disaster. *Nature* **271,** 321–325 (1978).
2. Joughin, I. & Alley, R. B. Stability of the West Antarctic Ice Sheet in a warming world. *Nature Geosci.* **4,** 506–513 (2011).
3. Schoof, C. Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *J. Geophys. Res.* **112,** F03S28 (2007).
4. Bamber, J. L., Riva, R. E. M., Vermeersen, B. L. A. & Le Brocq, A. M. Reassessment of the potential sea-level rise from a collapse of the West Antarctic Ice sheet. *Science* **324,** 901–903 (2010).
5. Scherer, R. P., Aldahan, A., Tulaczyk, S., Kamb, B., Engelhardt, H. & Possnert, G. Pleistocene collapse of the West Antarctic Ice Sheet. *Science* **281,** 82–85 (1998).
6. Anandakrishnan, S. & Alley, R. B. Stagnation of ice stream C, West Antarctica by water piracy. *Geophys. Res. Lett.* **24,** 265–268 (1997).
7. Rose, K. E. Characteristics of ice flow in Marie Byrd Land, Antarctica. *J. Glaciol.* **24,** 63–75 (1979).
8. Holt, J. W. *et al.* New boundary conditions for the West Antarctic Ice Sheet: Subglacial topography beneath Thwaites and Smith glacier catchments. *Geophys. Res. Lett.* **33,** L09502 (2006).
9. Pritchard, H. D., Arthern, R. J., Vaughan, D. G. & Edwards, L. A. Extensive dynamic thinning on the margins of the Greenland and Antarctic ice sheets. *Nature* **461,** 971–975 (2008).

10. Dowdeswell, J. A., Evans, J., Ó Cofaigh, C. & Anderson, J. B. Morphology of the continental slope off Pine Island Bay, Amundsen Sea, West Antarctica. *Bull. Geol. Soc. Am.* **118,** 606–619 (2006).
11. Siegert, M. J., Taylor, J., Payne, A. J. & Hubbard, B. P. Bed roughness of the Siple Coast ice streams in West Antarctica. *Earth Surf. Proc. Land.* **29,** 1591–1596 (2004).
12. Young, D. A. *et al.* A dynamic early East Antarctic Ice Sheet suggested by ice covered fjord landscapes. *Nature* **474,** 72–75 (2011).
13. Bo, S. *et al.* The Gamburtsev Mountains and the origin and early evolution of the Antarctic Ice Sheet. *Nature* **459,** 690–693 (2009).
14. Jankowski, E. J. & Drewry, D. J. The structure of West Antarctica from geophysical studies. *Nature* **291,** 17–21 (1981).
15. Joughin, I. *et al.* Integrating satellite observations with modeling: Basal shear stress of the Filcher–Ronne ice streams, Antarctica. *Phil. Trans. R. Soc. A* **364,** 1795–1814 (2006).
16. Joughin, I. *et al.* Basal conditions for Pine Island and Thwaites glaciers, West Antarctica, determined using satellite and airborne data. *J. Glaciol.* **55,** 245–257 (2009).
17. Brunt, K. E., Fricker, H. A. & Padman, L. Analysis of ice plains of the Filchner–Ronne Ice Shelf, Antarctica, using ICESat laser altimetry. *J. Glaciol.* **57,** 965–975 (2011).
18. Padman, L., Erofeeva, S. Y. & Fricker, H. A. Improving Antarctic tide models by assimilation of ICESat laser altimetry over ice shelves. *Geophys. Res. Lett.* **35,** L22504 (2008).
19. Shepherd, A., Wingham, D. & Rignot, E. Warm ocean is eroding West Antarctic Ice Sheet. *Geophys. Res. Lett.* **31,** L23402 (2004).
20. Nicholls, K. W., Østerhus, S., Makinson, K., Gammelsrød, T. & Fahrbach, E. Ice-ocean processes over the continental shelf of the southern Weddell Sea, Antarctica: A review. *Rev. Geophys.* **47,** RG3003 (2009).
21. Shepherd, A. *et al.* Recent loss of floating ice and the consequent sea level contribution. *Geophys. Res. Lett.* **37,** L13503 (2010).
22. Conway, H., Hall, B. L., Denton, G. H., Gades, A. M. & Waddington, E. D. Past and future grounding-line retreat of the West Antarctic Ice Sheet. *Science* **286,** 280–283 (1999).
23. Jenkins, A., Corr, H., Nicholls, K., Stewart, C. & Doake, C. Interactions between ice and ocean observed with phase-sensitive radar near an Antarctic ice shelf grounding line. *J. Glaciol.* **52,** 325–346 (2006).
24. Rignot, E. & Jacobs, S. S. Rapid bottom melting widespread near Antarctic ice sheet grounding lines. *Science* **296,** 2020–2023 (2002).
25. Scambos, T. A., Bohlander, J., Raup, B. & Haran, T. Glaciological characteristics of the Institute Ice Stream using remote sensing. *Antarct. Sci.* **16,** 205–213 (2004).
26. Holtedahl, H. Notes on the formation of fjords and fjord valleys. *Geogr. Ann.* **49,** 188–203 (1967).
27. Rignot, E., Mouginot, J. & Scheuchl, B. Ice flow of the Antarctic Ice Sheet. *Science* **333,** 1427–1430 (2011).
28. Pollard, D. & DeConto, R. M. Modelling West Antarctic Ice Sheet growth and collapse through the past five million years. *Nature* **458,** 329–332 (2009).
29. Bentley, M. J. *et al.* Deglacial history of the West Antarctic Ice Sheet in the Weddell Sea embayment: Constrains on past ice volume change. *Geology* **38,** 411–414 (2010).
30. Raymo, M. E. & Mitrovica, J. X. Collapse of polar ice sheets during the stage 11 interglacial. *Nature* **483,** 453–456 (2012).

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Author contributions

M.J.S. and N.R. wrote the paper. M.J.S., F.F., N.R., R.G.B., H.F.J.C., T.A.J., A.L.B. and D.M.R. planned the aerogeophysical survey. N.R. and T.A.J. collected the data. N.R., H.F.J.C. and T.A.J. processed the data. D.M.R. and R.G.B. led the basal roughness analysis of the Institute and Möller ice streams. D.Y. and D.D.B. produced bed roughness analyses for the Siple Coast and Thwaites Glacier. A.L.B. provided glaciological information on MISI. All authors commented on a draft of the paper. M.J.S. and N.R. provided the geomorphic interpretation. M.J.S. and F.F. led the project.

Additional information

The authors declare no competing financial interests. Supplementary information accompanies this paper on www.nature.com/naturegeoscience. Reprints and permissions information is available online at www.nature.com/reprints. Correspondence and requests for materials should be addressed to N.R. or M.J.S.

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SI 1. Basal shear stress

Although areas of low basal shear stress are present, the bed of the Thwaites Glacier trunk <200 km inland of the grounding line is highly heterogenous. Isolated areas of low basal shear stress are aligned in bands across present ice flow, which lie between bands of stronger bed [Joughin et al., 2009]. On average, the bed of the Thwaites Glacier trunk has a much higher average basal shear stress [Joughin et al., 2009] than the trunk of Institute Ice Stream, which is characterised by much more uniform basal shear stresses [Joughin et al., 2006]. In many ways, the Institute and Möller ice streams are more akin to Pine Island Glacier, the trunk of which has a uniformly weak basal shear stress [Joughin et al., 2009] and is near to floatation. Unlike Pine Island Glacier, however, the Institute and Möller ice streams are not "primarily restrained by their hard-bedded regions of resistive bed, which are tens of kilometres wide and located just above their respective grounding lines" [Joughin et al., 2006].

Joughin, I., Bamber, J.L., Scambos, T., Tulaczyk, S., Fahnestock, M., MacAyeal, D.R., Integrating satellite observations with modeling: basal shear stress of the Filcher-Ronne ice streams', Antarctica, Phil. Trans. Roy. Soc.,A 364, 1795-1814, (2006).

Joughin, I., Tulaczyk, S., Bamber, J.L., Blankenship, D.D., Holt, J.W., Scambos, T., Vaughan, D.G. Basal conditions for Pine Island and Thwaites Glaciers, West Antarctica, determined using satellite and airborne data. Journal of Glaciology, Vol. 55, No. 190, (2009)

SI 2. Calculation of basal roughness

In order to determine basal roughness, we applied the forward fast Fourier transform (FFT) approach used previously in several studies of the Antarctic subglacial bed [e.g., Taylor et al., 2004; Bingham et al., 2007; Rippin et al., 2011]. We use the FFT to translate bed elevations, in a moving window, into the frequency domain. We then determine roughness as the integral of the resultant power spectra in each moving window [cf. Taylor and others, 2004]. In more detail, the precise steps were as follows [after Rippin et al., 2011]:

(1) Missing data points are common in RES data because of a failure to identify the bed or equipment errors. Where such gaps are small (less than 10 consecutive missing

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points which equates to a gap of ~220 m at a nominal mean step-size of 22 m), we carried out an interpolation across these gaps so as to maintain continuity in the FFT analysis. Where there was a gap of ten or more points, we designated the line as 'broken' and no interpolation was carried out. FFT analysis then recommenced further along-track where there were sufficient continuous data.

(2) Continuous sections of line were then re-sampled at a regular ~22 m step size to account for the variation in the sample spacing, and to enable smoothing of the data (see step 3) to be carried out over a consistent distance.

(3) Large-scale variations in topography were removed by subtracting a running-mean topography over a moving window of 100 sample points (equating to 2200 m).

(4) An FFT was carried out over 2^N samples, where N=5. This is equivalent to an FFT window of 32 data points (704 m). This is the smallest recommended value of N to use [cf. Brigham, 1988; Taylor and others 2004; Bingham and others 2007]. Bed roughness is thus defined as the integral of the resultant FFT of power spectra within each window.

(5) Roughness measurements were then interpolated using Kriging onto a 1 km grid, before being smoothed over a circular neighbourhood with a radius of 8 cells

Bingham, R.G., M.J. Siegert, D.A. Young and D.D. Blankenship. Organized flow from the South Pole to the Filchner-Ronne ice shelf: An assessment of balance velocities in interior East Antarctica using radio echo sounding data, J. Geophys. Res., 112, F03S26, doi:10.1029/2006JF000556. (2007).

Brigham, E.O. The Fast Fourier Transform and its Applications, Prentice-Hall, Englewood Cliffs, N.J. (1998).

Rippin, D.M., D.G. Vaughan and H.F.J. Corr. The basal roughness of Pine Island Glacier, West Antarctica. J. Glac. 57(201), 67-76. (2011).

Taylor, J., M.J. Siegert, A.J. Payne and B. Hubbard. Regional scale roughness beneath ice masses: measurement and analysis, Comput. Geosci., 30(8), 899–908. (2004).

Supplementary Figure 1.

Comparison between the new bed elevation map of the Institute and Möller ice streams and the ALBMAP depiction of bed elevation. (a) ALBMAP bed elevation [LeBrocq et al., 2010] for the Institute and Möller ice streams; (b) New map of bed elevation for the Institute and Möller ice streams; (c) Difference between new map of bed elevation and ALBMAP.

Le Brocq, A.M., Payne, A.J., Vieli, A. An improved Antarctic dataset for high resolution numerical ice sheet models (ALBMAP v1), Earth Syst. Sci. Data, 2, 247-260, (2010).

Supplementary Figure 2.

Analysis of subglacial roughness. (a) Along-track measurements of basal roughness for the Institute and Möller ice stream survey (blue indicate low roughness, red indicates high roughness). Low roughness values are clearly apparent over the subglacial basin observed proximal to the grounding lines of these ice streams, whilst the upper catchments of these ice streams are generally characterised by a higher degree of roughness, although there is also a greater degree of heterogeneity. (b,c) Comparison of subglacial roughness in several WAIS regions, using two methods. Additional data used in the calculations are from University of Texas, Austin, surveys of Thwaites Glacier [Holt et al., 2006, Vaughan et al., 2006] and the Siple Coast [Blankenship et al., 2001; Luyendyk et al., 2003]. (b) Basal roughness using the FFT methods (described in Supplementary Information 1). (c) Basal roughness using the point-to-point RMS deviation [Shepard et al., 2001] (with an 800 meter baseline (lines were detrended over 8 km). Raw bed elevation data were interpolated from the 3.5-4 Hz sample rates to 20 meter spacing. For Thwaites Glacier, the RMS differences at the 800 meter length scale (typically 40 m) result in large local slopes opposing the lower regional MISI slopes, thus inhibiting ice retreat.

Blankenship, D.D., D.Morse, C.A. Finn, R.E. Bell, M.E. Peters, S.D. Kempf, S.M. Hodge, M.Studinger, J.C. Behrendt, and J.M. Brozena (2001), Geological controls on the initiation of rapid basal motion for West Antarctic Ice Streams: A geophysical perspective including new airborne radar sounding and laser altimetry results, in The West Antarctic Ice Sheet: Behavior and Environment, Antarctic Research Series, vol. 77, edited by R.B. Alley and R.A. Bindschadler, pp. 105–121, American Geophysical Union.

Holt, J.W., D.D. Blankenship, D.L. Morse, D.A. Young, M.E. Peters, S.D. Kempf, T.G. Richter, D.G. Vaughan, and H. Corr (2006), New boundary conditions for the West Antarctic ice sheet: Subglacial topography of the Thwaites and Smith Glacier catchments, *Geophysical Research Letters*, 33(L09502), 10.1029/2005GL025561.

Luyendyk, B.P., D.S. Wilson, and C.S. Siddoway (2003), Eastern margin of the Ross Sea Rift in western Marie Byrd Land, Antarctica: Crustal structure and tectonic development, *Geochemistry Geophysics Geosystems*, 4(10), 10.1029/2002GC000462.

Shepard, M.K., B.A. Campbell, M.H. Bulmer, T.G. Farr, L.R. Gaddis, and J.J. Plaut (2001), The roughness of natural terrain: A planetary and remote sensing perspective, *Journal of Geophysical Research*, 106(E12), 32,777–32,795, 1029/2001JE001429.

Vaughan, D.G., H.F.J. Corr, F.Ferraccioli, N.Frearson, A. O'Hare, D. Mach, J. Holt, D. Blankenship, D.Morse, and D.Young. (2006), New boundary conditions for the West Antarctic Ice Sheet: Subglacial topography beneath Pine Island Glacier, *Geophysical Research Letters*, 33(L09501), 10.1029/2005GL025588.

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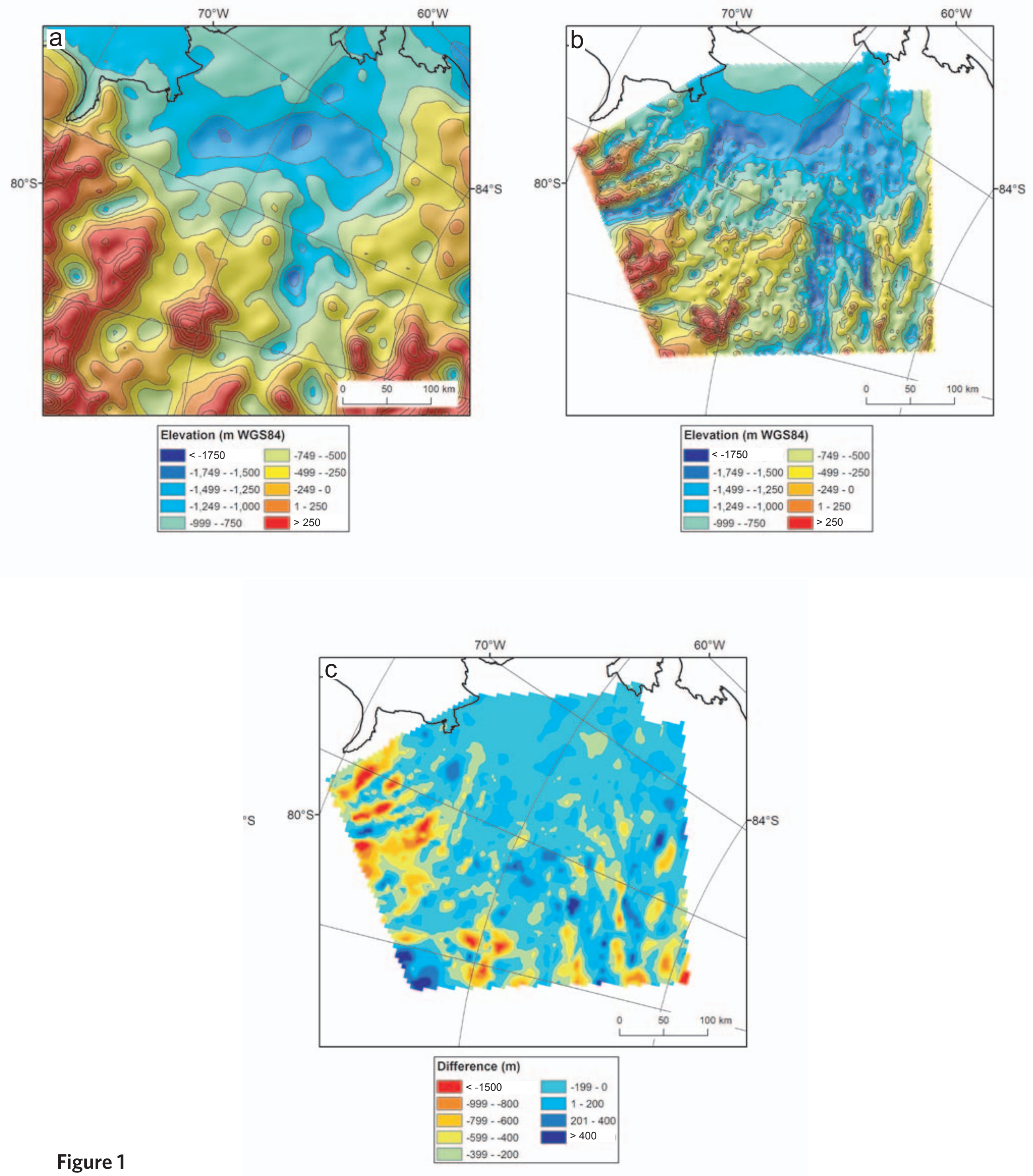

Figure 1

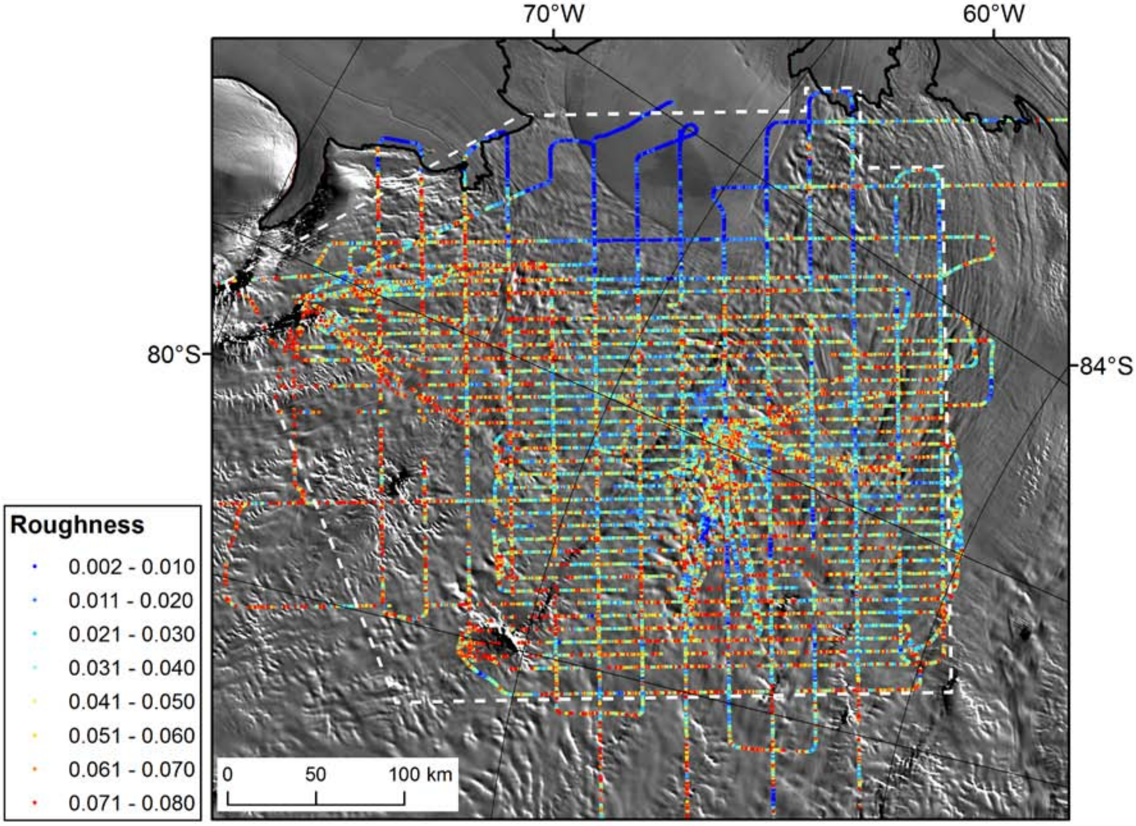

Figure 2a

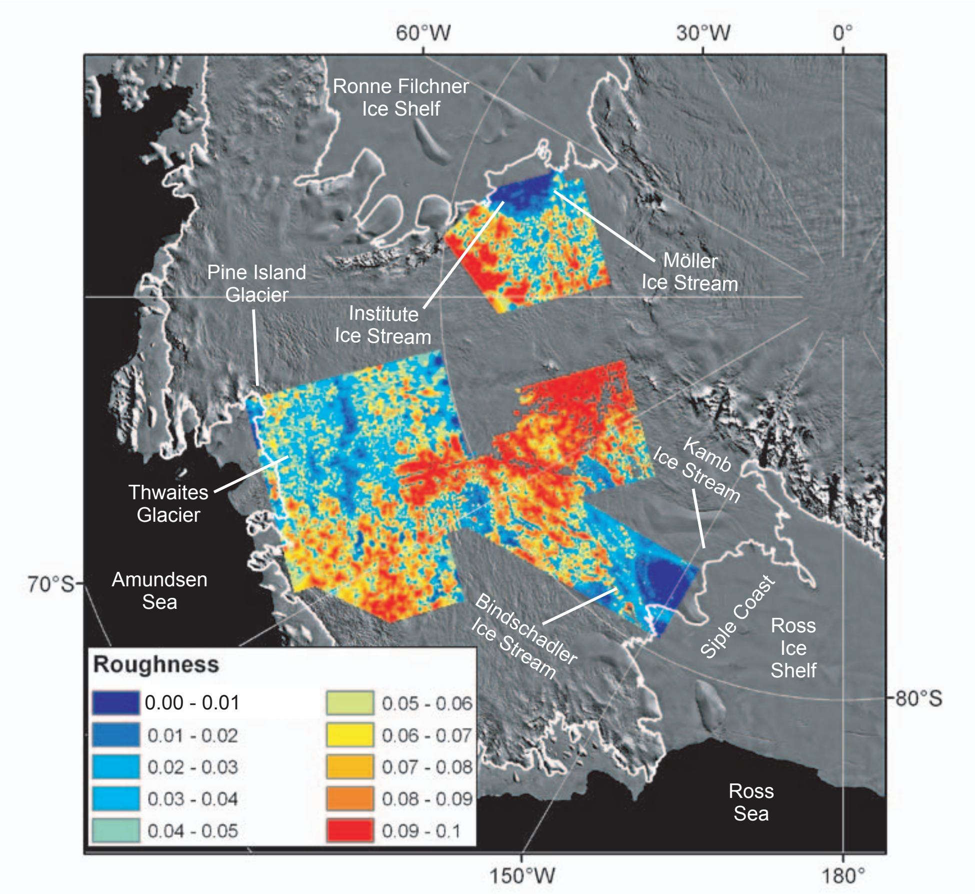

Figure 2b

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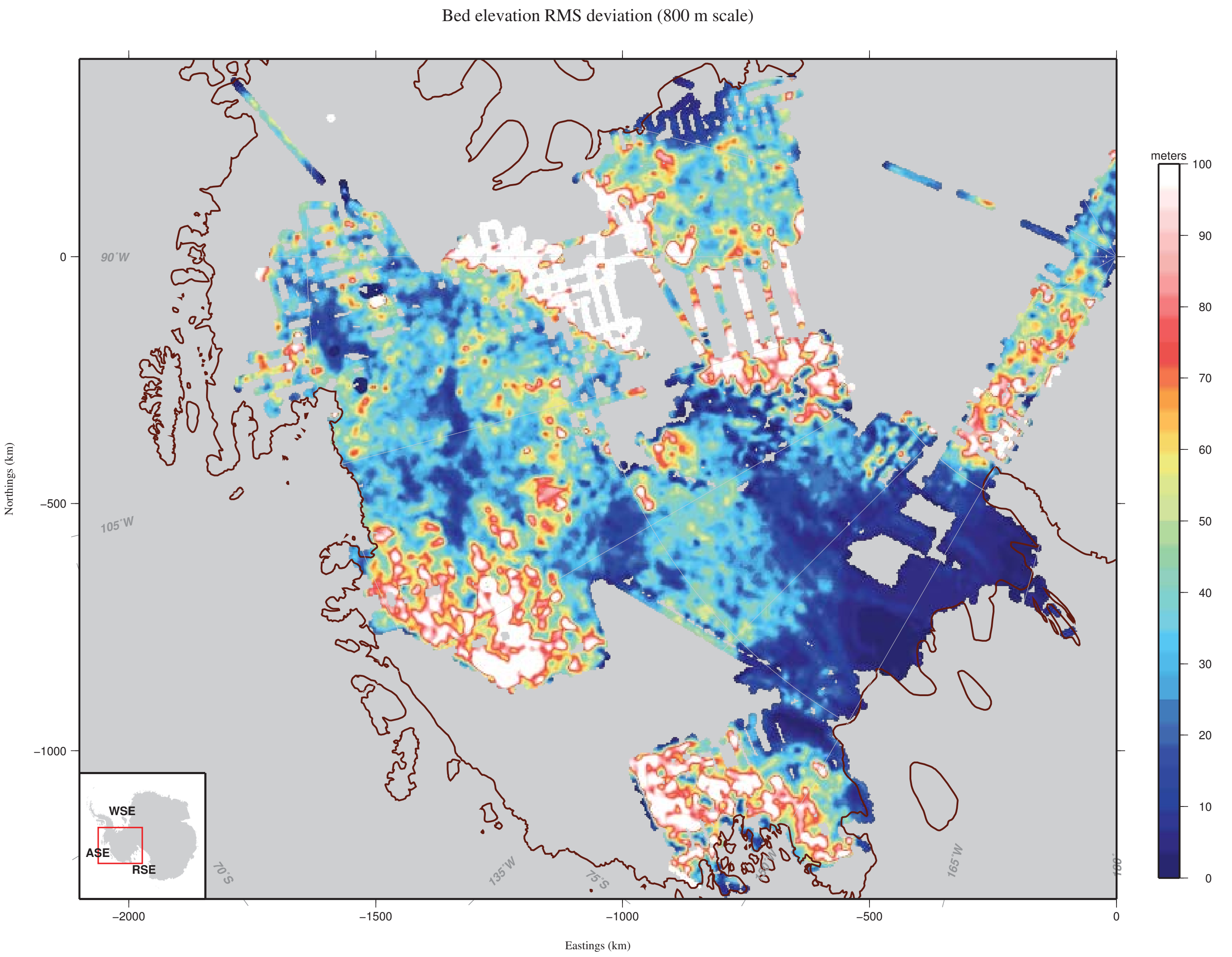

Figure 2c