Article:

Jamieson, S.S.R., Vieli, A., Livingstone, S.J. et al. (4 more authors) (2012) Ice stream grounding-line stability on a reverse bed slope. Nature Geoscience, 5. pp. 799-802.

https://doi.org/10.1038/ngeo1600

Ice stream stability on a reverse bed slope

Nature Geoscience (2012). doi:10.1038/ngeo1600

Stewart S.R. Jamieson[1], Andreas Vieli[1], Stephen J. Livingstone[2], Colm Ó Cofaigh[1], Chris Stokes[1], Claus-Dieter Hillenbrand[3], Julian A. Dowdeswell[4].

[1] Department of Geography, Durham University, South Road, Durham, DH1 3LE, United Kingdom.

[2]Department of Geography, University of Sheffield, Sheffield, S10 2TN, United Kingdom.

[3]British Antarctic Survey, Madingley Road, Cambridge, CB3 0ET, United Kingdom.

[4]Scott Polar Research Institute, University of Cambridge, Cambridge, CB2 1ER, United Kingdom.

Marine-based ice streams whose beds deepen inland are thought to be inherently unstable[1-3]. This instability is of particular concern because significant portions of the West Antarctic and Greenland ice sheets are marine-based and the resulting mass loss could contribute significantly to future sea-level rise[4-7]. However, the current understanding of ice-stream stability is limited by observational records that are too short to resolve multi-decadal to millennial-scale behaviour and validate numerical models[8]. Here we present a dynamic numerical simulation of Antarctic ice-stream retreat since the Last Glacial Maximum that, combined with bathymetric and marine geophysical mapping data, is consistent with the geomorphological record of palaeo-ice-stream retreat. We find that retreat of Marguerite Bay Ice Stream following the Last Glacial Maximum was highly non-linear and was interrupted by stabilisations on a reverse-sloping bed where rapid unstable retreat is expected from theoretical considerations. We demonstrate that these transient stabilisations were caused by enhanced lateral drag as the ice stream narrowed. We conclude that - in addition to bed topography - ice-stream width and long-term retreat history are crucial for understanding decadal- to centennial-scale ice-stream behaviour and ice sheet vulnerability.

Ice streams are fast-flowing arteries of ice sheets that dominate ice discharge into the oceans, impacting directly upon sea level. Many ice streams possess beds that are below sea level and typically deepen inland on a reverse-slope[9]. Theory suggests that ice discharge increases rapidly with water depth[3], and in the absence of lateral-drag induced buttressing from a floating ice shelf, grounding lines (marking the transition from grounded to floating ice) on reverse-bed slopes may be unstable[1-2]. Bed topography is therefore cited as a strong control on ice-stream retreat rate[3,10] and modern satellite observations of rapid ice-stream thinning and recession appear consistent with this theory[4-6]. However, with just two decades of data, these records are too short to identify the longer-term centennial to millennial-scale trends crucial for constraining future sea-level projections. Major uncertainties in predictions of ice-sheet vulnerability[11] relate to limitations in understanding processes controlling grounding-line motion and, importantly, to deficiencies in grounding-line treatment in ice-sheet models[12]. In recent years, significant advances in model development have been made[3,12-17], but tests have only been applied to simplified bed geometries or to steady-state conditions and lack validation against data over timescales longer than a few decades. We aim to understand the long-term controls and stability of marine ice streams and, for the first time, integrate a fully dynamic ice-stream model with the detailed geomorphological record of palaeo-ice-stream retreat imprinted on the sea-floor of Marguerite Bay, western Antarctic Peninsula (Fig. 1).

High-resolution mapping from swath bathymetry and analysis of subglacial landforms and sediments[18-21] (Supplementary Information; Fig. 1) identify the extent and evolution of the Marguerite Bay Ice Stream (MBIS) spanning several millennia following the Last Glacial Maximum (LGM). After 14 kyr the grounding line retreated rapidly along an overdeepened trough with a reverse-sloping bed from its LGM position at the continental shelf edge[20]. However, within this rapid retreat, short-term grounding-line stability is recorded by a series of eight major wedge-shaped sedimentary landforms (grounding-zone wedges: GZWs)[18-20,22]. This evidence for temporary stability on reversed beds challenges understanding of the controls on grounding-line behaviour. Indeed, palaeo-ice-stream retreat patterns are regionally inconsistent[20,23] indicating that although climate and ocean forcing are important, local controls such as basin geometry may modulate ice-stream retreat. We test this hypothesis using a time-dependent numerical ice-flow model.

The model[15,24] (Supplementary Information) considers variations in both along-flow bed geometry and ice-stream width, which are independently prescribed by marine geophysical observations. It includes resistive stresses from the bed and lateral margins, the transfer of stresses in up- and down-stream directions, a flotation criterion for calving and, crucially, a robust treatment of dynamic grounding-line behaviour which relies on a moving spatial grid. In order to assess ice-stream stability in the most unstable case, and because calving cannot be adequately constrained, the model has no ice shelf in the current treatment. A geophysically consistent steady-state LGM configuration extends to the continental shelf edge and provides the initial geometry (Supplementary Information; Supplementary Fig. S1) for a series of retreat experiments forced by simple sea-level and climate trends. Water depth (-100 to 0 m a.s.l.) and ice temperature (-20 to -15 °C) are increased linearly over an 8 kyr period to represent average trends during deglaciation. The imposition of this linear external forcing allows us to identify and understand the potential internal mechanisms controlling ice-stream retreat (it is not our intention to reproduce the precise chronology).

In reference experiment A (Fig. 2; Supplementary Video 1), linear external forcing results in highly non-linear stepped retreat of the grounding line across the outer and mid-shelf with several distinct phases of slow-down with retreat rates of only a few m yr^{-1}. Once the grounding line retreats from the continental shelf break (triggered by the prescribed water-depth increase), the sensitivity of ice discharge to water depth modulates the retreat pattern, partly explaining the contrasting non-linear response to linear forcing. For example, several retreat-rate slow-downs are consistent with the location of significant topographic highs, indicating the importance of bed morphology in defining 'pinning points' (e.g. at 500 km; Fig. 2). However, in the reverse-sloping portion of the bed and, significantly, in the absence of topographic highs, several slow-downs also occur. Moreover, the simulated stabilisations are broadly consistent with previously mapped GZW positions (Fig. 2).

These short-term changes in retreat-rate occur adjacent to GZWs 1, 2, 3, 6, 7 and 8 and prolonged grounding-line stability is modelled between GZWs 4 and 5 where the reverse slope steepens into the overdeepened portion of Marguerite Trough. Further sensitivity experiments that accelerate sea-level rise and temperature increases confirm that the spatial pattern of retreat is not controlled by external forcing because the locations of grounding-line stability remain fixed with only the timing of the retreat phases changing.

Importantly, bed slope alone cannot explain the temporary stabilisations identified by the modelling and geomorphological data, indicating that other factors dominate over water depth. In particular, GZWs and retreat slow-downs coincide spatially with narrowing of the trough, implying that variations in ice-stream width are a potentially important control (Fig. 1). It has been suggested that width variations may alter retreat rates in tidewater glaciers confined to fjords[25] but this has never been dynamically quantified on an ice-stream scale or over periods longer than a decade. To quantify the relative importance of the bed slope and width controls on ice-stream retreat behaviour we perform a series of experiments in which bed elevation and trough width are systematically modified.

Modelling experiment B is identical to A but imposes a horizontal bed over the outer 180 km of the continental shelf (Fig. 3). Again, simulated retreat behaviour is highly non-linear, but the grounding line experiences only two major stabilisations on the outer shelf (at GZW 2 and between GZWs 4 and 5). This non-linear response is surprising given the flat bed and linear forcing, but we note that these slow-downs again coincide with narrowing of the ice-stream trough (Fig. 3).

Experiment C combines the flat bed of experiment B with a straightening of the ice stream trough width over the outer 170 km of the continental shelf (Fig. 3). Nearly constant retreat rates of ca. 200 m yr^{-1} are obtained across the entire outer shelf, a pattern that is markedly different to the previous experiments and to the geomorphological evidence.

Model experiment D further tests the importance of width by applying a linear reverse-slope over the outer 180 km of the bed (Fig. 3). Although this induces rapid retreat over most of the outer shelf, a significant slow-down in retreat lasting ca. 50 years is nonetheless modelled between GZWs 4 and 6. This corresponds to the area of most significant trough narrowing (Fig. 3), demonstrating that unstable retreat associated with reverse beds can be temporarily overcome by constrictions in trough width.

These sensitivity experiments reveal that in addition to water depth, variations in ice-stream trough width play a fundamental role in controlling non-linear retreat behaviour. We explain the width-controlled grounding-line stabilisation effect via two mechanisms. Firstly, lateral resistance from the sides of the ice stream increases as its width narrows[26] (Supplementary Figs. S2-S3), thereby reducing the ice flux required to maintain a stable grounding line. This is particularly effective when the bed provides limited resistance, for example in the proximity of the grounding line. Lateral shear stress is consistently 50% higher over the narrow zone of GZWs 3-5 than in adjacent areas (Supplementary Figs. S4-S5) and evolves only marginally as the grounding-line retreats through this region. In contrast, basal shear stresses are significantly lower in this area and drop further as driving stresses are reduced and the ice stream thins and retreats. The enhanced zone of lateral shear stress is not present in constant-width Experiment C (Supplementary Figs. S2-S3) and thus stabilisations do not occur. Secondly, the principle of mass conservation requires ice-sheet thickening and consequent surface steepening where ice streams narrow which, for a given rate of surface thinning or sea-level change, leads to reduced retreat rates. The steeper surface enhances driving stress and thereby basal stress as suggested for tidewater systems[25] but, in our models, the reduction in basal stress during retreat (Supplementary Fig. S2) is a response to evolving surface geometry rather than a first order control for grounding-line retreat.

While the above two factors explain the mechanism for transient grounding-line stabilisation, the duration of stability is further dependent on the rate of inland-ice delivery to the grounding line. During stabilisations at width constrictions, thinning ceases near the grounding line but continues to propagate inland (Fig. 2) through direct longitudinal stress coupling and a time-transient feedback between accelerated ice flux and thinning[27] (Supplementary Figs. S4-S5). This upstream surface draw-down delivers extra ice to the grounding line which acts to slow the retreat and maintain grounding-line stability. When the inland ice storage is depleted, the next phase of rapid retreat is initiated. The timescale of stabilisation is therefore sensitive to both the volume of upstream ice within the catchment area and the degree of dynamic coupling between the grounding line and the upstream basin. Because upstream propagation of thinning increases with ice speed[27-28], the weak sedimentary bed beneath MBIS[19] explains the tight coupling between grounding-line stability and the upstream basin. In the context of assessing ice-sheet mass balance it is worth noting that even during phases of limited retreat, when short-term measurements would suggest ice-stream stability, mass loss through the grounding line is in reality enhanced significantly by several tens of percent over decades to centuries via drawdown of ice from upstream (Experiment A; Supplementary Figs. S4-S5).

This paper is the first to use a numerical ice-stream model to test and accurately replicate the 'known' dynamic behaviour of a palaeo-ice stream, including temporary stabilisations over timescales longer than decades. Our dynamic simulations provide an improved understanding of the controls of marine ice-stream retreat and show that grounding-lines can be close to stable on a reverse slope. Over and above the influence of bed topography described in the marine ice-sheet instability hypothesis[1-3], grounding-line stability is strongly controlled by subtle variations (e.g. 2-4 km narrowing over 20 km distance) in the width of the ice-stream trough. This effect would be enhanced by additional buttressing in the presence of an ice shelf[2] and would be strongest in basins that are relatively narrow[25] and/or underlain by weak sediments, such as the rapidly thinning Pine Island Glacier in West Antarctica[29].

We also show that ice-stream retreat is expected to be highly non-linear and asynchronous to its forcing. This helps explain temporal and regional variations in the rate of mass loss being observed on present-day ice sheets[6] and why retreat patterns are not regionally synchronous following the LGM[23]. The implications are that where modern grounding lines appear stable, mass loss may be continuing as part of a geometrically-controlled response to a longer-term phase of ice-stream retreat triggered decades to millennia earlier. Furthermore, the importance of width fundamentally questions the widely used assumption that future grounding-lines will retreat catastrophically across reverse-sloping beds[30] and provides a revised perspective on marine ice-stream stability. The lateral control and the longer-term upstream response suggest that interpretations of current rapid changes and future projections of ice-sheet stability should carefully integrate specific long-term ice-stream history and details of 3-dimensional trough geometry.

Figures:

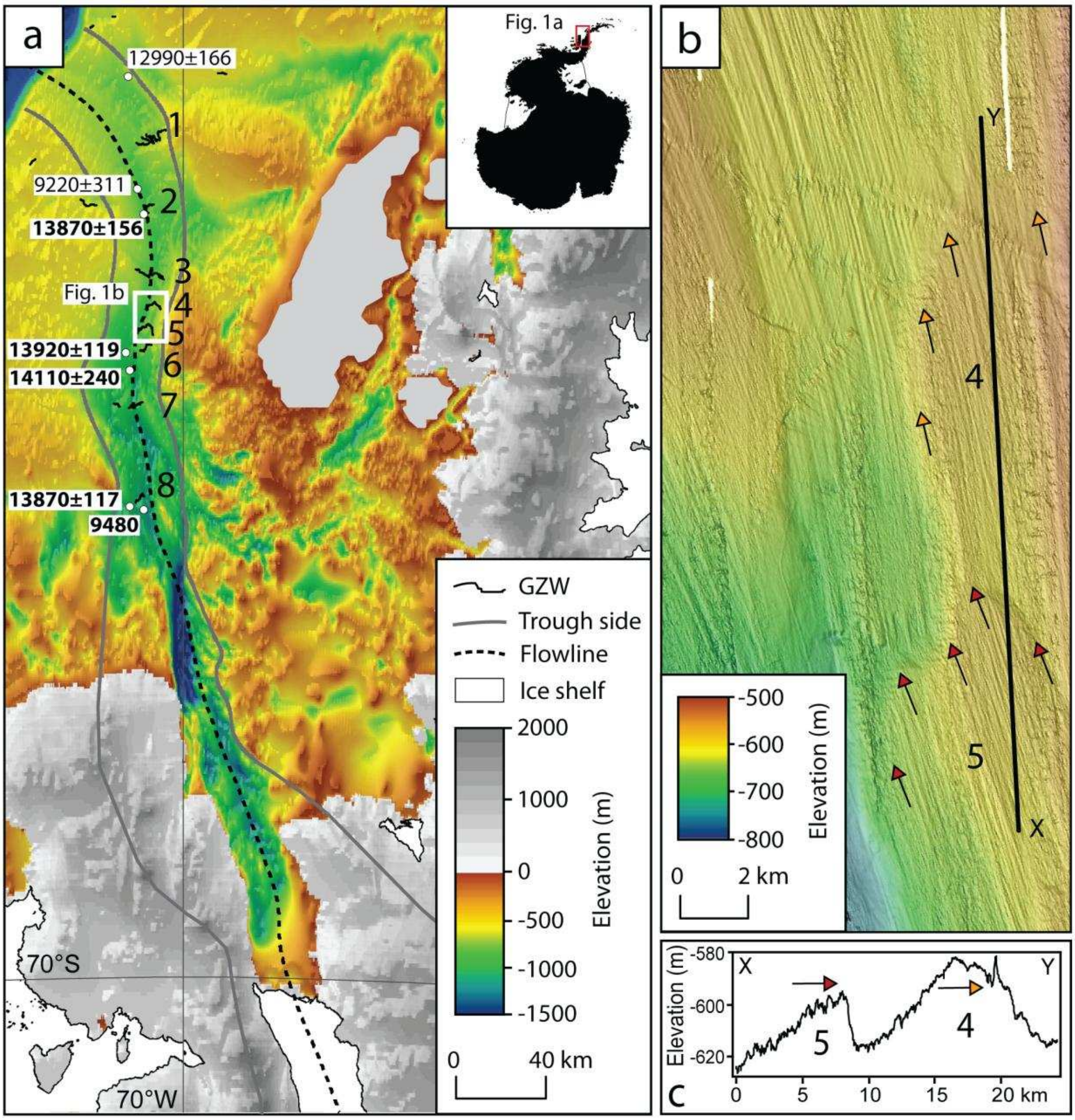


Figure 1: Topographic map and ice-stream landform assemblage in Marguerite Bay. a) Model domain and mapped geophysical retreat constraints. GZWs 1-8 indicate palaeo-grounding line stabilisations. White points show core locations[19,21,23] with corrected and calibrated AMS^{14}C minimum retreat ages in years before present ± 1σ error (bold are the most reliable dates[23]). b) Morphology of GZWs 4 and 5 with arrows indicating their downstream limits and the direction of streaming ice flow determined from superimposed mega-scale glacial lineations. c) Profiles showing scale of GZWs 4 and 5. Arrows indicate ice flow direction during separate depositional phases (red vs. orange).

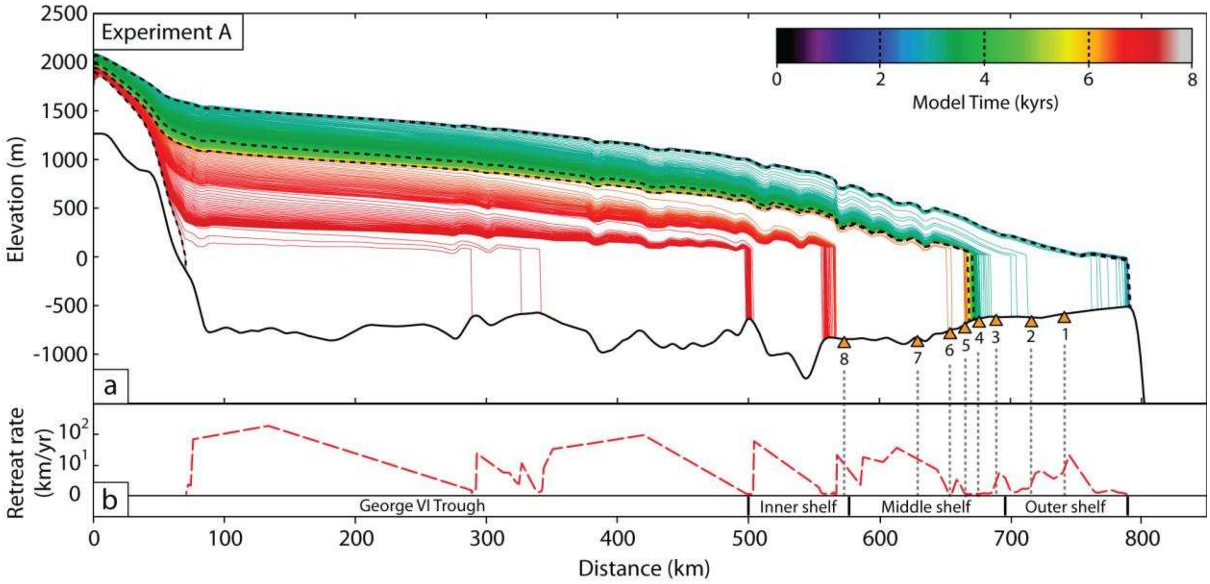


Figure 2: Modelled ice-stream retreat characteristics for reference Experiment A illustrating grounding-line stabilisations on a reverse bed slope. a) Coloured ice surface profiles shown at 5 year intervals over an 8 kyr period with dotted black lines highlighting ice surfaces every 2 kyrs. The continuous black line shows the bed geometry. b) Modelled grounding-line retreat rate (dashed red line) on a logarithmic scale. Orange triangles and grey dashed lines show mapped GZWs 1-8 (Fig. 1). Where a large number of palaeo-ice stream surfaces intersect the bed in panel a, slower retreat is observed in panel b.

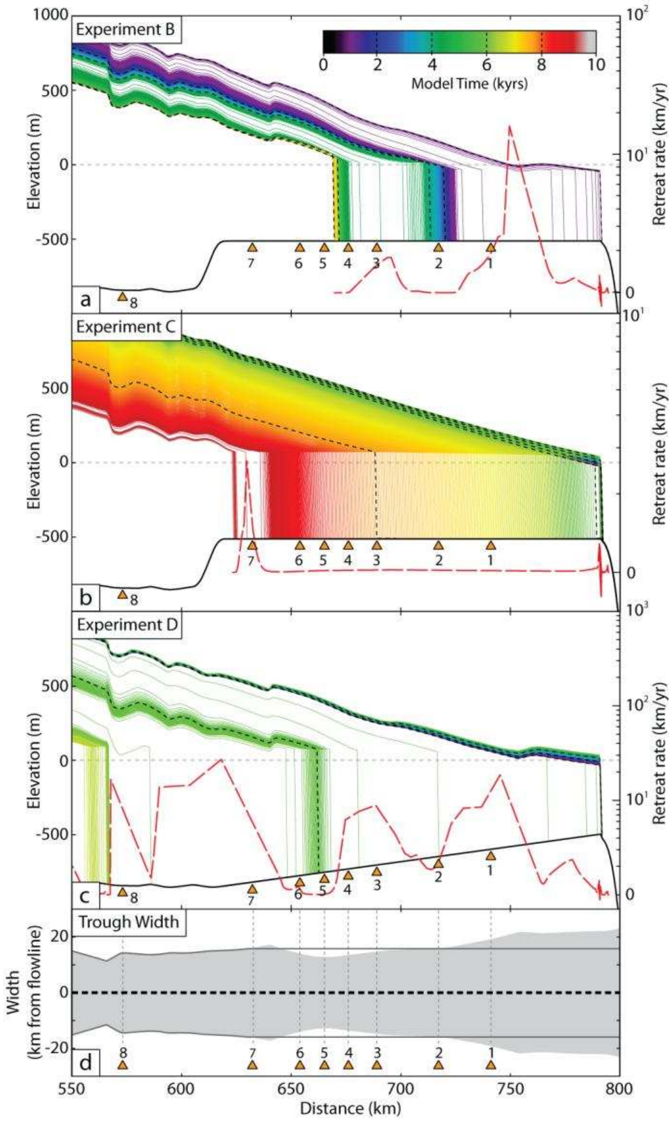


Figure 3: Ice-stream retreat behaviour and imposed geometries for Experiments B-D. a-c) Coloured ice surface profiles at 2 year intervals with dashed black lines every 2 kyrs. Black lines plot imposed bed topographies, dashed red lines show grounding-line retreat rates. Orange triangles represent GZWs 1-8 (Fig. 1). d) Mapped trough width (grey shading) applied in Experiments A, B and D and imposed straightening of trough width (dark grey lines) in Experiment C. Note that retreat onset timings vary because modification of trough width or bed depth alters ice-flux patterns. This does not affect the major finding of controls on retreat.

References:

1 Weertman, J. Stability of the junction of an ice sheet and an ice shelf. J. Glaciol. **13**, 3-11 (1974).

2 Thomas, R. The dynamics of marine ice sheets. J. Glaciol. **24**, 167-177 (1979).

3 Schoof, C. Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. J. Geophys. Res. **112**, F03S28 (2007).

4 Joughin, I., Smith, B. E. & Holland, D. M. Sensitivity of 21st century sea level to ocean-induced thinning of Pine Island Glacier, Antarctica. Geophys. Res. Lett. **37**, L20502 (2010).

5 Rignot, E. Recent Antarctic ice mass loss from radar interferometry and regional climate modelling. Nature Geosci. **1**, 106-110 (2008).

6 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 (2009).

7 Price, S. F., Payne, A. J., Howat, I. M. & Smith, B. E. Committed sea-level rise for the next century from Greenland ice sheet dynamics during the past decade. Proc. Natl. Acad. Sci. **108**, 8978-8983 (2011).

8 Joughin, I. & Alley, R. B. Stability of the West Antarctic ice sheet in a warming world. Nature Geosci. **4**, 506-513 (2011).

9 Mercer, J. H. West Antarctic ice sheet and CO_2 greenhouse effect: a threat of disaster. Nature **271**, 321-325 (1979).

10 Powell, R. D. in Glacial Marine Sedimentation: Paleoclimatic Significance Vol. 261 eds J.B. Anderson & G.M. Ashley) 75-94 (Geologcal Society of America, Special Paper, 1991).

11 Independent Panel on Climate Change (IPCC): Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA (2007).

12 Pattyn, F. et al. Results of the Marine Ice Sheet Model Intercomparison Project, MISMIP. The Cryosphere **6**, 573-588 (2012).

13 Pollard, D. & DeConto, R. Modelling West Antarctic ice sheet growth and collapse through the past five million years. Nature 458, 329-332 (2009).

14 Hindmarsh, R. C. A. The role of membrane-like stresses in determining the stability and sensitivity of the Antarctic ice sheets: back pressure and grounding line motion. Phil. Trans. Roy. Soc. A **364**, 1733-1767 (2006).

15 Vieli, A. & Payne, A. J. Assessing the ability of numerical ice sheet models to simulate grounding line migration. J. Geophys. Res. **110**, F01003 (2005).

16 Goldberg, D., Holland, D. M. & Schoof, C. Grounding line movement and ice shelf buttressing in marine ice sheets. J. Geophys. Res. **114**, F04026 (2009).

17 Katz, R. F. & Worster, M. G. Stability of ice-sheet grounding lines. Proc. Roy. Soc. A **466**, 1597-1620 (2010).

18 Ó Cofaigh, C., Pudsey, C. J., Dowdeswell, J. A. & Morris, P. Evolution of subglacial bedforms along a paleo-ice stream, Antarctic Peninsula continental shelf. Geophys. Res. Lett. **29**, 1199 (2002).

19 Ó Cofaigh, C. et al. Flow dynamics and till genesis associated with a marine-based Antarctic palaeo-ice stream. Quat. Sci. Rev. **24**, 709-740 (2005).

20 Ó Cofaigh, C., Dowdeswell, J. A., Evans, J. & Larter, R. D. Geological constraints on Antarctic palaeo-ice-stream retreat. Earth Surf. Proc. Land. **33**, 513-525 (2008).

21 Kilfeather, A. A. et al. Ice-stream retreat and ice-shelf history in Marguerite Trough, Antarctic Peninsula: Sedimentological and foraminiferal signatures. Geol. Soc. Amer. Bull. **123**, 997-1015 (2011).

22 Dowdeswell, J. A., Ottesen, D., Evans, J., Cofaigh, C. Ó. & Anderson, J. B. Submarine glacial landforms and rates of ice-stream collapse. Geology **36**, 819-822 (2008).

23 Livingstone, S. J. et al. Antarctic palaeo-ice streams. Earth Sci. Rev. **111**, 90-128 (2012).

24 Nick, F. M., Vieli, A., Howat, I. M. & Joughin, I. Large-scale changes in Greenland outlet glacier dynamics triggered at the terminus. Nature Geosci. **2**, 110-114 (2009).
25 O'Neel, S., Pfeffer, W.T., Krimmel, R., Meier, M. Evolving force balance at Columbia Glacier, Alaska, during its rapid retreat. J. Geophys. Res. 110, F03012 (2005).
26 Whillans, I. M. & van der Veen, C. J. The role of lateral drag in the dynamics of Ice Stream B, Antarctica. J. Glaciol. **43**, 231-237 (1997).
27 Payne, A. J., Vieli, A., Shepherd, A. P., Wingham, D. J. & Rignot, E. Recent dramatic thinning of largest West Antarctic ice stream triggered by oceans. Geophys. Res. Lett. **31**, L23401 (2004).
28 Price, S. F., Conway, H., Waddington, E. D. & Bindschadler, R. A. Model investigations of inland migration of fast-flowing outlet glaciers and ice streams. J. Glaciol. **54**, 49-60 (2008).
29 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).
30 Ross, N. et al. Steep reverse bed slope at the grounding line of the Weddell Sea sector in West Antarctica. Nature Geosci. **5**, 393-396 (2012).

Corresponding author:

Please send correspondence to Stewart S. R. Jamieson: Stewart.Jamieson@durham.ac.uk

Acknowledgements:

We thank Tad Pfeffer, Tony Payne and Richard Hindmarsh for constructive comments. This project was funded by the Natural Environment Research Council, UK, grant no. NE/G015430/1.

Contributions:

SSRJ and AV contributed equally to this work. SSRJ completed the numerical modelling and extended the model provided by AV. SSRJ wrote the first draft and all authors contributed to the analysis, interpretation and the writing of the paper. AV, CS, CÓC and CDH conceived the idea and designed the research. SJL carried out the mapping. CDH and JAD contributed chronological and geophysical data.

Supplementary Information

Model

The numerical model calculates the time-dependent evolution of ice flow, ice surface and internal stresses along an 800 km flowband of the Marguerite Bay palaeo ice stream in the western Antarctic Peninsula. The physics incorporated within the model are described extensively elsewhere[1-3] but their key features, in particular those relating to ensuring robust grounding-line behaviour, are outlined herein.

Calculations of ice flow balance driving stress τ_d against basal stress τ_b and lateral shear stress τ_{lat}, and against longitudinal stress gradients $\frac{\partial \tau_{xx}}{\partial x}$ in the direction of ice flow x are given by:

$$\frac{\partial \tau_{xx}}{\partial x} + \tau_b + \tau_{lat} = \tau_d \qquad (1)$$

For an ice stream of thickness H, half-width W and a Weertman-type non-linear sliding relation[4] that takes into account effective pressure N, the stress balance above (eqn. (1)) results in an equation for depth and width averaged ice flow u of the following form:

$$2\frac{\partial}{\partial x}\left(H\nu\frac{\partial u}{\partial x}\right) - \beta\left(\frac{u}{N}\right)^{1/m} + \frac{H}{W}\left(\frac{5u}{2AW}\right)^{1/n} = \rho_i g H \frac{\partial S}{\partial x} \qquad (2)$$

where the constants ρ_i and g refer to the density of ice (910 kg m^{-3}) and gravitational acceleration, S is the ice surface, A the rate factor, β the basal sliding coefficient , n and m the exponents for ice flow and sliding relations (both taken as 3). The effective viscosity ν is given by

$$\nu = A^{-1/n}\left|\frac{\partial u}{\partial x}\right|^{\frac{1-n}{n}} \qquad (3)$$

and is solved iteratively. The evolution of the ice surface explicitly accounts for the along-flow variation in ice stream width as determined by the glacial trough shape by using

$$\frac{\partial H}{\partial t} = a - \frac{1}{W}\frac{\partial (uHW)}{\partial x} \qquad (4)$$

The model allows time-dependent inputs for surface accumulation a, sea level, and flow rate factor A. The latter describes the rheological softness of the ice at a particular temperature. The ice body is isothermal during individual timesteps, but its temperature can be varied through time.

The model has a robust treatment of grounding-line behaviour[5] that is consistent with a boundary layer theory[6] which is important for avoiding errors imposed by model numerics[2]. Our treatment of grounding-line evolution relies on three components to allow the grounding line to move freely:

1) Calving front: A floatation criterion is applied to calculate the location of the grounding line for each time step. All ice beyond this location is calved off and therefore the model does not incorporate an ice shelf or its buttressing effects. Ignoring an ice shelf takes into account the lack of a robust calving model for floating ice-shelf termini but also the highly limited observational constraints on palaeo ice-shelf extent. Ignoring buttressing from an ice shelf means the grounding line is in general less stable and thus our model-based sensitivity analysis considers the most unstable case.

2) Moving spatial grid: The grounding line is tracked continuously using a stretched grid to allow precision without a dependency upon a specific grid resolution. This enables the model to more accurately evolve to changes at the ice-ocean boundary[2,5]. Initial grid resolution is ca. 900 m but reduces as the grounding line retreats inland.

3) Ice-ocean boundary stresses: We assume the longitudinal stress at the ocean boundary is balanced by the difference in hydrostatic pressure between the ice and ocean water [2,7]. This results in a boundary condition for the velocity gradient at the ice-stream front given by

$$\left.\frac{\partial u}{\partial x}\right|_{\text{Front}} = A\left[\frac{\rho_i g}{4}\left(1-\frac{\rho_i}{\rho_w}\right)\right]^n H^n \qquad (5)$$

where the density of the ocean water ρ_w is 1028 kg m^{-3}.

Model Constraints

Topographic, Basal and Climatic Boundary Conditions

Bed topography is constructed from 3 elevation datasets. High-resolution swath bathymetry data collected on the outer shelf were merged with a recent data 1 km resolution bathymetry compilation for the Bellingshausen Sea and Marguerite Bay[8] to produce a seamless digital elevation model of

Marguerite Trough and the area underneath the present-day George VI Ice Shelf. Above sea level in the upper reaches of the ice stream catchment the ALBMAP v1 dataset[9] is used. The bed is re-sampled to an initial resolution of 900 m and has the grounding-zone wedges (GZWs) removed to avoid pre-conditioning the grounding-line retreat pattern. The bed depth used in the model follows a centerline along the main portion of Marguerite Trough which then continues into George VI Sound and rises above sea level to the ice divide of central Palmer Land.

The glacial landform signature of Marguerite Bay reveals the composite extent of the palaeo-ice stream and is therefore a key constraint and allowed us to define its shape. These data are used to define the domain and basal conditions for the numerical model. The width of the modelled ice stream is defined by tracing the shoulder of the overdeepened trough in a manner consistent with the lateral extent and axis of a range of elongated glacial landforms delineating past ice flow that were mapped across the continental shelf. Glacial features were systematically mapped from high-resolution marine geophysical data of the continental shelf collected during James Clark Ross surveys JR59 [10], JR71[11-12] and JR157[13] and Nathaniel B. Palmer survey NBP0201[14-16]. Features included 5037 mega-scale glacial lineations which are diagnostic of rapidly flowing ice[17], 2761 drumlins, 452 crag and tail features, and 4000 locations at which bedrock surfaces are streamlined and gouged[18].

Sediment cores recovered from the outer shelf of Marguerite Bay demonstrate a weak bed unable to support high shear strengths during the presence of the ice stream. Measured shear strengths of subglacial till cored on the mid- to outer continental shelf are often below 10 kPa on the outer continental shelf[19]. The basal friction parameter in the model is prescribed as slippiest where the bed is below sea level, and becomes stiffer as the bed rises above 0 m. Below sea level, modelled basal friction is linked to effective pressure so that basal shear stresses remain low in the model in order to correspond with measurements (Figure S1).

Accumulation within this defined basin is from Arthern et al.[20] via the ALBMAP Antarctic dataset[9]. Mapping and bathymetric data also indicate that when ice reached the outer shelf, the outer part of Marguerite Bay palaeo-ice stream was fed by a tributary which originated on the northern edge of Palmer Land. To account for this additional mass input we integrate the total yearly accumulation experienced from this tributary basin[9,20]. This mass is added as accumulation over a 160 km section of the modelled flowline in a pattern assuming more ice would have entered via the deeper central portion of the tributary than at its shallower edges. Sensitivity tests which alternatively distributed this mass along the entire length of the ice stream indicate that the pattern of grounding-line retreat is insensitive to the location at which the mass was injected.

Constraints on Retreat Pattern

Grounding zone wedges are used as the key independent constraint for testing the ability of the model to reproduce grounding line retreat patterns in Marguerite Bay. These sedimentary wedges form at the grounding line and are interpreted as representing a period of relatively stable ice stream position[21-22] during which there is time to deposit significant volumes of sediment at the grounding line. Ten GZWs were identified in the outer and mid-shelf of Marguerite Bay using the swath bathymetry data although only 8 of these fall within the limits of the modelled ice stream extents. On the mid shelf, GZWs are mapped in the center of the trough and directly in the path of the modelled flowline. Towards the outer continental shelf, the GZWs are found on the lateral margins of the zone of fast-flowing ice as interpreted from the mega-scale glacial lineations. Importantly, and as indicated above, the topographic expression of the GZWs is removed from the bed topography and therefore does not pre-condition the modelled retreat pattern. Subtle shifts in ice-flow direction across the GZWs indicate minor re-organisations of ice-flow during these temporary stabilisations (Figure 1b). Moreover, the presence of these overlying lineations also indicates that at the time of deposition, ice was flowing rapidly at the grounding zone.

Experimental Design and Discussion

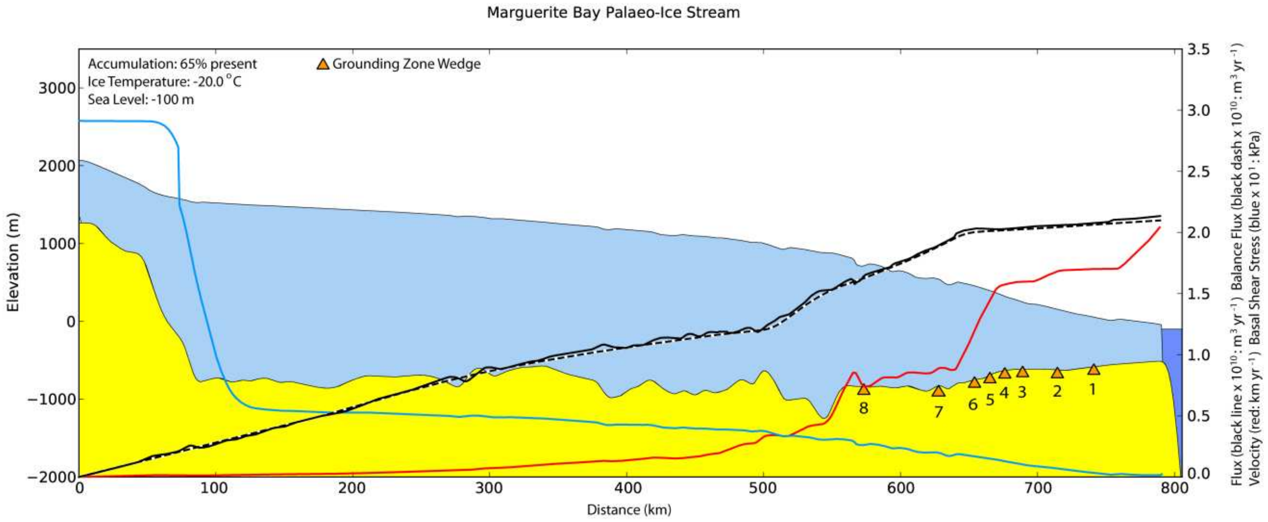
Our experiments are designed to understand how any internal factors may control grounding-line retreat patterns. We do not aim to reproduce the timing of retreat observed in Marguerite Bay, but merely wish to carry out sensitivity tests to understand controls governing the pattern of ice-stream retreat. The objective is to simulate a retreat pattern that is consistent with the glacial geomorphology, so that it will pause at, or near, the eight independently mapped GZWs, and in particular near those on the reverse-sloping portion of the bed. Changes in climate and sea level are used to initiate and drive retreat in our experiments but in order to avoid overprinting any climatic or sea-level controls on retreat pattern we impose simple linear rates of change for these parameters. The result is a system whose retreat rate can be disentangled from external forcing patterns.

The initial ice surface is consistent with previous geologically constrained reconstructions of the Last Glacial Maximum (LGM) Antarctic Peninsula Ice Sheet[23-24] and the grounding line is situated on the continental shelf edge as dictated by the bathymetric mapping. This initial configuration is in ‘steady-state’ and results in a modelled ice flux that agrees with the calculated balance flux given by the pattern of accumulation within the basin. Ice velocities stabilise at a maximum of 2 km yr^{-1} at the grounding line (Figure S1). Over the next 8 kyrs of model time, ice temperature is increased from -20 to -15 °C to soften the ice rheology over time, accumulation rates are increased from 65 %

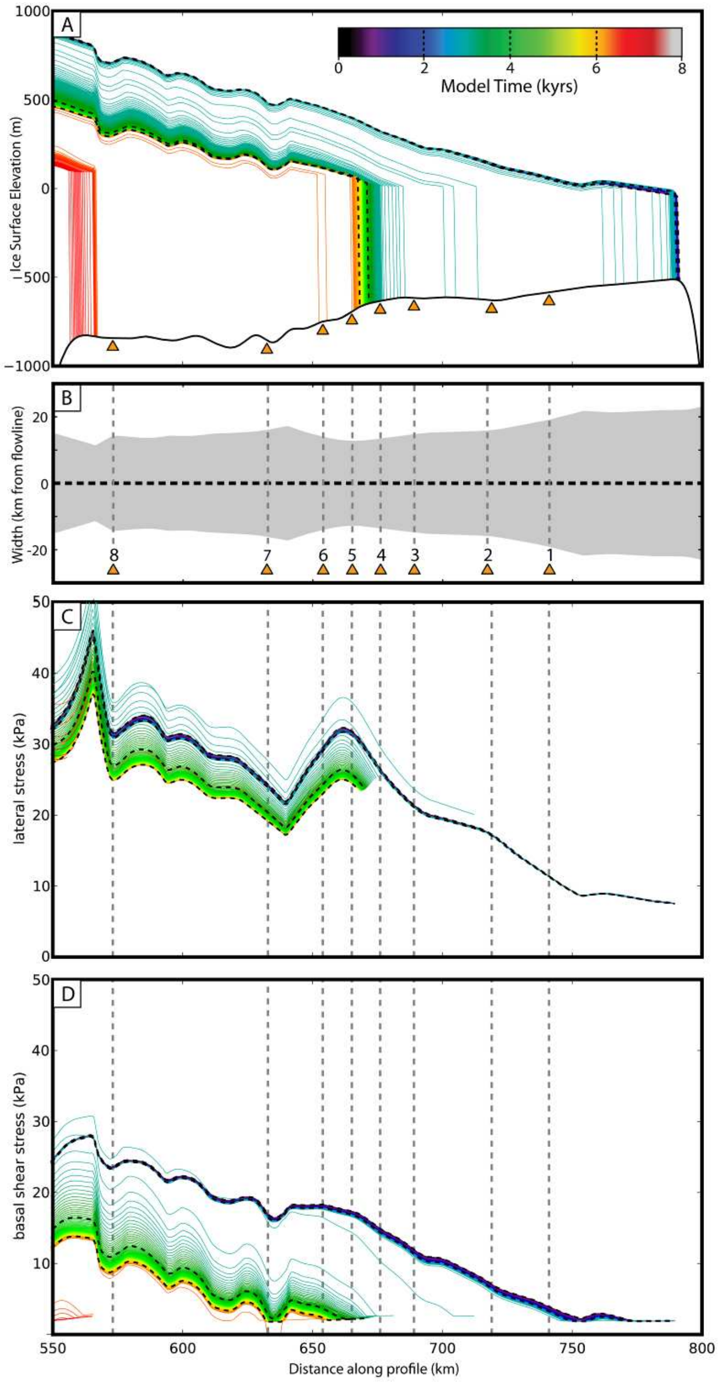
to 100 % of present day[20], and sea level is increased from -100 m to 0 m. These values represent approximations of environmental change between the LGM and present day in the Marguerite Bay area. A 2 kyr stabilisation period where all parameters are held constant is simulated at the end of each experiment. Given uncertainties about climate and sea level change, we test different rates of linear forcing but find that no significant differences in the pattern of grounding-line retreat or stability result from applying more rapid forcing. Experiment A is the baseline model from which we then systematically modify the width and bed geometries through Experiments B-D to test the hypothesis that internal geometric effects are important controls on the retreat patterns of ice streams. Supplementary Video SV1 shows the evolution of Marguerite Bay palaeo-ice stream along with its ice flux and ice velocity during Experiment A (described in the main text) and illustrates the temporary stabilization of the grounding line on the reverse sloping bed. Figures S4 and S5 show ice flux and ice surface elevation at a range of locations along the ice stream during the main period of grounding-line stabilization.

The evolution of basal (τ_b) and lateral (τ_{lat}) shear stresses in the modelled stress balance (Eqn. 1 and 2) in reference Experiment A are shown in Figure S2, indicating the changes in force balance through time. Changes in basal stress are most strongly related to ice thickness and driving stress (τ_d) and are systematically low. The lateral stress term is highly variable along flow, and is consistently and significantly elevated between GZWs 4-6 where the width is narrowest. The initial basal (τ_b) and lateral (τ_{lat}) stress conditions in Models A-D are compared in Figure S3 and demonstrate how the relationships between basal topography and basal stress, and lateral topography and lateral stress change in our sensitivity tests.

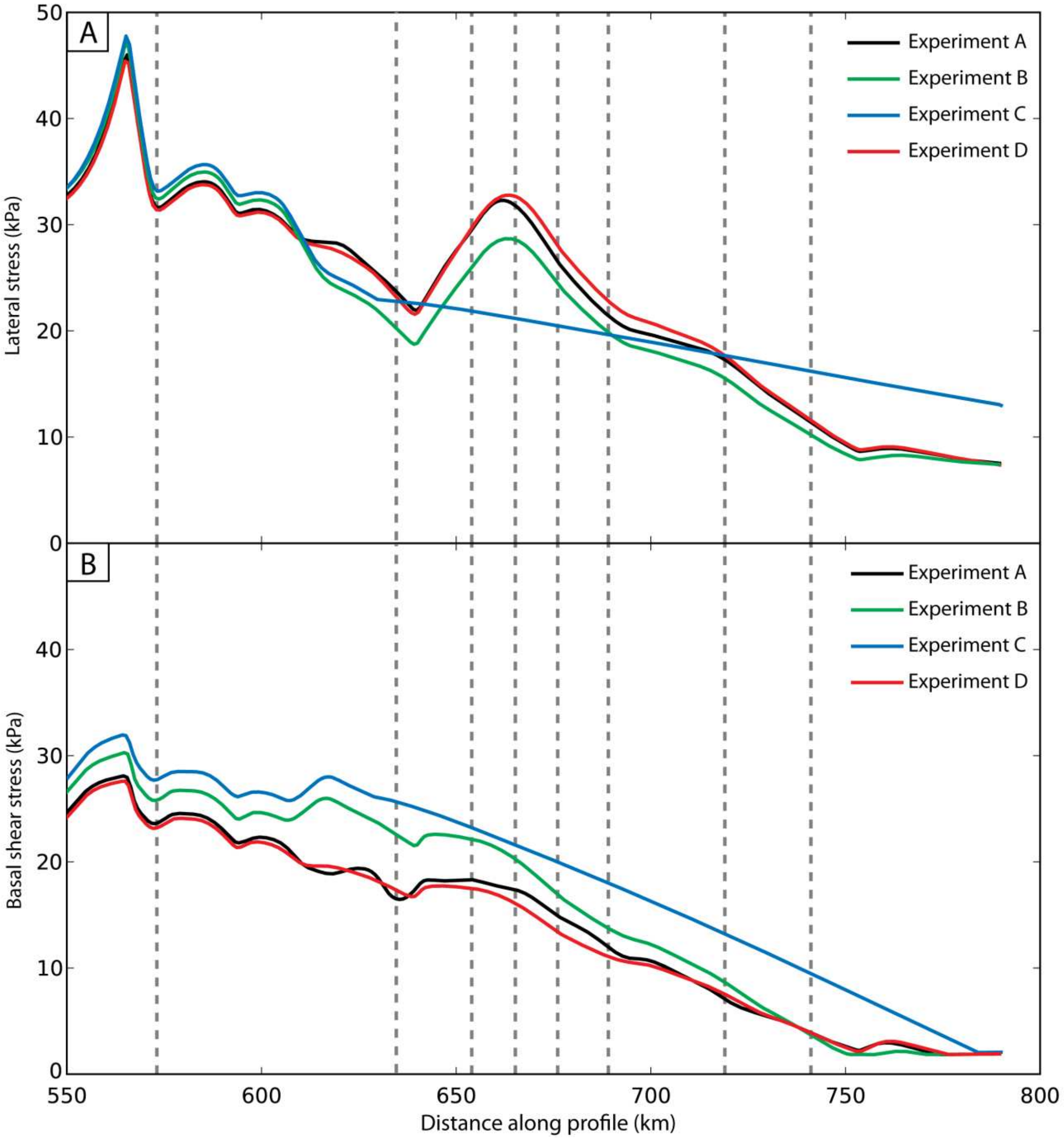
Figures:



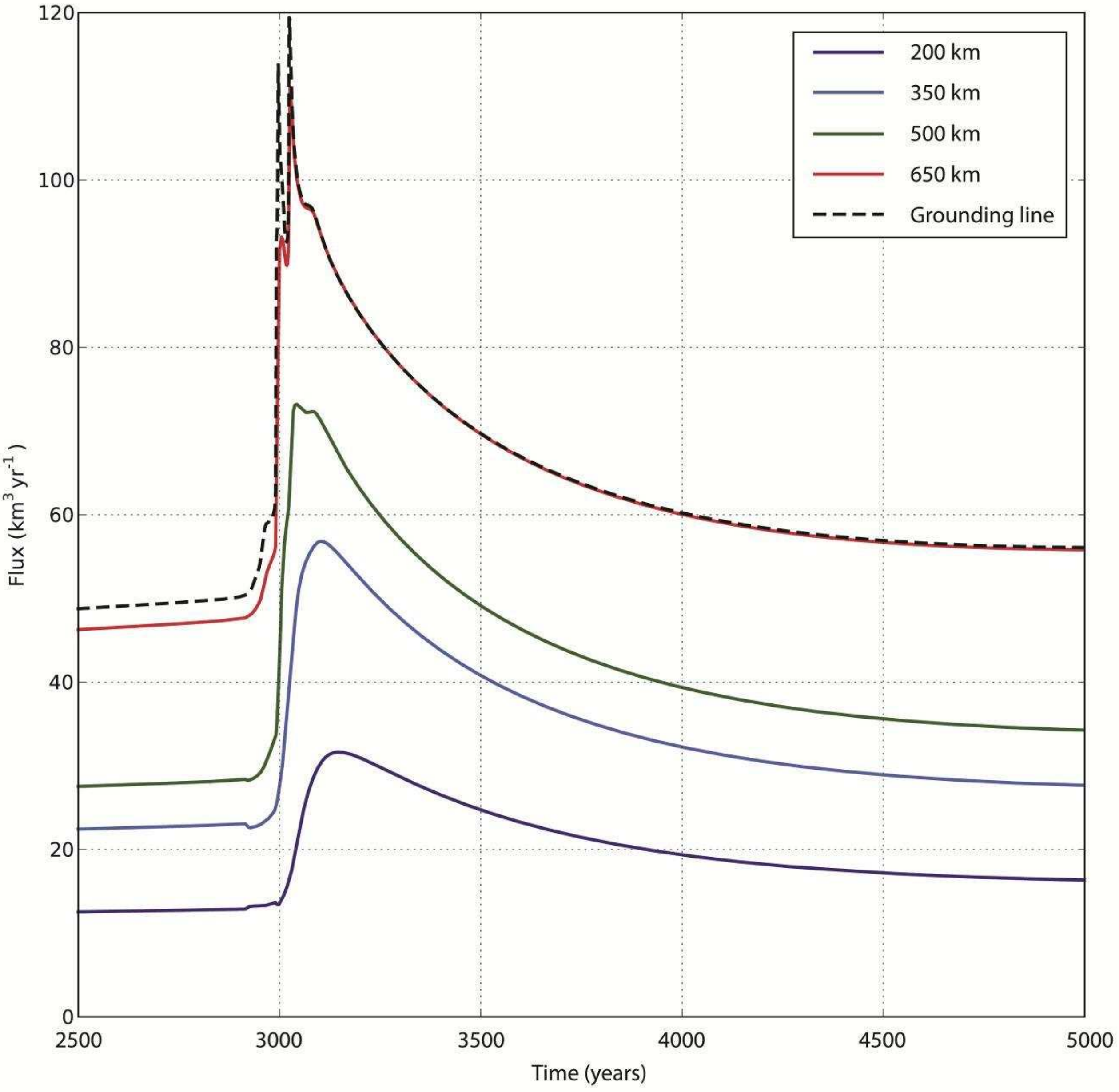
Supplementary Figure S1: Initial model conditions for experiments A-D equivalent to LGM configuration with the grounding line at the continental shelf edge (blue area shows ice body; yellow area shows bed, dark blue area shows ocean), rapid ice discharge into the ocean, and low basal shear stresses (blue line) in the submarine trough. Note that ice flux (solid black line) fits closely with the balance flux (integrated upstream accumulation; dashed black line) indicating that the ice stream is in a steady state and is flowing rapidly (red line shows ice velocity) prior to the beginning of the retreat experiments.



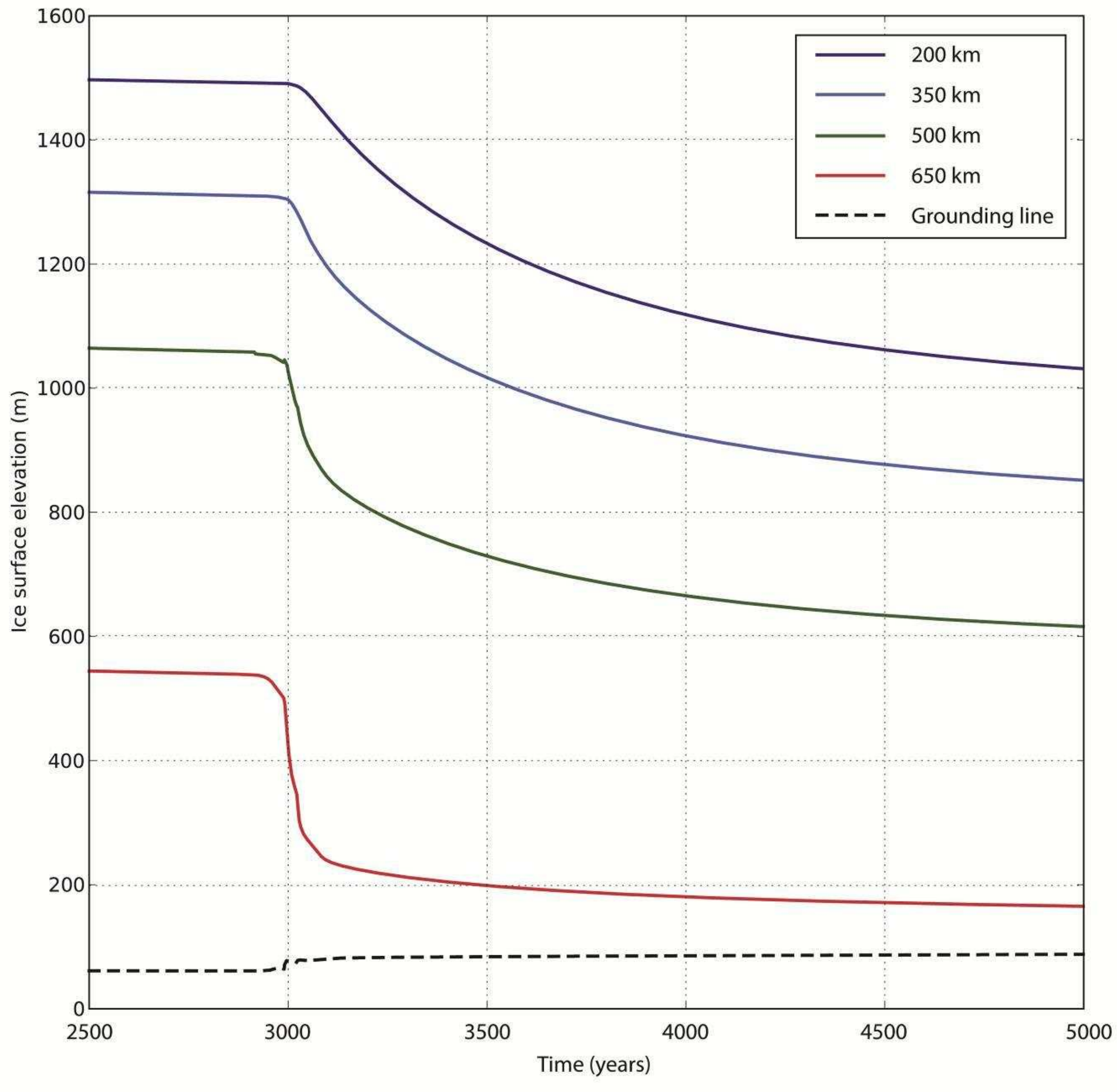
Supplementary Figure S2: Modelled evolution of ice surface (A), trough width (B) and lateral (C) and (D) basal shear stress for reference Experiment A. Panel A shows coloured ice surface profiles every 2 years with dashed black lines overprinted every 2 kyrs. Orange triangles indicate the locations of GZWs 1-8 (Figure 1) and correspond to dashed grey lines in all other panels. Panel B indicates the natural trough width used in Experiment A, B and D (grey shading) and the locations of the GZWs (grey dashed line, orange triangles) and the model flowline location (black dashed line). Coloured lines in panel C and D show the modelled evolution of lateral and basal shear stresses respectively (see Eqn. 1 and 2) over time in 50 year intervals.



Supplementary Figure S3: Initial (LGM) stress conditions for experiments A to D. Panel A shows the lateral stress term and panel B shows the basal shear stresses as modelled from Equations (1) and (2). Dashed grey lines indicate the location of GZWs 1-8 (see Figs. 1 and S2).



Supplementary Figure S4: Modelled ice flux at selected locations along the ice stream (see legend: distances are measured from ice divide) during main period of reverse-bed stabilization in Experiment A where the grounding line is between GZWs 4 and 5. This shows 1) that there is a delayed but significant response from the upstream ice flux, 2) that even after centuries, ice flux remains out of balance, and 3) that the ice flux does not return to pre-retreat values, although this is partly due to the increase in ice temperature and sea level during the period of stability.



Supplementary Figure S5: Modelled ice surface elevation at selected locations along the ice stream (see legend: distances are measured from ice divide) during main period of reverse-bed stabilisation in Experiment A where the grounding line is between GZWs 4 and 5. This shows that 1) the ice surface adjacent to the grounding line adjusts rapidly, but the response is damped in the upstream area, and 2) that the overall magnitude of thinning is more significant in the areas furthest away from the grounding line, thus illustrating the temporal and spatial scale of upstream adjustment when the grounding line stabilises in our experiments.

Caption for Supplementary Video 1 (JamiesonEtAlVideoSV1.mov): Modelled palaeo-ice stream retreat showing the outcome of Experiment A (see main text). The model starts from the initial condition illustrated in Figure S1 and the evolution of the forcing parameters are shown in the top left. Retreat is driven by linearly forcing sea level rise and ice temperature. The black line shows ice flux, and the red line indicates ice velocity. Note the nonlinear response to the linear forcing with retreat rate slowdowns occurring on the reverse bed slope.

References:

1 Nick, F. M., Vieli, A., Howat, I. M. & Joughin, I. Large-scale changes in Greenland outlet glacier dynamics triggered at the terminus. Nature Geosci. **2**, 110-114(2009).

2 Vieli, A. & Payne, A. J. Assessing the ability of numerical ice sheet models to simulate grounding line migration. J. Geophys. Res. **110**, F01003 (2005).

3 Nick, F. M., Van Der Veen, C. J., Vieli, A. & Benn, D. I. A physically based calving model applied to marine outlet glaciers and implications for the glacier dynamics. J. Glaciol. **56**, 781-794 (2010).

4 Weertman, J. On the sliding of glaciers. J. Glaciol. **3**, 33-38 (1957).

5 Pattyn, F. et al. Results of the Marine Ice Sheet Model Intercomparison Project, MISMIP. The Cryosphere **6**, 573-588 (2012).

6 Schoof, C. Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. J. Geophys. Res. **112**, F03S28 (2007).

7 Hindmarsh, R. C. A. The role of membrane-like stresses in determining the stability and sensitivity of the Antarctic ice sheets: back pressure and grounding line motion. Phil. Trans. Roy. Soc. A **364**, 1733-1767 (2006).

8 Graham, A. G. C., Nitsche, F. O. & Larter, R. D. An improved bathymetry compilation for the Bellingshausen Sea, Antarctica, to inform ice-sheet and ocean models. The Cryosphere **5**, 95-106 (2011).

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

10 Ó Cofaigh, C., Pudsey, C. J., Dowdeswell, J. A. & Morris, P. Evolution of subglacial bedforms along a paleo-ice stream, Antarctic Peninsula continental shelf. Geophys. Res. Lett. **29**, 1199 (2002).

11 Dowdeswell, J. A., Ó Cofaigh, C. & Pudsey, C. J. Continental slope morphology and sedimentary processes at the mouth of an Antarctic palaeo-ice stream. Marine Geol. **204**, 203-214 (2004).

12 Dowdeswell, J. A., Ó Cofaigh, C. & Pudsey, C. J. Thickness and extent of the subglacial till layer beneath an Antarctic palaeo-ice stream. Geology **32**, 13-16 (2004).

13 Noormets, R., Dowdeswell, J. A., Larter, R. D., Ó Cofaigh, C. & Evans, J. Morphology of the upper continental slope in the Bellingshausen and Amundsen Seas – Implications for sedimentary processes at the shelf edge of West Antarctica. Marine Geol. **258**, 100-114 (2009).

14 Anderson, J. B. & Fretwell, L. O. Geomorphology of the onset area of a paleo-ice stream, Marguerite Bay, Antarctic Peninsula. Earth Surf. Proc. Land. **33**, 503-512 (2008).

15 Pope, P. G. & Anderson, J. B. Late Quaternary glacial history of the northern Antarctic Peninsula’s western continental shelf: evidence from the marine record. Antarctic Research

Series **57**, 63-91 (1992).
16 Heroy, D. C. & Anderson, J. B. Ice-sheet extent of the Antarctic Peninsula region during the Last Glacial Maximum (LGM) - Insights from glacial geomorphology. Geol. Soc. Amer. Bull. **117**, 1497-1512 (2005).
17 Stokes, C. R. & Clark, C. Geomorphological criteria for identifying Pleistocene ice streams. Ann Glaciol. **28**, 67-74 (1999).
18 Livingstone, S. J. et al. in Fall Meeting, American Geophysical Union (San Francisco, 2010).
19 Ó Cofaigh, C. et al. Flow dynamics and till genesis associated with a marine-based Antarctic palaeo-ice stream. Quat. Sci. Rev. **24**, 709-740 (2005).
20 Arthern, R. J., Winebrenner, D. P. & Vaughan, D. G. Antarctic snow accumulation mapped using polarization of 4.3-cm wavelength microwave emission. J. Geophys. Res. **111**, D06107 (2006).
21 Dowdeswell, J. A., Ottesen, D., Evans, J., Cofaigh, C. Ó. & Anderson, J. B. Submarine glacial landforms and rates of ice-stream collapse. Geology **36**, 819-8221 (2008).
22 Dowdeswell, J. A. & Fugelli, E. M. G. The seismic architecture and geometry of grounding-zone wedges formed at the marine margins of past ice sheets. Geol. Soc. Amer. Bull. (In press).
23 Bentley, M. J., Fogwill, C. J., Kubik, P. W. & Sugden, D. E. Geomorphological evidence and cosmogenic Be-10/Al-26 exposure ages for the Last Glacial Maximum and deglaciation of the Antarctic Peninsula Ice Sheet. Geol. Soc. Amer. Bull. **118**, 1149-1159 (2006).
24 Bentley, M. J. & Anderson, J. B. Glacial and marine geological evidence for the ice sheet configuration in the Weddell Sea-Antarctic Peninsula region during the Last Glacial Maximum. Antarctic Sci. **10**, 309-325 (1998).