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Ambiguous stability of glaciers at bed peaks

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Abstract

Increasing ice flux from glaciers retreating over deepening (retrograde) bed topography has been implicated in the recent acceleration of mass loss from the Greenland and Antarctic ice sheets. We show in observations that some glaciers have remained at peaks in bed topography without retreating despite enduring significant changes in climate. Observations also indicate that some glaciers which persist at bed peaks undergo sudden retreat years or decades after the onset of local ocean or atmospheric warming. Using model simulations, we show that persistence of a glacier at a bed peak is caused by ice slowing as it flows up a reverse-sloping bed to the peak. Persistence at bed peaks may lead to two very different future behaviors for a glacier: one where it persists at a bed peak indefinitely, and another where it retreats from the bed peak after potentially long delays following climate forcing. However, it is nearly impossible to distinguish which of these two future behaviors will occur from current observations. We conclude that inferring glacier stability from observations of persistence obscures our true commitment to future sea-level rise under climate change. We recommend that further research is needed on seemingly stable glaciers to determine their likely future.

1. Introduction

Mass loss from the Greenland and Antarctic ice sheets has accelerated in recent decades, driven by increasing surface melt and discharge of ice from glaciers (Shepherd and others, 2018; Mouginot and others, 2019). The increase in glacier discharge is driven, in part, by glacier retreat over reverse-sloping bed topography (i.e. deepening toward the ice-sheet interior), which may initiate a positive feedback known as the 'marine ice-sheet instability' (Weertman, 1974). However, the climate forcing needed to initiate this positive feedback depends on a range of other processes including ice shelf buttressing and subglacial friction (Gudmundsson and others, 2012; Robel and others, 2016; Haseloff and Sergienko, 2018; Pegler, 2018; Sergienko and Wingham, 2019), which are not well represented in many theories of marine ice-sheet stability or in complex ice-sheet models used to project future ice-sheet changes. In particular, bed topography that fluctuates on length scales less than hundreds of kilometers leads to behaviors that are not accurately predicted using classical theories of marine ice-sheet stability (Sergienko and Wingham, 2021).

The response of glaciers to atmospheric and oceanic forcing is heterogeneous across neighboring catchments, partly due to the influence of underlying bed topography. As observations of subglacial bed topography and glacier retreat have improved, we are learning that bed topography is bumpy at a wide range of length scales (Jordan and others, 2017; Morlighem and others, 2017, 2020) and that many glaciers in Greenland and Antarctica have undergone large retreats in the past century (Tinto and Bell, 2011; Smith and others, 2017; Catania and others, 2018). Still, many glaciers have not retreated during the observational era, even while nearby glaciers have retreated in response to regional warming of the ocean and atmosphere. Geological evidence from Thwaites Glacier, West Antarctica recorded the persistence of the grounding line at a bed peak for hundreds to thousands of years (Tinto and Bell, 2011), even amidst significant fluctuations in ocean temperatures (Hillenbrand and others, 2017). Nearby, observations show that Pine Island Glacier persisted at a bed peak until the 1970s, even though regional warming of the ocean began in the 1940s (Smith and others, 2017). As we will discuss further in the next section, large portions of the Greenland coast have also been subject to incursions of warm ocean water, though different glaciers have responded to these incursions in different ways (Catania and others, 2018).

Here we demonstrate both observationally and using model simulations that retreating marine-terminating glaciers may pause at bed peaks for prolonged time periods even while the glacier continues to lose mass in response to a current or previous climate forcing. The persistence of glaciers at bed peaks ultimately leads to one of two very different future behaviors: one in which the glacier terminus continues to persist at the bed peak without losing mass, and another where retreat occurs suddenly without a concurrent change in climate and leads to a significant acceleration in mass loss. However, it is difficult to distinguish which of these two possible future behaviors will occur from current observations of persistent glaciers. Ultimately, this ambiguous behavior of seemingly 'stable' glaciers obscures the true commitment to future sea level rise under anthropogenic climate change.

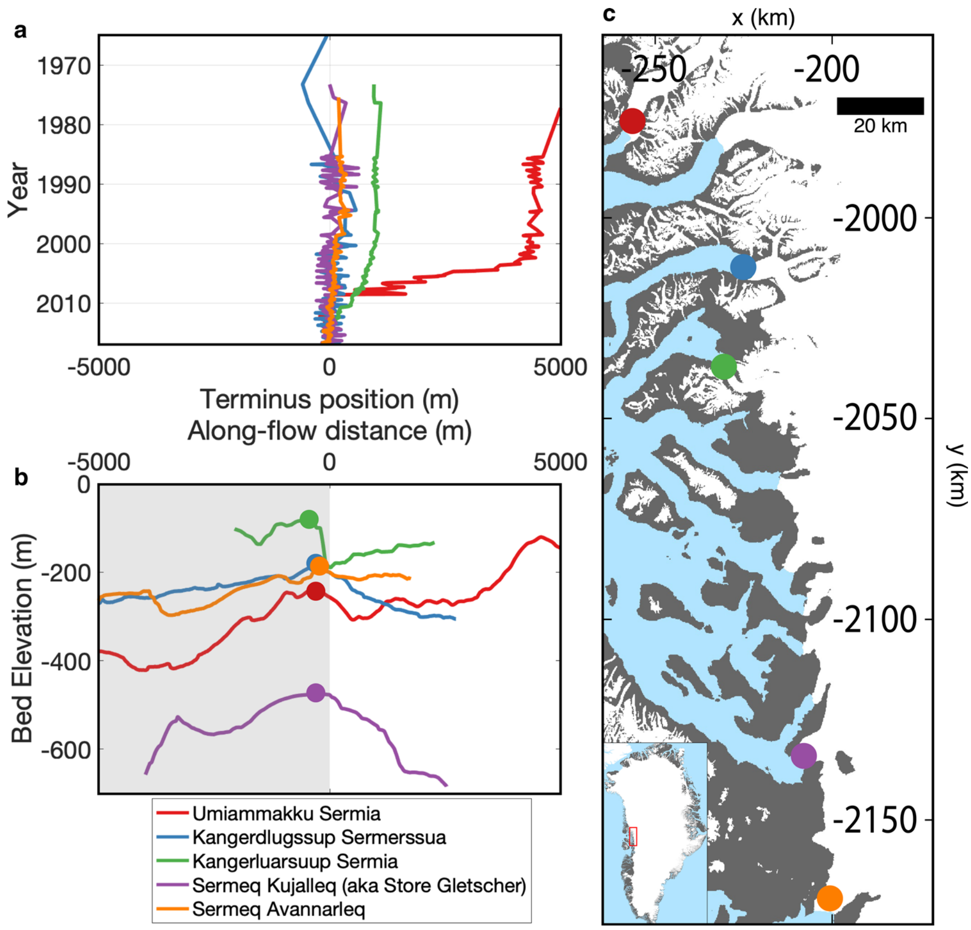

Fig. 1. Observational evidence of terminus and terminus persistence at bed peaks in Central West Greenland (CWG). (a) Terminus positions (*x*-axis) over time (*y*-axis) at five CWG glaciers derived from satellite-based sensors (Catania and others, 2018). (b) Along-flow bed topography at CWG glaciers in panel (a), with the *x*-axis is the along-flow distance relative to recent (2016) terminus position, with $x = 0$ representing the present position of the glacier termini and gray shading indicating where there is currently grounded glacier ice ($x < 0$). Nearest bed peaks upstream of the current terminus denoted by a filled circle in each case. For glaciers with strong cross-fjord variations in topography (Kangerluarsuup, Kujalleq), the deepest bed topography across the fjord is used; for the others, mean topography across the fjord is used. Bed topography error range is plotted in Figure S3, with typical error for proglacial fjords of < 10 m and typical error for near-terminus subglacial topography of 10–100 m. Bathymetry from BedMachine data compilation (panel b) (Morlighem and others, 2017). (c) Location of CWG glaciers in panels (a) and (b) on polar stereographic north projection (EPSG:3413). (https://epsg.io/3413).

2. Observations of glacier persistence

Central West Greenland (CWG) provides a particularly well-observed laboratory for understanding glacier retreat over bumpy beds. As in most of Greenland, surface melting has been persistently intensifying since the 1970s (or potentially earlier; Trusel and others, 2018). In the late 1990s an influx of warm water from the North Atlantic arrived in glacier fjords in this region (Holland and others, 2008). A compilation of terminus positions recorded by visible satellite imagery (Catania and others, 2018) show that many glaciers in CWG retreated between the late 1990s and the early 2000s when ocean temperatures were warm (Fig. S1). However, some glaciers in this region have not retreated during the observational record. Figure 1a shows observations of terminus positions at four such persistent glaciers, Kangerdlugssup Sermerssua (blue), Kangerluarsuup Sermia (green), Sermeq Kujalleq (purple, aka Store Gletscher) and Sermeq Avannarleq (orange). Figure 1b shows the along-flow bed topography at these same glaciers from the BedMachine v3 dataset (Morlighem and others, 2017). These glaciers have persisted <1 km downstream of bed peaks, indicating the critical importance of peaks in bed topography in potentially delaying or preventing rapid glacier retreat. This persistence likely cannot be explained by a heterogeneity in warm water reaching the termini of these glaciers, since warm water was pervasive below 150 mdepth throughout this region (Holland and others, 2008), and all of these glaciers are connected to the continental shelf through deep bathymetric troughs (Fig. S2; though submarine bathymetry is subject to uncertainty, see Fig. S3). Glaciers in this region that did retreat following the ocean warming event in the late 1990s mostly retreated away from bed peaks (on which they had previously persisted) and some have since ceased retreat upon reaching a new bed peak (Fig. S1; Catania and others, 2018). In Figure 1, we show one example of such a glacier, Umiammakku Sermia (red), which began rapidly retreating away from a bed peak ~5 years after the arrival of warm waters in the region, before ceasing retreat at a different bed peak around 2010.

Geological evidence from regions of past glacier retreat further demonstrates the importance of bed peaks in the response of glaciers to climate change. The bathymetry of the Ross Sea, Antarctica is composed of smooth, flat troughs separating large plateaus. Amid this smooth bathymetry, localized recessional moraines and grounding zone sediment wedges record locations where the deglacial retreat of glaciers in the Ross Sea embayment paused for prolonged time periods (Simkins and others, 2017; Greenwood and others, 2018). Figure 2 focuses on three particular locations in the Ross Sea where high-resolution multibeam bathymetric observations show pervasive grounding zone wedges

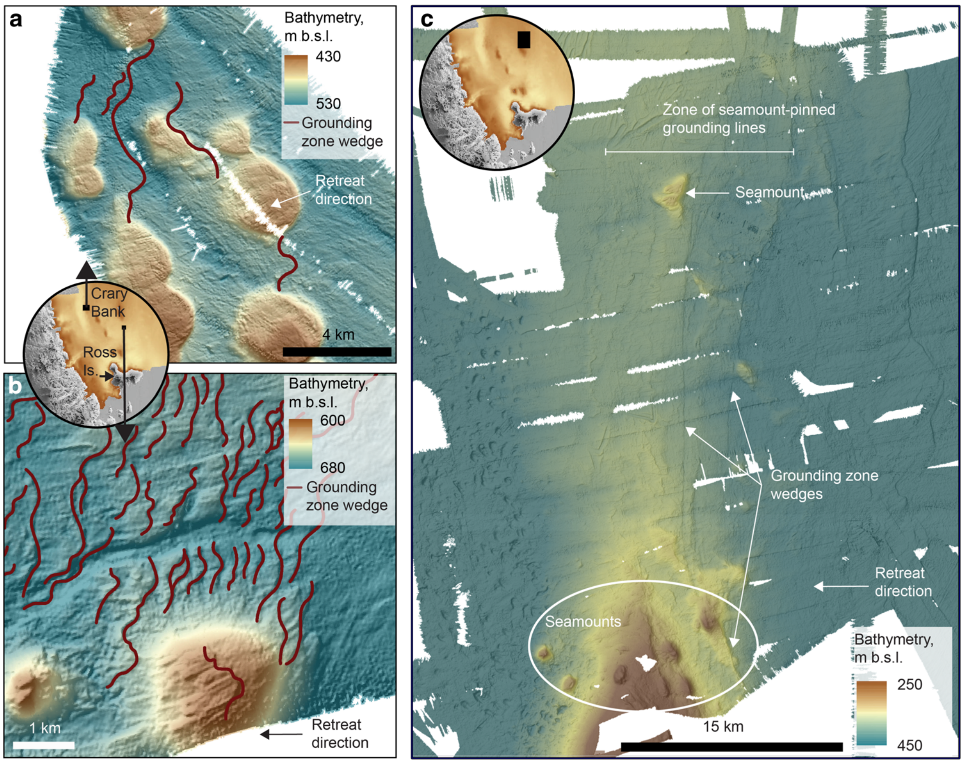

Fig. 2. Bathymetry (in meters below sea level, m b.s.l.) of the southwestern Ross Sea (inset) with linear to sub-linear grounding zone wedges (i.e.paleo-grounding lines) concentrated on and between isolated volcanic seamounts. (a) Paleo-grounding lines positioned (indicated by brown lines) between clustered, flat-topped seamounts. (b) Pinning of a paleo-grounding line (brown line) on and around a seamount. (c) Paleo-grounding line (indicated by a pointer) pinning on and between seamounts that are separated by several kilometers. Multibeam echo sounding bathymetry was collected on cruise NBP15-02A (Simkins and others, 2017; Greenwood and others, 2018).

connecting, parallel to, and on top of isolated seamounts on otherwise flat topography. Grounding zone wedges are formed when the grounding line of a glacier persists at a location for a sufficiently long time period that a significant sediment deposit forms at that location (Dowdeswell and Fugelli, 2012). These grounding zone wedges are not present in surrounding portions of the seafloor in the Ross Sea, indicating that bed peaks (which are generated by non-glaciological processes) at both of these locations in the Ross Sea exerted an important control on glacier retreat. Such grounding line persistence at isolated bed peaks also occurs in 3D numerical ice flow simulations of the retreat of West Antarctic ice streams through the Ross Sea following the last glacial period (e.g. Golledge and others, 2014). Other marine geophysical surveys of the seafloor in regions of past glacier retreat also reveal widespread geological evidence for prolonged periods of terminus persistence at bed peaks over a wide range of time periods and local conditions (Stoker and others, 2009; Todd and Shaw, 2012; Greenwood and others, 2017).

3. Model description

To simulate a typical marine-terminating glacier near a bed peak, we use a one-dimensional flowline model (in the x-direction) of a marine-terminating glacier (with the same equations and numerical methods as Schoof, 2007a, and many other studies). We simulate the glacier velocity (u), thickness (h) and terminus position (x_t). Velocity is determined from the shallow stream/shelf approximation (SSA) of the momentum balance

$$\frac{\partial}{\partial x}\left(2\bar{A}^{-1/n}h\left|\frac{\partial u}{\partial x}\right|^{1/n-1}\frac{\partial u}{\partial x}\right) - C|u|^{m-1}u - \rho_i g h\frac{\partial}{\partial x}(h-b) = 0, \tag{1}$$

where $\bar{A}$ is the depth-averaged flow law rate factor, C is the sliding law coefficient, m is the sliding law exponent, ρ_i is the density of ice, g is gravity, b is the bed topography. Ice thickness is determined from the mass conservation equation

$$\frac{\partial h}{\partial t} + \frac{\partial}{\partial x}(uh) = a, \tag{2}$$

where a is the surface mass balance (SMB, net annual snowfall and surface melt). In the cases we consider in this study, ice flow is dominated by sliding over a moderately slippery bed (parameters listed in Table 1), and the role of lateral shear stresses are not considered (in order to aid comparison with established theory). At the ice divide ($x = 0$), we set $u = 0$ and $\frac{\partial(h-b)}{\partial x} = 0$. We assume that the terminus is just at flotation (i.e. an unbuttressed grounding line), so we prescribe

$$h(x_t) = -\frac{\rho_w}{\rho_i} b(x_t), \tag{3}$$

where ρ_w is the density of seawater and bed topography is negative when below sea level. Ice velocity at the terminus is calculated (not prescribed) from the stress balance, assuming that the

Table 1. Parameter values for steady-state and transient retreat simulations (unless otherwise specified in text)

Parameter	Description	Value
A_g	Nye–Glen Law coefficient (Pa^{-n}s^{-1})	4.2×10^{-25}
b_x	Prograde bed slope	1×10^{-3}
C	Basal friction coefficient (Pam$^{-1/n}$s$^{1/n}$)	1×10^{6}
g	Acceleration due to gravity (ms^{-2})	9.81
m	Weertman friction law exponent	1/3
n	Nye–Glen Law exponent	3
Δt	Time step (yr)	1
ρ_i	Ice density (kgm^{-3})	917
ρ_w	Seawater density (kgm^{-3})	1028

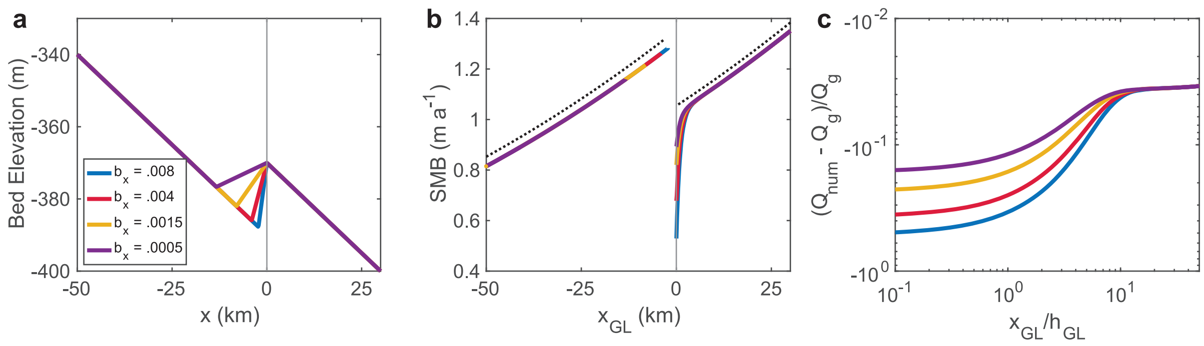

Fig. 3. Simulated stable terminus positions in the vicinity of a bed peak. (a) Four idealized bed topographies with differing bed slope just upstream of bed peak. (b) Bifurcation diagrams showing steady-state terminus positions over a range of surface mass balance. For each value of SMB, an initial guess on either side of the bed peak is used to determine if more than one steady-state exists. Lines are plotted from simulations of steady-state glacier state at 0.001 m a^{-1} increments of SMB (seaward of the bed peak simulated steady-state terminus positions are 1–10 m apart). The dotted line is the stable terminus positions calculated by solving $\alpha x_g = Q_g(x_g)$, with analytical approximation for ice flux from Schoof (2007b) (from topographies in panel a), derived by neglecting the effect of local slope. (c) The fractional difference between the ice flux at the terminus predicted from our numerical solution, Q_{num}, and the ice flux that would be predicted on neglect of the effect of local slope, $Q_g(x_g)$, as a function of distance from the bed peak (normalized by terminus ice thickness).

driving stress is fully supported by extensional stresses

$$\frac{\partial}{\partial x}\left(2\bar{A}^{-1/n}h\left|\frac{\partial u}{\partial x}\right|^{1/n-1}\frac{\partial u}{\partial x}\right) - \rho_i\left(1-\frac{\rho_i}{\rho_w}\right)gh\frac{\partial h}{\partial x} = 0. \qquad (4)$$

The momentum and mass conservation equations are solved using an implicit backward Euler method with a 1 year time step, and the constraint that the terminus is at flotation. The entire model is formulated on a moving mesh that is specified to be coarse (1–3 km) in the glacier interior and fine (50–100 m) in the grounding zone region, which is specified as the final 3% of the grounded glacier domain. The moving mesh is constructed so that the model domain always encompasses the entirety of the grounded glacier. This refined moving mesh numerical modeling approach has been shown to be a highly accurate method for simulating marine-terminating glaciers (Vieli and Payne, 2005).

The four idealized bed topographies we consider in this study all have a single sharp bed peak, but with different reverse bed slopes just upstream of the peak, and otherwise the same forward-sloping bed (i.e. shallowing toward the interior). We specify the bed topography through a piecewise-linear function that changes slope at the bed peak and trough. We design such idealized bed topographies to resemble the height and slopes around bed peaks in Greenland (Fig. 1b) and Antarctica (Fig. 2). More extensive high-resolution observations of subglacial bed topography indicate pervasive regions of Greenland and Antarctica where bed slope changes sign on horizontal length scales of less than a kilometer (Morlighem and others, 2017, 2020). We consider other versions of the bed peak topography in simulations plotted in the supplementary materials.

In the next two sections, we consider steady-state glacier configurations and transient simulations of glacier evolution. Glacier steady-states are determined by numerically solving for glacier states, with rates of change that are zero to within machine precision. To confirm stability of these steady-states, we perturb the SMB by 0.001 m a^{-1} and solve the transient glacier evolution, ensuring it does not evolve to a very different steady-state. Transient simulations are initialized from a steady-state and perturbed through a step reduction in SMB. In the supplementary materials, we consider various other forms of forcing (e.g. ocean melt and a linear trend in SMB) and find qualitatively similar results to the step reduction in SMB.

The type of model described here has been widely implemented, and shown to compare well to real glaciers in Greenland and Antarctica which flow primarily through sliding and float at their terminus (e.g. Enderlin and others, 2013; Jamieson and others, 2014; Huybers and others, 2017). We have also replicated the substance of the results described hereafter in simulations with a range of horizontal resolutions (Fig. S4) and high-resolution simulations of the Elmer/Ice Full-Stokes numerical ice-flow model (see Figs S5, S6), indicating that the resolution or SSA approximation does not affect the substance of the conclusions in this study.

4. Enhanced glacier stability at bed peaks

We first consider how bed peaks affect glacier stability over a range of climate forcing. Simulations show (Fig. 3b) that over a wide range of SMB, glacier termini persist indefinitely (i.e. reside at a stable steady-state) near bed peaks. We find that the sharp bed peaks we consider in this study, which entail an instantaneous transition (in space) from a forward-sloping bed to a reverse-sloped bed, lead to glacier stability over a wider range of SMB than what is predicted in prevailing theories of terminus stability (dotted line in Fig. 3b, reproduced from theory of Schoof, 2012). The steeper the reverse sloped bed upstream of the bed peak, the wider the range of SMB over which the glacier will remain stable. For the steepest reverse slope (shown in blue), the glacier remains stable a short distance downstream of the topographic high for a significant range of SMB from 0.5 to 1.0 m a^{-1}.

Ice must flow uphill to reach a terminus located at a bed peak. This slows the flow of ice approaching the bed peak and decreases the ice flux through a terminus retreating toward a bed peak. Figure 3c shows how ice flow near the terminus is changed by local bed slope by plotting the deviation of simulated steady-state terminus ice flux from the ice flux predicted by theories assuming negligible bed slope near the terminus (i.e. Schoof, 2007b). As the terminus approaches within $\sim$ 10 ice thicknesses of the bed peak, the ice flux decreases much more rapidly than is predicted under the assumption of negligible bed slope near the terminus. Indeed the magnitude of this reduction in ice flux near the bed peak (10–50% in these examples) is comparable to the effect of substantial ice shelf buttressing on grounding line ice flux (Reese and others, 2018; Mitcham and others, 2022). The cause of this rapid decline in ice flux is a lowered surface slope and driving stress and hence lowered velocity on the ice flowing up the bed peak, which then lowers terminus ice velocity, thickness and ice flux through longitudinal viscous stresses. Figure 4 shows surface elevation profiles for three steady-state simulations with termini just downstream of the bed peak, demonstrating how the surface slope flattens where ice flows up to the bed peak. Seen another way, ice that slows as it flows up the bed peak thickens, increasing the gradient in height above buoyancy near the terminus, which reduces the potential ability for changes in SMB or ocean melt at the terminus to affect terminus position.

The enhanced stability of simulated termini near bed peaks, as compared to prior theory, explains counterintuitive aspects of observations. There is a wide range of external forcing over which a terminus will persist at a bed peak, explaining why so many glacier termini are observed at bed peaks on bumpy bed topography. Indeed, repeating these steady-state simulations for corrugated bed topography (a repeated series of peaks and troughs) indicates stable glacier terminus positions exist almost exclusively at bed peaks for retreating glaciers (Fig. S7, though advancing glaciers may have stable configurations away from bed peaks). The reduced glacier sensitivity to climatic changes at bed peaks also explains why many glaciers are observed to persist, seemingly on the precipice of instability, even while experiencing substantial fluctuations in local climate (Tinto and Bell, 2011; Hillenbrand and others, 2017; Catania and others, 2018). Such enhanced stability of glaciers at bed peaks is in contrast to the prevailing idea that glaciers at bed peaks are necessarily 'vulnerable' to even small changes in climate due to their spatial proximity to reverse-sloping beds over which the marine ice-sheet instability occurs (Ross and others, 2012; Morlighem and others, 2020). As we have shown here, glaciers can be close to a bed peak, but are only 'vulnerable' in the sense in that they may initiate a rapid retreat in response in a large change in climate forcing.

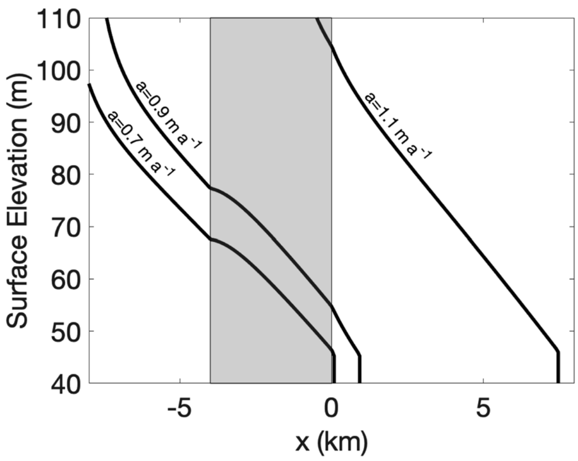

Fig. 4. Simulated near-terminus surface elevations for three stable glacier configurations near the bed peak for the $b_x = 0.004$ bed topography (red lines in Fig. 3). Gray shading indicates region of reverse-sloping bed.

The enhanced range of stability near points of destabilization (in state space) is a hallmark of a 'crossing-sliding bifurcation' (di Bernardo and others, 2008). The system behavior in the vicinity of such bifurcations is different from the canonical 'saddle-node bifurcation', which has previously been identified as the route through which grounding lines lose stability (Mulder and others, 2018; Pegler, 2018). In a saddle-node bifurcation, the loss of stability occurs suddenly in state space, without a change in the sensitivity of the stable branch (i.e. lines in Fig. 3b) to parameter changes upon approach to the bifurcation point. In a crossing-sliding bifurcation, the stable branch instead 'slides' along the bifurcation manifold with parameter variation (i.e. the system state remains close to the bifurcation without crossing it). Crossing-sliding bifurcations typically arise in system with non-smooth variations in system properties coinciding with bifurcations. In the case of a glacier retreating toward a sharp bed peak, as the system approaches the bifurcation point due to parameter variation (i.e. SMB or ocean melt) the stable glacier state (Fig. 3b) becomes much less sensitive to parameter variation, before eventually crossing the bed peak and initiating a large change in glacier state. This distinction in the type of bifurcation is important because it leads to much larger jumps in the system state (i.e. ice volume loss) upon crossing the bifurcation. As the forcing changes (i.e. SMB decreases), the onset of rapid ice loss is delayed, leading to a higher rate of ice loss if and when the terminus crosses the bed peak. In Figure 3b, this amounts to the difference between a 25 km retreat in terminus position for a relatively smooth bump (i.e. the dotted line), compared to a retreat of 40–100 km for sharper peaks (purple, yellow, red, blue lines).

5. Distinguishing glacier stability from transient persistence

In transient simulations of terminus retreat over sharp bed peaks (those plotted in Fig. 3a), a glacier is initialized at a steady state with its terminus just downstream of a bed peak, and is then subjected to a 40% step reduction in SMB uniformly over the glacier catchment. Figure 5a shows that some of the simulated glaciers retreat up to, then transiently persist just downstream of the bed peak for a period of time spanning decades to centuries (yellow and red lines), before eventually crossing the bed peak and rapidly retreating over the reverse-sloping bed. There are also cases where there is merely a brief slowdown in the rate of retreat at the bed peak (purple line), and other cases where the persistence continues indefinitely (blue line). We define such an indefinitely persistent case as stable in the same mathematical sense that we do in the previous section, where a system state persists forever with no change in forcing. Similar behaviors of transient and indefinite persistence of glaciers at bed peaks also occur in equivalent full-Stokes simulations of glacier retreat over bed peaks (Figs S5, S6) and in SSA simulations of glacier retreat

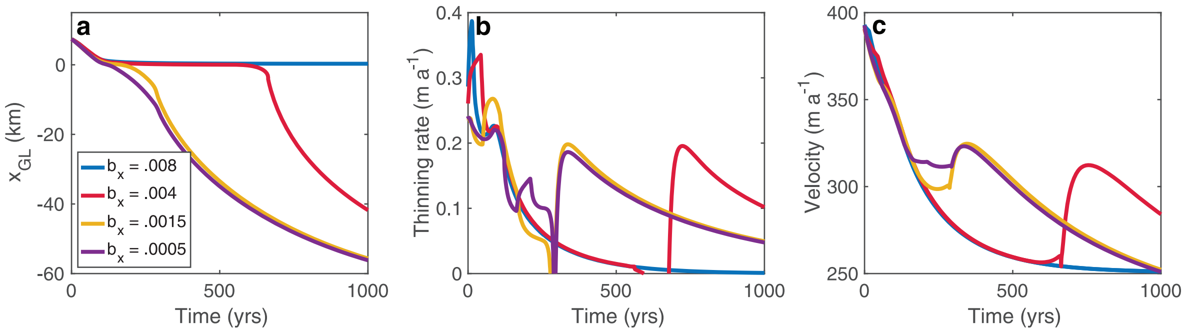

Fig. 5. Simulated terminus retreat in the vicinity of a bed peak. (a) Evolution of a terminus from steady state, in response to an instantaneous 40% reduction in surface mass balance over the glacier catchment (1.1–0.66 m a^{-1}), for a variety of upstream bed slopes. Terminus position (y-axis) is relative to bed peak location as in Figure 3a. (b) Thinning rate 50 km upstream of terminus in transient simulations. (c) Ice velocity 50 km upstream of terminus in transient simulations.

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with different types of forcing and smoothed bed peaks (Figs S8–S10). These transient simulations show that even for SMB values that do not correspond to a stable glacier configuration (plotted in Fig. 3b), there may still be prolonged periods of transient terminus persistence. Longer periods of transient persistence lead to more rapid subsequent retreat, which continues even as the terminus encounters forward-sloping bed topography.

Glaciers that persist at bed peaks continue to lose mass through thinning upstream of the terminus following changes in climate forcing, as seen in observations of recent thinning upstream of the terminus at persistent glaciers throughout Greenland (Kjeldsen asnd others, 2015; Felikson and others, 2017; Mouginot and others, 2019; Shepherd and others, 2020). At persistent glaciers in CWG, this thinning is mostly being driven by negative SMB anomalies, which are largely offset by dynamic thickening bringing ice from upstream portions of glacier catchments (Felikson and others, 2017). Ultimately, this upstream-intensified thinning leads a decrease in ice surface slope and upstream slowing, which is captured in observations of persistent CWG glaciers (Joughin and others, 2010) and our simulations (Figs 5b-c). Though such thinning is less than that occurring at retreating glaciers through dynamic thinning, it nonetheless shows that persistence of a glacier terminus is not necessarily indicative of a glacier in mass equilibrium.

We can compare our simulated glaciers which stabilize at bed peaks to those which merely pause at bed peaks to ascertain whether observations of persistent glaciers may provide evidence of their eventual fate. We find that, regardless of whether they ultimately remain at the bed peak or retreat from it, the glaciers we simulate have upstream thinning rates within millimeters per year of each other (Fig. 5b and Fig. S11), and ice velocities within meters per year of each other (Fig. 5c and Fig. S12) while they persist at a bed peak. It would thus be exceedingly difficult to observationally distinguish glaciers that are merely paused from those that have stabilized indefinitely at bed peaks. Other studies have also found that, in realistic simulations of the future retreat of glaciers away from bed peaks, small uncertainties in the observed glacier state, bed topography or the climate forcing produce large uncertainties in the timing of the onset of rapid glacier retreat which is then amplified by the divergence of retreat predictions due to marine ice-sheet instability (Gladstone and others, 2012; Robel and others, 2019). Ultimately, the delicate balance between advection and thinning at persistent glaciers makes it exceedingly difficult to project retreat of glaciers over bumpy bed topography, and further emphasizes the need for more accurate observational constraints on glacier state and rate of change, bed topography and local climate change.

6. Discussion

We carefully construct idealized bed topographies in this study (plotted in Fig. 3a) to test the effect of bed slope upstream of the bed peak while keeping the rest of the topographic profile identical. This latter constraint ensures that initial glacier steady-states far upstream from the bed peak and at the terminus are nearly identical across simulations. However, the result is that the full bed peak feature is narrower and taller for more reverse-sloping beds. Thus, in interpreting these simulations, these multiple geometric effects are challenging to disentangle from one another. We conduct variations on the transient simulations plotted in Figure 5 smoothing the bed peak (Fig. S10) and keeping the bed trough to peak height (Fig. S13) and width (Fig. S14) constant. Though the exact timing of glacier retreat over a bed peak is dependent on the details of bed peak topography (Sun and others, 2014; Castleman and others, 2022; Christian and others, 2022), ultimately we reconfirm the main conclusions of this study that the bed slope just upstream of a sharp bed peak is the primary (though not only) determinant of whether a retreating glacier persists at the bed peak and the duration of this persistence.

In addition to the role of along-flow bed topography in glacier retreat considered in this study, it may be that other abrupt changes in the subglacial or lateral boundary conditions could cause similar enhanced stability or transient persistent as described here. Previous studies have shown that lateral narrowing of glacier troughs can delay or halt retreat of glaciers (Jamieson and others, 2014; Åkesson and others, 2018). Similarly, rapid transitions in basal friction have also been argued to be potential locations of stability for future glacier retreat (Schroeder and others, 2013). As we have shown in Figure 2, there is observational evidence that even narrow bed peaks in wide glacier troughs can cause sufficient slowdown (or pause) in retreat for the deposition of a grounding zone wedge. Simulations using 3D ice flow models have also found that isolated bed peaks can slow or stop glacier retreat at a range of settings in Greenland and Antarctica (Gladstone and others, 2012; Morlighem and others, 2016; Waibel and others, 2018; Robel and others, 2019). Nonetheless, we expect that there is a minimum lateral extent of bed peaks beyond which they are unlikely to affect glacier retreat, in line with studies showing that bed topography below an ice thickness in horizontal extent have limited effect on ice flow (Gudmundsson, 2008).

Though we impose idealized SMB forcing in the transient simulations considered here (and ocean forcing in supplementary simulations, see Fig. S8), marine-terminating glaciers are subject to forcing from both the ocean and atmosphere on a wide range of time scales (Christian and others, 2020). Glaciers that succeed in retreating over bed peaks may then experience less ocean melt as the subglacial bed peaks we consider in this study become proglacial submarine sills which can block the influx of warm water to the glacier terminus (e.g. Jakobsson and others, 2020). Additionally, ocean melt that undercuts a glacier terminus or SMB changes concentrated closer to terminus may induce other dynamic changes in the glacier state beyond those considered here. Further work may consider how the rate and spatial distribution of climate forcing interacts with bathymetric effects to determine the progression of glacier retreat.

7. Conclusions

We have shown that glaciers observed at bed peaks have two possible future behaviors: they may remain at the bed peak indefinitely (i.e. stabilize) or initiate retreat, potentially long after the onset of a change in local climate. Glaciers persisting at bed peaks may continue to lose mass in response to a previous or sustained climate change, though there will be an increasing ‘disequilibrium’ between this mass loss and the total committed glacier mass loss implied by contemporaneous climate forcing (Christian and others, 2018). If the terminus does eventually cross the bed peak, terminus retreat and total glacier mass loss accelerates rapidly, relaxing the glacier disequilibrium between instantaneous and total committed mass loss. Eventually, the total sea level contribution from non-persistent and transiently persistent glaciers may be similar, though the timing and rate of peak mass loss may be very different (e.g. Fig. 5).

In attempting to infer the future behavior of glaciers persisting at bed peaks, observations can be deceptive. We have shown that at retreating glaciers, ice flux and thickness may change considerably with relatively little change in the terminus position, due to the longitudinal transmission of lowered driving stress on the reverse-sloping slide of the bed peak. Thus, interpreting observations of terminus change requires accurate measurements of bed topography and the critical context of changes in other aspects of glacier state (particularly interior thickness and velocity) to assess whether the glacier is in balance. Additionally, the slow response time scale of glaciers, particularly those that have encountered bed peaks, indicates that the utility of ‘stability’ as a tool for categorizing observed glacier changes is limited without the critical context of multi-centennial (or millennial) glacier changes, and the climate forcing over that time period. The scope of these challenges and potential impacts indicate that we should direct a similar degree of attention and resources to closely observing and carefully simulating persistent glaciers as we do to rapidly changing glaciers, as it is possible and perhaps likely that they will eventually contribute just as much to future sea level rise.

Supplementary material. The supplementary material for this article can be found at https://doi.org/10.1017/jog.2022.31.

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Data Availability. Model codes used for conducting numerical experiments are available as persistent Zenodo repositories (MATLAB SSA model: https://doi.org/10.5281/zenodo.5245271, Julia SSA model: https://doi.org/10.5281/zenodo.5245331). Full-Stokes simulations were conducted with Elmer/Ice which is openly available at: http://elmerice.elmerfem.org/.

References

Åkesson H, Nisancioglu KH and Nick FM (2018) Impact of Fjord geometry on grounding line stability. *Frontiers in Earth Science* **6**, 71. doi: 10.3389/feart.2018.00071

Castleman BA, Schlegel NJ, Caron L, Larour E and Khazendar A (2022) Derivation of bedrock topography measurement requirements for the reduction of uncertainty in ice-sheet model projections of Thwaites glacier. *The Cryosphere* **16**(3), 761–778. doi: 10.5194/tc-16-761-2022

Catania G and 7 others (2018) Geometric controls on tidewater glacier retreat in central western Greenland. *Journal of Geophysical Research: Earth Surface* **123**(8), 2024–2038. doi: 10.1029/2017JF004499

Christian JE and 5 others (2020) The contrasting response of outlet glaciers to interior and ocean forcing. *The Cryosphere* **14**(7), 2515–2535. doi: 10.5194/tc-14-2515-2020

Christian JE, Koutnik M and Roe G (2018) Committed retreat: controls on glacier disequilibrium in a warming climate. *Journal of Glaciology* **64** (246), 675–688. doi: 10.1017/jog.2018.57

Christian JE, Robel AA and Catania G (2022) A probabilistic framework for quantifying the role of anthropogenic climate change in marine-terminating glacier retreats. *The Cryosphere Discussions* 1–28. doi: 10.5194/tc-2021-394

di Bernardo M, Budd C, Champneys AR and Kowalczyk P (2008) *Piecewise-smooth Dynamical Systems: Theory and Applications*. Vol. **163** SEP. London, UK: Springer Science & Business Media.

Dowdeswell J and Fugelli E (2012) The seismic architecture and geometry of grounding-zone wedges formed at the marine margins of past ice sheets. *Bulletin* **124**(11–12), 1750–1761. doi: 10.1130/B30628.1

Enderlin EM, Howat IM and Vieli A (2013) The sensitivity of flowline models of tidewater glaciers to parameter uncertainty. *The Cryosphere* **7**(5), 1579–1590. doi: 10.5194/tc-7-1579-2013

Felikson D and 9 others (2017) Inland thinning on the Greenland ice sheet controlled by outlet glacier geometry. *Nature Geoscience* **10**(5), 366–369. doi: 10.1038/ngeo2934

Gladstone RM and 9 others (2012) Calibrated prediction of Pine Island Glacier retreat during the 21st and 22nd centuries with a coupled flowline model. *Earth and Planetary Science Letters* **333**, 191–199. doi: 10.1016/j.epsl.2012.04.022

Golledge N and 6 others (2014) Antarctic contribution to meltwater pulse 1a from reduced Southern Ocean overturning. *Nature Communications* **5**(1), 1–10. doi: 10.1038/ncomms6107

Greenwood SL, Clason CC, Nyberg J, Jakobsson M and Holmlund P (2017) The Bothnian Sea ice stream: early Holocene retreat dynamics of the south-central Fennoscandian Ice Sheet. *Boreas* **46**(2), 346–362. doi: 10.1111/bor.12217

Greenwood SL, Simkins LM, Halberstadt ARW, Prothro LO and Anderson JB (2018) Holocene reconfiguration and readvance of the East Antarctic Ice Sheet. *Nature Communications* **9**(1), 1–12. doi: 10.1038/s41467-018-05625-3

Gudmundsson G (2008) Analytical solutions for the surface response to small amplitude perturbations in boundary data in the shallow-ice-stream approximation. *The Cryosphere* **2**(2), 77–93. doi: 10.5194/tc-2-77-2008

Gudmundsson G, Krug J, Durand G, Favier L and Gagliardini O (2012) The stability of grounding lines on retrograde slopes. *The Cryosphere* **6**(4), 2597–2619. doi: 10.5194/tc-6-1497-2012

Haseloff M and Sergienko OV (2018) The effect of buttressing on grounding line dynamics. *Journal of Glaciology* **64**(245), 417–431. doi: 10.1017/jog.2018.30

Hillenbrand CD and 9 others (2017) West Antarctic Ice Sheet retreat driven by Holocene warm water incursions. *Nature* **547**(7661), 43–48. doi: 10.1038/nature22995

Holland DM, Thomas RH, De Young B, Ribergaard MH and Lyberth B (2008) Acceleration of Jakobshavn Isbrae triggered by warm subsurface ocean waters. *Nature Geoscience* **1**(10), 659–664. doi: 10.1038/ngeo316

Huybers K, Roe G and Conway H (2017) Basal topographic controls on the stability of the West Antarctic ice sheet: lessons from Foundation Ice Stream. *Annals of Glaciology* **58**(75pt2), 193–198. doi: 10.1017/aog.2017.9

Jakobsson M and 9 others (2020) Ryder glacier in northwest Greenland is shielded from warm Atlantic water by a bathymetric sill. *Communications Earth & Environment* **1**(1), 1–10. doi: 10.1038/s43247-020-00043-0

Jamieson SS and 5 others (2014) Understanding controls on rapid ice-stream retreat during the last deglaciation of Marguerite Bay, Antarctica, using a numerical model. *Journal of Geophysical Research: Earth Surface* **119**(2), 247–263. doi: 10.1002/2013JF002934

Jordan TM and 6 others (2017) Self-affine subglacial roughness: consequences for radar scattering and basal water discrimination in northern Greenland. *The Cryosphere* **11**(3), 1247–1264. doi: 10.5194/tc-11-1247-2017

Joughin I, Smith BE, Howat IM, Scambos T and Moon T (2010) Greenland flow variability from ice-sheet-wide velocity mapping. *Journal of Glaciology* **56**(197), 415–430. doi: 10.3189/002214310792447734

Kjeldsen KK and 9 others (2015) Spatial and temporal distribution of mass loss from the Greenland Ice Sheet since AD 1900. *Nature* **528**(7582), 396–400. doi: 10.1038/nature16183

Mitcham T, Gudmundsson GH and Bamber JL (2022) The instantaneous impact of calving and thinning on the Larsen C Ice Shelf. *The Cryosphere* **16**, 883–901. doi: 10.5194/tc-16-883-2022

Morlighem M and 6 others (2016) Modeling of Store Gletscher’s calving dynamics, West Greenland, in response to ocean thermal forcing. *Geophysical Research Letters* **43**(6), 2659–2666. doi: 10.1002/2016GL067695

Morlighem M and 9 others (2017) Bedmachine v3: complete bed topography and ocean bathymetry mapping of Greenland from multibeam echo sounding combined with mass conservation. *Geophysical Research Letters* **44**(21), 11051–11061. doi: 10.1002/2017GL074954

Morlighem M and 9 others (2020) Deep glacial troughs and stabilizing ridges unveiled beneath the margins of the Antarctic ice sheet. *Nature Geoscience* **13**(2), 132–137. doi: 10.1038/s41561-019-0510-8

Mouginot J and 8 others (2019) Forty-six years of Greenland Ice Sheet mass balance from 1972 to 2018. *Proceedings of the National Academy of Sciences* **116**(19), 9239–9244. doi: 10.1073/pnas.1904242116

Mulder T, Baars S, Wubs F and Dijkstra H (2018) Stochastic marine ice sheet variability. *Journal of Fluid Mechanics* **843**, 748–777. doi: 10.1017/jfm.2018.148

Pegler SS (2018) Suppression of marine ice sheet instability. *Journal of Fluid Mechanics* **857**, 648–680. doi: 10.1017/jfm.2018.742

Reese R, Winkelmann R and Gudmundsson GH (2018) Grounding-line flux formula applied as a flux condition in numerical simulations fails for buttressed Antarctic ice streams. *The Cryosphere* **12**(10), 3229–3242. doi: tc-12-3229-2018

Robel AA, Schoof C and Tziperman E (2016) Persistence and variability of ice-stream grounding lines on retrograde bed slopes. *The Cryosphere* **10** (4), 1883. doi: 10.5194/tc-10-1883-2016

Robel AA, Seroussi H and Roe GH (2019) Marine ice sheet instability amplifies and skews uncertainty in projections of future sea-level rise. *Proceedings of the National Academy of Sciences* **116**(30), 14887–14892. doi: 10.1073/pnas.1904822116

Ross N and 9 others (2012) Steep reverse bed slope at the grounding line of the Weddell Sea sector in West Antarctica. *Nature Geoscience* **5**(6), 393–396. doi: 10.1038/ngeo1468

Schoof C (2007a) Ice sheet grounding line dynamics: steady states, stability, and hysteresis. *Journal of Geophysical Research - Earth Surface* **112**(F3), 1–19. doi: 10.1029/2006JF000664

Schoof C (2007b) Marine ice-sheet dynamics. Part 1. The case of rapid sliding. *Journal of Fluid Mechanics* **573**, 27–55. doi: 10.1017/S0022112006003570

Schoof C (2012) Marine ice sheet stability. *Journal of Fluid Mechanics* **698**, 62–72. doi: 10.1017/jfm.2012.43

Schroeder DM, Blankenship DD and Young DA (2013) Evidence for a water system transition beneath Thwaites Glacier, West Antarctica. *Proceedings of the National Academy of Sciences* **110**(30), 12225–12228. doi: 10.1073/pnas.1302828110

Sergienko OV and Wingham D (2019) Grounding line stability in a regime of low driving and basal stresses. *Journal of Glaciology* **65**(253), 833–849. doi: 10.1017/jog.2019.53

Sergienko OV and Wingham DJ (2021) Bed topography and marine ice-sheet stability. *Journal of Glaciology* 1–15. doi: 10.1017/jog.2021.79

Shepherd A and 9 others (2018) Mass balance of the Antarctic Ice Sheet from 1992 to 2017. *Nature* **558**, 219–222. doi: 10.1038/s41586-018-0179-y

Shepherd A and 9 others (2020) Mass balance of the Greenland Ice Sheet from 1992 to 2018. *Nature* **579**(7798), 233–239. doi: 10.1038/s41586-019-1855-2

Simkins LM and 8 others (2017) Anatomy of a meltwater drainage system beneath the ancestral East Antarctic ice sheet. *Nature Geoscience* **10**(9), 691–697. doi: 10.1038/ngeo3012

Smith JA and 9 others (2017) Sub-ice-shelf sediments record history of twentieth-century retreat of Pine Island Glacier. *Nature* **541**(7635), 77–80. doi: 10.1038/nature20136

Stoker MS, Bradwell T, Howe JA, Wilkinson IP and McIntyre K (2009) Late glacial ice-cap dynamics in NW Scotland: evidence from the Fjords of the Summer Isles region. *Quaternary Science Reviews* **28**(27-28), 3161–3184. doi: 10.1016/j.quascirev.2009.09.012

Sun S, Cornford SL, Liu Y and Moore JC (2014) Dynamic response of Antarctic ice shelves to bedrock uncertainty. *The Cryosphere* **8**(4), 1561–1576. doi: 10.5194/tc-8-1561-2014

Tinto K and Bell RE (2011) Progressive unpinning of Thwaites Glacier from newly identified offshore ridge: constraints from aerogravity. *Geophysical Research Letters* **38**(20), 1–6. doi: 10.1029/2011GL049026

Todd BJ and Shaw J (2012) Laurentide Ice Sheet dynamics in the Bay of Fundy, Canada, revealed through multibeam sonar mapping of glacial landsystems. *Quaternary Science Reviews* **58**, 83–103. doi: 10.1016/j.quascirev.2012.10.016

Trusel LD and 8 others (2018) Nonlinear rise in Greenland runoff in response to post-industrial Arctic warming. *Nature* **564**(7734), 104–108. doi: 10.1038/s41586-018-0752-4

Vieli A and Payne AJ (2005) Assessing the ability of numerical ice sheet models to simulate grounding line migration. *Journal of Geophysical Research: Earth Surface* **110**(F1), 1–18. doi: 10.1029/2004JF000202

Waibel M, Hulbe C, Jackson C and Martin D (2018) Rate of mass loss across the instability threshold for Thwaites Glacier determines rate of mass loss for entire basin. *Geophysical Research Letters* **45**(2), 809–816. doi: 10.1002/2017GL076470

Weertman J (1974) Stability of the junction of an ice sheet and an ice shelf. *Journal of Glaciology* **13**, 3–11. doi: 10.3189/S0022143000023327