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Understanding and Modelling Rapid Dynamic Changes of Tidewater Outlet Glaciers: Issues and Implications

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Abstract Recent dramatic acceleration, thinning and retreat of tidewater outlet glaciers in Greenland raises concern regarding their contribution to future sea-level rise. These dynamic changes seem to be parallel to oceanic and climatic warming but the linking mechanisms and forcings are poorly understood and, furthermore, large-scale ice sheet models are currently unable to realistically simulate such changes which provides a major limitation in our ability to predict dynamic mass losses. In this paper we apply a specifically designed numerical flowband model to Jakobshavn Isbrae (JIB), a major marine outlet glacier of the Greenland ice sheet, and we explore and discuss the basic concepts and emerging issues in our understanding and modelling ability of the dynamics of tidewater outlet glaciers. The modelling demonstrates that enhanced ocean melt is able to trigger the observed dynamic changes of JIB but it heavily relies on the feedback between calving and terminus retreat and therefore the loss of buttressing. Through the same feedback, other forcings such as reduced winter sea-ice duration can produce similar rapid retreat. This highlights the need for a robust representation of the calving process and for improvements in the understanding and implementation of forcings at the marine boundary in predictive ice sheet models. Furthermore, the modelling uncovers high sensitivity and rapid adjustment of marine outlet glaciers to perturbations at their marine boundary implying that care should be taken in interpreting or extrapolating such rapid dynamic changes as recently observed in Greenland.

Keywords Tidewater outlet glaciers · Cryosphere · Greenland · Calving · Ice sheet modelling

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1 Introduction

Tidewater outlet glaciers are narrow, fast flowing major drainage channels of ice sheets and polar ice caps that terminate in ocean fjords where the ice is calved off as icebergs. They flow through deep channels with beds well below sea-level and widths of only a few kilometers and they have speeds of several hundreds of meters to a few kilometers per year (Joughin et al. 2010a). In the recent decade many of such tidewater outlet glaciers in Greenland have strongly increased their flow speed (Rignot and Kanagaratnam 2006; Howat et al. 2007; Joughin et al. 2010a) and their surfaces started to thin rapidly (Pritchard et al. 2009; Thomas et al. 2009; Krabill et al. 2004; Abdalati et al. 2001). For example the three major Greenland outlet glaciers of Jakobshavn Isbrae (JIB) in the West and Helheim Glacier and Kangerdlugssuaq Glacier in the southeast have almost doubled their speed in recent years, they thinned dramatically with rates of tens of meters per year and their terminus retreated by several kilometers (Howat et al. 2007; Stearns and Hamilton 2007; Joughin et al. 2004, 2008a, 2008c, 2010a). The latter two, however, recently started to slow-down and readvance again (Howat et al. 2007; Murray et al. 2010).

These dramatic dynamic changes raised concern regarding Greenland's contribution to future sea level rise (IPCC 2007; Alley et al. 2005). Such dynamic mass loss has been estimated to contribute to sea level at a rate of 0.25 mm a^{-1} between 2003 and 2008 which is comparable with the mass loss through surface melt of the entire ice sheet and is significant in the sea level budget (Van den Broeke et al. 2009). Although oceanic warming has been suggested as the dominant trigger of such changes (Straneo et al. 2010; Holland et al. 2008; Rignot et al. 2010), the forcing mechanisms and controlling processes are not well understood and current large scale ice-sheet models are unable to realistically simulate these rapid dynamic changes (Alley et al. 2005; Bamber et al. 2007). In the Fourth IPCC Assessment Report, the contribution from dynamic mass loss have therefore been excluded and consequently sea-level projections appear to be underestimated but such dynamic change from marine outlet glaciers has been identified as a major limitation in our ability to assess future sea level (IPCC 2007).

Large scale ice sheet models that are typically used in prognostic assessments of Greenland's contribution to future sea-level have spatial resolutions of 5–10 km, with recent efforts pushing towards 1 km (Seddik et al. 2010; Gillet-Chaulet et al. 2011). This, is simply not enough to spatially resolve most of these very narrow and deep outlet glacier channels and, furthermore, the depth and detailed basal topography is still unknown for many of them. In recent years, clear advances have been made in terms of process representation within these models such as a more robust treatment of grounding line motion (Docquier et al. 2011; Pollard and DeConto 2009; Schoof 2007) and including fast ice flow and ice streaming by considering higher-order stresses to include longitudinal stress transfer (Bueler and Brown 2009; Pattyn 2003; Price et al. 2011; Pattyn et al. 2008), but most of these models are not yet fully operational on the required spatial resolution of outlet glaciers (Stone et al. 2010). And, importantly, major deficiencies remain in their representation of processes acting at the marine boundary such as calving and ocean melt. Some of these issues have been overcome in spatially reduced models that are specifically designed for single outlet glacier basins (Nick et al. 2009; Dupont and Alley 2005; Thomas 2004; Joughin et al. 2010b); however, such models cannot be generalized to an ice-sheet wide application and uncertainties in the forcing processes remain.

On the basis of such a reduced 1-dimensional flowband model applied to the example of JIB, this paper aims to explore and discuss the basic concepts and emerging issues in our understanding and predictive ability of the dynamics of tidewater outlet glaciers. We do not attempt here to give a complete review on this topic but, rather, we will focus on issues and implications investigated and informed by our modelling experiments, and with a particular view on the recent rapid changes and their potential contribution to future sea level rise.

1.1 Jakobshavn Isbrae

JIB's acceleration was one of the earliest of the outlet glaciers in Greenland to be detected in the satellite era and was in parallel with the onset of a rapid retreat of its 15 km long floating ice tongue. Before, its flow of about 6 km a^{-1} near the grounding line (Fig. 1c) and its terminus position were relatively stable with only small seasonal fluctuations (Joughin et al. 2008c; Sohn et al. 1998; Echelmeyer et al. 1990; Luckman and Murray 2005). This changed in 1997 when a rapid retreat of the terminus set in, with some re-advance in 2001, and reaching almost complete disintegration of the floating ice tongue

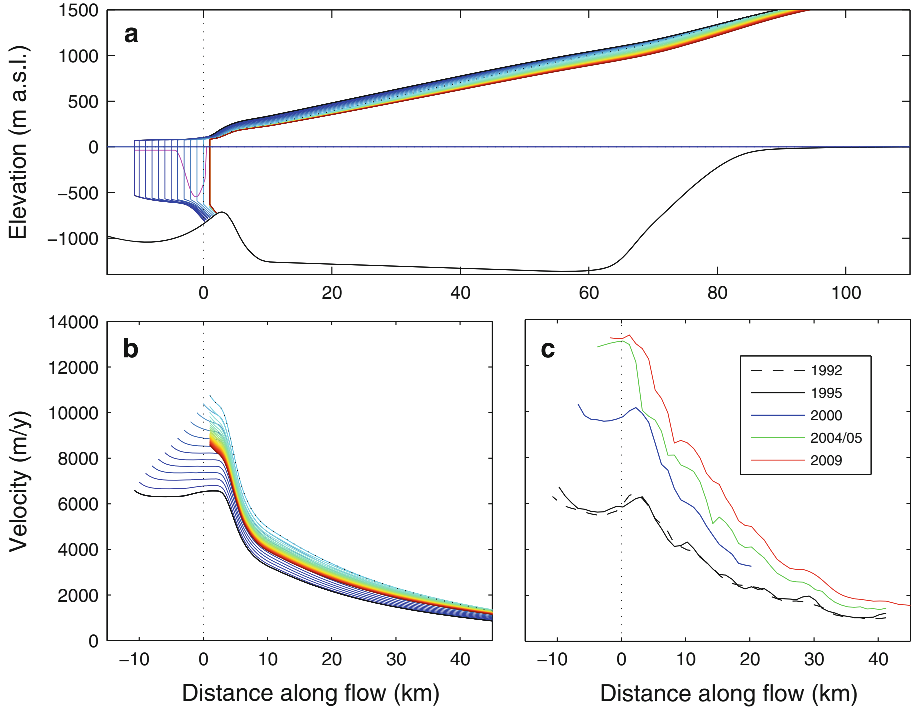

Fig. 1 **a** Profiles along the main channel of JIB showing **a** geometry and **b** modelled and **c** observed flow speed for a prescribed gradual retreat of 12 km over 6 years. The along flow distance is measured from the pre-retreat grounding line position. In **a** the *black line* shows the smoothed basal topography used in the model and the *coloured lines* show the modelled surface with time and go from *black* (initial unperturbed steady state surface) over *dark blue*, *light blue*, *green*, *yellow* to *red*, with time intervals of 0.5 years and over a total period of 20 years. The *dotted line* marks the time at 6 years when the retreat in the experiment was stopped. The *magenta line* indicates the prescribed basal melt rate pattern at the underside of the floating ice tongue in (m a^{-1}) with the origin at sea-level. **b** Profiles of modelled centreline flow speed over time using the same colour-coding as in **a**. **c** Observed profiles of along flow speed from satellite interferometry (from Joughin et al. 2004, 2010a and updated using recent TerraSAR-X observations, I. Joughin, personal comm.)

in 2004. During this retreat, JIB almost doubled its speed (Fig. 1c) and induced surface thinning with rates of up to 20 m a^{-1} near the terminus (Joughin et al. 2008c; Thomas et al. 2003, 2009). Further a strong seasonal variation in speed (Luckman and Murray 2005) and terminus position developed with summer retreat and winter re-advance in the order of 5–6 km (Joughin et al. 2008c). The thinning and speed-up have been observed to successively propagate and diffuse inland but is most pronounced in the main deep channel (Joughin et al. 2008c; Thomas et al. 2009). Currently the terminus position and grounding line is still near a relatively shallow sill but, behind, the bed deepens rapidly into a deep bedrock trough that reaches 70 km inland (Fig. 1a) and carries the potential for rapid unstable retreat in the near future (Hughes 1986; Weertman 1974; Vieli et al. 2001) As a trigger for its rapid retreat, ocean related processes such as changes in sea-ice concentration or subsurface melt at the ice ocean contact seem currently most plausible (Holland et al. 2008; Joughin et al. 2008c; Lloyd et al. 2011), but the linking processes at this marine boundary are still not well understood.

2 The Model

The model is described in detail in Nick et al. (2009, 2010) and is applied to an approximate case of the main channel of Jakobshavn Isbrae. It simulates the flow and surface evolution of JIB in a flowband along the main channel and includes basal and lateral resistance and transfer of stresses along the glacier through longitudinal stresses. The lateral resistance is parametrized by integrating the horizontal shear stress over the channel width (Van der Veen and Whillans 1996). For basal flow a non-linear Weertman-type sliding relation is assumed that includes effective pressure dependency (Fowler 2010). The model further includes a robust treatment of grounding line motion and a dynamic calving model based on a criterion of crevasse depth which in turn is a function of along flow strain rate (Nick et al. 2010; Benn et al. 2007). For basal topography a smoothed version of the high resolution dataset of CReSIS (https://www.cresis.ku.edu/plummer/jakob.html) is used and a fixed spatial basal melt pattern beneath the floating ice tongue is assumed based on estimates by Motyka et al. (2011) and outlined in Fig. 1a. A rate factor corresponding to −5 degree ice temperature has been assumed but has been softened by a factor of 2.5 in the side drag term to account for lateral shear softening. The model uses a moving grid that follows its ice front continuously and has an approximate along flow grid size of 250 m. The basal friction parameter has different values in the bed trough and upstream under the main ice sheet and additionally depends on effective pressure. It has been been tuned to correspond to the pre-1997 flow and surface geometry when the model is run to steady state. Thus, for the model experiments that follow below, the initial geometry (surface, ice front and grounding line position) is stable without any change in the boundary conditions.

3 Dynamic Adjustment

Recent rapid dynamic changes of marine based outlet glaciers around Greenland have the typical characteristics of calving front retreat in combination with strong flow acceleration and rapid surface thinning that propagates upstream with time (Joughin et al. 2008c, 2010a; Murray et al. 2010; Howat et al. 2007). Similarly, as a response to the collapse of the Larsen B ice shelf in the Antarctic Peninsula, strong acceleration and thinning have been observed for its former tributary glaciers (Rignot et al. 2004; Scambos et al. 2004).

This pattern of change suggests an initiation at the terminus with the most likely causing process being terminus retreat and the concurrent loss of buttressing from the terminus area (Nick et al. 2009; Howat et al. 2007; Joughin et al. 2008c; Thomas 2004). Recent research suggests that enhanced lubrication of the glacier bed from increased surface melt due to atmospheric warming is for fast-flowing outlet glaciers no major control (Joughin et al. 2008b; Nick et al. 2009; Murray et al. 2010; Schoof 2010), although it may be relevant for slower flowing grounded ice sheet margins (Zwally et al. 2002; Bartholomew et al. 2010) and therefore this process is in this paper not discussed much further. The observed acceleration and surface thinning that propagate upstream can in simple terms therefore be considered as a dynamic adjustment to a change in the boundary condition at the marine terminus. Regarding the abruptness in onset and the high rates of mass loss it is crucial in the wider context of potential sea-level contribution to discuss the process of dynamical adjustment and, importantly, the involved time scales.

Any retreat of the calving front leads to an instantaneous redistribution of stresses as the removed part of the terminus can no longer provide resistance to the upstream ice. This loss in buttressing induces an instantaneous acceleration in flow at the terminus which is transfered some distance upstream through longitudinal stresses. This longitudinal coupling distance is typically about 10 ice thicknesses and for outlet glaciers in the order of 5–20 km. The acceleration at the terminus increases the ice flux and initiates a thinning, which steepens the surface, increases the driving stress and leads to further acceleration. Through this feedback between acceleration and driving-stress the thinning is then propagating upstream with time (Payne et al. 2004; Joughin et al. 2003, 2008c). This thinning is a direct consequence of mass conservation and readjusts the flux back towards the pre-perturbation state in order to compensate for the mass loss. As a result, the flow and thinning start to slow down rapidly from the terminus after the retreat ceased whereas upstream it is still increasing although with reduced rates (Nick et al. 2009).

This dynamic adjustment is illustrated in Fig. 1 at the example of a prescribed gradual retreat of the floating ice tongue of JIB from its stable pre-1997 position by an annual rate of 2 km a^{-1} over a total period of 6 years. The successive retreat provides a gradual reduction of buttressing from the floating ice tongue and leads to a continuous increase in speed that is transferred upstream through longitudinal stresses and surface adjustment. The rapid slow down of flow and thinning at the terminus after stopping the retreat at 6 years illustrates well this rapid dynamic adjustment at the terminus and is supported by the modelled adjustment in flux and thinning rates with time (Fig. 2). Only 4 years after stopping the retreat, the mass loss decayed near the terminus more than a factor 2 from above 40 to 20% of its original balance flux of 24 km^3a^{-1} and the thinning rate drops there within half a year from around 20 m a^{-1} to almost zero. Upstream, where the flow speed is strongly reduced, the peak in flux change and thinning rate is delayed but also much wider and significantly reduced in magnitude (Fig. 2c+d). At 80 km behind the grounding line the peak in flux is reached about 8 years after ceasing the retreat and the amplitude is reduced to 7% of the initial balance flux at the grounding line, and both flux and thinning rates are maintained at such slightly enhanced levels for time-scales of decades.

Theoretical considerations and modelling suggest that the speed of upstream propagation of acceleration and thinning is a multiple of the flow-speed which is confirmed by our modelling experiment (Howat et al. 2007; Payne et al. 2004; Bindschadler 1997; Nye 1963). This is in simple terms explained by the fact that faster flowing systems can much quicker redistribute their mass, and it has important consequences regarding interpretation and assessment of rapid dynamic changes. It means that, although changes can propagate

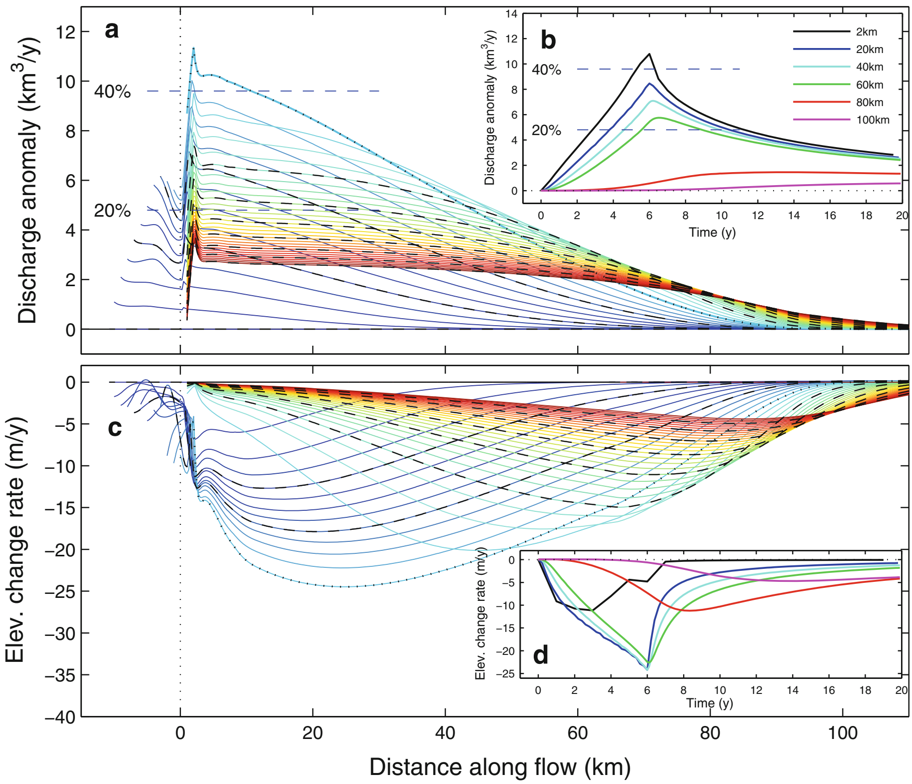

Fig. 2 Modelled along flow profiles of **a** ice discharge anomaly with reference to the pre-retreat discharge and **c** surface elevation change rates for the gradual retreat experiment corresponding to Fig 1. The *lines* are *colour-coded* as in Fig. 1 and are shown at 0.5 year time steps for *coloured lines* and 2 years for *dashed black lines* and span a total period of 20 years. The *dotted line* marks 6 years, the time when the retreat in the experiment was stopped. The *dashed blue lines* in **a** mark the relative percentage to the steady state pre-retreat grounding line discharge. The *insets* **b** and **d** show the evolution of the same quantities as in **a** and **c** but at fixed locations (legend in **c**) against time and again with the prescribed gradual retreat starting at time zero and stopping at year 6

upstream rapidly, they affect in the first few years only fast flowing areas whereas the slowly flowing inland ice acts almost as a buffer and changes will be much delayed and damped (see slowly flowing areas upstream of deep channel; Figs. 1a and 2). On the other hand, it also means slowly flowing upstream areas will continue to respond and contribute to mass loss at a reduced rate for the longer-term future as also demonstrated by Price et al. (2011) and further, that changes, currently observed upstream, may still be affected by terminus perturbations that happened several decades ago. This is important regarding interpretation of the present observations of thickness change in context of the longer-term terminus retreat of JIB since the Little Ice Age, when the terminus position was 35 km further advanced then at present (Csatho et al. 2008; Young et al. 2011).

Such rapid slow-down and readjustment at the terminus has been observed for many tidewater outlets around Greenland (Howat et al. 2007; Howat et al. 2008; Moon and Joughin 2008; Murray et al. 2010; Nick et al. 2009) and has several important consequences regarding the assessment of dynamic mass loss.

Firstly, it means that without any further perturbations, e.g., retreat or additional feedbacks at the terminus, such glaciers cannot keep their high speeds for long (more than a few years) and they start immediately readjusting and slowing down and thus the mass loss will decrease rapidly. However it is important to note that, after this initial adjustment phase, mass loss at a reduced level (order of few % of balance flux) may prolong for several decades as a result of the longer-term adjustment of slow flowing inland ice areas (Fig. 2; Price et al. 2011).

Secondly, estimations of mass loss through monitoring ice flux near the terminus are strongly affected by short-term perturbations and therefore may record outlet glacier *weather* rather than longer-term trends. Moving the flux gates somewhat upstream would provide more robust trends, but may not be practical, as relative observational errors increase significantly for reduced flow speeds (<100 ma^{-1}).

Thirdly, an obvious but important point in this context is that the flow speed will not go fully back to its initial pre-retreat level, even when as in our case the surface mass balance and consequently the steady state flux is kept the same, as the strong terminus thinning reduced the thickness significantly near the final grounding line and therefore higher flow speeds are needed to recover the pre-retreat flux. This effect is important in the interpretation of repeated flow speed observations and if neglected it may overestimate mass loss significantly. In our example, this overestimation in mass loss would be around 30% of the peak flux anomaly, as similarly determined for Helheim and Kangerdlugssuaq Glacier in SE-Greenland (Howat et al. 2007).

In our first prescribed retreat simulation, the patterns of flow acceleration and thinning are in good agreement with observations from the recent retreat phase of JIB (Figs. 1, 2; Joughin et al. 2004, 2008c; Thomas et al. 2009). The absolute increase in speed, however, is underestimated by the model and, more importantly, JIB has not been observed to slow down after the terminus ceased retreating in 2004, although the observed speed stagnated (Fig. 1c). This suggests that further feedbacks must be continuing to perturb this glacier which will be discussed in the section below.

4 Feedback Mechanisms

4.1 Grounding Line Retreat

A first major feedback is the retreat of the grounding line or grounded terminus as it removes areas of flow resistance at the base and if retreating into deeper water it may trigger unstable retreat as a result of increased ice flux with increasing water depth. This latter effect is well known from theoretical studies (Schoof 2007; Katz and Worster 2010) and studies on grounded tidewater glacier termini (Vieli et al. 2001; Meier and Post 1987) and has been observed for many tidewater glaciers such as Helheim (Howat et al. 2007) or Columbia Glacier (Van der Veen 1996). This effect can be enhanced by the positive feedback mechanism between subglacial motion and water pressure at the glacier bed. Any thinning brings the surface near the terminus closer to flotation and basal water pressure will approach the hydrostatic pressure of the overlying ice and therefore reduce frictional forces at the glacier bed (Pfeffer 2007; Vieli et al. 2000). It is important to note here that the effect of bed topography on grounding line stability is however significantly reduced in the case of a floating ice tongue or ice shelf through buttressing the upstream ice at their sides and stable positions are possible even on reversed slopes (Nick et al. 2010; Dupont and Alley 2006; Thomas 1978; Weertman 1974). This is particularly relevant for ice tongues with a narrow width as in the case of outlet glaciers and makes their existence or disintegration a crucial control in outlet glacier stability. Recent modelling work also demonstrated that for calving criteria that are not directly water-depth dependent the stability of the grounding line is less sensitive to the sign of the bed slope (Nick et al. 2010).

The strong overdeepening behind the sill at JIB shows after removal of the floating tongue potential for such an instability; however, it is currently not clear whether the grounding line already retreated over the sill into deeper water or not, as basal topography there is not well known. Although in our model the grounding line retreats slightly (Fig. 1a), it is unable to retreat upstream over the shallow sill. An additional experiment with a more extreme and dynamic calving model that always removes all floating ice and therefore its buttressing (using a flotation criterion; Vieli et al. 2001) also showed stabilization of the grounding line downstream of the sill (Fig. 3). However, when we run the same experiment for an artificially lowered bed sill by 100 m, the grounding line eventually retreats over the bed sill and a catastrophic retreat through the basal trough is initiated before it stabilizes again on a shallower bed about 80 km upstream (Fig. 3). These rather extreme experiments illustrate the importance of detailed knowledge of basal topography of tidewater glaciers and how in any numerical model relatively small uncertainties in water depth around such a sill may decide between enhanced stability or catastrophic retreat. For few cases, including the major outlet glaciers Helheim, Kangerlugdlugssuaq and JIB, detailed basal topography of the marine based channels became recently available (Operation IceBridge; Koenig et al. 2010), but for many tidewater outlet glaciers in Greenland it is largely unknown and provides a major limitation in assessing and modelling their future behaviour. Another important point regarding basal topography is that the bed for most of these outlets eventually rises upstream to above sea-level and therefore limits this potential instability arising from reverse bed slopes for the longer-term

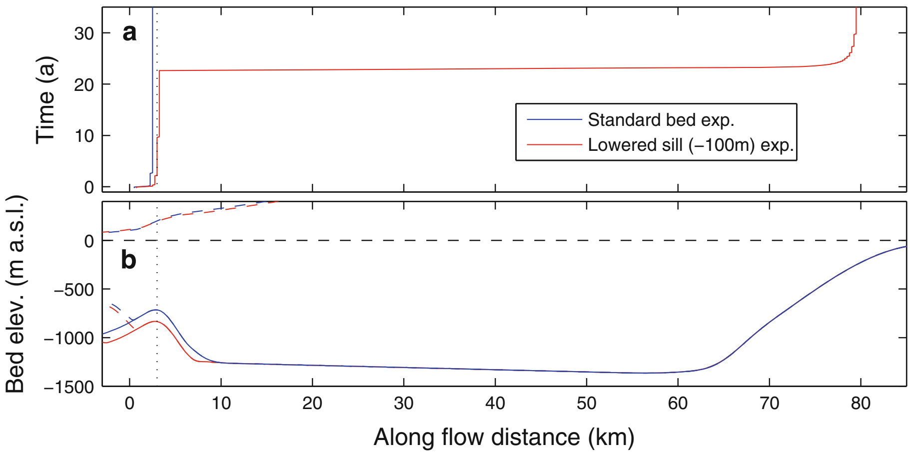

Fig. 3 **a** Modelled grounding line position against time for two retreat experiments with different basal topography (as shown in **b**) using a dynamic flotation criterion for which from time zero onwards any floating ice is always removed. The *black dotted line* refers to the location of the sill crest. **b** Basal topography showing the original bed (*blue line*) and the bed with a 100 m lower sill (*red line*). The *dashed coloured lines* in **b** refer to the initial *upper* and *lower* ice surface geometry before the removal of the floating ice tongue

future. This is, however, not the case for JIB and for a few other major outlets in the North and Northeast which have marine basins that extend inland by several tens of kilometres (Thomas et al. 2009).

There are further important issues in modelling grounding line motion which are largely of numerical nature and affect in particular large scale ice sheet models of relatively sparse resolutions and fixed grids (Schoof (2007); Vieli and Payne (2005); and see detailed discussion in this issue in Docquier et al. 2011). However, major advances have recently been made in the treatment of such grounding lines towards resolving these issues (Docquier et al. 2011; Gladstone et al. 2010; Pollard and DeConto 2009; Durand et al. 2009; Goldberg et al. 2009; Nowicki and Wingham 2008; Schoof 2007; Pattyn et al. 2006).

4.2 Calving Retreat Feedback

In the previous main experiment the retreat history of the terminus has been prescribed, however, in reality the position of the terminus is controlled by the interaction between the calving process and the flow dynamics and should be an output of a numerical model rather than an input.

The process of calving is still not well understood and rather poorly represented in numerical models (Benn et al. 2007; Bassis 2011). Most, larger scale ice sheet models prescribe calving rates using empirical relationships for calving flux which may constantly offset (over or underestimate) mass loss at the marine terminus or they do not evolve the terminus position with time and use a free flux condition and thereby ignore potential calving feedbacks. Recently, significant advances in calving model development have been made such as using strain rate criteria for determining terminus position (Benn et al. 2007; Nick et al. 2010) or calving rates (Alley et al. 2008; Albrecht et al. 2011; Amundson and Truffer 2010) or using a stochastic approach to calving (Bassis 2010). Implementation of such calving approaches in large scale 3-dimensional ice sheet models is, however, lacking behind and still in its infancy (Albrecht et al. 2011; Bassis 2011).

As explained above, after initiating retreat, the flow accelerates and leads to increased longitudinal stretching at the terminus which may enhance crevassing and therefore calving and further retreat. This positive feedback could therefore lead to rapid retreat of a terminus. We demonstrate this effect by including a dynamic, physically based calving criterion (Nick et al. 2010) that retreats the terminus back to a position up to where the longitudinal extension rate penetrates crevasses through the full depth. The initial retreat is here triggered by an increase in ocean melt beneath the floating tongue of 20% (corresponding to a warming of about 1°C, Motyka et al. 2011). This small increase in basal melt rate leads to a slight thinning which when applying the calving criteria triggers a continous retreat and initiates the strong acceleration and thinning as discussed above (Fig. 4). After 8.5 years the terminus retreat and flow start to slow down and stabilise automatically, probably as a result of the steepening of the surface in this location. Although in this case ocean melt triggered the retreat, the main work and cause for the dynamic changes is from the loss of buttressing as a result of a highly sensitive feedback between calving and retreat. Indeed, for an additional model run by keeping the terminus fixed but applying the same increase in ocean melt, the acceleration in flow is insignificant (below 1%). Thus, the feedback between calving and retreat seems crucial for the large scale dynamics and is therefore likely sensitive to the used calving parametrization. This implies that any prediction of dynamic changes of outlet glaciers hinges on a robust treatment of calving and thus there is an urgent need for development and validation of such calving models.

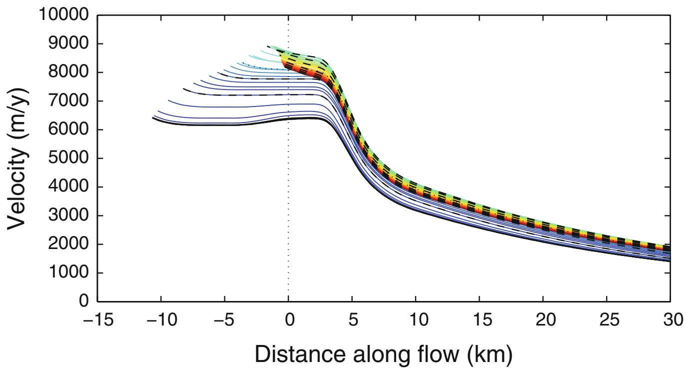

Fig. 4 Modelled along flow profiles of speed for an increase in ocean melt of 20% and using a dynamic calving model. The *lines* are colour-coded as in Fig. 2 (0.5 year time steps over 20 years). Note the automatic onset of slow down after 8.5 years

4.3 Lateral Shearing and Rifting

The very high flow speeds of such glaciers produce strong shearing along the lateral margins where shear stresses are highest. This shearing rheologically weakens the ice through strain heating (softening of ice by enhanced heat dissipation) and more effectively mechanically, through damaging the ice by fracture. This mechanical weakening then produces heavily crevassed bands along the lateral margins which are well visible in the field or on imagery and are characteristics of marine outlet glaciers. A further acceleration of the glacier will increase the shearing and therefore further breakup and weaken these margins. This weakening will therefore reduce the resistance provided by the sides (lateral drag) which in turn leads to further acceleration. This feedback has been suggested to play an important role in controlling Antarctic ice stream dynamics (Echelmeyer et al. 1994) and the pre-collapse changes of the Larsen B ice shelf (Vieli et al. 2007). At JIB shortly after the initiation of the recent retreat of the ice tongue, large rifts formed in these lateral shear margins and enhanced crevassing then propagated upstream (Joughin et al. 2008c). These rifts mechanically decouple the ice from its margin and the buttressing effect of the floating tongue is highly reduced. This may partly explain the continued acceleration between 1998 and 2001 despite a brief still stand of retreat or even slight re-advance of the terminus. With acceleration propagating upstream, this weakening of the margins will play a role some way upstream of the grounding line, as indicated in the case of JIB from imagery and in changed cross flow profiles in speed (Joughin et al. 2008c). Further, this feedback may explain the underestimated and continuing acceleration after the retreat ceased down.

In Fig. 5 we illustrate this potential effect in a very simple model experiment in which the softness of the ice is reduced proportionally to the excess of flow speed beyond an arbitrary critical speed of 7,500 m a^{-1} and additionally limiting this softening to a factor 20. For the same prescribed retreat as in the first experiment, this feedback triggers an earlier and far more abrupt acceleration and reaching higher speeds which are similar to those observed. However, the flow starts now to slow down even before the retreat ceases which is a simple consequence of the remaining floating terminus providing hardly any buttressing as the sides are very weak.

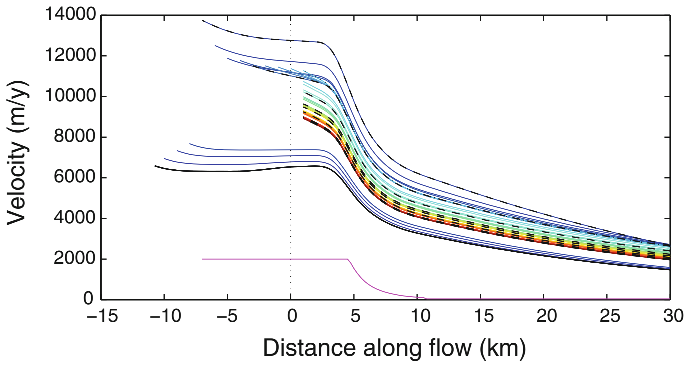

Fig. 5 Modelled along flow profiles of speed for the same prescribed retreat experiment as in Fig. 2 but including a feedback of lateral shear softening for flow above 7.5 km a^{-1}. The *lines* are colour-coded as in Fig. 1 (0.5 year time steps spanning 20 years). The *magenta line* indicates the factor of rheological lateral shear softening (100 times exaggerated) that is obtained 2 years after onset of the retreat

The example of JIB demonstrates the potential importance of this feedback in understanding and predicting the rapid mass loss from tidewater outlet glaciers, but also indicates the difficulties in including such effects into numerical flow models. The process of strain heating is relatively well understood and included in most thermo-mechanical ice sheet models. In the case of outlet glaciers, it is, however, the rheological damage through fracture such as rifting that is the dominant process which is not well understood in this context. Attempts of modelling the evolution of damage or rifting in ice have been undertaken (Pralong et al. 2003; Larour et al. 2004; Sandhaeger 2003), but they require high resolution and computational effort and are so far limited to small and well constrained cases. Larger scale models are unlikely to spatially resolve these narrow lateral shear zones sufficiently and therefore a more realistic approach may be to use a Coulomb-Plastic friction relation for side drag as similarly introduced for sliding at the glacier bed (Schoof 2005; Gagliardini et al. 2007). For 3-dimensionsal ice sheet models this parametrization would technically be included in the basal sliding boundary condition.

4.4 Vertical Deformation

A further mechanism that may explain the underestimation in flow acceleration by the model may be the neglect of vertical variations in horizontal stresses and flow. Our model and most existing large scale ice sheet models solve for vertically averaged flow and longitudinal stresses in areas of floating or rapidly flowing ice and thus vertical shear is ignored. However, preliminary experiments of ice tongue removal at JIB using a finite-element model that solves the full 3-D momentum equations indicate that even for an ice tongue with no lateral drag, terminus retreat resulted in flow acceleration (Luethi et al. 2009). Although this needs further investigation, this could imply that in order to include realistic boundary conditions at the marine terminus more sophisticated 3-D models that include the higher-order stresses (Blatter/Pattyn type models, Blatter 1995; Pattyn 2003) or solve the full-Stokes equation are needed. Such models are becoming readily available now (Pattyn et al. 2008) but are computationally far more expensive and therefore unlikely to be used for longer-term simulations on an full ice sheet scale in the near future. Thus, it will be crucial to rigorously test and assess the validity of different approximations of flow physics such as the shallow-shelf approximation (MacAyeal (1989); Bueler and Brown (2009); and as used here), a depth-integrated hybrid approximation (Bassis 2010; Schoof and Hindmarsh 2010; Goldberg 2011) or the first-order Blatter/Pattyn type approximation (Blatter 1995; Pattyn 2003) against the full-Stokes case for a marine terminating boundary and in view of computational expense. Importantly, this requires also validation against real world data.

5 Forcing

As any natural system, tidewater outlet glaciers are constantly exposed to variations in climate and ocean conditions and, thus, they will dynamically respond to these external forcings. Indeed, the recent dynamic changes of tidewater outlet glacier coincide in rough terms with the general pattern of atmospheric and oceanic warming in Greenland and further such changes seem to be synchronous on a regional spatial scale (Howat et al. 2008; Murray et al. 2010; Moon and Joughin 2008; Rignot and Kanagaratnam 2006). The linking processes between the climate, oceanic forcing and dynamic changes are still not well understood and are only crudely represented in current ice flow models, including the model used in this study. This implies that even if we would have detailed knowledge of oceanic and atmospheric forcing we are still unable to predict accurately predict their impact and response in terms of outlet glacier dynamics. Below we outline our current understanding and issues related to these linking processes between forcing and ice dynamics.

5.1 Oceanic Forcing

In recent years enhanced oceanic melt at the glacier terminus has been suggested as a major trigger for dynamic changes of Greenland outlet glaciers. The onset of acceleration at JIB in the West and of several marine outlet glaciers in the Southeast time well with warmer subsurface waters from the Irminger current reaching these Greenland coasts (Holland et al. 2008; Lloyd et al. 2011; Murray et al. 2010). Recent observations of ocean conditions within the fjords of major outlet glaciers such as JIB, and Helheim and Kangerdluggssuaq Glacier in the Southeast show clear evidence of such warm waters and indicate that their ice termini are in contact with these warm water masses and that significant ice melt occurs at depth (Holland et al. 2008; Rignot et al. 2010; Straneo et al. 2010, 2011). Estimates of melt-rates beneath the floating ice termini of JIB (Motyka et al. 2011) and at vertical grounded calving fronts elsewhere (Rignot et al. 2010; Motyka et al. 2002) suggest melt rates in the order of several 100s of meters per year, in particular at depth near the grounding line. These high melt rates suggest that the increase in oceanic melt may be sufficient to trigger a dynamic acceleration through modifying terminus geometry. In case of a floating ice tongue, enhanced melt will lead to a thinning which then initiates the calving retreat feedback as demonstrated in our ocean melt experiment for JIB (Fig. 4). At JIB, the thinning may additionally have led to an ungrounding of a pinning point and further enhance acceleration (Thomas et al. 2003; Joughin et al. 2008c). An additional experiment with overlaying seasonal variations (sinusoidal) in oceanic melt, in addition to the step change (Fig. 6), indicates that higher melt rates beneath the grounding-line may explain the increased seasonal variation in flow speed after the major retreat of the ice tongue (Echelmeyer et al. 1990; Joughin et al. 2008c; Luckman and Murray 2005).

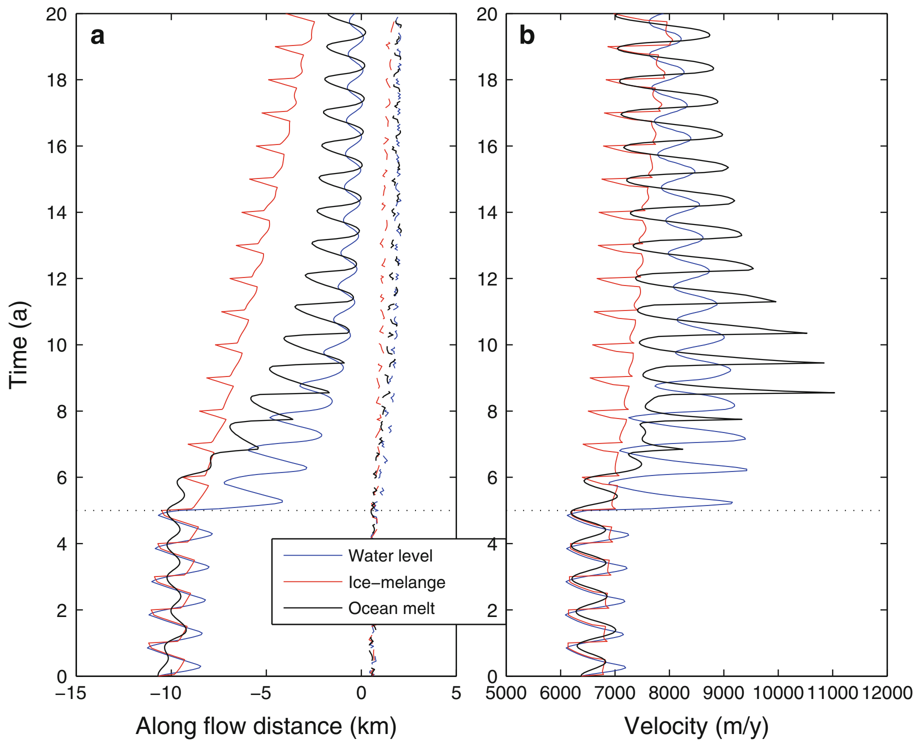

Fig. 6 **a** Position of the calving terminus (*solid*) and grounding line (*dashed*) with time for experiments of seasonal variations in water level within crevasses (*blue*), buttressing from ice-mélange (*red*) and ocean melt at the base of the floating ice tongue (*black*). After 5 years a step change in waterlevel (+1 m), in buttressing (reduction of ice-mélange season by 2 months) and in ocean melt (increase of 20%) has been applied. **b** Flow speed at pre-retreat grounding line position (0 km) with time

For grounded vertical calving faces which are most common for Greenland tidewater outlet glaciers, thinning of the terminus from underneath is not possible and ocean melt has to act in horizontal direction. Maximum rates, although high, are still several times below the typically observed calving rates, suggesting that basal melt must rather act through changing the shape of the front thereby influencing the stress field, such as through undercutting, and consequently affecting the calving rates (Rignot et al. 2010; Motyka et al. 2002; Vieli et al. 2002). This provides a major challenge for forcing ice flow models with ocean data as it is in this case unclear how to link ocean melt rates to the process of calving. Larger ice flow models use spatially fixed grids; thus directly applying ocean melt in horizontal direction is not possible and requires a sub-grid parametrization or a more sophisticated and high resolution treatment of the marine boundary. A further complication is added by the observation that melt rates are unlikely to be uniform along the width of a terminus as indicated by localised upwelling of freshwater plumes at many calving termini (Extreme Ice Survey: http://www.extremeicesurvey.org/).

Unlike prescribing a spatially fixed pattern of ocean melt as done in this study, a realistic oceanic forcing in a model should determine ice melt rates from the given conditions in temperature and salinity of the ambient fjord water and the given geometry. Although such ocean models are becoming available now (Thoma et al. 2010; Jenkins, in review; Holland et al. 2010; Payne et al. 2007) their application is still limited and importantly they lack a direct coupling with dynamic ice flow models. Examples of such limitations are (1) the poorly known fjord bathymetry and vertical stratification of waters which are crucial for mixing the waters at the ice interface (Straneo et al. 2011), (2) lack of information of the ambient fjord water temperature and salinity, and, importantly, (3) the lack of understanding of the replacement of fjordwater with coastal ocean water. On the latter, the intensity of along shore winds has been suggested as a crucial control for the fjord circulation and transporting the warm subsurface waters to the ice face (Straneo et al. 2010) which in Greenland seems to be affected by atmospheric circulation in the North Atlantic (Holland et al. 2008). The effect of fresh water from ice melting at the terminus or icebergs and from surface melt water run off further adds to the complexity of fjord circulation (Straneo et al. 2011).

5.2 Atmospheric Forcing

The effect of atmospheric warming on surface melt and surface mass balance is well understood and reasonably well represented in large scale ice sheet models; however, its potential impact on ice dynamics is still uncertain and rather indirect through triggering some of the feedback mechanisms discussed above, with the calving-retreat feedback probably being the most effective.

As mentioned in Sect. 3 the effect of enhanced basal lubrication by increased surface melting appears in the context of fast flowing outlet glaciers of secondary importance, although it may on short time scales still be significant (Howat et al. 2010). The timescales of adjusting the drainge system at the base of a glacier to enhanced delivery of surface melt water are known to be extremely short (days to weeks) which means that enhanced surface melt may in the long-term not affect the flow much or even slow it down (Sundal et al. 2011; Schoof 2010; Van de Wal et al. 2008). In any case, the relationship between melt water input at the surface and basal lubrication at the glacier bed is far from straight forward and therefore linking atmospheric forcing with basal sliding conditions involves currently high uncertainties.

Enhanced surface melt in the terminus region may increase water delivery into crevasses and thereby enhance crevasse penetration though hydro-fracturing (Weertman 1973; Van der Veen 1998). This may weaken the ice and therefore ease calving at the terminus. This process has earlier been suggested for explaining the seasonal variations in calving rates and terminus positions of tidewater glaciers (Sohn et al. 1998); however, a quantitative assessment of its role is still lacking and requires detailed data and observations. The warming or enhanced melt acts in this case only as a trigger for calving retreat which then causes the actual dynamic changes. As our dynamic model for calving is based on a criterion of critical crevasse depth, we can directly force it with water level within crevasses.

This effect is illustrated by varying in our calving criterion the water level within the crevasses seasonally (sinusoidal), with an amplitude of 1 m and then performing a step increase in water level of 1 m after 5 years (Fig. 6). This produces seasonal variations in front position and flow speed, and the step change in water-level is able to initiate a rapid dynamic retreat and acceleration which is both in rough agreement with observations at JIB (Joughin et al. 2008b, 2008c; Sohn et al. 1998). This seems encouraging regarding prognostic modelling of such systems as it allows a direct forcing with atmospheric data. The very high sensitivity to small changes in water level, however, indicates a strong reliance on the calving model and any parameterization that link surface melt to water level within crevasses. Further, the onset of terminus retreat in spring observed at JIB is typically well before surface temperatures reach above zero degrees and melting starts (Joughin et al. 2008b, 2008c).

Through a similar process, atmospheric warming may induce rheological weakening in lateral shear zones and thereby reduce buttressing from the sides, but this effect is probably of secondary importance. A further related effect could be rheological softening by warming up the ice from the surface through enhanced penetration of surface meltwater into the ice in crevassed areas. Although, this process has been suggested to be significant in slowly flowing ice sheet margins in Greenland (Phillips et al. 2010), its effect on outlet glacier dynamics remains speculative.

5.3 Sea-Ice and Ice-Mélange

A further potential forcing which is interlinked with both ocean and atmospheric forcing is the extent and duration of the seasonal *ice-mélange* in front of the calving terminus. At JIB, during winter when the sea-ice and icebergs freeze together to a rigid mass, calving activity has been observed to be highly reduced or even vanish and therefore the terminus advances in winter (Joughin et al. 2008c) at about the speed of ice flow. This ice-mélange is believed to exert a small force that is sufficient in preventing the already almost detached vertical ice-lamellas at the calving face from topping over and breaking off and therefore acts to reduce calving rates (Amundson et al. 2010). This effect provides a potential explanation of seasonal variations in calving rate (Joughin et al. 2008a, 2008c; Sohn et al. 1998; Howat et al. 2010; Reeh et al. 2001). The fact that for Greenland the onset of spring terminus retreat is correlated well with the disintegration of the ice-mélange supports this mechanism, in particular as the onset is well before surface melt starts. For JIB, a significantly reduced length of winter sea-ice in the outer fjord also seems to correlate well with the start of the terminus retreat in 1997 and the subsequent flow acceleration (Joughin et al. 2008c). A stabilizing effect of such ice-mélange on calving is also relevant for Antarctic outlet glaciers and ice shelves (Larour et al. 2004) where accretion of marine ice in rifts may act to strengthen such ice-mélange (Khazendar et al. 2009).

Including the stabilizing effect from ice-mélange is in our model implemented by adding a small longitudinal stress (40 kPa) at the marine boundary with a simple seasonal pattern of 6 months of ice free conditions (stress set to zero) followed by 6 months of compact ice-mélange (set to 40 kPa). After 5 years the duration of winter ice-mélange is reduced to 4 months. The resulting seasonal fluctuations in terminus position are again a consequence of changing calving rates from the ice-mélange forcing (Fig. 6). For the initial stable phase, our modelled terminus position and velocity variations at the grounding line are consistent with observations and we are able to trigger a rapid retreat by reducing the seasonal period of ice-mélange. However, the seasonal amplitude in terminus position variations is after the retreat far below the observed 5-6 km and the overall retreat is much less abrupt than observed which may both indicate a mis-representation of the ice-mélange forcing mechanism in our model. Further, it remains to be tested whether the longitudinal stress value of 40 kPa, required in our model to produce significant retreat and seasonal variations, is realistic or not.

5.4 Discussion on Forcing

All three different examples of seasonal forcings applied to our model above are able to produce similar patterns of seasonal front variations and abrupt retreat (Fig. 6). This is not surprising, given that they rely on the same calving retreat feedback and use the same calving criteria. Further, atmosphere, ocean and sea-ice are not independent and temporal variability of these forcings will inherently be in phase. However, during the rapid retreat period we obtain some notable phase shifts in seasonal variations of speed and terminus position (Fig. 6), which may be used to identify controlling forcings. The above experiments also stress the need for further development and robust validation of parameterizations of atmospheric and in particular oceanic forcings within numerical ice flow models. The above forcings should of course not be considered in isolation. Sea-ice and ice-mélange are obviously placed right at the interface between ocean and atmosphere and fjord circulation seems to be controlled by along shore winds and therefore regional scale atmospheric circulation and surface runoff may further affect water exchange within the fjord (Holland et al. 2008; Straneo et al. 2010, 2011).

The high temporal variability in forcing is well reflected in observed outlet glacier dynamics and in our model results (Fig. 6) which underpins the extreme sensitivity of these systems to variation in external forcing. The dynamic feedbacks discussed, such as calving retreat, act here again as amplifiers for the applied forcing. The observed episodic nature in mass loss of such outlet glaciers can be interpreted as a non-linear threshold response to a fluctuating external forcing signal. The recent record of dynamic changes of Greenland outlet glaciers is therefore not a reliable measure of longer-term records of ice sheet mass balance.

Further, on longer time-scales of several decades to millenia, a longer-term deficit in surface mass balance leads to a general slow thinning trend which through interaction with basal topography can produce episodic rapid retreat phases, as similar suggested for valley tidewater glaciers (Meier and Post 1987; Vieli et al. 2001). For the recent rapid changes in Greenland this seems, however, unlikely to be a direct forcing as it would not explain the strongly synchronous behaviour of outlets on a regional scale. However, regarding the longer term trends involved in future projections of sea level, this effect may well be relevant. Even in the current example of JIB, ’hidden’ within the current rapid retreat may still be a slow adjustment signal of retreat from the Little Ice Age (Csatho et al. 2008; Lloyd et al. 2011).

6 Implications and Conclusions

This paper explores and illustrates the major concepts and issues regarding our understanding and predictive ability of tidewater outlet glacier dynamics on the example of a numerical model application to the recent rapid changes of JIB. Our modelling shows that the dramatic dynamic changes at JIB are triggered from the marine terminus and supports the loss of buttressing from the disintegration of the floating ice tongue as a major cause. Furthermore, enhanced ocean melt is confirmed as a possible triggering mechanism for initiating retreat but other forcings such as reduced winter sea-ice coverage produce in our model similar dynamic behaviour. Importantly, we find that the modelled dynamic changes crucially depend on the calving-retreat feedback mechanism which implies a strong dependency of predictive outlet glacier models to the detailed treatment of calving. As demonstrated in our modelling, some advances have been made on this issue in recent years; however, the process of calving and importantly its link to atmospheric and oceanic forcing is still poorly understood and its representation and implementation remains a major limitation in current generation ice sheet models. In particular, regarding the likely oceanic warming as a trigger for outlet glacier acceleration, we still seem a long way away from being able to force dynamic ice sheet models with ocean data. Major efforts are needed in monitoring and understanding fjord circulation and in numerical model development, in particular in the coupling of ocean melt with ice dynamics and calving.

This also includes the need for increased spatial resolution in ice sheet models in order to resolve the deep bedrock channels, but also the involved processes such as grounding line retreat or lateral shear weakening. Such increased resolution requires, however, accurate knowledge of basal topography which is for most Greenland outlet glaciers still rare but critical for the stability of grounding lines and calving termini. Therefore efforts in collecting basal data such as the recent Operation IceBridge (Koenig et al. 2010) are vital for improving predictions of future ice sheet change and should also include bathymetric data in ocean fjords as they are important for their circulation and therefore heat transport from the ocean to the ice contact.

Unlike in our modelling, JIB continues its high flow speed at peak-level even after the terminus stopped retreating which suggests that other feedback mechanisms than loss of buttressing may play an important role such as rheological weakening in lateral shear zones. Alternatively, this discrepancy may indicate that the assumption of negligible vertical shear at fast flowing marine termini, as commonly used in ice sheet models, requires careful re-consideration of its validity. This means that, for the calculation of flow, additional higher-order stress terms may have to be included or even the full-Stokes equations solved. Although such models are becoming available (Pattyn et al. 2008) they still lack a realistic dynamic treatment of marine boundaries (calving) and importantly their computational-time is dramatically increased which makes predictions on the required time scales of decades to centuries challenging.

Our modelling further confirms the extreme sensitivity of tidewater outlet glaciers to perturbations at their marine boundary and their ability to adjust rapidly as indicated by observations. This implies that the dramatic accelerations in speed as recently observed for many marine outlet glaciers in Greenland cannot be maintained for long without additional perturbations or feedback mechanisms, but also that on a reduced but significant level mass loss can continue for decades. Furthermore, these rapid changes reflect rather short-term fluctuations in climate and oceanic forcing that are amplified through feedback mechanisms such as calving retreat. This is crucial in interpreting the dynamic changes as it means we are recording outlet glacier *weather* rather than longer-term trends. Thus, assessments and projections of dynamic mass loss based on our existing relatively short observational record (Rignot et al. 2011) or based on scenarios such as doubling flow speed over decadal or century time-scales (Pfeffer et al. 2008) are unrealistic and may drastically overestimate the contribution to future sea-level rise.

Considering the limitations and issues outlined in this paper, it is not realistic to expect that ice sheet scale models will in the near future be able to predict the exact magnitude and timing of change of individual outlet glaciers around Greenland. Regarding projections for future sea-level rise, longer-term trends (decades, centuries) in mass loss should therefore strategically be considered in ice sheet model development and importantly their validation. This requires a better understanding of the involved processes and forcings and the development of more sophisticated models with full process representation and high spatial and temporal resolution that then can be reduced and simplified with confidence. Validation of such simplified models will therefore require data constraints that span time scales well beyond the recent decade and therefore should include reconstructions of past outlet glacier change (Roberts et al. 2008; Csatho et al. 2008; Alley et al. 2010; Young et al. 2011). In any case, such a longer time perspective from the palaeo-record would be beneficial for the interpretation and current discussion of dynamic mass loss from Greenland tidewater outlet glaciers.

Overall, the reasonable performance of our numerical model, despite being relatively simple, is encouraging for future development of full ice sheet models. However, its application to JIB also outlines some important limitations and issues that need addressing, of which implementation of ocean forcing, a robust calving model and an improved dataset of basal geometry being the most prominent ones. We should therefore in the near future not expect to fully solve this issue of rapid dynamic ice mass loss. The fact that most outlets have beds that are upstream eventually above sea-level indicates, however, that for longer-term future warming scenarios (centuries to millenia) the process of surface melt may dominate the overall mass loss from Greenland (Van den Broeke et al. 2009), which is reasonably well represented in existing ice sheet models, but introduces additional uncertainties (Stone et al. 2010).

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