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Evolution of basal crevasses links ice shelf stability to ocean forcing

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ABSTRACT

Iceberg calving is one of the primary mechanisms responsible for transferring ice from the Antarctic ice shelves to the ocean, but remains poorly understood. Previous theories of calving have sought to explain the calving process as a brittle phenomenon that occurs rapidly when surface or basal crevasses penetrate the entire ice thickness. Here we show that the strain-rate-weakening nature of ice permits initially narrow basal crevasses to seed an instability that gives rise to locally enhanced ductile deformation and thinning over length scales that are large compared to the ice thickness. This ductile failure process, called necking, amplifies long wavelength features of bottom topography and allows basal crevasses to penetrate an increasing fraction of the ice thickness as they advect downstream. Application of the model to the four largest Antarctic ice shelves shows that necking occurs downstream of pinning points and sharp protrusions in the ice shelf embayment where stress is highly concentrated. However, model predictions are sensitive to assumptions about basal melting and refreezing within crevasses, suggesting that the combination of mechanical instability and ice–ocean interaction on the scale of an individual crevasse may play a leading role in controlling ice shelf stability.

1. Introduction

Mass loss from the Antarctic Ice Sheet occurs primarily through its ice shelves, freely floating platforms of ice that surround the ice sheet. This mass loss process is controlled by the gradual erosion of ice by basal melting combined with the sudden, sporadic detachment of blocks of ice in a process called iceberg calving. Basal melting accounts for slightly more than half of the total mass lost (Rignot et al., 2013), but partitioning between basal melt and iceberg calving is highly variable between ice shelves with iceberg calving accounting for more than 70% of mass loss from the three biggest ice shelves, the Ross, Filchner–Ronne and Amery ice shelves (Rignot et al., 2013). Furthermore, although mass loss by basal melting is known to be directly related to ocean forcing, the effect of climate forcing on iceberg calving is less well understood with existing parameterizations of calving providing conflicting predictions about whether a warming ocean will increase or decrease future iceberg production rates (e.g., Benn et al., 2007; Alley et al., 2008; Bassis, 2011; Bassis and Jacobs, 2013).

Attempts to develop parameterizations of iceberg calving have traditionally relied on empirical parameterizations (e.g., Brown et al., 1982; Alley et al., 2008) or the assumption that a calving event occurs when the combined depth of surface and basal crevasse approaches the ice thickness (Weertman, 1980; Benn et al., 2007; Nick et al., 2010; Bassis, 2011; Bassis and Walker, 2012; Bassis and Jacobs, 2013). Uncertainties remain in estimating crevasse penetration depths, but most researchers rely on one or more flavor of fracture mechanics, treating iceberg calving as a nearly instantaneous brittle process (e.g., Weertman, 1980; Van der Veen, 1998; Rist et al., 2002; Benn et al., 2007; Bassis and Walker, 2012). Crevasse depth based calving laws have been successfully applied to simulate the response of grounded outlet glaciers in Greenland to various perturbations (e.g., Nick et al., 2010), but fail when applied to the cold ice shelves of Antarctica, like the Ross, Filchner–Ronne and Amery ice shelves where surface ponding of water is rare. In these cases the predicted depth of basal and surface crevasses is rarely sufficient to penetrate the entire ice thickness (Bassis and Walker, 2012).

Recent observations of pervasive systems of basal crevasses with width comparable to or even exceeding the ice thickness (Humbert and Steinhage, 2011; Luckman et al., 2011; McGrath et al., 2012) suggest that brittle theories of crevasse inception may provide an inadequate description of the calving process. That wide basal crevasses are not purely brittle features can be seen by observing that the aspect ratio of brittle-elastic fractures scales like the ratio of stress opening crevasses to Young's modulus. Because the tensile stress opening crevasses within ice shelves is of the order 0.1–1 MPa and Young's modulus is of the order 10 GPa, the aspect ratio of brittle fractures should be many orders of magnitude smaller than the observed aspect ratio of wide basal crevasses. This

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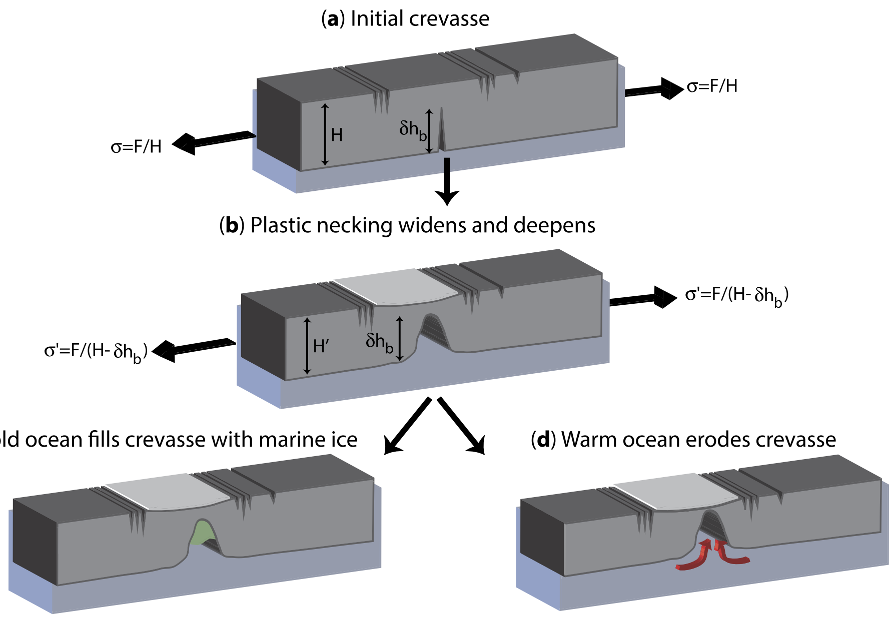

Fig. 1. Schematic showing the necking instability. Initially, basal crevasses are narrow (a). If longitudinal tensile stresses exceed the gravitational restoring forces then perturbations to the ice shelf bottom deepen and widen over time (b) causing a depression in the ice surface. This process can be modulated by snow or water filling the depressions at the ice shelf surface (white shaded region). Once crevasses are wide, crevasse morphology will be modulated by the formation of marine ice within basal crevasses in a cold ocean (green shaded region, (c)) or the excavation of basal crevasses by warm ocean waters (d).

suggests that the evolution of crevasses after inception through ductile deformation or through thermodynamic ice–ocean interaction controls crevasse depth and potentially, ice shelf stability.

In this letter we show that pseudo-plastic necking, a viscous instability associated with strain-rate-weakening materials such as ice, promotes locally enhanced ductile deformation and ice shelf thinning over length scales that are larger than the ice thickness. The necking instability proposed here is analogous to theories used to explain passive rifting in the lithosphere (e.g., Zuber and Parmentier, 1986), hinting that aspects of lithospheric rifting may provide an apt analogy to ice shelf rifting and *vice versa*.

2. Model description

2.1. Model overview

We seek to determine how crevasses evolve after inception. Building on previous efforts to model necking in metal bars (Hart, 1967) and lithospheric rifting (e.g., Smith, 1977; Emerman and Turcotte, 1984; Fletcher and Hallet, 1983; Ricard and Froidevaux, 1986; Zuber and Parmentier, 1986), the conceptual basis for the model is sketched in Fig. 1. The depth-integrated tensile force causing the ice shelf to spread is $F = H\tau_{xx}$ where τ_{xx} is the depth averaged deviatoric stress (Fig. 1a). A perturbation or bottom crevasse that decreases the ice thickness by δh_b causes the same depth integrated force F to be distributed over a smaller thickness and this enhances the stress acting on the thinned portion of the ice shelf (Fig. 1b). Because ice is a strain-rate-weakening material, the enhanced tensile stress will lead to enhanced ductile deformation and an increased rate of dynamic thinning. The necking instability, however, will be opposed by gravitational restoring forces causing ice to flow into depressions. The balance between horizontal stretching and gravitational restoring forces determines whether ductile flow will enhance or diminish crevasse penetration heights and will depend not only on the spatial wavelength of the perturbation, but also on how melting and marine ice accretion modifies crevasse penetration (Fig. 1c–d). The goal of the analysis that follows is to use a linear stability analysis to determine the rate at which individual wavelengths grow or decay (Fletcher and Hallet, 1983). The model that we present is similar to models used to explain lithospheric rift morphology (Zuber and Parmentier, 1986; Fletcher, 1995). Moreover, we find an explicit long wavelength approximation for the evolution of the ratio of crevasse penetration height to ice thickness that is valid for perturbations with length scales that are large compared to the ice thickness.

2.2. Governing equations

The ice shelf is idealized as a layer of uniform ice thickness H overlying an inviscid ocean. The coordinate system we use assumes the bottom of the ice shelf is located at $z = 0$ and the ice shelf surface is located at $z = H$ with the x and y axes aligned with the directions of the horizontal principal stresses. Denoting the stresses σ_{ij} and pressure p, the flow of the ice must satisfy the stress equilibrium equations, whence

$$\frac{\partial \sigma_{ij}}{\partial x_j} = \rho_{ice} g \delta_{iz}, \tag{1}$$

where $\rho_{ice} = 920$ kg/m^3 is the density of ice and $g = 9.8$ m/s^2 represents the acceleration due to gravity (pointing downwards) and δ_{ij} is the Kronecker delta function. Ice is assumed to behave as an incompressible fluid with a power-law rheology of the form

$$\sigma_{ij} = -p\delta_{ij} + 2\mu\varepsilon_{ij}, \tag{2}$$

where ε_{ij} denotes strain rates (symmetric part of the velocity gradient), and $\mu(\theta) = B\theta^{\frac{1}{2n}-\frac{1}{2}}$ is the effective viscosity of ice with flow law exponent $n \sim 3$. The parameter B is a temperature dependent stiffness parameter, which we assume is constant and $\theta = \frac{1}{2}\varepsilon_{ij}\varepsilon_{ij}$ is the second strain rate invariant. For simplicity, we assume ice temperature is constant and neglect spatial variations of B. To close the system we specify that traction is continuous across both the ice–atmosphere and ice–ocean interfaces. Because the atmosphere and ocean are both well approximated as inviscid fluids this implies that shear tractions at the interfaces vanish and continuity of normal traction across the top and bottom of the ice shelf is determined by a balance between normal traction of the ice and the hydrostatic pressure of the atmosphere or ocean.