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The effect of basal friction on melting and freezing in ice shelf–ocean models

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ABSTRACT

The ocean is an important control on the mass budget of the Antarctic ice sheet, through basal melting and refreezing underneath the floating extensions of the ice sheet known as ice shelves. The effect of the ice surface roughness (basal roughness) on melting and refreezing is investigated with idealised ice shelf–ocean numerical simulations. Both "hot" ocean forcing (e.g. Pine Island Glacier; high basal melting) and "cold" ocean forcing (e.g. Amery Ice Shelf; low basal melting, stronger refreezing) environments are investigated. The interaction between the ocean and ice shelf is further explored by examining the contributions to melt from heat exchange across the ice–ocean interface and across the boundary layer–ocean interior, with a varying drag coefficient. Simulations show increasing drag strengthens melting. Refreezing increases with drag in the cold cavity environment, while in the hot cavity environment, refreezing is small in areal extent and decreases with drag. Furthermore, melting will likely be focussed where there are strong boundary layer currents, rather than at the deep grounding line. The magnitude of the thermal driving of the basal melt decreases with increasing drag, except for in cold cavity refreeze zones where it increases. The friction velocity, a function of the upper layer ocean velocity and the drag coefficient, monotonically increases with drag. We find friction-driven mixing into the boundary layer is important for representing the magnitude and distribution of refreezing and without this effect, refreezing is underestimated. Including a spatially- and temporally-varying basal roughness (that includes a more realistic, rougher refreezing drag coefficient) alters circulation patterns and heat and salt transport. This leads to increased refreezing, altered melt magnitude and distribution, and a pattern of altered vertical flow across the entire ice shelf. These results represent a summary of melting and freezing beneath ice shelves and strongly motivate the inclusion of appropriate vertical mixing schemes and basal roughness values that vary spatially and temporally in ocean models of ice shelf cavities.

1. Ice shelf basal roughness

Ice shelves form around the Antarctic coastline where the ice sheet flows into the ocean. Hydrostatic pressure from the ocean lifts the ice off the bedrock at the 'grounding line', forming an ice shelf and ocean-filled cavity beneath (see Fig. 1(a and b)). The flow of the ice sheet into the ocean is controlled in part by the buttressing effect of ice shelves (Paterson, 2002; Dupont and Alley, 2005). Understanding the dynamics and mass loss from ice shelves is important for projecting future ice sheet flow and Antarctic mass balance (Pritchard et al., 2012).

Approximately half of the mass loss from the Antarctic Ice Sheet is from basal melting (1454 ± 174 Gt/yr. Depoorter et al., 2013), where warm (relative to the local pressure freezing point) ocean temperatures drive melting at the base of ice shelves. Increases in the rate of thinning of Antarctic ice shelves attributed to increased basal melting (Pritchard et al., 2012) suggests increased heat delivery to sub-ice shelf cavities. There are several different mechanisms hypothesised to be increasing the delivery of warmer water to ice shelf cavities, including changing wind regimes (Dinniman et al., 2012), polynya and sea ice interactions (Holland et al., 2010; Cougnon et al., 2013; Gwyther et al., 2014), thermocline shoaling (Hattermann et al., 2014) and coastal current redirection

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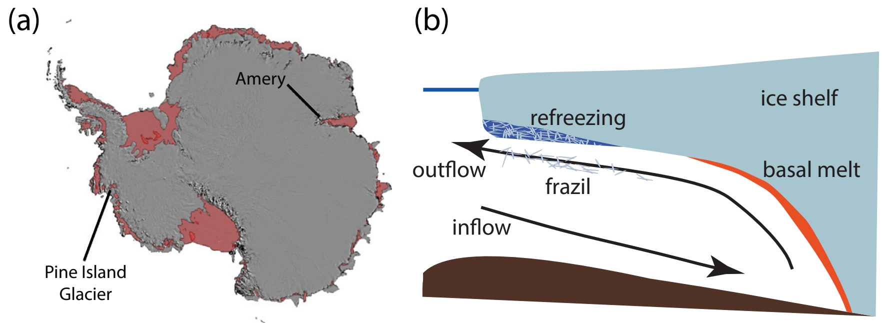

Fig. 1. (a) Antarctic ice shelves, including Pine Island Glacier Ice Shelf and Amery Ice Shelf, are marked in red on a MODIS MOA basemap (Scambos et al., 2007; Greene et al., 2013). (b) The ice shelf environment is illustrated, showing the grounding line, ocean inflow and outflow, basal melting and refreezing. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

(Jacobs et al., 2011; Hellmer et al., 2012). The plethora of different driving mechanisms suggests regionally-varying factors are important for controlling delivery of oceanic heat.

Basal melting at the ice-ocean interface is a function of the ocean circulation and amount of heat within the boundary layer, which can be subdivided into three layers. Direct interaction between the flow and the surface roughness occurs in the viscous sublayer, where molecular viscous forces are dominant ($\mathcal{O}(1)$ cm thick; Soulsby, 1983). This lies within the logarithmic layer (log layer; $\mathcal{O}(1)$ m), where the vertical velocity profile can generally be described with a simple logarithmic relationship or 'law of the wall' (Soulsby, 1983). Together with the outer layer (extending out to $\mathcal{O}(10)$ m; Soulsby, 1983), which is most influenced by the free-stream flow, these comprise the boundary layer. Heat and salt enter the boundary layer from the ocean below and mixing carries heat and salt to the ice interface. Basal roughness controls the turbulent exchange of heat to the ice–ocean interface and changes the thickness of the boundary layer (which affects entrainment and delivery of heat from below). Basal melting is also a function of hydrostatic pressure (through the pressure dependence of the freezing point temperature), ice shelf basal slope (through the velocity and thickness of the buoyant boundary layer) and driving temperature (Holland et al., 2008).

Using observations of the boundary layer shear profile to determine the coefficient of drag (C_D) has only been accomplished beneath relatively thin sea ice (McPhee et al., 1987; McPhee, 1992) but it is likely that this environment is different from the sub-ice shelf environment. Nevertheless, values of C_D used for the ice shelf–ocean interface in various numerical studies typically range between $C_D = 0.0015$ (e.g. Millgate et al., 2013), $C_D = 0.0025$ (e.g. Hunter, 2006; De Rydt et al., 2014) or $C_D = 0.003$ (e.g. Timmermann, 2002; Dinniman et al., 2007; Klinck and Dinniman, 2010). As C_D is the least observed parameter, it is often tuned to reduce the mismatch between the simulated and observed melt rates (e.g. Jenkins et al., 2010). However, all ice shelf–ocean models to date have used a single C_D for the entire ice–ocean interface, where in reality C_D will vary both spatially and temporally. It is likely that C_D would vary between zones of melting (where ablation of ice would lead to a hydraulically smoother interface and low C_D) and refreezing (high C_D due to the porous and flaky nature of marine-accreted ice (Craven et al., 2009)).

The roughness of the ice shelf-ocean interface affects melting at two different levels; by affecting turbulent flux of heat across the ice shelf-ocean interface (Section 1.1), and by changing the shear profile, boundary layer thickness and consequently entrainment of heat into the boundary layer (Section 1.2).

1.1. Ice–ocean heat flux

The roughness of the ice shelf interface affects basal melting by influencing the transfer of heat (and salt) via turbulence, across the

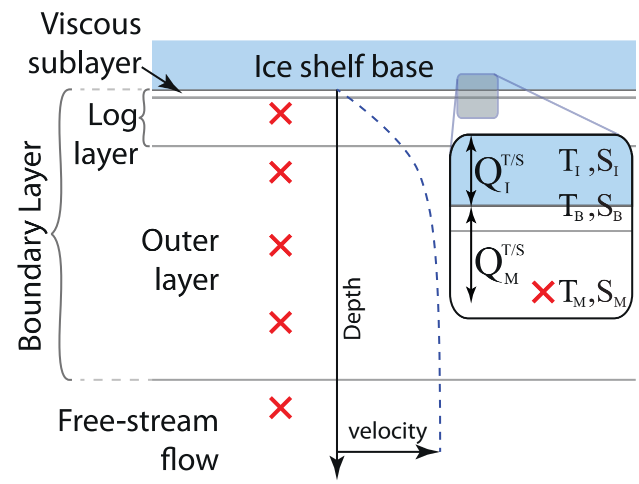

Fig. 2. The ice shelf–ocean interface is illustrated, with a vertical velocity profile (blue dashed line) and example model grid points (red crosses). Inset shows the ice shelf–ocean interface with temperature (T), salinity (S), heat flux (Q^T) and salt flux (Q^S) shown. Subscripts I, B and M refer to the ice shelf, ice shelf base and ocean model top cell, respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

ice shelf-ocean interface, as in Fig. 2. Most ocean models use a simple '3-equation parameterisation' of the ice-ocean interface (Hellmer and Olbers, 1989; Scheduikat and Olbers, 1990; Holland and Jenkins, 1999). See Appendix A for details.

In the Holland and Jenkins (1999) parameterisation, T_M and S_M refer to the mixed layer temperature and salinity. Defining a mixed layer of a single density class below the ice shelf interface is appropriate in an isopycnal coordinate model (e.g. Holland and Jenkins, 2001). However, we define the ice shelf–ocean interface in terms of the momentum boundary layer, which is critical for transferring the temperature and salinity properties from the interior cavity to the ice shelf. To implement the parameterisation, it is standard practice to use the values of T and S in the top model cell (e.g. Dansereau et al., 2014). However, as we are able to resolve the outer layer explicitly, T_M and S_M taken from the upper-most model cells are within the log layer, rather than from the outer layer. Therefore models that resolve $\leq\mathcal{O}(1)$ m at the ice shelf–ocean interface may provide different estimates of the gradients of temperature and salinity across the boundary layer and hence, different melt rates. However, a large proportion of the temperature and salinity changes occur over the viscous sublayer, which can lead to relatively well-mixed conditions through the rest of the boundary layer (Steele et al., 1989). In such cases, the standard practice for implementing the Holland and Jenkins (1999) parameterisation will produce results approximately consistent with