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Active water exchange and life near the grounding line of an Antarctic outlet glacier

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ABSTRACT

The grounding line (GL) of the Antarctic ice sheet forms the boundary between grounded and floating ice along the coast. Near this line, warm oceanic water contacts the ice shelf, producing the ice sheet's highest basal-melt rate. Despite the importance of this region, water properties and circulations near the GL are largely unexplored because in-situ observations are difficult. Here we present direct evidence of warm ocean-water transport to the innermost part of the subshelf cavity (several hundred meters seaward from the GL) of Langhovde Glacier, an outlet glacier in East Antarctica. Our measurements come from boreholes drilled through the glacier's ∼400-m-thick grounding zone. Beneath the grounding zone, we find a 10–24-m-deep water layer of uniform temperature and salinity (−1.45 °C; 34.25 PSU), values that roughly equal those measured in the ocean in front of the glacier. Moreover, living organisms are found in the thin subglacial water layer. These findings indicate active transport of water and nutrients from the adjacent ocean, meaning that the subshelf environment interacts directly and rapidly with the ocean.

1. Introduction

The Antarctic ice sheet drains ice into the ocean through floating ice shelves and outlet glaciers, which account for 74% of the Antarctic coastline (Bindschadler et al., 2011; Rignot et al., 2011). Here, the flowing ice separates from the underlying bed at the grounding line (GL), entering the grounding zone, a km-scale transition zone between the grounded ice and the freely floating ice shelf in hydrostatic equilibrium (Fricker and Padman, 2006). Physical conditions beneath the grounding zone greatly influence the ice-sheet mass budget because the subshelf melt rate is greatest near the GL (Jacobs et al., 1992; Jenkins and Doake, 1991; Rignot and Jacobs, 2002). And, due presumably to ocean forcing (Dutrieux et al., 2014; Jacobs et al., 2011; Thoma et al., 2008), this basal-melt rate is thought to be increasing (Pritchard et al., 2012), leading to the recent ice mass loss in Antarctica.

The structure of the grounding zone influences the ice-sheet dynamics as well as the subshelf ocean circulation. The bed slope near the GL affects the stability of marine-terminating glaciers and ice streams (Weertman, 1974), with a reverse-sloping bed (i.e., lower elevation inland) thought unstable (Schoof, 2007; Jamieson et al., 2012). Moreover, bed geometry is a control of seawater intrusion and tidal mixing beneath the grounding zone, and thus affects the efficiency of heat transport from the ocean to ice-shelf base (Holland, 2008; MacAyeal, 1984). Researchers have used satellite remote sensing (Bindschadler et al., 2011; Rignot et al., 2011; Fricker and Padman, 2006; Rignot and Jacobs, 2002), and radar/seismic soundings (Anandakrishnan et al., 2007; Christianson et al., 2013; Horgan et al., 2013a, 2013b) to locate the GL and investigate the structure of the grounding zone, but have made few direct observations (Fricker et al., 2011; Powell et al., 1996). Because in-situ measurement data are unavailable, validation of remote sensing is difficult and the detailed structure of the grounding zone is poorly understood.

To investigate the physical structure and hydrological environment beneath the grounding zone of an Antarctic outlet glacier, we drilled boreholes near the GL of Langhovde Glacier. Using these boreholes, we directly measured subshelf water temperature, salinity, and current within several hundred meters from the GL as well as recorded video within the boreholes. Our data indicate active transport of warm water from the ocean to the grounding zone, and borehole video images reveal biological activities in the subglacial environment.

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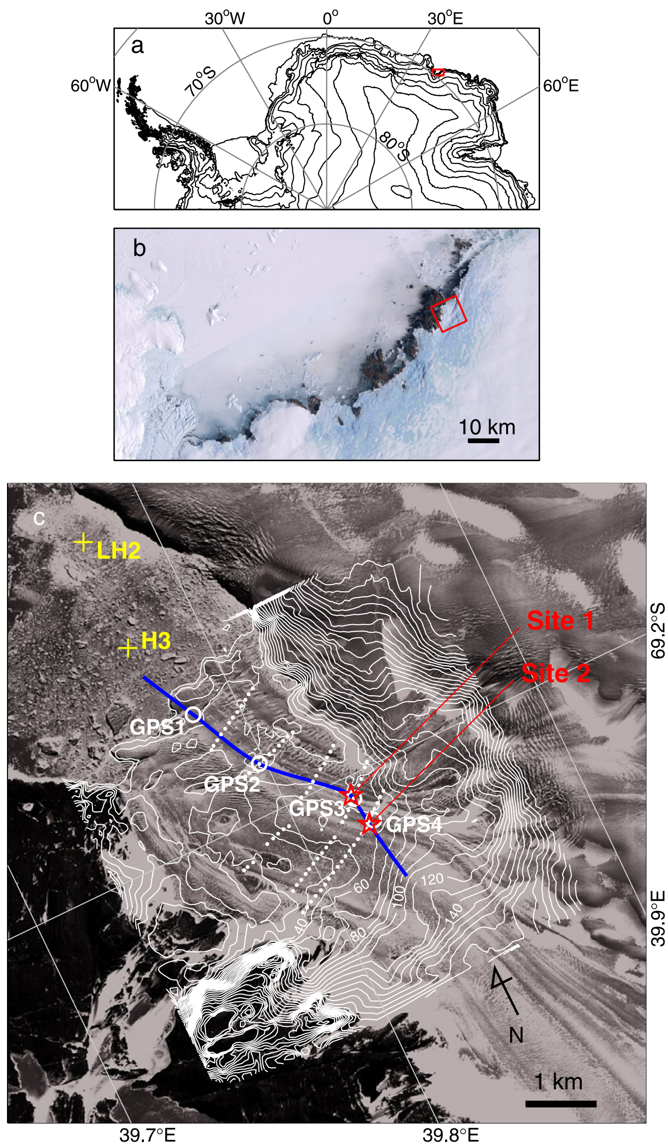

Fig. 1. The study area and measurement sites. (a) Antarctica and the location of Lützow–Holm Bay (red box, upper right). (b) Satellite image (Landsat Imagery Mosaic Antarctica) of Lützow–Holm Bay, covering the region of the box in (a). The dark color regions are ice-free land surface and the white region generally above the ice-free area is sea ice. The red box is the region in (c). (c) Locations of the drilling (stars), GPS (○), ice radar (•), and ocean measurement (+) sites. The blue line shows the location of the cross-section in Fig. 3. Ice surface contour intervals are 10 m. The background image is from ALOS PRISM, taken on 10 November, 2010. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

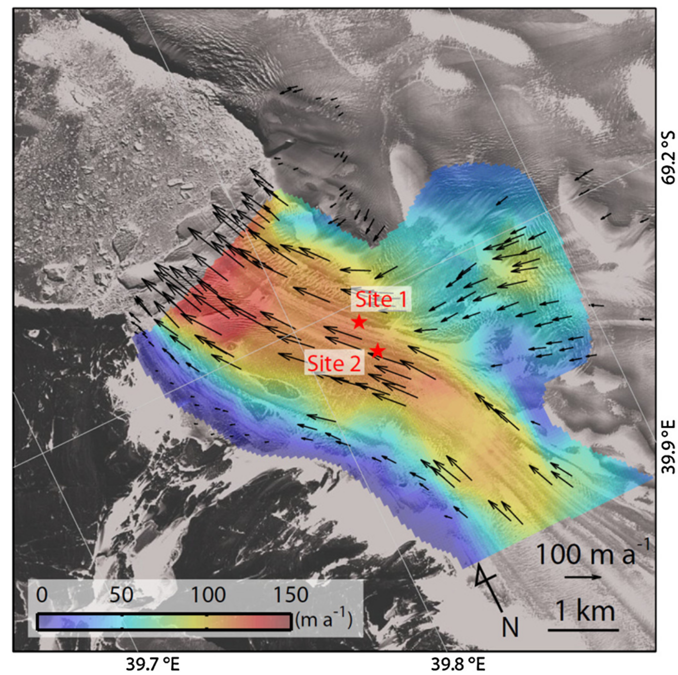

Fig. 2. Surface ice speed of Langhovde Glacier (color scale). The flow vectors were obtained from the displacement of surface features from 16 November 2006 to 19 November 2007. The color scale was determined by interpolating the flow vectors. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

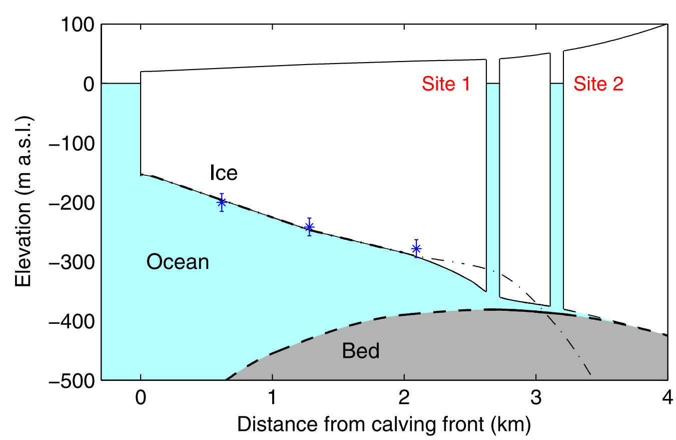

Fig. 3. Longitudinal cross-section of Langhovde Glacier along the blue line in Fig. 1(c). The ice bottom profile is based on ice radar data marked by ∗, as well as the hydrostatic equilibrium assumption (dash-dotted line) and the ice thicknesses at Sites 1 and 2. The bed elevation is based on the borehole measurements and known bathymetry near the calving front (Moriwaki and Yoshida, 1990). The ice bottom and bed elevation are extrapolated upglacier from Site 2.

2. Study site

Langhovde Glacier is a 3-km wide outlet glacier draining into Lützow–Holm Bay in East Antarctica (Fig. 1). It is a relatively small, but typical Antarctic outlet glacier. The lower 10 km of the glacier flows in a channel, which is bound by bedrock to the west and by relatively slowly moving ice to the east. In this region, ice flows at a rate up to 150 $\mathrm{m\,a^{-1}}$ as shown in Fig. 2 (see Section 3.4 for the method used to obtain the velocity map). According to our radar measurements (see Section 3.5) and a digital elevation model generated by a photogrammetric analysis of satellite images (Fukuda et al., 2014), the ice near the glacier terminus forms a floating ice shelf extending at least 2 km from the calving front. Ice in this region is 170–320 m thick, and the thickness and freeboard show the ice is in hydrostatic equilibrium (Fig. 3). Here, the glacier surface is flat and level in the lower reaches, but inland its slope increases at about 2.5 km from the terminus (Figs. 1(c) and 3). The floating ice is one to two orders of magnitude smaller in length than previously studied large ice shelves in West Antarctica (e.g. Ross, Filchner-Ronne, and Larsen Ice Shelves). The glacier terminus showed no significant retreating or advancing trend, i.e. frontal variation during 2000–2012 was within +300/−200 m (Fukuda et al., 2014). At the calving front, a bathymetric map (Moriwaki and Yoshida, 1990) shows a deep submarine trough with water depth of 602 m (and deeper offshore), but the seabed topography beneath the ice shelf is unknown.