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Evidence of unfrozen liquids and seismic anisotropy at the base of the polar ice sheets

Gérard Wittlinger [a,*], Véronique Farra [b]

[a] EOST, UMR-CNRS 7516, Université de Strasbourg, 5 rue René Descartes, 67084 Strasbourg, France
[b] IPG Paris, Paris Sorbonne Cité, UMR-CNRS 7154, 1 rue Jussieu, 75005 Paris, France

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Abstract

We analyze seismic data from broadband stations located on the Antarctic and Greenland ice sheets to determine polar ice seismic velocities. P-to-S converted waves at the ice/rock interface and inside the ice sheets and their multiples (the P-receiver functions) are used to estimate in-situ P-wave velocity (Vp) and P-to-S velocity ratio (Vp/Vs) of polar ice. We find that the polar ice sheets have a two-layer structure; an upper layer of variable thickness (about 2/3 of the total thickness) with seismic velocities close to the standard ice values, and a lower layer of approximately constant thickness with standard Vp but ~25% smaller Vs. The lower layer ceiling corresponds approximately to the −30 °C isotherm. Synthetic modeling of P-receiver functions shows that strong seismic anisotropy and low vertical S velocity are needed in the lower layer. The seismic anisotropy results from the preferred orientation of ice crystal c-axes toward the vertical. The low vertical S velocity may be due to the presence of unfrozen liquids resulting from premelting at grain joints and/or melting of chemical solutions buried in the ice. The strongly preferred ice crystal orientation fabric and the unfrozen fluids may facilitate polar ice sheet basal flow.

Keywords: Polar ice; Seismic anisotropy; Unfrozen liquids

1. Introduction

The polar ice mass balance is to first order governed by the amount of snow accumulation and the amount of ice ablation. The ice stored inland is driven by flow to the coastal line where its removal occurs mainly by basal melting and iceberg calving. The amount of flow in the polar ice sheets is thus a key parameter to estimate the ice mass balance (Rignot and Thomas, 2002; Rignot, 2006). The ice flow is preferentially vertically downward in the upper part of the ice sheets, especially near the ice divides, and turns progressively to a nearly horizontal basal flow in the lower part of the ice sheets. The basal flow depends on the mechanical properties of the ice sheet, but also on the nature of the contact between the ice and the bedrock. The polar ice sheets often lie directly on the rocky basement, but at many places, for example near the Wilkes basin, a sediment layer (~200–600 m thick) is inserted between the ice and the basement (Anandakrishnan and Winberry, 2004.). Till layers are also observed, especially beneath the outlet glaciers

* Corresponding author.
E-mail addresses: gerard.wittlinger@unistra.fr (G. Wittlinger), farra@ipgp.fr (V. Farra).

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(Walter et al., 2014 in Greenland and Alley et al., 1986 in Antarctica) and the existence of morainic blocks mixed with the basal ice is also reported (Bentley, 1971). Numerous subglacial lakes (Siegert et al., 2005; Studinger et al., 2003) some of them of huge extension seem to play an important role in ice sheet evolution (Flament et al., 2014; Stearns et al., 2008). Seismic observations may help to evaluate two key parameters controlling the flow of the ice: the crystal orientation fabric via its proxy the amount of seismic anisotropy, and the shear stiffness via its proxy the shear-wave velocity.

The viscoplastic properties of single ice-crystals are anisotropic with hexagonal symmetry (Mangeney et al., 1996; Castelnau et al., 1996). Under applied stress, the deformation rate of a single crystal is anisotropic: the ice crystal deforms more easily (about ten times, Gagliardini et al., 2009) in the directions parallel to the basal plane (i. e. in the plane perpendicular to the hexagonal symmetry axis, which is parallel to the crystal c-axis) than in other directions (Duval et al., 1983). At the scale of an aggregate of ice crystals (also called a polycrystal) the degree of viscoplastic anisotropy depends on how similar the orientations of the c-axes are. If all c-axes in the polycrystal are aligned, the viscous anisotropy is high (Lliboutry, 1993). The polycrystal then has a similar anisotropic deformation rate under applied stress to that of the single crystal (Duval et al., 1983). Under the vertical compression induced by the ice column weight, the single crystals' c-axes rotate toward the direction of compression, i. e. toward the vertical (Durand et al., 2009; Gagliardini et al., 2009). The viscoplastic anisotropy of the polycrystals gradually increases with depth. Near the free surface, the ice is macroscopically isotropic since the c-axes of the single crystals are randomly oriented; Toward the deeper part of the ice sheet, the macroscopic anisotropy is the strongest when all the c-axes are aligned in the same direction (Castelnau et al., 1996; Durand et al., 2009; Gagliardini et al., 2009).

The elastic properties of the single ice crystal are anisotropic with hexagonal symmetry (also called transverse isotropy or TI), the symmetry axis being parallel to the crystal c-axis (Gammon et al., 1983; Gagnon et al., 1988; Gagliardini et al., 2009). As for viscous anisotropy, the macroscopic elastic anisotropy of ice gradually increases from the free surface, where it is zero (isotropic ice), to the deeper part of the ice sheet. At depth it is the strongest and is close to vertical transverse isotropy (transverse isotropy with vertical symmetry axis, also called VTI) since most of the c-axes are vertically oriented (Gillet-Chaulet et al., 2005). We will retain the following parameters: for macroscopically isotropic ice, a P-wave velocity Vp of ~3.96 km/s, an S-wave velocity Vs of ~2.0 km/s, and a density of 0.92 g/cm^3 (Gagnon et al., 1988); for anisotropic ice and propagation along the c-axis direction, a P-wave velocity Vp of ~4.12 km/s, an S-wave velocity Vs of ~1.85 km/s, and a density of 0.92 g/cm^3 (Gagnon et al., 1988).

In-situ measurements of the seismic velocities are difficult to perform and are rather scarce due to the unfriendly conditions prevailing on the polar ice caps. The first measurements (Joset and Holtzscherrer (1953), Thiel and Ostenso (1961), Robinson (1968) and Kohnen (1972)), were dedicated to the determination of the P velocity in the upper part of the polar ice sheets, mainly the firn layer (King and Jarvis, 2007). The following measurements (Bentley, 1971; Blankenship et al., 1986; Anandakrishnan et al., 1998) were devoted to resolving the nature of the ice-bed rock or ice-sediments contact rather than to measuring the seismic velocities in the ice. In addition, most of these measurements were performed using active seismic sources generating only compressional waves and thus only Vp could be determined. Surface wave analysis using the ambient noise correlation technique, as applied to a Western Antarctica ice shelf by Zhan et al. (2013), is a promising method for determining the S-wave velocity of the polar ice. The P and/or S receiver function analysis, mostly used to determine the characteristics of the sediment layers inserted between the ice and the bed rock (Anandakrishnan and Winberry, 2004; Winberry and Anandakrishnan, 2003; Hansen et al., 2010) are alternative methods to measure the P and S-wave in-situ velocities.

In this paper, we revisit the seismic observations presented in our previous paper (Wittlinger and Farra, 2012) in which we used the P receiver functions (here after PRFs). The paper is constructed as follows. In Section 2, we confirm our previous seismic observations by adding several stations in Greenland and by using also S receiver functions (here after SRFs) at a few stations. In Section 3, using synthetic modeling of the PRFs, we show that the strongly preferred ice crystal orientation fabric in the lower part of the ice sheet alone cannot explain the observed low S velocity. As another mechanism must be present, we identify it to be the presence of unfrozen liquids and in Section 4 we discuss some plausible processes that may produce these liquids at very low temperatures.