# Interfacial water in polar glaciers and glacier sliding at -17 °C

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**Abstract.** We have observed sliding at a cold (-17 °C) ice-rock interface beneath Meserve Glacier, Antarctica, and the segregation of ice into clean lenses amidst the dirty basal layers of this glacier. We interpret these as manifestations of thin water films at ice-rock interfaces. We use Shreve's theory for sub-freezing sliding to estimate the nominal film thickness to be at least tens of nanometers. Such water films should exist around rocks in most polar ices, and likely have high solute concentrations due to solute rejection during regelation and due to exchange with veins and grain boundaries where impurities reside.

## Introduction

The presence of interfacial water films at the surfaces of rock particles embedded in polar glaciers and ice sheets may have important consequences for research on these ice bodies. Interfacial films can complicate the interpretation of ice core paleoclimate records by, for example, facilitating diffusion of isotopes and chemicals through the polycrystalline ice, providing a medium for chemical reactions and ion storage, and causing size-dependent migration of particles due to temperature gradients (Gilpin, 1979). Interfacial films may also affect bulk mechanical properties and help explain why the impure basal layers of ice sheets deform so easily (Dahl-Jensen and Gundestrup, 1987; Echelmeyer and Zhongxiang, 1987). The role of interfacial films in these problems remains unexplored.

Interfacial films exist due to a reduction of the chemical potential of water very close to the surface of a foreign solid (Gilpin, 1979; Wettlaufer et al., 1996), which depresses the melting point as an inverse function of the film thickness. Soluble impurities can depress the melting point and induce film growth (Beaglehole and Wilson, 1994). There is abundant experimental evidence for the presence of interfacial films between ice and foreign solids (Dash et al., 1995).

Here we report and interpret observations of two macroscopic manifestations of interfacial water in the basal layers of an active polar glacier, at a temperature of ~-17 °C. Our first observation is of glacier sliding over basal boulders, apparently facilitated by a high solute concentration in the interfacial water. In addition to allowing us to estimate the interfacial film thickness, this measurement is interesting because until recently glacier sliding was thought not to occur under cold-based glaciers. There are only two reported measurements of such sliding, at considerably higher temperatures of -4 °C and -1 °C (Echelmeyer and Zhongxiang, 1987 and Hallet et al., 1986, respectively). The second observation is of ice segregated into clean lenses amidst rock-rich layers, which we infer to result from natural filtration of rock particles by mass transfer through a very thin film at the surfaces of englacial boulders. We hope that these observations will encourage glaciologists and paleoclimatologists to consider the role of interfacial films in current analyses of ice cores and glacier deformation.

## Subfreezing Sliding: Theory

Water films at an ice-rock interface enable sliding by allowing the ice to negotiate very small roughness elements on the bed by regelation. Shreve (1984) has shown that the rate-controlling process in subfreezing sliding is the viscous flow of water in this film; the sliding rate is therefore sensitive to film thickness, $h$, the flux of water scaling with $h^3$.

Gilpin's (1979) interfacial film theory, which he calibrated using measurements of motion of wires and particles through ice, implies a melting point depression of $bh^{-\alpha}$, where $b$ and $\alpha$ are empirical constants. Shreve (1984) blended Gilpin's results with the theory of glacier sliding over a bed whose topography has an arbitrary roughness (Nye, 1969). Shreve also suggested that to incorporate the effects of dissolved solutes one can assume the contributions of interfacial forces and dissolved solutes to the chemical potential are additive. Thus, the melting point depression, $\tilde{\theta}$, *relative* to the bulk pressure-melting point for clean ice is

$$\tilde{\theta} \approx bh^{-\alpha} + M(C) \qquad (1)$$

in which the function $M$ defines the liquidus in terms of solute concentration $C$. Note that $C$ may be written

in terms of solute per area of interface $C^*$ as $C^*h^{-1}$. Though physical models of interfacial films show that effective values for $b$ and $\alpha$ will change with solute concentration (Wettlaufer, 1999), the additive form of (1) is supported by direct measurements of film thickness at a salty interface (Beaglehole and Wilson, 1994) and of unfrozen water content of salty frozen soils (Tice et al., 1984). Shreve's expression for the sliding rate $u_b$ as a function of shear stress $\tau_b$, ice viscosity $\eta_i$, and the spectral power density of bed topography $S_b$ is

$$u_b = \pi\tau_b \left[\eta_i \int_0^\infty \Omega(k)S_b(k)dk\right]^{-1} \qquad (2)$$

where the function $\Omega$, which is integrated over all wave numbers $k$, is most strongly controlled by the thickness of the interfacial water layer (Shreve's equation 12a). The derivation of (2) is entirely independent of values for Gilpin's parameters $b$ and $\alpha$, and the inferred $u_b$ is negligibly sensitive to them if one calculates the integral in terms of a specified $h$. Thus we consider equation (2) useful for quantitative estimates insofar as its assumptions are met. Two assumptions which are violated at our study site (small bedrock roughness and linear rheology) strengthen the conclusions below because they cause equation (2) to over-predict sliding rate for a given $h$ (cf. Gudmundsson, 1997).

## Sliding Observation

In the austral summer of 1995-96 we excavated a tunnel along the bed of the Meserve Glacier to expose impure basal layers for study, following Holdsworth (1974). Here we observed cavities in the lee of boulders on the bed, as reported by Holdsworth. The cavity roofs are striated, with a wavelength correlated to the grain roughness of the crystalline plutonic rocks at the up-flow margins of the cavities. This suggested sliding



Figure 1. a. Example of continuous time displacement measurement spanning four months of 1996, using an armature LVDT, anchored in the basal boulder and inserted 1 mm above ice/rock interface. The curvature is probably due to displacement of the ice into the tunnel and toward the transducer. LVDT measurements showed noise of magnitude 0.015 mm about the mean. b. The net slip during 1996 at four locations on the ice rock interface, as a function of distance upglacier from the cavity edge.

occurs at the ice-rock interface up-flow of the cavities and motivated us to install (by freeze-on) displacement markers on the ice at the interface whose positions could be measured relative to fixed bolts on the rock, using a micrometer, and to insert a displacement transducer 1 mm above the bed, and another at 30 mm. These were all just upstream of a prominent cavity, at a distance of approximately 16 meters from the ice margin, which is a 20-meter high cliff. We also inserted a displacement transducer 1 mm above the bed at a location on the opposite side of the tunnel, not upstream of a cavity. All these markers moved downglacier over the periods of measurement (up to ten months) at a rate more than an order of magnitude faster than that due to shear strain (which has a maximum observed value of $0.2a^{-1}$ here), and therefore were measuring sliding (Figure 1). Enhanced shear at the interface due to a more favourable orientation of ice crystals there (a possibility raised by Barnes et al., 1971) cannot explain our data because the fabric throughout the ice where strain rate measurements were made is already a tight, vertical single maximum and therefore already as soft as possible.
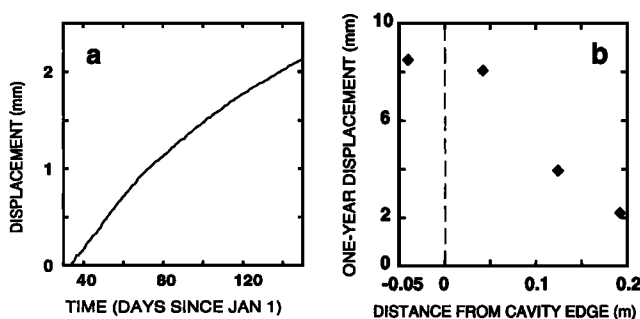
The markers showed a sliding rate of about 8 mm·a$^{-1}$ (parallel to the tunnel) near the lip of the prominent cavity, decreasing to 2 mm·a$^{-1}$ some 20 cm up-flow, away from the stress perturbation associated with the cavity. This is quite similar to the 3 mm·a$^{-1}$ measured on the opposite side of the tunnel, and so is an appropriate value for sliding rate driven by the average basal shear stress.
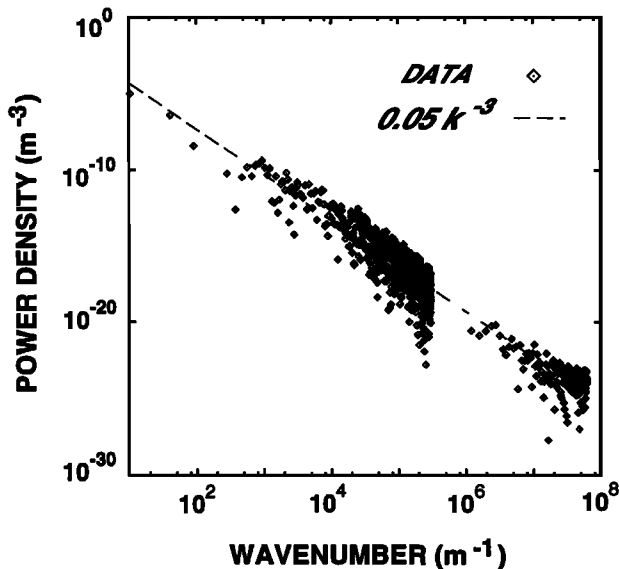
## Interpretation of Sliding

The measured sliding rate is considerably higher than one would estimate from Gilpin's results (0.2 mm·a$^{-1}$ according to Shreve (1984)) and deserves explanation. The principal unknown quantity is the mean thickness of the interfacial film, and we now use equation (2) to estimate it. We first measured the detailed topography of the rock surface beside the displacement markers and from this calculated the spectral power density $S_b$. The resulting $S_b$ (Figure 2) approximates a power-law over seven orders of magnitude, and is on average $S_b(k) = 0.05k^{-3}$. This is rougher than the example bed used by Shreve, and so the observed rapid sliding is not a consequence of an unusually smooth interface.

The actual measurement of rock topography was done (mostly on two separate samples of the surface) using four methods at different spatial scales: tape and compass survey ($k < 10^2$ m$^{-1}$), contour tool profiling (for $10^2 < k < 10^3$), optical interferometry (Zygometrics NewView 200; $10^3 < k < 3 \times 10^5$), and contact-mode Atomic Force Microscopy ($10^6 < k < 5 \times 10^7$). Wavenumbers $10^2 < k < 10^5$m$^{-1}$ proved to be the only significant ones in the integral (2).

For the $\eta_i$ it is not appropriate to use standard tabulated values as the basal ices of the Meserve Glacier

**Figure 2.** Measured spectral power density of rock surface. For consistency with Nye (1969), wavenumbers $k$ are defined unconventially as the inverse of wavelength, without a coefficient of $2\pi$.

are unusually soft (Holdsworth, 1974). Instead we use $\eta_i = \tau_b/n\dot{\epsilon}_b$, in which $\dot{\epsilon}_b$ is our in-situ measurement of shear strain rate for ice of this impurity content $(0.1a^{-1})$, and $n=3$ is the non-linearity of ice creep. The down-flow (tunnel-parallel) shear stress is a multiple $E$ of the weight-slope product. $E$ is an enhancement term accounting for the presence of the ice cliff, and is 1.4 to 2 according to two-dimensional finite element calculations we have made for non-linear creep of a sloping ice body with a terminal cliff. We use the low estimate $E = 1.4$ because this gives a lower limit for $\eta_i$ and hence $h$. We find $\eta_i \approx 4 \times 10^{12}$ Pa s. This substitution for $\eta_i$ eliminates $\tau_b$ from the numerator in (2), but $\eta_i$ remains an important parameter in $\Omega$.

For these values, and the measured spectral power density, (2) indicates that a liquid film of thickness approximately 20 to 40 nm is necessary for sliding to occur at the measured rate (with estimated error bars of -10 and +150). All plausible thicknesses are sufficiently large to make the interfacial term in (1) very small (20 nm is one order of magnitude larger than the film thickness predicted from Gilpin, 1979). Consequently the interfacial film likely has a very high solute content (a situation analogous to that produced in experiments of Beaglehole and Wilson, 1994). If this is the primary reason for the thick film, its solute concentration must be $C \approx \bar{\theta}/M$, which is approximately 3 moles/L based on the NaCl liquidus (70% of soluble impurities here are NaCl). The relatively rapid sliding of this polar glacier in this view is made possible by a brine layer at the ice-rock interface.

Though high, a 3 molar solute concentration is plausible; the total salt per area of bed in the layer, approximately $9 \times 10^{-5}$ moles $m^{-2}$, is equivalent to the

salt per unit area in a $10^{-3}$m thickness of ice, which is approximately 1/3 the diameter of ice crystals here (the bulk solute concentration of ice just above the sliding interface is $10^{-4}$ molar, as measured by the U. New Hampshire glaciochemistry lab (Buck et al., 1992)).

The impurity concentration in grain boundaries and veins in polycrystalline ice is several orders of magnitude higher than that of the crystal lattice (Alley et al., 1986). Further, as ice freezes from water, as during regelation cycles, it rejects solutes at a very high efficiency (Gross et al., 1975). Interfacial films will generally intersect grain boundaries, allowing solutes to exchange. Because of this, and the refreezing accompanying regelation, the solute concentration of interfacial films will generally be orders of magnitude higher than the bulk concentration of adjacent ice, and this makes for generally thicker films, especially in ices with a temperature not far from the bulk phase boundary.

Thus solutes are also a likely explanation for the high sliding rate measured by Echelmeyer and Zhongxiang (1987). Without solutes they could only account for 10% of their measured 0.18 m·a$^{-1}$ slip rate, despite the relatively low roughness of this interface $(0.004\ k^{-3})$. At -4 °C, solutes will dominate film thickness at concentrations of $\sim 0.5$ mol/L (for NaCl), which is $10^4$ times greater than typical solute concentrations of glaciers in the Tien Shan (Wake et al., 1989). A $10^4$ concentration difference between grain boundaries and bulk ice is expected (Gross et al., 1975; Alley et al., 1986).

## Ice Segregation

Water films also can explain structures observed within the glacial ice. The dirtiest basal layers of the Meserve Glacier contain $\sim 2\%$ by volume of finely dispersed silt-sized rock particles, which give these layers a pronounced yellow-brown color (Holdsworth, 1974). Within these layers are sparsely distributed larger rock clasts. The shear of ice around these clasts induces two pressure shadows on opposite sides of the rock surface, and it is common to observe small cavities separating the clast and ice in these positions (Holdsworth, 1974).

If an interfacial film is present, the pressure gradient along the rock surface will drive a net flow toward the cavities. A small rate of net melt must occur adjacent to the interface to maintain film thickness. At the cavity, the water will accumulate as ice, either by direct refreezing or by evaporation into the cavity and subsequent deposition. The small thickness of the water film will prevent transport of all but the finest particles so the accumulating ice will be free of them; the rock surface and its interfacial film act as a natural filtering system.

We have observed very pronounced clear ice zones, which stand in stark contrast to the adjacent amber-colored ice, adjacent to the low-pressure regions of clasts, and tailing off as thin zones far up- and down-flow (similar to S-shaped strain figures seen in metamor-

phic rocks). Gow et al. (1997) report observing similar features in the basal ices of central Greenland (temperature $\approx$ -9 °C). It is likely that these clear ice zones are accumulations of the sort envisioned here. This process may have an important role in the formation of foliation in creeping dirty glacier ice and other heterogeneous media.

## Conclusion

Observations of glacier sliding and ice segregation beneath Meserve Glacier are best explained as manifestations of unfrozen water in films at ice-rock interfaces, despite the low temperature. Estimated film thickness is at least 20-40 nm. Here these water-dependent phenomena are strengthened and made easily discernible by the unusually high solute content of this ice, but interfacial films and resultant phenomena should be present in most polar ice masses. Films in most natural glacier ices are likely thicker than implied by Gilpin (1979) due to concentrated impurities.

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