Hydraulic conditions at the base of Jakobshavn Isbræ, West Greenland

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DRAFT: May 17, 2004, Version: 1.4, 2004/05/14

Abstract

Large areas surrounding Jakobshavn Isbrae move faster than can be explained by ice deformation alone. Borehole deformation measurements at our drill site adjacent to the ice stream revealed that up to 65 % of the surface velocity is due to basal motion. All eight boreholes immediately connected to a well developed subglacial drainage system at, or close to, overburden pressure. Pressure pulses from draining boreholes propagated to neighboring holes within minutes. Measured variations of subglacial water pressure during borehole drainage and pump tests were interpreted with standard methods. The derived hydraulic transmissivity is about $7.5 \cdot 10^{-5}\,\mathrm{m}^2\,\mathrm{s}^{-1}$. This value corresponds to either a coarse grained sediment, or alternatively to a system of gaps or linked cavities. In contrast to the distributed drainage system inferred from these data, earlier measurements indicated a channelized drainage system at the ice stream margins.

1 Introduction

The fast flow of ice streams is often attributed to high rates of basal motion. This assumption is certainly true for the Siple Coast ice streams in West Antarctica (e.g. Alley and Whillans, 1991; Engelhardt and Kamb, 1998). In contrast, ice streams that flow through deeply incised bedrock troughs show a large fraction of internal deformation (e.g. Truffer and Echelmeyer, 2003). In situ borehole measurements and numerical modeling lead to the conclusion that the fast flow of Jakobshavn Isbræ (West Greenland) is mainly due to internal ice deformation (Iken and others, 1993; Funk and others, 1994; Lüthi and others, 2002).

Unlike the Antarctic Siple Coast ice streams that plow through nearly stagnant ice, the ice sheet surrounding Jakobshavn Isbræ moves at high velocities of up to $500\,\mathrm{m\,a^{-1}}$ (Fastook and Brecher, 1995). The fast flow of the ice sheet is only partly due to internal deformation, and cannot be explained without motion of the ice over the base, or the deformation of the underlying sediments. Both processes require melting point temperatures and the presence of liquid water at high pressure. Indeed, temperate bed conditions in the Jakobshavn basin were predicted by all thermo-mechanically coupled models as an effect of the geometry, the surface temperature and the mass balance distribution alone (Budd and others, 1982; Letreguilly and others, 1991; Funk and others, 1994; Greve, 1997; Wang and others, 2002). The areas in close neighborhood of Jakobshavn Isbræ experience high lateral drag from the ice stream. The stress transfer takes mainly place in the cold, stiff upper 60% of the ice. Model calculations showed that the ice is partly pulled over the base. There is also indication of internal fracturing along the highly stressed horizontal interface separating Holocene and Wisconsin ice (Lüthi and others, 2003).

No seasonal changes of surface velocity could be observed on the ice stream (Echelmeyer and Harrison, 1990). In contrast, marked seasonal variations were measured at site D (Lüthi and others, 2002), as well as in the ice sheet 40 km north of the ice stream (Zwally and others, 2002). These velocity variations are thought to be closely linked to the input of meltwater from the surface to the subglacial hydraulic system. In this paper we present measurements of water pressure at the bottom of boreholes drilled at site D. Interpretation of these data allows us to draw conclusions on the nature of the subglacial hydraulic system.

2 Background

Jakobshavn Isbræ is one of the fastest moving outlet ice streams. It drains about 6.5% of the Greenland ice sheet into Disko Bay (Carbonell and Bauer, 1968; Fastook and others, 1995). The ice fluxes at the calving front are estimated to be in the order of $25 - 28\,\mathrm{km}^3\,\mathrm{a}^{-1}$ (Carbonell and Bauer, 1968; Lingle and others, 1981; Echelmeyer and Harrison, 1990; Echelmeyer and others, 1992). Jakobshavn Isbræ flows through a deeply eroded bedrock trough which extends about 80 km inland of the calving front, as indicated in Figure 1a (Clarke and Echelmeyer, 1996).

Recently, a marked increase of flow velocities of the floating terminus from about $7\,\mathrm{km\,a^{-1}}$ to $10\,\mathrm{km\,a^{-1}}$, together with a rapid retreat of the calving front, has been observed (Abdalati et al., 2001; Thomas et al., 2003). The cause for this large change in dynamics is subject of several current studies (e.g. Thomas, in press).

3 Observations

The measurements on the subglacial hydraulic conditions presented here are a by-product of a deep-drilling effort aimed at the dynamics of Jakobshavn Isbræ. These measurements, the borehole instrumentation, and conclusions on many aspects of the flow of sheet and stream have been described in detail in Lüthi and others (2002).

To obtain information about an inaccessible aquifer, the most common method is to disturb the water level in a borehole and to measure the response in the same borehole or in adjacent locations (e.g. Marsily, 1986). The velocity and form of pressure pulses traveling to neighboring sites provides information about the effective hydraulic conductivity of the glacier substrate and the type of drainage system. Great depths and the very cold ice in polar ice streams render this kind of experiment a non-trivial task. Theoretical formulations for the special setting of glaciers and ice streams as well as measurements in such areas have been published (e.g. Stone and Clarke, 1993; Waddington and Clarke, 1997; Engelhardt and Kamb, 1997).

3.1 Instrumentation

During our 1995 field program, eight boreholes were drilled to the bedrock at site D, north of the crevassed margin of Jakobshavn Isbræ (Fig. 1). Two boreholes, spaced 15 m at the surface, were permanently instrumented with tilt cells, thermistors and with pressure sensors P1 and P2 which were in contact with the subglacial water drainage system. The sensing units in both boreholes were operated with identical, but completely independent digital data transmission and logging systems (details are given in Lüthi and others, 2002).

3.2 Borehole water levels

Water pressure was registered for five days in a 10 or 20 minutes interval at both pressure sensors P1 and P2. Additionally, water levels were recorded in irregular intervals with a float-switch suspended from a cable while the boreholes were open. On 25.7.1995 the measurement interval was switched to three days in order to provide the data logger storage capacity for one year of continuous measurement.

The water levels measured at pressure sensors P1 and P2 are shown in Figure 3. No free water surface existed after refreezing of the boreholes some hours after installation of the instruments. Nevertheless, all pressure readings are given in units of meter water level below surface, since the pressure transducers were calibrated during lowering into the boreholes with depth readings of the marked cable. Offset water levels measured in the open borehole with the float-switch were corrected for in the data processing. The resolution of the water level measurements with the digital transmission system was ~ 0.5 m and the linear calibration curve has a standard deviation of 0.7 m, so that the absolute accuracy of the measurements is about 1 m. Water levels measured with the float-switch are accurate to some centimeters.

Pressure measurements in both boreholes are nearly identical until the end of

August (Fig. 3a). Three pressure peaks on July 20, 22 and 24 were caused by draining nearby boreholes, after drilling reached near bedrock (holes 5 (I1), 6 (I2) and 7 in Fig. 2). The rise of water level is very fast, followed by a slow decrease, until after 14 hours the initial water level is precisely recovered. The equilibrium water level of $75.0 \pm 0.2\,\mathrm{m}$ below the surface corresponds to the flotation level at the ice thickness of $h_{\mathrm{ice}} = 831 \pm 2\,\mathrm{m}$, if the average ice density is $\rho_{\mathrm{ice}} = (h_{\mathrm{w}}/h_{\mathrm{ice}})\,\rho_{\mathrm{w}} \simeq 910\,\mathrm{kg\,m^{-3}}$, where the density of water at 0° C of $\rho_{\mathrm{w}} = 999.841\,\mathrm{kg\,m^{-3}}$ has been used and the compressibility of water has been neglected. Assuming on the other hand an ice density of $917\,\mathrm{kg\,m^{-3}}$, a borehole water level of 68.8 m is calculated at floating equilibrium. Therefore the final borehole water levels of 75 m would correspond to an an ice overpressure of roughly 70 kPa, or 99 % of the overburden pressure.

A steady increase in basal water pressure of $\sim 0.15\,\mathrm{m\,d^{-1}}$ water level was observed during August at both sensors. This is likely to be caused by the thickening of the overlaying ice column as the instruments wander downslope a subglacial hill. Under the assumption of flotation pressure, a basal sliding velocity of about $350\,\mathrm{m\,a^{-1}}$ (measured by Lüthi and others, 2002) and a surface slope of 1° this corresponds to a local bedrock inclination of about 10°.

The large pressure variations of $1.8 \cdot 10^5$ Pa registered at pressure sensor P1, starting on August 29, would lower the freezing temperature by 0.014 K. The fact that all thermistors near the bed in the same borehole show a temperature decrease of this order indicates that the registered event is not an artefact caused by an instrument error. Temperatures measured at thermistor T12 near the bed and converted to pressure (using the Clausius-Clapeyron equation of melt point depression) are indicated with a dotted line in Figure 3. The cable broke after this sudden pressure increase on September 10, and the cable of sensor P2 broke several days later.

3.2.1 Water pressure variations

The sharp increase in pressure is either caused by some bedrock irregularity or by large water pressure variations in the subglacial system. Such a pressure variation could occur when a lake at the ice surface suddenly drains through freshly opened crevasses (Echelmeyer and Harrison, 1990), releasing huge amounts of water to the glacier base and thus inducing a local strong increase of sliding motion. The drainage of one nearby lake was witnessed at the end of the field campaign, however, the recording interval of the pressure sensors was already set to three days, so that an occasional pressure pulse could not be registered.

During the quiescent phases between two breakthrough events small diurnal variations in basal pressure are visible in the Figure 3b. The amplitude of these variations is less than 1 m water level, but clearly follows a daily pattern with the lowest pressures recorded during the night. Attempts to correlate the data with tidal data from the Godhavn gauge station failed (even with a varying time lag of up to one day).

3.3 Breakthrough events

All boreholes connected immediately to the subglacial drainage system when they reached the bedrock. The fast drop of borehole water level was also recorded with a force meter on the winch measuring the tension of the drilling hose. These events, releasing water to the glacier base, will be further referred to as breakthrough events. The borehole water level dropped to a depth between 67 and 69 m below the surface within several minutes where it remained approximately constant until the water supply into the hole stopped. This primary water pulse is estimated to be between 5 and 8 m^3, calculated with a uniform borehole radius near the surface of $r_s = 0.15$ to $0.2\,\mathrm{m}$. After this initial phase a slow decrease in water level was observed until equilibrium was reached.

Four breakthrough events are plotted in Figures 4 which were recorded in sufficient detail to provide information on the nature of the subglacial water system. The water levels measured in the open boreholes with a float-switch are indicated with dots, the moment of breakthrough with a dotted line and the shutdown of the pumps (stop of water supply) is marked with an arrow. Pressure sensor P1 installed at the base of borehole 5 (I1) recorded water levels in a 10 or 20 minutes interval, starting two hours after the breakthrough. Similarly pressure sensor P2 in borehole 6 (I2) started recording in a 20 minutes interval two hours after breakthrough.

Pressure sensors P1 and P2 installed at the base of boreholes spaced some 20 m at the surface registered simultaneously all breakthrough events in less than a sampling interval of 20 minutes. Closer inspection of Figure 4 shows that the pressure pulse propagated within two minutes to both pressure sensors in some tens of meters distance after breakthrough of holes 6 and 7. While the distance of the holes at the glacier base is not known, the chance of three holes joining in close neighborhood is small. We thus prefer the explanation of a rapid pressure pulse in a developed subglacial drainage system. Similar observations of the rapid propagation of a pressure pulse to neighboring boreholes within some minutes have been made on Trapridge glacier in boreholes spaced some 20 m (Stone and Clarke, 1993) and ice stream B, Antarctica, in holes spaced 14 and 39 m (Engelhardt and Kamb, 1997).

4 Interpretation

Pressure variations recorded at both sensors P1 and P2 are in close accordance during the whole measurement period. The maximum observed difference is 0.5 m water level and corresponds to the resolution of the data transmission system. The only exception is the large pressure pulse recorded only at sensor P1 at the end of the measurement period. The spacing of the boreholes at the glacier bed is unknown, but it seems reasonable to assume the same distance as at the surface ($\sim$ 20 m). The close similarity of both pressure logs is indicative of a hydraulical connection between the sensors. This requires an active subglacial water system, consisting either of a porous sediment layer, a linked cavity system, or a single larger channel.

The rapid initial drop in borehole water levels is only possible if the water system beneath the glacier provides the required storage capacity. The fast propagation of the pressure pulse to the neighboring pressure sensors is strongly suggestive of some preexisting, hydraulically active subglacial drainage system or, alternatively, of a rapid separation of ice from the glacier bed in the sense of the gap opening model proposed by Engelhardt and Kamb (1997).

The observation that the water level stabilized at overburden pressure after each breakthrough event and in all boreholes excludes the existence of a large conduit passing near the drill site. Such a conduit would lower the water level below overburden pressure and would also drain the water away faster than observed. The absence of pressure variations is unlikely to occur in a large channel, which is thought to be influenced by a short term variation of water flux. Such a large channel was hypothesized in order to explain the water level record at the 1989 ice stream drill site A (Iken and others, 1993).

In the following sections an attempt is made to decide on the nature of the subglacial drainage system. This is achieved by interpretation of the borehole water level logs in terms of steady or transient responses to the water input during drilling.

Simple conceptual models for a spatially uniform subglacial drainage system are either a porous sediment layer or a local small gap between the base of the ice and bedrock. In reality, a uniform gap would be unstable and would evolve to a system of linked cavities (Walder, 1986; Kamb, 1987). Such a system could also be expected at the prevailing high sliding velocities of about $350\,\mathrm{m\,a^{-1}}$ measured at the drill site (Lüthi and others, 2002).

Both conceptual models can be treated with the same methods if the water flow in the hypothetical gap or sediment is laminar. The apparent hydraulic transmissivity T_{gap} of a uniform horizontal gap of width δ is (de Marsily, 1986)

$$T_{\mathrm{gap}} = \frac{F\delta^3}{12}\frac{\rho_{\mathrm{w}} g}{\eta_{\mathrm{w}}} = F\delta^3 \cdot 4.5 \cdot 10^5\,\mathrm{m^{-3}}\,, \tag{1}$$

where $\rho_{\mathrm{w}} = 999.841\,\mathrm{kg\,m^{-3}}$ and $\eta_{\mathrm{w}} = 0.0018\,\mathrm{Pa\,s}$ are the density and viscosity of water at $0°$ C, $g = 9.82\,\mathrm{m\,s^{-2}}$ is the gravitational acceleration and F is the fraction of the glacier bed occupied by the gap. This relation allows to interpret the measurements in terms of both simplified models and therefore to obtain rough estimates of the dimensions and the nature of the subglacial drainage system.

4.1 Pumping test

During the recovery of the drilling hose the water flux into the borehole was approximately $80\,\mathrm{l\,min^{-1}}$, provided by four regulated pumps. The constant water level until the shutdown of the drilling system (marked with arrows in Figures 4) corresponds to a pumping test under nearly steady conditions, from which information on the nature of the subglacial hydraulic system can be drawn. The small increase in water level at the moment when pump operation stopped is due to our emptying the drill hoses into the boreholes, thus releasing additional $0.25\,\mathrm{m}^3$ of water.

Under the simplifying assumption of spatial uniformity (i.e. a sediment layer or a gap of constant width), the variations in water level can be interpreted in terms of laminar flow if the flow velocities are small. The Reynold number (indicating turbulent flow for $Re > 2000$) can be calculated for the initial constant water input of $Q = 80\,\mathrm{l\,min}^{-1} = 1.33 \cdot 10^{-3}\,\mathrm{m}^3\,\mathrm{s}^{-1}$. If the borehole is considered as a long cylindrical pipe of radius r_h, the Reynolds number at the water flow velocity v is (de Marsily, 1986)

$$Re = \frac{v\,2r_h\rho_\mathrm{w}}{\eta_\mathrm{w}} = \frac{2Q\rho_\mathrm{w}}{\pi\eta_\mathrm{w}r_h} = \frac{475\,\mathrm{m}}{r_h}\,. \tag{2}$$

In the last equality the actual water flux and the physical properties of water were inserted, and the borehole radius r_h at the bed is about $0.05\,\mathrm{m}$. Thus the Reynolds number is of the order 9500 and the flow is turbulent. Similarly in an open gap of thickness δ between the glacier and the bedrock, the Reynolds number in a distance R from the borehole is approximately (de Marsily, 1986; Iken and others, 1996)

$$Re = \frac{2QD_h\rho_\mathrm{w}}{2FR\pi\delta\eta_\mathrm{w}} \simeq \frac{Q\rho_\mathrm{w}}{FR\pi\eta_\mathrm{w}} \simeq \frac{240\,\mathrm{m}}{FR}\,. \tag{3}$$

Here $D_h \sim 2\delta$ is the hydraulic diameter of the water film and F designates the hydraulic active fraction of the glacier bed. If $F = 0.1$ (a rather low value) the flow is turbulent within a critical distance $R_\mathrm{crit} = 1\,\mathrm{m}$ from the borehole. Outside of this turbulent zone the flow is laminar.

The interpretation of borehole water levels during constant water input Q in terms of a sediment layer of thickness b and hydraulic conductivity K_s allows to determine the hydraulic transmissivity $T_s = bK_s$. Under the assumption that the system has reached a steady state, the water level in the open borehole will stabilize at a depth h below the surface, which will be higher than the equilibrium level h_{eq}. The hydraulic head $s := h - h_{\mathrm{eq}}$ observed in the borehole is given by Dupuit's formula (de Marsily, 1986, p. 165)

$$s = \frac{Q}{2\pi T_s} \ln \frac{R}{r_0} \tag{4}$$

where r_0 is the effective radius of the borehole and R is the radius of action, i.e. the zone affected by the water input. In practice r_0 will deviate from the borehole radius as it includes the effects of turbulence. With $s = 7\,\mathrm{m}$ (the mean hydraulic head during the pumping tests) and $Q = 1.33 \cdot 10^{-3}\,\mathrm{m}^3\,\mathrm{s}^{-1}$, the hydraulic transmissivity is

$$T_s = \frac{Q}{2\pi s} \ln \frac{R}{r_0} \sim 3 \cdot 10^{-5} \ln \frac{R}{r_0}\,\mathrm{m}^2\,\mathrm{s}^{-1}\,. \tag{5}$$

Assuming a borehole radius at the glacier base of $r_0 = 0.1\,\mathrm{m}$ and a radius of action $R = 10$ to $1000\,\mathrm{m}$, reasonable choices of R/r_0 are $1 \cdot 10^2$ to $1 \cdot 10^5$ and the last factor is between 4.6 and 11. Thus the hydraulic transmissivities are of the order $1.4 - 3.5 \cdot 10^{-4}\,\mathrm{m}^2\,\mathrm{s}^{-1}$.

4.2 Recovery phase

After the shutdown of the pumps the water input stopped and the undisturbed borehole water level was slowly recovered. This phase of the pumping test is conveniently interpreted with an exponential decay in the early recovery phase and with Jacob's method in later stages.

Under the assumption of a constant water flux from the borehole into the sediment for some initial time interval, the drop in water level after shutdown of the pumps can be evaluated as an exponential decay. The drop in hydraulic head s then is

$$s(t) = s_0 \exp \frac{-t}{D} \tag{6}$$

with a time constant D and the initial hydraulic head s_0 attained during pumping. Plotting the borehole water level records on a logarithmic vertical axis in Figure 5 (left plots), the initial water level variation can be fitted with a time constant of $D \sim 9000\,\mathrm{s}$.

The hydraulic conductivity can be calculated from this value with the expression (Engelhardt and Kamb, 1997)

$$T_s = \frac{r_s^2}{2D} \ln \frac{R}{r_0}\,. \tag{7}$$

Assuming a borehole radius near the surface of $r_s = 0.2\,\mathrm{m}$ and an effective radius of the same order as in equation (5), a transmittivity of 1 to $2.5 \cdot 10^{-5}\,\mathrm{m^2\,s^{-1}}$ results. These values are an order of magnitude lower than those of the steady pumping test.

For the late stages of the recovery phase, Jacob's method is used (de Marsily, 1986, p. 168). The difference in hydraulic head at two times t_1 and t_2 is given by

$$s_2 - s_1 = \frac{Q}{4\pi T_s} \ln \frac{t_2}{t_1}\,. \tag{8}$$

Plotting the hydraulic head against time on a logarithmic scale in Figure 5 (right plots) allows to determine the transmissivity T_s which is about $7.5 \cdot 10^{-5}\,\mathrm{m^2\,s^{-1}}$ for all three boreholes. These values are 2 to 4 times smaller than those obtained for the constant flux (Eq. 5) and the assumed ratio of radii. Matching the transmissivity obtained for the recovery phase would give $R/r_0 = 12$ and thus a very localized influence of the draining water.

5 Discussion

Based on the observations of borehole water levels the existence of a large subglacial channel at drill site D was excluded. Simple conceptual models for a uniform sediment layer or a gap conduit both give realistic values for the hydraulic conductivity or the gap width. Hydraulic transmissivity derived from borehole water level drop using Jacob's method is about $7.5 \cdot 10^{-5}\,\mathrm{m}^2\,\mathrm{s}^{-1}$ which corresponds to the conductivity of medium sized sand. These values compare well to hydraulic conductivities measured at several other glaciers (Table 1 in Fischer and others, 1998). Interpreting the above results in terms of a gap of width δ with Equation (1) yields gap widths of $0.7 - 0.9\,\mathrm{mm}$ for the steady water input and $0.5\,\mathrm{mm}$ for the recovery phase. The additional assumption that the gaps are only operative at a fraction F of the glacier base alters the above results, resulting in gap widths of about $5 - 9\,\mathrm{mm}$ for $F = 0.1$.

We have no means to finally decide on the nature of the drainage system. The high observed basal sliding velocities are compliant with both hypotheses. On the other hand the large storage capacities needed to absorb the initial water pulse after the breakthrough events, as well as the fast propagation of the pressure pulse to neighboring boreholes, are conveniently explained with a system of gaps or linked cavities (Kamb, 1987).

6 Conclusions

High rates of basal motion ($350\,\mathrm{m\,a}^{-1}$, or 65 % of the surface velocity) have been inferred from borehole inclination measurements at site D (Lüthi and others, 2002). These velocities are due to basal processes that rely on a highly pressurized subglacial drainage system. We found a well developed subglacial drainage system, operating at 99 to 100 % of the overburden pressure. Small daily variations of water pressure, and seasonal variations of velocity are indicative of the high susceptibility of the fast-moving ice sheet. This means that large areas of the ice sheet margin might be sensitive to increased meltwater influx, as the melt season gets longer and/or more intense due to a changing climate. Interseasonal and interannual variations of surface velocities observed at Swiss Camp (located outside the influence of the ice stream near the equilibrium line) have been interpreted in that sense (Zwally and others, 2002).

On the other hand, high melt water influx does not necessarily lead to high rates of basal motion for extended periods of time. Large moulins drain meltwater streams at the glacier surface in the ablation area (Thomsen and others, 1988). This water is most likely to flow in large intra- or subglacial streams to the calving terminus (Echelmeyer and Harrison, 1990). As is well known from mountain glaciers (and is likely to be true for the ice sheet) the subglacial drainage system evolves to a channelized system with lower water pressure, and therefore reduces basal motion (e.g. Paterson, 1994).

Long term measurements of basal water pressure in the temperate ice sheet are desirable to resolve seasonal and interannual changes. These are, however, difficult to achieve, since cables will break due to the large ice deformation. Alternatively, much information could be gained by disturbing the basal hydraulic system. This happens naturally when large lakes at the glacier surface drain (e.g. Echelmeyer and Harrison, 1990). Controlled drainage of a lake into a borehole with a flow control gage would allow to gradually change the hydraulic conditions. By monitoring closely the basal water pressure in neighboring sites, horizontal and vertical surface displacements, ice deformation and seismicity, the nature of the basal hydraulic system could be explored. This would allow to test theories of basal motion, and constrain the sensitivity of the ice sheet to increased meltwater input.

Acknowledgements

We express our thanks to all who contributed to this project. The electronic transmission system was designed and built by H.P. Hächler and C. Senn. Assistance in the field was given by M. Truffer, H. Bösch, P. Haffner, B. and H. Jenny, J. Luthiger, and C. Senn. We acknowledge the support by KNI and the pilots and staff of Greenlandair.

This project was funded by Swiss National Science Foundation Grant No 21-39582.93 and ETH Zurich Grant No 0-20-982-95.

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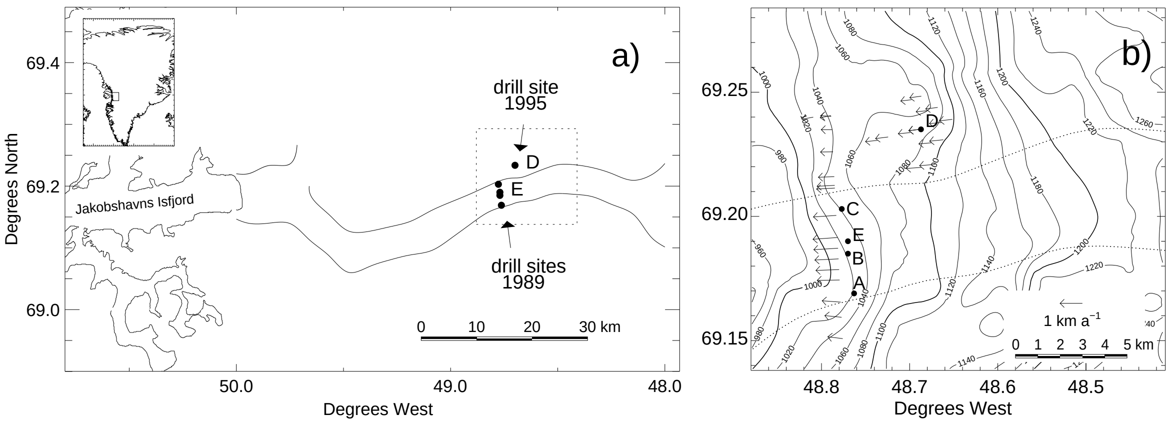

Figure 1: *a) Location of the drill site D north of Jakobshavn Isbræ. The ice stream is indicated with solid lines inland from Jakobshavn Isfjord. b) Enlargement of the dotted frame. The surface topography has been interpolated from laser altimeter data of NASA (see text for references). Contour lines are given in meters a.s.l. and ice stream margins are indicated with dotted lines. Velocities measured at stakes on the glacier surface are indicated along a transverse profile over the ice stream (Clarke and Echelmeyer, 1996) and around drill site D.*

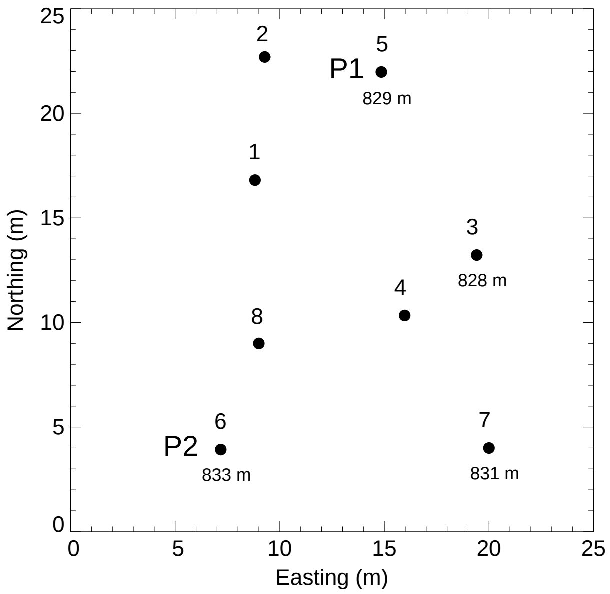

Figure 2: *The locations of boreholes 1 to 8 are indicated with dots in a local coordinate system. Pressure sensors were installed at the bottom of boreholes P1 and P2. Borehole depths are given where measured.*

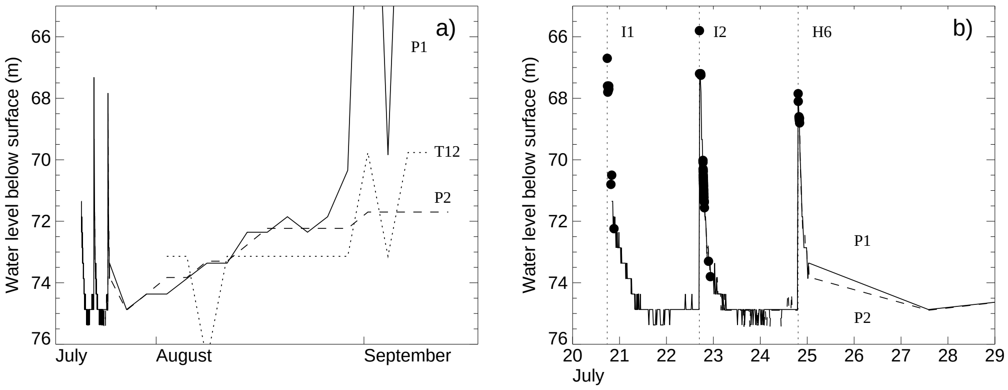

Figure 3: *a) Water pressure was measured in two boreholes (P1, P2) for nearly two months. The initial spikes stem from breakthrough of several nearby boreholes. A slow increase in water pressure is terminated by a large spike, prior to the rupture of the sensor cables. The dotted line shows the temperature variation measured at thermistor T12, rescaled to water level with the dependence of melting temperature on pressure. b) The water pressure stabilizes at the same level after breakthrough of boreholes 5, 6, and 7, marked with vertical dotted lines. Small diurnal variations of the pressure are also visible in the pressure record. Water levels measured with the float switch are indicated with points.*

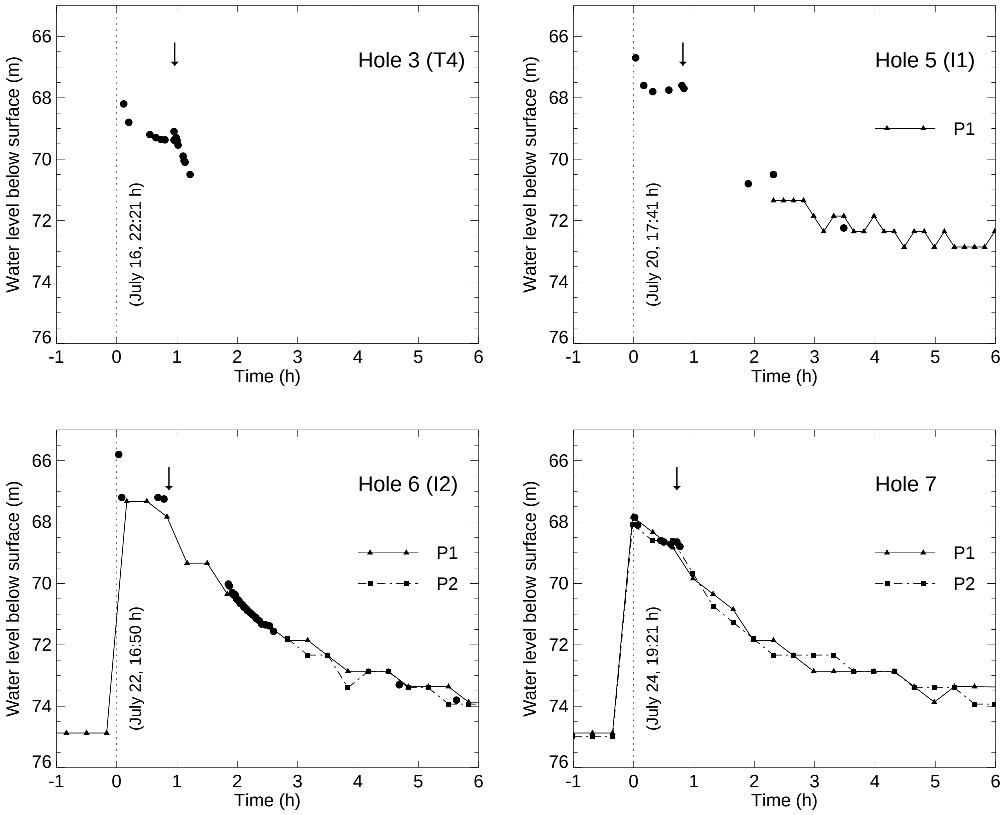

Figure 4: Water breakthrough events were recorded in four open boreholes with a float-switch (black dots). The water level in the open borehole is plotted versus time in hours after breakthrough (date and time of breakthrough are also given). The pressure sensor P1 installed in hole 5 (I1, the relative location of the boreholes is shown in Fig. 2) registered two additional breakthrough events from neighboring boreholes 6 and 7, and pressure sensor P2 (installed in hole 6) the breakthrough of hole 7. Initial drilling water input into the holes was $80\,\mathrm{l\,min^{-1}}$ until the supply stopped at the moment indicated with a vertical arrow.

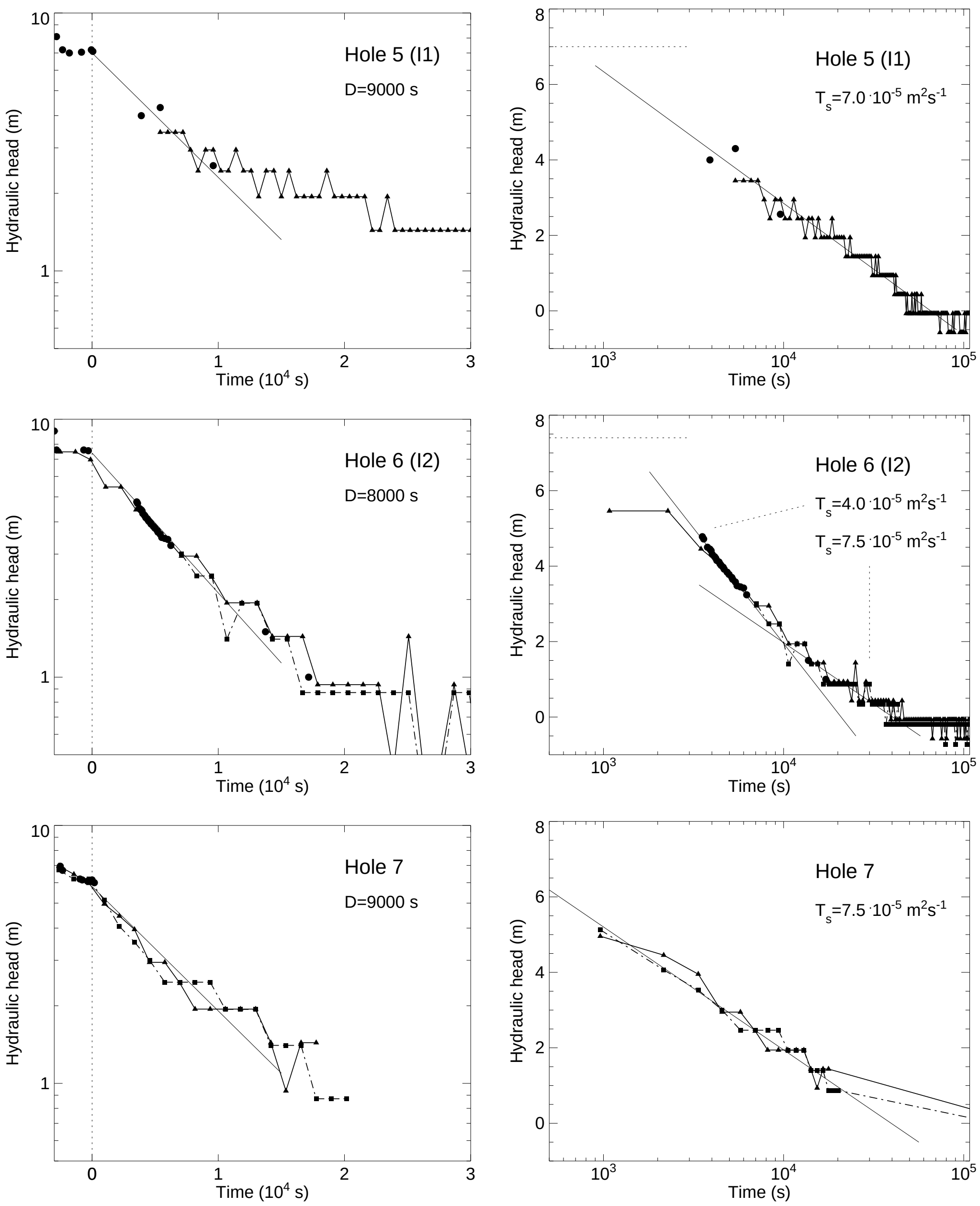

Figure 5: Interpretation of the water breakthrough events in boreholes 5, 6 and 7. Left figures: The borehole water level drop is interpreted as an exponential decay with time constant D. Right figures: Apparent transmittivity based on Jacob's method.