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Hydraulic and mechanical properties of glacial sediments beneath Unteraargletscher, Switzerland: implications for glacier basal motion

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Abstract:

The force on a 'ploughmeter' and subglacial water pressure have been measured in the same borehole at Unteraargletscher, Switzerland, in order to investigate ice–sediment coupling and the motion at the base of a soft-bedded glacier. A strong inverse correlation of the recorded pressure and force fluctuations, in conjunction with a significant time lag between the two signals, suggests that pore-water pressures directly affect the strength of the subglacial sediment. The lag is interpreted to reflect the time required for the water-pressure wave to propagate through the pores of the sediment to the depth of the ploughmeter. Analysis of the propagation velocity of this pressure wave yielded an estimate of the hydraulic diffusivity, a key parameter necessary to characterize transient pore-water flow. Furthermore, the inferred inverse relationship between pore-water pressure and sediment strength implies that Coulomb-plastic deformation is an appropriate rheological model for the sediment underlying Unteraargletscher. However, the sediment strength as derived from the ploughmeter data was found to be one order of magnitude smaller than that calculated for a Coulomb-frictional material using the water-pressure measurements. This significant discrepancy might result from pore-water pressures in excess of hydrostatic down-glacier from the ploughmeter. As the ploughmeter is dragged through the sediment, sediment is compressed. If the rate of this compression is large relative to the rate at which pore water can drain away, excess pore-water pressures will develop that have the potential to weaken the sediment. The same process could lead to highly fluid sediment down-glacier from clasts that protrude into the glacier sole and thus would otherwise provide the roughness to couple the glacier to its bed (Iverson, 1999). Rapidly sliding glaciers overlying sediments might therefore move predominantly by 'ploughing', which tends to focus basal motion near the glacier sole rather than at depth in the bed. Copyright © 2001 John Wiley & Sons, Ltd.

KEY WORDS subglacial water pressure; pore pressure diffusion; pressure-wave propagation; hydraulic diffusivity; subglacial sediment strength; ploughing; excess pore pressure; basal dynamics

INTRODUCTION

Since geophysical investigations on Ice Stream B, West Antarctica (Blankenship *et al.*, 1987) led to the provocative suggestion that the fast motion of the ice stream might be facilitated by pervasive deformation of the underlying sediment (Alley *et al.*, 1987), much research has focused on efforts to advance our understanding of the dynamics of glaciers overlying deformable beds (see reviews by Murray (1997) and Fischer and Clarke (in press)). Although it is widely recognized that the motion at the base of a glacier and the drainage of water beneath the ice are strongly interdependent, the exact nature of the relationship between subglacial hydrological conditions and mechanisms of glacier basal motion is not fully understood.

In general, basal motion of a glacier over a sedimentary bed can arise from sliding between ice and bed, ploughing of clasts through the upper layer of the bed, pervasive deformation of the bed or shearing across

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discrete planes in the bed (Alley, 1989). However, whether a glacier deforms its bed, ploughs it or slides over it depends on the degree of coupling at the ice–bed interface. Strong coupling between the glacier sole and the underlying sediment can result from a high density of clasts at the bed surface. If these roughness elements approximate the controlling obstacle size of sliding theories (e.g. Weertman, 1964), sliding motion is inhibited. In this case, pervasive bed deformation is favoured, if high pore-water pressures cause the sediment yield strength to drop below the shear stress that can be supported by the ice–bed interface (Boulton and Hindmarsh, 1987; Alley, 1989). Alternatively, complete decoupling of ice and sediment can result from increased ice–bed separation owing to the presence of a layer of highly pressurized water at the interface. If this water layer is thick enough to drown clasts near the controlling obstacle size, the contribution of sliding to glacier motion may increase (Iken and Bindschadler, 1986; Alley, 1996). Incomplete coupling of a glacier to the bed can lead to a transitional state between sliding and pervasive bed deformation entitled 'ploughing', in which clasts that protrude across the ice–bed interface are dragged through the upper layer of the sediment. This ploughing process can occur at basal shear stresses that are too small to deform the bed at depth but requires that pore-water pressures are high enough to cause the bed to weaken such that the local deviatoric stress developed in front of these clasts is sufficient to deform the sediment locally (Brown *et al*., 1987; Alley, 1989).

The foregoing discussion highlights the important role of the subglacial hydrological system in controlling the basal motion of soft-bedded glaciers. The ability of the sediment to deform depends critically on the sediment pore-water pressure, especially where overpressurisation occurs. Therefore, the efficiency with which water is drained and excess pore pressures are dissipated diffusively in subglacial sediments has become a key issue in current investigations of glacier motion over deformable beds.

Instruments suited to the investigation of hydraulic and mechanical properties of basal sediments include ploughmeters: steel rods that are driven vertically into the sediment bed through boreholes and that measure the bending force as they are dragged through the sediment during basal motion (Fischer and Clarke, 1994, 1997; Fischer *et al*., 1998, 1999). In this paper we present measurements of force variations on a ploughmeter and fluctuations in subglacial water pressure from the same borehole at Unteraargletscher, Switzerland. We use these data to estimate hydraulic parameters that regulate the water flow through the subglacial sediment. Furthermore, the strength of the basal sediment is assessed using two approaches, which yield considerably different results. We discuss this difference with reference to excess pore pressures that might develop in front of the ploughmeter as it is dragged through the sediment and subsequently conclude with potential implications for glacier basal motion.

STUDY AREA

Unteraargletscher is a temperate valley glacier, situated in the Bernese Alps, Switzerland, emanating from the confluence of its two main tributaries Finsteraar- and Lauteraargletscher (Figure 1). Glaciological studies on Unteraargletscher can be traced back to the works of Franz Josef Hugi and Louis Agassiz in the nineteenth century (Hugi, 1830, 1842; Agassiz, 1840, 1847). Since 1924 systematic measurements of surface topography changes and velocities have been made every year (Flotron, 1924–1998). The glacier continues to remain a focal point of glaciological investigations in the Swiss Alps. The main emphasis of current research activity is the study of the general flow characteristics of the glacier (e.g. Iken *et al*., 1983; Funk and Röthlisberger, 1989; Gudmundsson *et al*., 1997; Gudmundsson, 1999).

Unteraargletscher, which extends roughly 6 km eastwards from the confluence zone, has a mean width of *c*. 1 km and a mean surface slope of approximately 4°. A prominent feature of the glacier is its extensive debris cover of typically 5 to 15 cm thickness. Seismic reflection soundings (Knecht and Süsstrunk, 1952) suggest that Unteraargletscher is predominantly underlain by a layer of unconsolidated material (Röthlisberger and Vögtli, 1967; Funk and Röthlisberger, 1989). Recent studies of the glacier's hydrology and dynamics have included investigations of the substrate by observations in subglacial meltwater tunnels at the glacier terminus

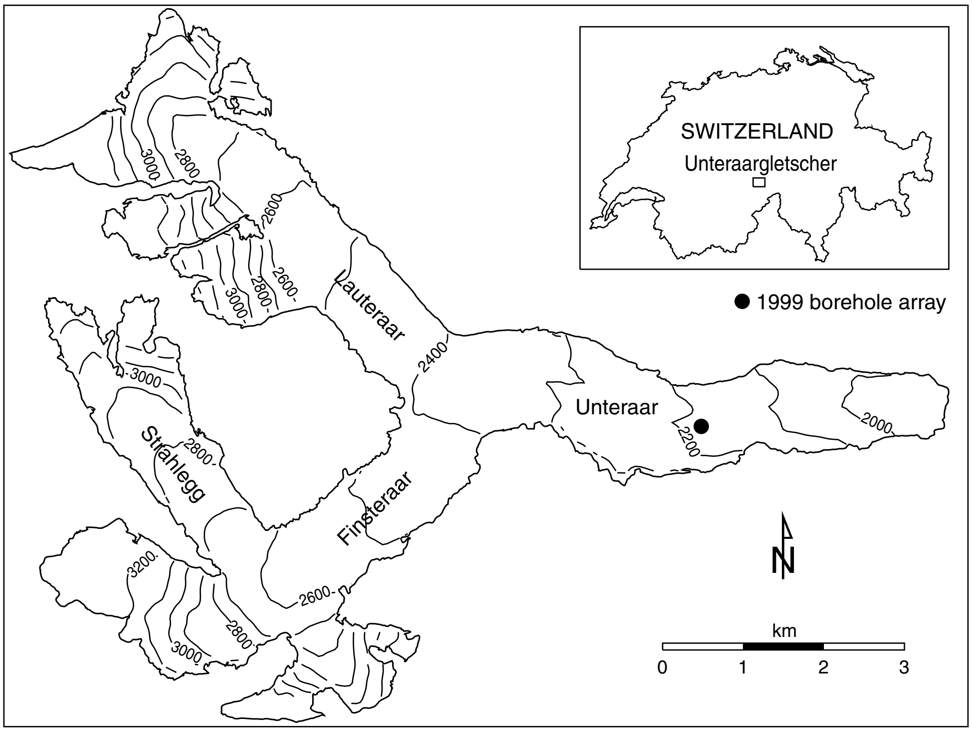

Figure 1. Map of Unteraargletscher showing the surface topography and the location of the 1999 borehole array. Inset shows the study area in the Bernese Alps, Switzerland

and by bed penetrometry, video observations and direct sampling through boreholes. These investigations demonstrate that, at least in part, Unteraargletscher rests on a soft sediment bed.

BACKGROUND

Pore pressure diffusion

The equation governing pore pressure diffusion in a saturated compressible aquifer in one dimension is

$$\frac{\partial}{\partial z}\left(K\frac{\partial p_{\mathrm{w}}}{\partial z}\right) = \rho_{\mathrm{w}} g\,(\alpha + n\beta)\,\frac{\partial p_{\mathrm{w}}}{\partial t} \tag{1}$$

(e.g. Bear and Verruijt, 1987), where z is the vertical coordinate, K is the hydraulic conductivity, p_{w} is the pore-water pressure, ρ_{w} is the density of water, g is the gravitational acceleration, n is the sediment porosity, α and β are the compressibility coefficients for the porous medium and water respectively, and t denotes time. Assuming constant K, Equation (1) reduces to the standard one-dimensional diffusion equation

$$\frac{\partial p_{\mathrm{w}}}{\partial t} = D\frac{\partial^2 p_{\mathrm{w}}}{\partial z^2}, \tag{2}$$

where $D = K/\rho_{\mathrm{w}} g(\alpha + n\beta)$ corresponds to the hydraulic diffusivity. From the solution of Equation (2) (e.g. Carslaw and Jaeger, 1959), applying to a harmonically varying pressure at the surface of a semi-infinite aquifer, it follows that pressure fluctuations with angular frequency ω are propagated into the aquifer with velocity

$$v = \sqrt{2\omega D}. \tag{3}$$

Sediment strength

For a sediment that deforms as a Coulomb-plastic material, the shear strength τ is independent of the rate of deformation and depends linearly on the effective pressure p_e (the difference between the overburden pressure and the pore-water pressure), such that

$$\tau = c + p_e \tan \phi, \tag{4}$$

where c is the cohesion and ϕ is the angle of internal friction (e.g. Lambe and Whitman, 1979). If sheared to a sufficiently large strain, the sediment will deform at a steady porosity (Skempton, 1985) and is said to have reached its residual state. Cohesion becomes negligible in the residual state (Mitchell, 1976; Head, 1994), so Equation (4) reduces to

$$\tau_r = p_e \tan \phi_r, \tag{5}$$

where ϕ_r is the residual friction angle and τ_r is referred to as the ultimate or residual strength.

Measurement methods

Local subglacial water pressures were measured with the use of pressure transducers submerged in boreholes. Typically, these sensors were suspended 250 m below the glacier surface. Thus, for holes that are *c.* 252 m deep and assumed to be straight and close to vertical, the sensors are positioned roughly 2 m above the bed.

Sediment strength was measured with ploughmeters installed at the bottom of boreholes (Figure 2). A detailed description of this instrument has been given by Fischer and Clarke (1994). Briefly, ploughmeters are steel rods, *c.* 1·5 m long, on to which strain gauges have been bonded (Figure 2b). The ploughmeters are installed at the glacier bed such that the tip protrudes into the subglacial sediment. Similar to an ice-entrained clast, the immersed tip is dragged through the sediment as the glacier slides forward (Figure 2a). Elastic bending of the ploughmeter is recorded by the strain gauges and is converted into a force on the tip with the use of a laboratory calibration (Fischer and Clarke, 1994).

Studies of the penetration of cones, plates and piles through water-saturated fine-grained sediment provide a framework for the calculation of the force on the immersed tip of a ploughmeter as it is dragged through a soft sediment bed. Consistent with such penetration studies, we assume that subglacial sediment behaves as a Coulomb-plastic material. We follow the analysis of Iverson *et al.* (1994) and use the penetration model of

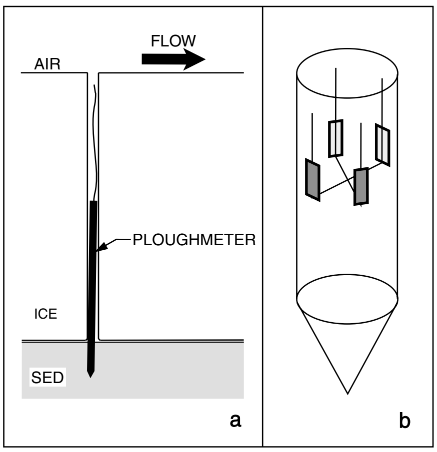

Figure 2. Schematic diagram of ploughmeter. (a) Installation and operation at the base of the glacier. (b) Arrangement of strain gauges near the tip of the steel rod

Senneset and Janbu (1985), which is explicitly formulated in terms of the effective pressure. This model is appropriate for our calculation because it is broadly applicable to ploughing objects of different shapes. Consider a ploughmeter being dragged through subglacial sediment (Figure 3). A zone of compression develops down-glacier from the ploughmeter. There, the sediment yields plastically along slip planes that are oriented at an angle γ with a normal to the direction of motion. The force per unit length F on the leading surface of the ploughmeter is given by

$$F = 2a\left(N_f\left(p_e + \frac{c}{\tan\phi}\right) - \frac{c}{\tan\phi}\right), \tag{6}$$

where a is the radius of the ploughmeter and N_f is a dimensionless bearing-capacity factor

$$N_f = \tan^2\left(\frac{\pi}{4} + \frac{\phi}{2}\right) e^{(\pi-2\gamma)\tan\phi} \tag{7}$$

(Janbu and Senneset, 1974; Senneset and Janbu, 1985). For deforming sediment in its residual state, $c = 0$ and $\phi = \phi_r$. Thus, substituting Equation (7) into Equation (6), the residual force per unit length F_r on the leading surface of the ploughmeter is

$$F_r = 2ap_e\tan^2\left(\frac{\pi}{4} + \frac{\phi_r}{2}\right) e^{(\pi-2\gamma)\tan\phi_r}. \tag{8}$$

Empirical support for Equation (8) stems, for example, from Iverson *et al.* (1994), who found good agreement between calculated forces and those measured on cones that were dragged through sediment in the laboratory.

FIELD OBSERVATIONS

As part of our efforts to elucidate how the seasonal evolution of the glacial hydrological system influences patterns of subglacial water pressures and basal drag at Unteraargletscher, six holes were drilled through the ice to the bed near the central flowline, roughly 3 km up-glacier from the terminus during June 1999 (Figure 1). At this site, the ice was *c.* 252 m thick. With one exception, all boreholes remained full of water when the drill reached the bed, indicating that the holes were not connected with the subglacial drainage system at that time. However, about two weeks after the holes had been drilled, one of the unconnected boreholes established hydraulic communication with the subglacial water system, whereas the one that was initially connected lost its connection. Eleven weeks later another borehole also became connected. Water pressures recorded in these

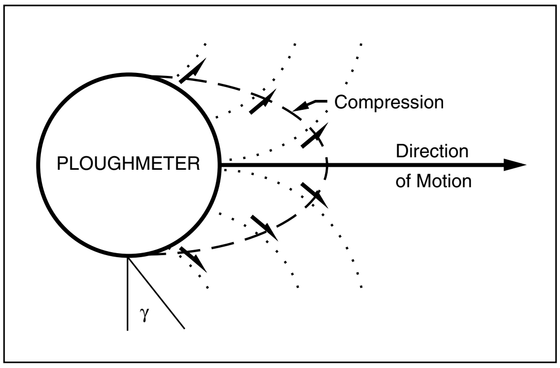

Figure 3. Plan view of ploughmeter moving from left to right through subglacial sediment. Slip planes lie approximately paralled to a plane defined by γ

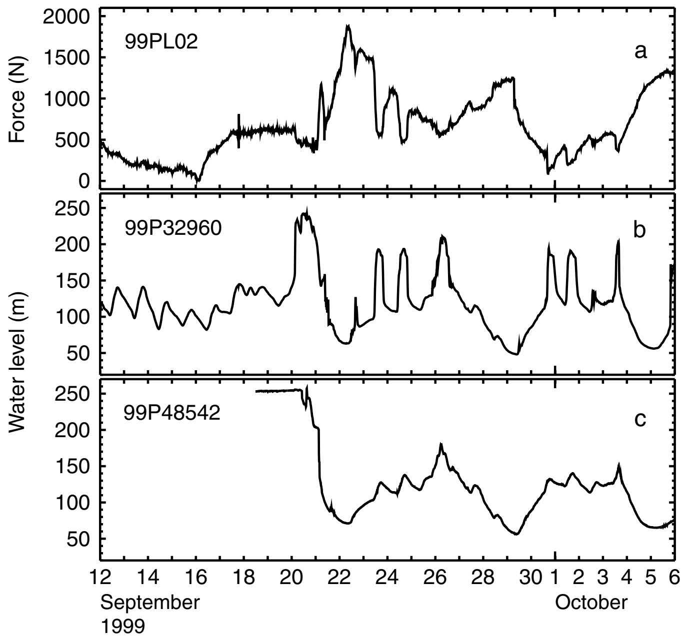

Figure 4. Records showing (a) the force applied to the immersed tip of a ploughmeter (99PL02) and subglacial water pressure in the same (99P32960) (b) and a nearby (99P48542) (c) borehole

two connected holes subsequently tracked each other quite closely for about two weeks (Figures 4b and c), which suggests that the two holes were connected to a part of the subglacial drainage system that allowed an efficient hydraulic communication between the two sensors during this period.

Four holes were instrumented with ploughmeters to measure the strength of the subglacial sediment. Figure 4a shows roughly 26 days of observations for ploughmeter 99PL02. We estimate that this ploughmeter was inserted *c.* 0·1 m into the basal sediment. The records of subglacial water pressure measured in the same borehole (99P32960), and in another that was located about 74 m down-glacier (99P48542), are also included (Figures 4b and 4c, respectively) and plotted along the same time axis. During the second half of the observation period, variations in the ploughmeter signal (Figure 4a) are inversely correlated with fluctuations in borehole water level (Figure 4b) such that high forces experienced by the ploughmeter coincide with low water pressures and vice versa.

RESULTS

Estimation of hydraulic diffusivity

The observed inverse relationship of water pressure and ploughmeter response (Figures 4a and 4b) indicates that increased water pressures weaken the sediment, resulting in less force on the ploughmeter. This observation implies that pressure changes in the subglacial drainage system drive pore-water pressure variations at depth in the sediment and thereby affect its strength. A potential gradient therefore may exist across the water–sediment interface at the base of the glacier, which reverses temporally in accordance with variations in subglacial water pressure (Fischer *et al.*, 1998). As a result, pressure waves are driven downwards into the sediment bed when the subglacial water pressure is high and upwards when the water pressure is low. Fluctuations in pore-water pressure in the sediment should lag those recorded with the pressure transducer in the borehole, as it takes time for the pressure wave to propagate through the sediment. Because pore-water pressures directly affect the sediment strength (Equation (5)), variations in the force response of the ploughmeter also should lag those in subglacial water pressure.

In close analogy to the analysis of Fischer *et al.* (1998), we estimate the hydraulic diffusivity of subglacial sediment on the basis of the velocity of water-pressure waves as they propagate through it. A reasonable idealization is to represent the glacier bed as a semi-infinite aquifer. A steady water pressure is assumed at infinity and the pressure variation in the subglacial drainage system is treated as a boundary condition at the top of the sediment bed. With $\omega = 2\pi f$ and $v = z_1/t$ substituted into Equation (3), where f is the frequency of the periodic boundary condition and z_1 is the distance over which diffusion of pore pressures is considered, we can compute the hydraulic diffusivity as

$$D = \frac{z_1^2}{4\pi f t^2}. \tag{9}$$

In our derivation we have assumed implicitly that the hydraulic diffusivity $D = K/\rho_w g(\alpha + n\beta)$ is a constant. In reality, however, both K and α change with porosity and hence with effective pressure. We justify our assumption by noting that K and α decrease as effective pressure increases. Because these parameters appear, respectively, in the numerator and denominator of the expression for D, their variations with effective pressure may to some extent balance, minimizing the variation in D.

Visual inspection of the time-series shown in Figures 4a and 4b suggests that the records of the force on the ploughmeter and the subglacial water pressure are dominated by variations with periods of one day and approximately five days. Power spectral analysis (Press *et al.*, 1992) was applied to both time-series and supports our suggestion. A dominant peak is present in both force- and pressure-power spectral density functions (Figure 5a for the 99PL02 data and Figure 5b for the 99P32960 data) at frequency 0·2 day^{-1} and thus indicates that the ploughmeter clearly responds to water-pressure variations with a period of five days. In addition, a secondary peak in both spectra at frequency 1 day^{-1} also points to a ploughmeter response to diurnal water-pressure fluctuations.

The velocity with which pore-pressure waves propagate through the subglacial sediment is dependent on frequency (Equation (3)). Therefore, before calculating the cross-correlation between the record of ploughmeter 99PL02 (Figure 4a) and the record of pressure transducer 99P63290 (Figure 4b) in order to estimate the time lag between the ploughmeter response and water-pressure forcing, we applied bandpass filtering (Press *et al.*, 1992) to retain narrow frequency bands around 0·2 day^{-1} and 1 day^{-1}. These filtered records are shown in Figures 6a and 6b as solid and dashed lines for the 0·2 day^{-1} and 1 day^{-1} frequency variations, respectively. Subsequent calculation of the cross-correlation suggests that the time lag between ploughmeter response and water-pressure forcing is *c.* 0·19 days for variations with frequency 0·2 day^{-1}. For the 1 day^{-1} frequency variations the cross-correlation analysis yields a lag of *c.* 0·06 days.

Following the analysis of Fischer *et al.* (1998), we assume that the depth to which the pressure wave has to propagate corresponds to the insertion depth of the ploughmeter. Although the ploughmeter response

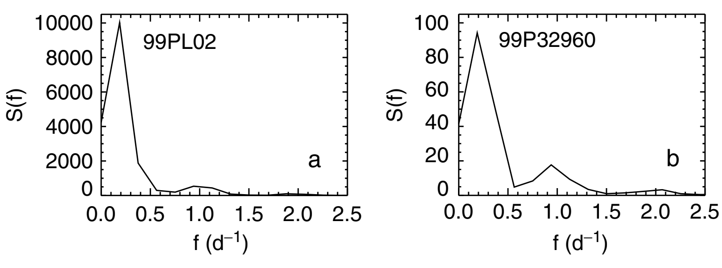

Figure 5. Power spectral density functions for (a) the ploughmeter record shown in Figure 4a and (b) the subglacial water-pressure record shown in Figure 4b. Preprocessing of the time-series included removing the trend and offset by subtracting out a regression line followed by padding the resulting data with zeros up to the next power of 2

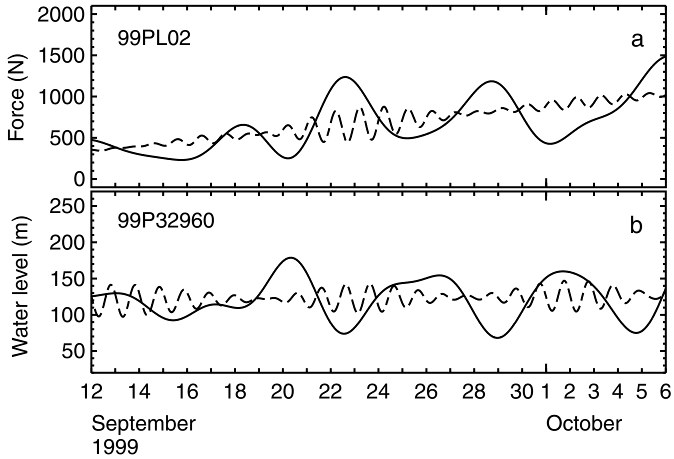

Figure 6. Bandpass-filtered records of (a) the ploughmeter data shown in Figure 4a and (b) the subglacial water-pressure data shown in Figure 4b. The solid and dashed lines show the 0·2 day^{-1} and 1 day^{-1} frequency variations, respectively

scales with the integral of the force along its length (Fischer and Clarke, 1994), we argue that the sediment becomes weakened within a sufficiently thick layer for the ploughmeter to experience reduced forces only when the pressure wave reaches the tip. This argument is based on our finding that the ploughmeter response is insensitive to high-frequency water-pressure fluctuations, suggesting that pressure fluctuations felt at smaller depths do not affect the ploughmeter (Fischer *et al.*, 1998). Therefore, substituting $z_1 = 0{\cdot}1$ m (insertion depth of ploughmeter), $t = 0{\cdot}19$ days and 0·06 days (lags from the cross-correlation analysis) and $f = 0{\cdot}2$ day^{-1} and 1 day^{-1} (dominant frequencies from the power spectral analysis) into Equation (9), we compute hydraulic diffusivity values D for Unteraargletscher sediment of $1{\cdot}3 \times 10^{-6}$ m^2 s^{-1} and $2{\cdot}3 \times 10^{-6}$ m^2 s^{-1}, respectively.

Estimation of sediment strength

The observation that the force on the ploughmeter varies directly with the effective pressure beneath Unteraargletscher (Figures 4a and 4b) adds to the body of evidence from other field and laboratory studies (e.g. Hooke *et al.*, 1997; Iverson *et al.*, 1998; Kamb, 1991; Truffer *et al.*, 2000) that a Coulomb-plastic rheology is appropriate for basal sediments. With knowledge of the subglacial water pressure (here taken as a proxy indicator for pore-water pressure in the sediment), the overburden pressure (from the ice thickness) and the angle of internal friction, we can calculate the residual strength of Unteraargletscher sediment from Equation (5). The overburden pressure of an ice layer 252 m thick corresponds to a water head of about 227 m. The subglacial water pressure is taken from Figure 4b. However, 2 m of water head was added to this record because the pressure transducer was suspended in the borehole roughly this distance above the bed. Friction angles vary between about 19° for clays and 40° for sands (Senneset and Janbu, 1985). We assume $\phi_r = 25°$ which is typical for a silty material such as a glacial sediment. Figure 7 shows the resultant variations in residual strength with time.

We also can compute the residual strength from the force plotted in Figure 4a. The force measured with the ploughmeter in the field F_{pl} is related to the force per unit length F_r (Equation (8)) by

$$F_{pl} z_g = F_r \int_{z_1 - z_g}^{z_1} (z - (z_1 - z_g))\mathrm{d}z, \tag{10}$$

where z_g is the distance from the tip of the ploughmeter to the point where the strain gauges were bonded. Here, the coordinate system is taken with the z axis positive downward (Figure 8). The integral term sums the bending moments applied to the end of the ploughmeter as it is dragged through the sediment. It is these

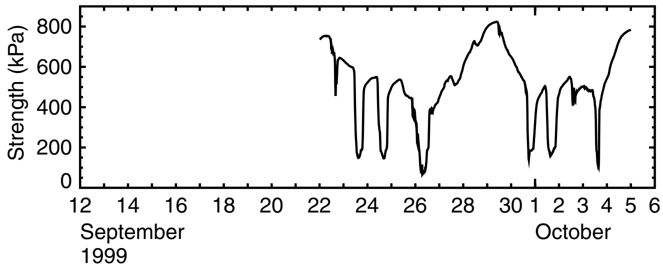

Figure 7. Sediment strength as calculated for a Coulomb-frictional material using the water-pressure measurements shown in Figure 4b

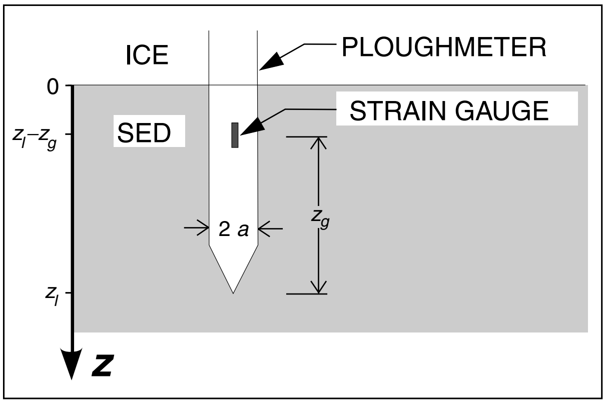

Figure 8. Definition sketch for Equation (10) showing a close-up view of a ploughmeter installed at the base of the glacier

moments that are sensed by the strain gauges bonded to the steel rod. Combining Equations (5), (8) and (10) yields

$$\tau_r = \frac{F_{pl} z_g \tan \phi_r}{2aN_f \int_{z_l - z_g}^{z_l} (z - (z_l - z_g))\mathrm{d}z}, \tag{11}$$

where N_f is given by Equation (7). The distance of the position of the strain gauges from the tip of the ploughmeter z_g is 0·1 m and the radius of the ploughmeter a is 16 mm. The orientation of the slip planes in the sediment down-glacier from the ploughmeter given by the angle γ (needed to calculate N_f in Equation (7)) depends on the sediment compressibility and particle size. Experiments with cones, plates and piles indicate that γ ranges from $+15°$ in compressible, fine-grained sediment to $-30°$ in compact, coarse-grained sediment (Senneset and Janbu, 1985). Basal sediments should be compressible in the residual state and are typically composed of silt and fine sands. On this basis, we assume $\gamma = +15°$. Furthermore, we again use $\phi_r = 25°$. The resultant residual strength variations with time that are based on the measured force F_{pl} taken from Figure 4a are shown in Figure 9.

DISCUSSION

Our estimates of the hydraulic diffusivity for the sediment beneath Unteraargletscher are comparable to values gained from other glaciers (Table I). Our analysis is based on standard diffusion theory, and with it we determine the hydraulic diffusivity from the time lag between fluctuations in borehole water level and force variations on the ploughmeter. We feel confident that the time lag between the ploughmeter response and the water-pressure forcing is real and not some artefact of the measurement. With a data-collection interval

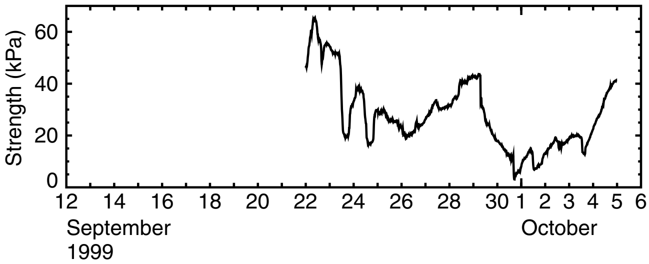

Figure 9. Sediment strength as derived from the ploughmeter data shown in Figure 4a

of 5 min, lag times in the order of hours between different signals are easily resolved. Furthermore, clock drift between different data loggers that could result in a lag between instrument responses can be ruled out because data from both pressure transducer and ploughmeter were recorded by the same logger. The exact insertion depth of the ploughmeter in the subglacial sediment, and thus the depth to which the pressure wave is thought to propagate, is uncertain. However, based on our experience with inserting similar instruments (Blake *et al*., 1994; Fischer and Clarke, 1994), we are able to reasonably constrain the insertion depth to within ± 2 cm. Fortunately, when accounting for this uncertainty in Equation (9), we find that the order of magnitude of the hydraulic diffusivity estimate (10^{-6} m^2 s^{-1}) is not affected.

Although values of the shear strength for subglacial sediments are found to vary widely (Table II), it is interesting that the strength of the sediment beneath Unteraargletscher as derived from the ploughmeter data (Figure 9) is about one order of magnitude smaller than that calculated using the water-pressure measurements (Figure 7). This raises the question whether this discrepancy has a physically based explanation or merely arises from uncertainties in the parameters used in the sediment strength calculations (Equations (5) and (11)).

Table I. Reported values of hydraulic diffusivity for subglacial sediments

Hydraulic diffusivity (m^2 s^{-1})	Location	Method	Source
10^{-6}	Storglaciären	Pressure diffusion	Fischer *et al*. (1998)
10^{-6}	Storglaciären	Consolidation test	Iverson *et al*. (1997)
10^{-7}	Trapridge Glacier	Consolidation test	Murray and Clarke (1995)
10^{-6}	Bakaninbreen	Pressure diffusion	Porter and Murray (in press)
10^{-7}	Black Rapids Glacier	Pressure diffusion	Truffer *et al*. 2000

Table II. Reported values of shear strength for subglacial sediments

Sediment strength (kPa)	Location	Driving stress (kPa)	Source
5·5–13	Columbia Glacier	100	Humphrey *et al*. (1993)
55	Storglaciären		Iverson *et al*. (1994)
48–57	Trapridge Glacier	77	Fischer and Clarke (1994)
17–87	Bakaninbreen	26	Porter *et al*. (1997)
2–4	Ice Stream B	20	Kamb (1991), Engelhardt *et al*. (1990b)

To assess the latter we computed percentage changes of the sediment strength in response to variations in ice thickness (and therefore ice overburden pressure), position of the pressure transducer above the bed, friction angle and angle of slip planes in the sediment (Tables III and IV). The ice thickness at our study site was estimated by measuring the total length of the boreholes from the surface to the bed of the glacier. For nearly vertical and straight holes, these borehole lengths closely approximate the ice thickness. Although, in general, the ice thickness determined by this method is in good agreement with that obtained by radio-echo soundings (Funk *et al*., 1994), the individual lengths of the six boreholes were not the same but ranged between about 248 and 255 m. We attribute these differences in borehole length to the difficulty of drilling vertical, straight holes. This uncertainty in determining the ice thickness also introduces an uncertainty in the exact position of the pressure transducer above the bed. The least well-constrained parameters in the sediment strength calculation are the friction angle and the angle of slip planes in the sediment, for which there are no direct measurements. Therefore, values were taken from the literature and uncertainties were assumed that seemed appropriate for a compressible subglacial sediment composed of silt and fine sand. The sensitivity analysis (Tables III and IV) shows that although neither the ice thickness nor the pressure sensor position are critical parameters, there is a large sensitivity of the sediment strength to uncertainties in friction angle and slip-plane angle. However, even with these uncertainties accounted for, it remains difficult to explain the one-order-of-magnitude difference in sediment strength as calculated with Equations (5) and (11).

Excess pore pressure

The discrepancy between the sediment strength as derived from the ploughmeter data and that calculated using the water-pressure measurements is significant, and cannot be accounted for by uncertainties in the parameters used in the calculations. We, therefore, now explore the process of excess pore-pressure generation as a possible explanation. Calculating the force on the ploughmeter using Equation (8) (and hence the sediment strength using Equation (11)) is appropriate only if the compression of the sediment down-glacier from the ploughmeter did not significantly perturb the pore-water pressure there. However, Iverson *et al*. (1994) suggested that it is possible for the rate of sediment compression in front of ploughing objects to exceed the rate at which pore pressure diffusively dissipates, which thus would lead to the generation of pore pressure

Table III. Sensitivity of the sediment strength as calculated for a Coulomb-frictional material (Equation (5)) to uncertainties in ice thickness, pressure sensor position and friction angle

Parameter	Uncertainty	Sensitivity
Ice thickness (252 m)	+3 m	+3%
	−3 m	−3%
Sensor position above bed (2 m)	+2 m	−2%
	−2 m	+2%
Friction angle ($\phi_r = 25°$)	+5°	+24%
	−5°	−22%

Table IV. Sensitivity of the sediment strength as derived from ploughmeter data (Equation (11)) to friction angle and slip-plane angle

Parameter	Uncertainty	Sensitivity
Friction angle ($\phi_r = 25°$)	+5°	−24%
	−5°	+23%
Angle of slip planes ($\gamma = 15°$)	−15°	−22%

in excess of hydrostatic, weakening the sediment. As such, the pore-water pressure down-glacier from the ploughmeter is a function of the velocity of the ploughmeter and the hydraulic diffusivity of the sediment. We closely follow the analysis of Iverson *et al.* (1994) and use a simple scaling argument to assess the tendency for the generation of excess pore pressure in front of the ploughmeter beneath Unteraargletscher. This tendency can be expressed by a single dimensionless parameter R

$$R = \frac{D}{v_{pl}\delta} \tag{12}$$

(adapted from Iverson and LaHusen, 1989), where v_{pl} is the velocity of the ploughing object (i.e. the ploughmeter) through the sediment and δ is the characteristic length of the zone of compression down-glacier from the ploughmeter. The parameter R represents the ratio of the time-scale of generation of excess pore pressure (δ/v_{pl}) to the time-scale for diffusive pore-pressure equilibration across $\delta(\delta^2/D)$. Consequently, if R is large, excess pore pressure dissipates more rapidly than it is generated by sediment compression, and significant excess pore pressure is unlikely. If R is small, there is insufficient time for pore pressures to equilibrate as the sediment is compressed, and the potential for local excess pore pressure is enhanced.

The threshold value for R, below which excess pore pressure should arise, can be estimated from a cone penetration study conducted by Campanella *et al.* (1983), in which a cone was forced downward through a silty clay at different but constant velocities, and pore pressure was recorded by a transducer in the cone. All parameters relevant to Equation (12) were either measured or could be estimated reliably. Using this study, Iverson *et al.* (1994) determined a threshold range for R between 0·15 and 15 and concluded that when $R \gg 15$, generation of excess pore pressure is unlikely.

Laboratory studies of sediment deformation during cone penetration (e.g. Malyshev and Lavisin, 1975) suggest that δ is approximately equal to the diameter of the ploughing object. Thus, for the ploughmeter (radius $a = 16$ mm), $\delta = 0{\cdot}032$ m is a good approximation. The velocity with which the ploughmeter is dragged through the subglacial sediment should, approximately, be equal to the basal velocity of the glacier, which cannot exceed the surface velocity. Automated surveying (Gudmundsson *et al.*, 2000) of the displacement of stakes drilled into the surface of the glacier indicate that the surface velocity of Unteraargletscher in the vicinity of the 1999 borehole array (Figure 1) was on average 0·07 m day^{-1} during the period July–September 1999. Continuous tilt measurements conducted in a borehole *c.* 700 m up-glacier from the ploughmeter measurement suggest that basal motion contributes up to 60% to the total forward motion of the glacier (Gudmundsson *et al.*, 1999). Thus, we assume $v_{pl} = 0{\cdot}04$ m day^{-1}. As described above, the hydraulic diffusivity of the subglacial sediment is about 10^{-6} m^2 s^{-1}. Using these values in Equation (12) shows that $R = 67$ for a ploughmeter being dragged through sediment beneath Unteraargletscher. This value of R is of the same order of magnitude as the threshold range mentioned above. Therefore, we cannot rule out that pore-water pressures in excess of hydrostatic could have developed in front of the ploughmeter, thereby weakening the sediment. Thus, it is possible that this effect may account for the significant discrepancy between the sediment strength as derived from the ploughmeter data (Figure 9) and that calculated using the water-pressure measurements (Figure 7).

Because the pore-water pressure in front of a ploughing object is coupled to the velocity with which this object is dragged through the sediment, this effect of local sediment weakening should be more pronounced beneath rapidly sliding glaciers. Humphrey *et al.* (1993) reported on an unplanned experiment in the subglacial shear zone of fast-moving Columbia Glacier, Alaska, which bears some similarity to our ploughmeter measurements at Unteraargletscher described above. During hot-water drilling on Columbia Glacier, the drill became inadvertently stuck in the bed and was dragged for 5 days through the basal sediment. After its subsequent retrieval, analysis of the bent drill stem yielded estimates of the sediment strength that were about one order of magnitude smaller than the applied shear stress (Table II). This large difference led Humphrey *et al.* (1993) to conclude that the basal sediment does not contribute significantly to the drag at the bed and plays a minor role in controlling the flow of Columbia Glacier. We now revisit this experiment to reassess the conclusions of Humphrey *et al.* (1993) in light of our findings discussed in this paper. As in Iverson *et al.* (1994), we apply the scaling argument presented above (Equation (12)) to test the likelihood for the generation of excess pore pressures in front of the drill stem stuck in the basal sediment. Unfortunately, the hydraulic diffusivity of the sediment beneath Columbia Glacier is not known. We assume a value of 10^{-6} m^2 s^{-1}, which corresponds to the upper limit of the range of hydraulic diffusivities derived for sediments beneath different glaciers (Table I), minimizing the potential for excess pore pressures. Further, according to Humphrey *et al.* (1993) the diameter of the drill stem was 0·034 m and the sliding velocity at the site of the experiment was about 3·5 m day^{-1}. Substituting these values into Equation (12), we find that $R = 0{\cdot}7$ for the drill stem being dragged through sediment beneath Columbia Glacier. This value of R indicates that pore pressures in excess of hydrostatic down-glacier from the drill stem were likely to have developed, a conclusion that was also reached by Iverson *et al.* (1994). Therefore, owing to excess pore-water pressures, the sediment in front of the drill stem may have been weakened substantially. In this case, the strength estimates as obtained from the analysis of the bent drill stem would be too small to represent the shear strength of the sediment beneath Columbia Glacier. As such, the results of the drill stem analysis also may be interpreted to lead to a conclusion that is contrary to that of Humphrey *et al.* (1993). Because the sediment weakening down-glacier from the drill stem is only a local effect, the strength of the sediment bed as a whole may well be comparable in magnitude to the applied shear stress, implying that the basal sediment may be important in resisting the flow of the glacier.

Implications for glacier basal motion

The hydraulic diffusivity is a parameter that scales with the ratio of hydraulic conductivity to sediment compressibility (Freeze and Cherry, 1979). Therefore, the generation of excess pore-water pressure and associated sediment weakening become more likely for less hydraulically conductive and more compressible sediments. As was pointed out by Iverson (1999), this effect is clearly demonstrated by the cone-penetration tests conducted by Campanella *et al.* (1983). As a cone was forced through a coarse sand with a high hydraulic conductivity, the pore pressure in front of the cone was hydrostatic and the resistive stress on the cone increased owing to the increase in overburden pressure with depth. However, when the cone reached a more compressible clayey silt layer of much lower hydraulic conductivity, excess pore pressure developed in front of the cone, which reduced the stress on the cone by more than an order of magnitude (Campanella *et al.*, 1983).

Iverson (1999) recognized that pore-water pressures in excess of hydrostatic in front of ploughing clasts of any reasonable size may have fundamental implications for the ploughing process (Brown *et al.*, 1987; Alley, 1989) and thus for the basal motion of glaciers. This effect may be of particular importance for rapidly sliding glaciers underlain by compressible sediments of low hydraulic conductivity, as examined by Iverson (1999) for the case of Ice Stream B. We now repeat this calculation. Recalling that $D = K/\rho_w g\alpha$ (note that $\beta \ll \alpha$), the minimum clast size required to generate excess pore pressure can be calculated from Equation (12). Tests performed on samples recovered from the bed of Ice Stream B indicate that the hydraulic conductivity K of the basal sediment is 2×10^{-9} m s^{-1} (Engelhardt *et al.*, 1990a). Furthermore, the sediment was found to consist of an extremely poorly sorted, clay-rich diamicton (Tulaczyk *et al.*, 1998). Thus, the compressibility α should be approximately 10^{-6} Pa^{-1} (Freeze and Cherry, 1979). Measurements of sliding at the base of the ice stream suggest an average sliding velocity of *c.* 1 m day^{-1} (Engelhardt and Kamb, 1998). Finally, remembering that δ is approximately equal to the diameter of the ploughing particle, Equation (12) yields a minimum clast diameter required to generate significant excess pore pressure of about 0·01 m. As a consequence, down-glacier from ice-entrained clasts that are dragged through the bed and are larger than 0·01 m, basal sediment is expected to be weaker than elsewhere. Iverson (1999), therefore, concluded that these clasts cannot provide the roughness necessary to couple the ice stream to its bed, thereby effectively suppressing pervasive deformation of the underlying sediment. Instead, excess pore pressures and consequent sediment weakening may help to increase the likelihood that the motion beneath Ice Stream B occurs near the base of the ice by ploughing. Such ploughing therefore may account for the findings of Engelhardt and

Kamb (1998) that 83% of the basal motion of the ice stream is focused within the top 30 mm of the subglacial sediment layer.

CONCLUSION

In this paper, we have shown how hydraulic and mechanical properties of basal sediments can be estimated from the force on a ploughmeter and subglacial water pressure measured in the same borehole. By applying results from standard diffusion theory, we were able to determine the hydraulic diffusivity from the propagation velocity of pressure waves through the subglacial sediment. Furthermore, a strong inverse relationship between the recorded pressure and force variations implies that the sediment beneath Unteraargletscher is best characterized by a Coulomb-plastic rheology. We calculated a shear strength for this sediment using two approaches and found that when we based our calculation on the ploughmeter data, the sediment strength is about one order of magnitude smaller than that derived for a Coulomb-frictional material using the water-pressure measurements. We subsequently showed that we cannot dismiss the possibility that the generation of excess pore-water pressure in front of the ploughmeter may have been responsible for a local reduction in sediment strength and hence may provide an explanation for the discrepancy in our results (Figures 7 and 9).

Although we were not able to show conclusively that excess pore pressures developed in front of the ploughmeter that led to a local weakening of the sediment, this process may nevertheless be important, particularly beneath glaciers that slide rapidly over sediments of low hydraulic conductivity. One therefore should be cautious when interpreting sediment strength measurements in cases where excess pore pressures induced by the ploughmeter cannot be precluded. An obvious implication is that pore-water pressure should be measured directly. By housing a miniature pressure transducer within the tip of the ploughmeter, the measurements of pore pressure, when linked to those of subglacial water pressure, would provide the means to estimate pressures in excess of hydrostatic.

The process of local sediment weakening owing to the generation of excess pore-water pressure in front of ploughing objects (Iverson *et al*., 1994; Iverson, 1999) raises an important and significant issue regarding the role of subglacial hydraulic conditions in the basal motion of sediment-based glaciers. For rapidly sliding glaciers underlain by sediments of low hydraulic conductivity, excess pore-water pressure may develop such that sediment down-glacier from ploughing clasts provides little resistance to glacier motion. Thus, instead of these clasts leading to a strong coupling at the ice-bed interface, the ploughing process may in fact help decouple the ice from the bed. Furthermore, provided that there is a high abundance of clasts protruding across the ice-sediment interface that are large enough to generate significant excess pore pressure, ploughing may be a dominant mode of glacier basal motion.

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