



THIS MANUSCRIPT HAS BEEN SUBMITTED TO THE JOURNAL OF GLACIOLOGY AND HAS NOT BEEN PEER-REVIEWED.

A rheology for temperate ice in glaciers (RTIG) based on borehole deformation experiments

Journal:	<i>Journal of Glaciology</i>
Manuscript ID	Draft
Manuscript Type:	Article
Date Submitted by the Author:	n/a
Complete List of Authors:	Lüethi, Martin; University of Zurich, Department of Geography Funk, Martin; ETH Zürich D-BAUG, VAW-Glaciology Bauder, Andreas; ETH Zürich D-BAUG, VAW-Glaciology
Keywords:	Ice rheology, Basal ice, Glacier flow
Abstract:	The rheology of temperate glacier ice, which contains small amounts of liquid water, largely controls the fast dynamics of glaciers and polar ice streams. Recent laboratory experiments suggest a constant-viscosity ice rheology when the water content between ice grains exceeds 0.6%. Such a linear flow law, where the deformation rate is proportional to the applied stress, is in contrast to the power-law relations usually assumed for glacier ice. Here, we use in-situ deformation measurements from 16 boreholes to derive a new rheology for temperate ice in glaciers (RTIG). Our analysis of borehole tilt data from Greenland, the Alps and Canada suggests two distinct deformation regimes depending on stress magnitude. At intermediate stresses, a linear rheology fits the measurements, with a viscosity that is 40 times higher than in the laboratory experiments. At stresses exceeding 0.157 MPa the data fits a pseudo-plastic rheology with a power $n \sim 22$. The combination of both deformation regimes can be described with the Ellis flow rule. Applying this rheology for temperate glacier ice has profound consequences for our understanding of the flow of large ice masses, calling for a radical reassessment of how such ice is modeled.



A rheology for temperate ice in glaciers (RTIG) based on borehole deformation experiments

Martin P. LÜTHI,¹ Martin FUNK,² Andreas BAUDER²

¹*Department of Geography, University of Zürich, Switzerland*

²*Laboratory of Hydraulics, Hydrology and Glaciology (VAW), ETH Zürich, Switzerland*

Correspondence: Martin Lüthi <martin.luethi@geo.uzh.ch>

ABSTRACT. The rheology of temperate glacier ice, which contains small amounts of liquid water, largely controls the fast dynamics of glaciers and polar ice streams. Recent laboratory experiments suggest a constant-viscosity ice rheology when the water content between ice grains exceeds 0.6 %. Such a linear flow law, where the deformation rate is proportional to the applied stress, is in contrast to the power-law relations usually assumed for glacier ice. Here, we use in-situ deformation measurements from 16 boreholes to derive a new rheology for temperate ice in glaciers (RTIG). Our analysis of borehole tilt data from Greenland, the Alps and Canada suggests two distinct deformation regimes depending on stress magnitude. At intermediate stresses, a linear rheology fits the measurements, with a viscosity that is 40 times higher than in the laboratory experiments. At stresses exceeding 0.157 MPa the data fits a pseudo-plastic rheology with a power $n \approx 22$. The combination of both deformation regimes can be described with the Ellis flow rule. Applying this rheology for temperate glacier ice has profound consequences for our understanding of the flow of large ice masses, calling for a radical reassessment of how such ice is modeled.

INTRODUCTION

The dynamics of glaciers and polar ice sheets is largely controlled by processes in ice that contains small amounts of liquid water (e.g., Haseloff and others, 2019; Meyer and others, 2024; Lilien and others, 2026). Such watery ice, termed *temperate*, contributes to the fast ice flow speeds of many alpine-type glaciers and ice sheet outlet ice streams, and is ubiquitous at the base of extended areas of the Greenland Ice Sheet (GrIS) (e.g., MacGregor and others, 2016; Law and others, 2023) and in fast-moving ice streams in Antarctica (e.g., Joughin and others, 2003; Gudmundsson, 2007). Laboratory experiments have established that interstitial water between ice grains lowers the ice viscosity, and therefore leads to an increase in strain rate that is linear with the water content (Lliboutry, 1976; Duval, 1977). Recent laboratory experiments confirm this effect, but further suggest that a switch in ice rheological behavior occurs when the interstitial water content exceeds a threshold of 0.6% (Adams and others, 2021; Schohn and others, 2025). Above this threshold the strain rates scale linearly with the applied stress, which is equivalent to a constant viscosity ($n = 1$ in Eq. 1).

Water contents exceeding the 0.6 % threshold have been observed in glaciers and ice sheets in many settings (e.g., Lüthi and others, 2002; Hubbard and others, 2003; Pettersson and others, 2004; Navarro and others, 2005; Bradford and others, 2009; Ryser and others, 2013), and on Storglaciären (Sweden) a water content about 0.6 % was found throughout the temperate ice (Gusmeroli and others, 2010). Using a constant-viscosity rheology for such ice would limit run-away ice deformation, and therefore alter the results of continental-scale ice flow models, which are crucial components of earth system models predicting changes in sea level.

As ice flow models become more detailed even for continent-scale domains (e.g., Aschwanden and others, 2016; Leger and others, 2025) the need for an accurate ice flow law

is increasing. The often used power-law rheology with $n = 3$ is referred to as Glen's flow law (Glen, 1955; Nye, 1953)

$$\dot{\epsilon} = E A \sigma^n \quad (1)$$

has proven to be a useful approximation to reality, although by its simplicity it cannot account for important processes such as fabric evolution, anisotropic preferred deformation, and interstitial water, which are often lumped into an enhancement factor E that has little physical meaning. For temperate ice at the bottom of the GrIS, enhancement factors much smaller than unity are required to describe the shear deformation (Lüthi and others, 2002; Ryser and others, 2014b; Doyle and others, 2018; Maier and others, 2019), hinting to a different rheology of such ice.

Here, we use several own unpublished data sets from borehole deformation experiments in Alpine glaciers, which are complemented with a re-analysis of published borehole deformation data, to constrain the ice rheology of temperate ice. We use 63 data points from 16 boreholes at 12 sites in Greenland, the Alps and Canada. All of these experiments were conducted in-situ under high confining pressure in deep boreholes to the bedrock. One set of data points was derived from time series of two-axis tilt sensors that were installed at cold and temperate conditions and which deformation was measured for months. A second set of data points was acquired by lowering an inclinometer into the borehole, yielding borehole inclination angle at several times.

Our re-analysis of own and published borehole deformation data corroborates that temperate glacier ice follows a linear-viscous rheology similar to the one observed in laboratory ring-shear experiments, although with a different viscosity. In addition, we find strong indication for a pseudo-plastic stress limit, that not only applies in temperate ice, but seems to be a universal upper bound for shear stress in glaciers.

DATA AND METHODS

Field data

Our data set consists of results from own published and unpublished experiments from deep boreholes in several glaciers in the Alps and from drill sites on the GrIS. These data are complemented by borehole deformation data taken from the literature. A list of all boreholes and their characteristics is given in Table 1.

Our own data sets were recorded with different versions of a wired digital borehole sensor system (DiBOSS, Ryser and others, 2014b) with sensors for temperature, tilt and pressure, that were installed at several drill sites between 2006 and 2014. On polythermal Grenzgletscher (Zermatt, Switzerland) three boreholes to the bedrock (Hut, Center, Lake) were equipped with strings with 8–13 digital sensor units that sampled continuously for two years (Ryser and others, 2013, and unpublished). At three drill sites in West Greenland (DUCK, FOXX, GULL; Lüthi and others, 2002; Ryser and others, 2014a,b) sensor systems were deployed throughout the full ice thickness to depths reaching from 630 to 830 m. Single tilt sensor units were also installed in boreholes in the Swiss Alps on Unteraargletscher (Gudmundsson and others, 1999; Bauder, 2001) and Rhonegletscher (Keller and Blatter, 2012).

The above data sets consist of time series from discrete sensor units that were installed at different depths in the boreholes. The tilt angle variation of each sensor unit on two perpendicular axes was used to obtain surface-parallel shear strain rates. Such tilt angle time series often show phases with erratic motion which are caused by two classes of processes: either the sensor moves within the borehole (rotation, sliding along the hole), or the motion is due to short-term ice deformation caused by changes in basal or internal processes. For the present analysis, only time periods that were not affected by short-term phenomena were selected. For most sensors this amounts to 20–50 consecutive days during winter (the data sets and selected phases are given in the Supporting Material). With an optimization procedure (described below) the best-matching velocity field components were determined, which were then used to calculate strain rates. The depth of the sensors below the surface, the ice temperatures and the surface slope were taken from the publications, or from the original data sets of the same or neighboring boreholes.

These own data sets were complemented with published results from similar studies from Greenland (Doyle and others, 2018; Maier and others, 2019), Canada (Savage and Paterson, 1963; Paterson and Savage, 1963) and from alpine-type glaciers (Gudmundsson and others, 1999; Bauder, 2001; Helbing, 2006; Keller and Blatter, 2012; Roldán-Blasco and others, 2025). For publications without data tables the graphics of borehole deformation were digitized and data points were extracted manually.

The resulting data set used in this study consists of 63 data points from 16 boreholes at 12 sites. The shear deformation rates, temperatures, and derived quantities are shown in Tables 2 and 3. Depth profiles of the measured vertical shear deformation rates for each borehole are shown in Figure 1.

Borehole Deformation Analysis

To obtain reliable deformation rates for the tilt sensors installed in the boreholes, we developed a method that improves upon the analytic method used before (Keller and Blatter, 2012; Ryser and others, 2014b), and builds on the algorithms of Gudmundsson and others (1999) and Ryser

and others (2014a). The principle is to track the motion of a rigid long sensor unit, containing two tilt sensors, through a spatially varying velocity field. The sensor unit is represented by a central point at which three orthogonal vectors of unit length are attached. The central point and each tip of the three vectors are moved by the local velocity field during one time step. Subsequently, a plane perpendicular to the vector representing the long axis of the sensor unit is constructed, and the two other vectors are mapped onto this plane. In a next step, these two vectors are symmetrically rotated along the plane such that they become orthogonal. After this process, all three vectors are orthogonal. Finally, all three vectors are normalized to unit length. This process is repeated for each time step, such that the motion of a sensor through a spatially and temporally varying flow field can be accurately tracked.

Synthetic tilt curves for the two tilt sensor axes were generated with help of the above algorithm. These tilt curves depend on three initial values of the sensor unit orientation: two initial tilt angles and the rotation around the sensor unit long axis. Depending on the complexity of the assumed ice flow field, there are several quantities characterizing the velocity field. For example, a velocity field with horizontal uniaxial extension and vertical simple shear is described by three quantities u_0 , c_x and c_z

$$u(x, y, z) = u_0 + c_x x + c_z z. \quad (2)$$

For the analysis of the tilt variation of an individual sensor the background velocity u_0 is arbitrary, and therefore set to 0. Together with the initial condition these are quantities that need to be optimized. This optimization was performed using the BOBYQA and COBYLA algorithms implemented in the NLOpt optimization library (Johnson, 2012). The above fitting procedure was applied to each tilt sensor individually.

From the optimization procedure we obtained the vertical gradient of horizontal velocity and the longitudinal and vertical extension rates. Under the assumption of plane strain (no across-flow extension) these components are equal and opposite in sign due to ice incompressibility

$$c_z = \frac{\partial u}{\partial z} \quad \text{and} \quad c_x = \frac{\partial u}{\partial x} = -\frac{\partial w}{\partial z}. \quad (3)$$

The quantities of interest are the strain rates which are calculated by

$$\begin{aligned} \dot{\epsilon}_{xz} &= \frac{1}{2} \frac{\partial u}{\partial z} = \frac{1}{2} c_z \\ \dot{\epsilon}_{xx} &= -\dot{\epsilon}_{zz} = \frac{\partial u}{\partial x} = c_x. \end{aligned} \quad (4)$$

The standard deviation between fitted and measured velocity gradients were less than $5 \cdot 10^{-4} \text{ a}^{-1}$ for the sensors used in this study. Several sensors showed erratic behaviour throughout the measurement periods, and consequently could not be used for our purpose.

For the data taken from literature the above method could not be applied and the values reported in the publications were used. Methods and error estimates vary strongly and no attempt was made to consolidate these data.

Shear stress

The shear stress at each tilt sensor is difficult to assess. To obtain the relevant components of the stress tensor analytical methods with simplifying assumptions or an ice flow model are required. The latter is based on a rheology, a detailed bed geometry and poorly constrained boundary conditions.

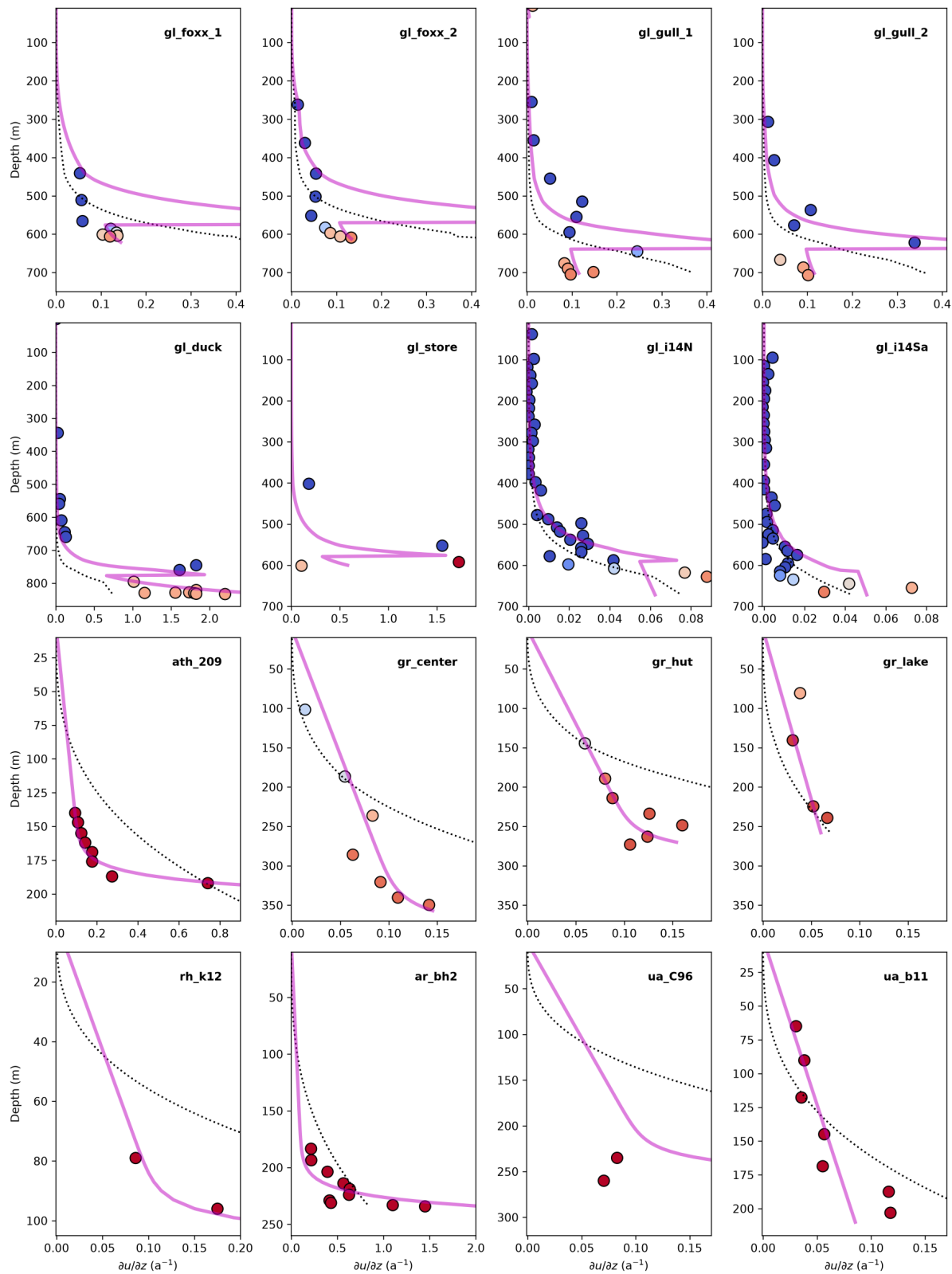


Fig. 1. Vertical shear rates of tilt sensors and inclinometers from 16 boreholes in Greenland (gl), the Alps (gr, rh, ar, ua) and Canada (ath). Shear strain rates calculated with Glen's flow law ($n = 3$) are indicated by a black dotted line, those with the new RTIG rheology with a wide purple line. All cold portions of the curves are corrected for ice temperatures measured at the sensors or in nearby boreholes. Data sources are given in Table 1, the values in Tables 2 and 3.

Drillsite	own	Location	Borehole (m)	Depth (deg)	Slope	
FOXX	*	Paakitsoq, GrIS	gl_foxx_1	624	1.55	Ryser and others (2014b)
FOXX	*	Paakitsoq, GrIS	gl_foxx_2	614	1.55	Ryser and others (2014b)
GULL	*	Paakitsoq, GrIS	gl_gull_1	703	1.3	Ryser and others (2014b)
GULL	*	Paakitsoq, GrIS	gl_gull_2	705	1.3	Ryser and others (2014b)
DUCK	*	Sermeq Kujalleq, GrIS	gl_duck	834	1.4	Lüthi and others (2002)
Store		Store Glacier, GrIS	gl_store	601	1.8	Doyle and others (2018)
I14N		Issunguata Sermia, GrIS	gl_i14N	675	0.8	Maier and others (2019)
I14S		Issunguata Sermia, GrIS	gl_i14Sa	675	0.65	Maier and others (2019)
BH209		Athabasca Glacier, Yukon	ath_209	209	5.7	Savage and Paterson (1963)
Center	*	Grenzgletscher, Alps	gr_center	351	2.4	this publications
Hut	*	Grenzgletscher, Alps	gr_hut	271	3.6	this publications
Lake	*	Grenzgletscher, Alps	gr_lake	248	2.0	this publications
K12		Rhonegletscher, Alps	rh_k12	119	9.7	Keller and Blatter (2012)
BH2		Glacier d'Argentière, Alps	ar_bh2	237	4.9	Roldán-Blasco and others (2025)
C96	*	Unteraargletscher, Alps	ua_c96	300	4.15	Gudmundsson and others (1999)
D211		Unteraargletscher, Alps	ua_b11	211	3.5	Helbing (2006)

Table 1. The 16 boreholes used for the analysis of inclination data. The borehole depth, the surface slope, and the data reference are given. The borehole designations match the labels in the graphs. Greenland Ice Sheet is abbreviated by GrIS. Own boreholes with tilt sensor time series are marked by a star.

To avoid these difficulties, we used the driving stress which is readily calculated from ice thickness and surface slope

$$\sigma_{xz} = \rho gh \tan \alpha, \tag{5}$$

where α is the surface inclination angle and h is the depth of the sensor below the surface. We used a gravity acceleration of $g = 9.81 \text{ ms}^{-2}$ and a density of $\rho = 900 \text{ kg m}^{-3}$ for all glaciers, which is a value that agrees with density measurements on natural sedimentary glacier ice with air bubbles. We expect large errors on the shear stress since neither the ice density nor the relevant surface slope are known precisely. For sensors in deep boreholes the relevant surface slope is usually derived from surface elevations within a distance of one to three ice thicknesses. Surface undulations and bedrock shape add to the uncertainty of the relevant shear stress. For measurements close to the glacier surface longitudinal stress gradients add to the uncertainty and difficult to asses.

RESULTS

We analyzed the strain rates of sensor units with the optimization procedure described above for the own boreholes (marked in Tab. 1). For the remaining boreholes the data were copied from tables, or read out from graphics in the respective publications. The corresponding stress fields at depth are not known, but have been calculated with approximate formulas or numerical models for several sites. Since the methodologies of these stress calculations vary strongly, and often depend on assumptions about the flow law, they are not useful for our purpose. Throughout this study, we assume that the relevant shear stress is the driving stress calculated from depth below the surface and average surface slope according to Equation (5). This assumption is approximately valid for the sites on the GrIS where along-flow topography varies slowly, but is likely overestimating the relevant shear stress in valley glaciers.

All data points used in this study, together with depth, temperature, calculated driving stress, viscosity and dissipation rate are listed in Tables 2 and 3. In our analysis we plot for reference the often used Glen-Nye flow law (Eq. 1) with

standard parameters. Written for plane strain and simple shear it reads

$$\dot{\epsilon}_{xz} = \frac{1}{2} \frac{\partial u}{\partial z} = A \sigma_{xz}^n, \tag{6}$$

with the parameters $A = 75.7 \text{ MPa}^{-3} \text{ a}^{-1}$ and $n = 3$ recommended by Cuffey and Paterson (2010).

Data from ice deformation experiments is often quantitatively analyzed by plotting strain rate versus stress on a double-logarithmic plot (e.g., Glen, 1955; Goldsby and Kohlstedt, 2001). In such a graphics the slope of a straight line corresponds to the power n of the flow law. The resulting strain rates from the analysis of the tilt sensor time series are displayed in Figure 2. It is noteworthy that most data points are located within one cluster, whereas the data points for low and high stress are from only two boreholes (LAKE and DUCK). The conventionally used Glen flow law (Eq. 6) lies between the data points, but under- or overestimates the deformation rates.

A systematic relation between shear stress and strain-rate appears using the full data set with 63 values from ice with temperature within 0.1 K from the pressure melting temperature (PMT). In Figure 3 the data points cluster along two straight lines which are tentatively fitted with power laws for the regimes, corresponding to $n = 1$ and $n \simeq 22$. The $n = 1$ line shows that over a stress range between 0.03 and 0.157 MPa the borehole tilt rates are compatible with a linear flow relation with constant viscosity. The numerical value of viscosity $\eta \simeq 40 \cdot 10^{12} \text{ Pa s}$ is 40 times larger than the one determined in laboratory ring-shear experiments performed at similar stresses on ice with more than 0.6 % water content (Adams and others, 2021; Schohn and others, 2025).

At stresses exceeding a threshold of $\sigma_t = 0.157 \text{ MPa}$ the ice deformation rates cluster along a straight line with a very high slope. In this regime the measurements are approximated by a power law with $n = 22 \pm 5$. Such a strong nonlinearity is reminiscent of plastic deformation. We refer to this regime as shear-thinning, and the stress states are at the pseudo-plastic limit. It is important to note that these measurements stem from several sites with very different conditions (DUCK, GULL, Store Glacier, Athabasca Glacier and Glacier d'Argentière). The power 22 is a reasonable

approximation to the data set, but different values are possible, as inspection of the red lines for 5° difference in Figure 3 shows.

The determination of the fitting curves was performed visually, since data quality precludes a meaningful result with a fitting algorithm. The accuracy of the borehole velocity gradients is 0.005 a^{-1} for our own time series, but unknown for the literature values. The shear stresses were calculated with a crude approximation which relies on the poorly constrained relevant surface slope. Horizontal stress gradients pose an additional challenge. Furthermore, the difficulty to determine the relevant surface slope on an uneven glacier surface leads to a systematic error for all stresses in one borehole that is difficult to quantify.

Despite these difficulties of assessing data accuracy the pattern laid out by the data points appears robust. The rheology of temperate glacier ice suggested by the two lines displayed in Figure 3 that can be summarized as

$$\dot{\epsilon}_{xz} = \begin{cases} A_{lin} \sigma_{xz}, & 0.03 \leq \sigma_{xz} \leq 0.157 \text{ MPa} \\ A_{pp} \sigma_{xz}^{22}, & \sigma_{xz} > 0.157 \text{ MPa} \end{cases} \quad (7)$$

with parameters

$$\begin{aligned} A_{lin} &= \frac{1}{2\eta} = 1.25 \cdot 10^{-8} \text{ MPa}^{-1} \text{ s}^{-1}, \\ \eta &= 40 \cdot 10^{12} \text{ Pa s}, \\ A_{pp} &= 1.11 \cdot 10^9 \text{ MPa}^{-22} \text{ s}^{-1}. \end{aligned} \quad (8)$$

The flow law parameters are readily converted to glaciologically more meaningful units

$$\begin{aligned} A_{lin} &= 0.4 \text{ MPa}^{-1} \text{ a}^{-1}, \\ A_{pp} &= 3.5 \cdot 10^{16} \text{ MPa}^{-22} \text{ a}^{-1}. \end{aligned} \quad (9)$$

DISCUSSION

Measured ice deformation rates from 16 boreholes show that temperate glacier ice exhibits shear deformation rates that are significantly different from those calculated with the standard parameters of Glen's flow law (Eq. 1). Instead, the temperate ice can be classified into two flow regimes with a transition at $\sigma_t = 0.157 \text{ MPa}$.

Constant-viscosity regime

A constant-viscosity rheology ($n = 1$) explains the borehole tilt data in an intermediate range of stresses that are relevant to most temperate portions of glaciers and ice sheets. This observation supports the findings from laboratory experiments that temperate ice with high water content has a linear (constant-viscosity) rheology (Adams and others, 2021; Schohn and others, 2025). The spread of values around the red line in Figure 3 is to be expected since the ice at different sites and depths contains varying amounts of interstitial water, and therefore has a softness and deformation rate which varies up to a factor three (Duval, 1977; Morgan, 1991). How much grain size affects these values cannot be assessed since it has not been determined for any of the boreholes.

The factor 40 difference between viscosities from field measurements and laboratory experiments merits further investigation. One possible explanation is a grain size effect. The laboratory grains had an average size of 8 mm (Schohn and others, 2025, supp. inf.), whereas grains in glacier ice are often an order of magnitude larger. If the linear flow relation is mainly due to diffusional flow (as suggested by Schohn and others, 2025), deformation depends with a power $p = 2$ or

$p = 3$ on grain diameter (Nabarro-Herring and Coble creep). Therefore, the factor 40 can be explained by grain sizes within the glacier of 50 or 27.5 mm. No direct information on grain size or ice fabrics exists for the boreholes considered. However, ice from the glacier bottom exposed at ice margins or in tunnels in alpine glaciers often features grains in this size range. For example, grain diameters of temperate ice reached 10 – 50 mm at 60–80 m depth in Rhonegletscher (Hellmann and others, 2021) and 15–30 mm at 80–180 m depth in Vallée Blanche (Vallon and others, 1976), and up to 90 mm in Storglaciären (Disbrow-Monz and others, 2024).

The linear rheology with $n = 1$ is not unique to our data set. Several earlier studies analyzing borehole data found values of $n \sim 1$ at stresses up to 0.1 MPa (Meier, 1960; Marshall and others, 2002) or $n = 1.3$ (Colbeck and Evans, 1973) or $n = 1.5$ (Gerrard and others, 1952). An optimization analysis of borehole deformation measurements yielded a best fit for $n = 1$ (Chandler and others, 2008), although with a viscosity that is 5 times higher than the one determined here. For laboratory deformation tests at very low stresses (below 0.01 MPa) the best fits also indicated $n \sim 1$ (Butkovich and Landauer, 1960).

Pseudo-plastic regime

The pseudo-plastic limit apparent for temperate ice at stresses exceeding $\sigma_t = 0.157 \text{ MPa}$ is most noteworthy. There are two obvious explanations for such a stress limit: Either it is caused by accelerated deformation or failure within the polycrystalline temperate ice, or alternatively by a strongly non-linear behavior of water-saturated subglacial sediments. Given the arguments below we believe that the first option explains the phenomenon.

The pseudo-plastic limit of Figure 3 (orange line) is traced out by points from five drill sites with different conditions: DUCK (Sermeq Kujalleq; Jakobshavn Isbræ), GULL (Paakitsoq area), BH14b (Store Glacier), borehole 209 (Athabasca Glacier), and borehole B2 (Glacier d'Argentière). The wide variety of basal conditions occurring at these sites makes it unlikely that the stress limit is due to failure of subglacial sediments.

A similar limiting stress, above which the strain rate increases rapidly, was also found in several laboratory experiments. The ring-shear experiments (Schohn and others, 2025) show an upper limit for wet temperate ice, although at a higher shear stress of 0.24 MPa. The ice deformation experiments by Steinemann (1958) conducted at -1.9°C showed that the tertiary creep was strongly stress dependent, with a regime change between 1.2 – 1.6 MPa, and a stress exponent of the order 10 (his Fig. 17). Invoking again a grain size effect would approximately explain the different limiting stresses. From the ratio of limiting stresses a grain size of $\sim 20 \text{ mm}$ can be derived using the grain size dependence with power $p = -\frac{1}{2}$ for tensile strength (Frost, 2001; Ranganathan and others, 2021).

At the scale of ice sheets and glaciers an upper limit to the bed-parallel shear stress has been proposed for decades and for many settings. Such a stress limit has proven to be a useful approximation to explain large-scale ice flow behavior. The relevant limiting stress, often called yield stress, is of the order $\sigma_t = 0.157 \text{ MPa}$. The overall shape of ice sheets, as well as smaller basins, could be predicted using a perfect plastic yield stress of the order 0.09 – 0.1 MPa (Vialov, 1958; Reeh, 1982). A plastic yield strength in the range of 0.11 – 0.18 MPa was obtained by inverting surface topography for basal yield stress

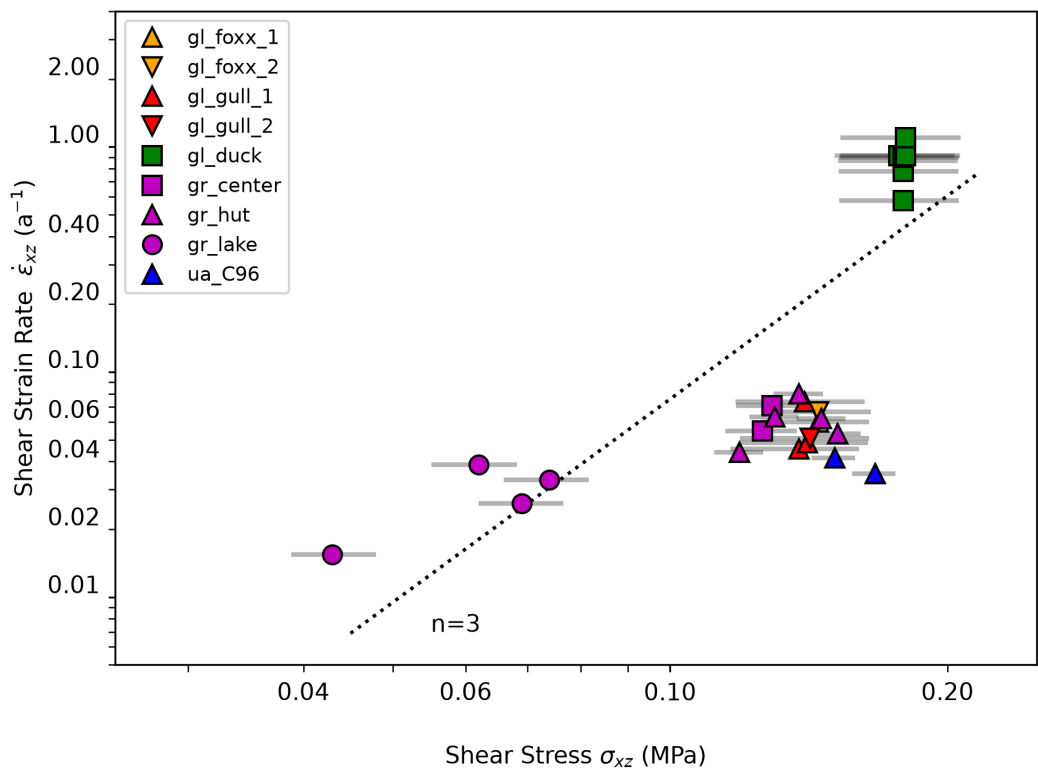


Fig. 2. Shear strain rates of 32 individually fitted tilt sensors in temperate ice in 9 boreholes. Error bars on shear stress indicate an uncertainty in surface slope angle of 0.2° . Error bars for strain rates are smaller than the symbols. The dotted line shows Glen’s flow law ($n = 3$) with standard parameters for temperate ice (Cuffey and Paterson, 2010).

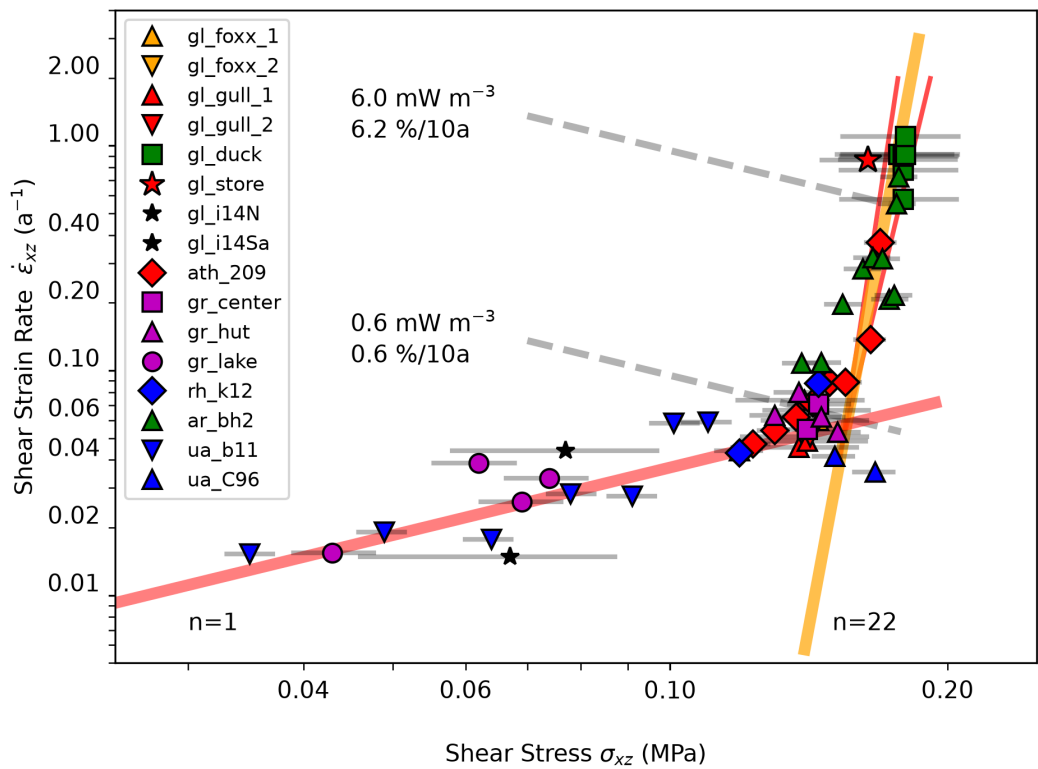


Fig. 3. Shear strain rates of 63 tilt sensors in temperate ice in 16 boreholes. Error bars on shear stress indicate an uncertainty in surface slope angle of 0.2° . The wide red and orange lines show a linear ($n = 1$) and a pseudo-plastic flow relation ($n = 22$), while thin red lines indicate $n = 22 \pm 5$. Dashed gray lines are contours of dissipation rate, labeled by the amount of water that is melted within 10 years.

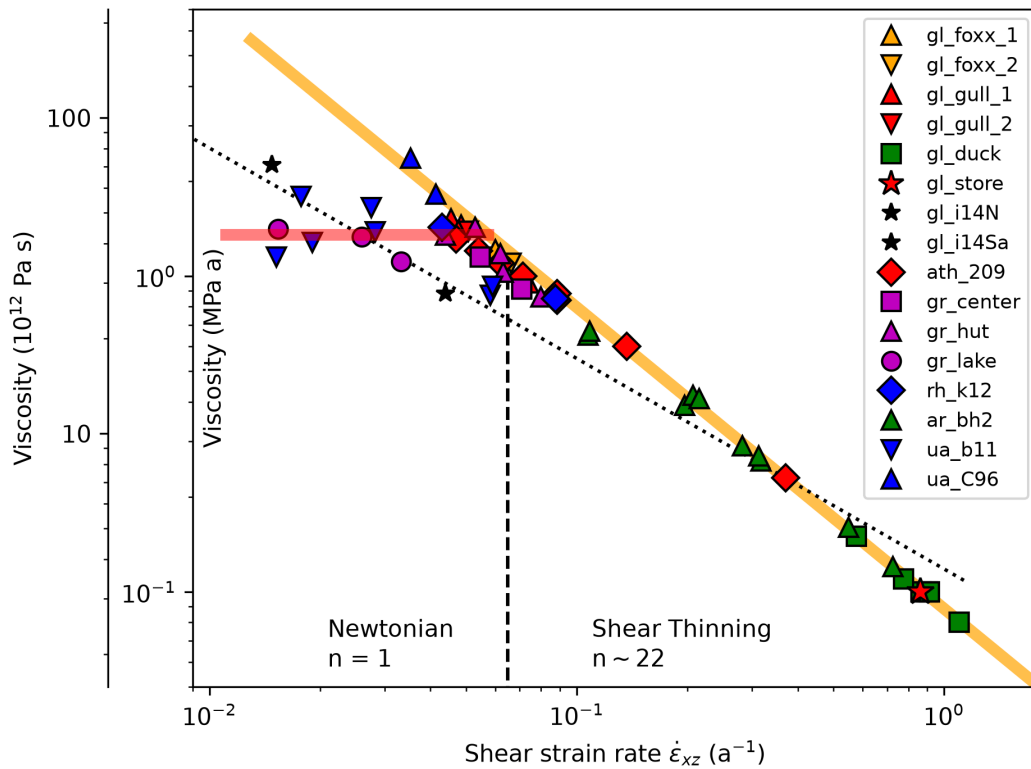


Fig. 4. The data points and lines from Figure 3 are displayed as shear viscosity. The black dotted line indicates the viscosity calculated from Glen's flow law. The deformation regimes are indicated at the bottom.

(May and Bassis, 2025). Investigating the conditions for basal motion, Krabbendam (2016) invoked a stress limit of 0.1 MPa to explain the formation and persistence of a temperate basal layer. Recent studies of tidewater glacier dynamics (Ultee and Bassis, 2016) and calving dynamics (Bassis and Ultee, 2019) rely on a perfectly plastic rheology with a yield stress of the order 0.15 MPa, or a threshold stress of 0.17 MPa for calving of tidewater glaciers (Mercenier and others, 2018).

Deformation regimes and time scales

Our data shows two distinct rheological regimes for the deformation of temperate glacier ice. The transition occurs at a shear stress of $\sigma_t = 0.157$ MPa, with corresponding shear strain rate of $\dot{\epsilon}_t = 0.061 \text{ a}^{-1}$. It is instructive to calculate the dissipation rate at the regime transition

$$P = \dot{\epsilon}_{ij} \sigma'_{ij} \approx 2 \dot{\epsilon}_{xz} \sigma_{xz}, \quad (10)$$

which gives a heat production rate of 0.6 mW m^{-3} . If the entire dissipative heat is used to melt ice, the meltwater production rate is high enough to reach a water content of 0.6 % in the ice matrix within 10 years. For context, contour lines of dissipation rate are shown in Figure 3.

Since temperate glacier ice is subject only to tiny heat fluxes caused by the Clausius-Clapeyron gradient, the ice is under nearly adiabatic conditions without heat exchange with the surroundings. The consequence is that all dissipative heat production is used locally to drive phase change, and is stored as latent heat in form of water between the ice crystals, at least as long as the pore space precludes water percolation. Above the regime transition it only takes a few decade to reach water contents exceeding 2 % when percolation becomes important (Fowler and Iverson, 2023). As our measurements

suggest, such high water contents activate a pseudo-plastic rheology which is likely due to granular physics of the wet grains under high fluid pressure. This pseudo-plastic limit for glacier ice is reminiscent of the dynamics of partially molten rocks which experience large strength drops with increasing liquid content (e.g., Rosenberg and Handy, 2005). Their study finds a transition of the number of wetted grain boundaries that coincides with a strength reduction at a melt fraction of 0.07. Given the orders of magnitude contrast in melt viscosity of water versus magma, it is likely that many of the relevant processes are of similar origin. Ice could therefore be regarded as an analogue to high-temperature deformation of rocks, as has been advocated by numerous authors (Wilson and others, 2014, and references therein).

Ice viscosity

Plotting the data points from Figure 3 in terms of shear viscosity

$$\eta = \frac{1}{2} \frac{\sigma_{xz}}{\dot{\epsilon}_{xz}} \quad (11)$$

provides further insights into the relevant processes. Figure 4 shows the two distinct flow regimes. In the Newtonian regime the viscosity is a constant and in the shear-thinning regime the viscosity decreases with deformation rate.

The combined linear and pseudo-plastic behavior can be described with the Ellis model (Boger, 1977; Chhabra and Richardson, 2008) which is often used in material science

$$\eta = \frac{\eta_0}{1 + (\tau/\tau_m)^{\gamma-1}}. \quad (12)$$

The Ellis model with parameters $\eta_0 = 1.33 \text{ MPa a}$, $\tau_m = \sigma_t = 0.157 \text{ MPa}$ and $\gamma = 22$ was used to replace the Newtonian and pseudo-plastic lines with one single curve shown in Figure 5.

The viscosity in the Newtonian regime is spread out considerably to values that lay above the horizontal line which appears to delineate a lower bound for viscosity. Our interpretation is that in this realm (highlighted as blue area) the ice is increasingly watery towards lower viscosity values, until it reaches the limit of 0.6 % where constant-viscosity behavior occurs. Higher viscosity values would therefore correspond to ice with a low water content, as implied by the dependence of viscosity on interstitial water (Duval, 1977; Adams and others, 2021; Schohn and others, 2025).

An upper bound of temperate ice viscosity is apparent in Figure 5. This strain rate dependent upper bound seems to hold for a large range of deformation rates, spanning three orders of magnitude in strain rate and viscosity. Even at the high strain rates of ring-shear laboratory experiments (Duval, 1977; Schohn and others, 2025) the viscosities are close to this limiting viscosity (Fig. 7A). Fitting (again visually) a power law through all temperate data points results in a power $n \sim 11$ for the limiting viscosity (Fig. 7B). The spread of possible values for this important quantity is large, but cannot be better constrained with the present data set.

The limiting viscosity represented by the orange curve in Figure 5 can be expressed in function of strain rate by

$$\eta = B \dot{\epsilon}_{xz}^{-\mu} = 0.088 \dot{\epsilon}_{xz}^{-0.955} \text{ MPa a}, \quad (13)$$

where $\mu = 1 - 1/n \sim 0.955$. For the blue curve of Figure 7B we calculate $\mu = 0.91$, which results in an uncertainty range that appears more acceptable. These values for a limiting viscosity are similar to findings from deep boreholes on Athabasca Glacier (Raymond, 1969). That analysis (his Figure 36) yielded a viscosity that was systematically decreasing with strain rate in the lower parts of the boreholes. Fitting those curves with the expression (13) yielded values of B between 0.068 and 0.182 with a mean of 0.085, and values of μ from 0.56 to 0.98 with a mean of 0.81 (his Table 16). Such a good qualitative, and even quantitative, agreement between analyses performed with completely different methodologies suggests an universal phenomenon.

There are other geophysical systems that show a similar rheological behavior. In experimental data for silica melts (magma) with suspended crystals a strain-rate dependent viscosity that is similar to our Figure 5 was found (Webb and Dingwell, 1990; Wadsworth and others, 2018; Vasseur and others, 2023). Their hypothesis that the flow of liquid melt in the gaps between moving crystals may be the source of the observed shear thinning is analogous to the composite system of ice crystals with interstitial water. Concurrently, they found a limiting value for shear stress τ_m that is universal to all silicate liquids. Such a plasticity effect seems to be strongly influenced by the shapes of the crystals in the system (Deubelbeiss and others, 2011).

RTIG rheology for temperate ice

The main result of this study is a new rheology for shear deformation of temperate ice in glaciers (RTIG). Equations (10) and (12) with relevant parameters are repeated here in a form that can be easily implemented in numerical codes

$$\begin{aligned} \dot{\epsilon}_{xz} &= \frac{1}{2\eta} \sigma_{xz}, \\ \eta &= \frac{\eta_0}{1 + (\sigma_{xz}/\sigma_t)^{21}}, \\ \dot{\epsilon}_{xz} &= \frac{1}{2\eta_0} \left[1 + \left(\frac{\sigma_{xz}}{\sigma_t} \right)^{21} \right] \sigma_{xz}, \end{aligned}$$

$$\dot{\epsilon}_{xz} = A_R \left[1 + \left(\frac{\sigma_{xz}}{\sigma_t} \right)^{21} \right] \sigma_{xz}, \quad (14)$$

with parameters

$$\begin{aligned} \eta_0 &= 1.33 \text{ MPa a}, \\ \sigma_t &= 0.157 \text{ MPa}, \\ A_R &= 0.376 \text{ MPa}^{-1} \text{ a}^{-1}. \end{aligned} \quad (15)$$

The last of these expressions is in a form familiar to glaciologists. Writing this equation with scales for stress and strain rate facilitates an easier comparison between Glen's flow law and RTIG. With a reference strain rate $\dot{\epsilon}_t = 0.061 \text{ a}^{-1}$ the equations read

$$\frac{\dot{\epsilon}_{xz}}{\dot{\epsilon}_t} = 4.8 \left(\frac{\sigma_{xz}}{\sigma_t} \right)^3, \quad (16)$$

$$\frac{\dot{\epsilon}_{xz}}{\dot{\epsilon}_t} = \left[1 + \left(\frac{\sigma_{xz}}{\sigma_t} \right)^{21} \right] \frac{\sigma_{xz}}{\sigma_t} = \frac{\sigma_{xz}}{\sigma_t} + \left(\frac{\sigma_{xz}}{\sigma_t} \right)^{22}. \quad (17)$$

The last expression clearly shows the importance of σ_t as a transition stress between the linear and the strongly nonlinear regimes. For $\sigma_{xz}/\sigma_t < 1$ the second term is small whereas it dominates for $\sigma_{xz}/\sigma_t > 1$. At the transition stress Glen's flow law overestimates deformation rates by a factor 5.

How well this new rheology reproduces the measured borehole deformation rates is visualized in Figure 1. For all 16 investigated boreholes the data are shown together with the borehole deformation curves calculated with Glen's flow law and the RTIG rheology. To facilitate plotting deformation curves over the whole ice thickness, we complement RTIG with a tentative power law for cold ice, which is corrected for temperature with factor given by Equation (20) (Smith and Morland, 1981). For this we use Equation (1) and assume, without much justification or tuning, $A_4 = 2500 \text{ MPa}^{-4} \text{ a}^{-1}$ and $n = 4$. This choice is in line with recent studies which find $n = 3.5 - 4$ for cold ice in laboratory studies (e.g., Treverrow and others, 2012) and from remote sensing data (Bons and others, 2018; Millstein and others, 2022; Ranganathan and Minchew, 2024), although our softness parameter A_4 is higher than in these studies. Further support stems from the analysis of the GrIS flow patterns by Bons and others (2018). They minimized inferred basal motion by assuming an ice rheology for the mainly cold ice with $n = 4$, a transition stress of 0.03 MPa, and a maximum stress of about 0.15 MPa.

Within the temperate ice (red dots in Fig. 1) the agreement between RTIG and the data is satisfactory in most boreholes. The improvement of the shapes of deformation profiles with respect to Glen's flow law is considerable. Clearly visible is the strong increase in deformation rates in ice close to bedrock where the transition stress σ_t is exceeded. In contrast, discrepancies between measurements and the modeled curves in the cold ice are much larger. This is expected for an ad-hoc power-law rheology without much tuning. Adding to the complexity of the problem, all Greenland drill sites feature thick layers of ice from the Wisconsin ice age which is known for high variability in grain size, fabric, and therefore strain rates (e.g., Dahl-Jensen and others, 1997; Thorsteinsson and others, 1999). These data should be analyzed with a self-consistent approach such as the Wattmeter or an adapted composite rheology that also includes grain size (Behn and others, 2021; Kuiper and others, 2020).

We should emphasize again the large uncertainties in relevant shear stress that certainly contribute to the misfits in Figure 1. For example, using a lower surface slope for

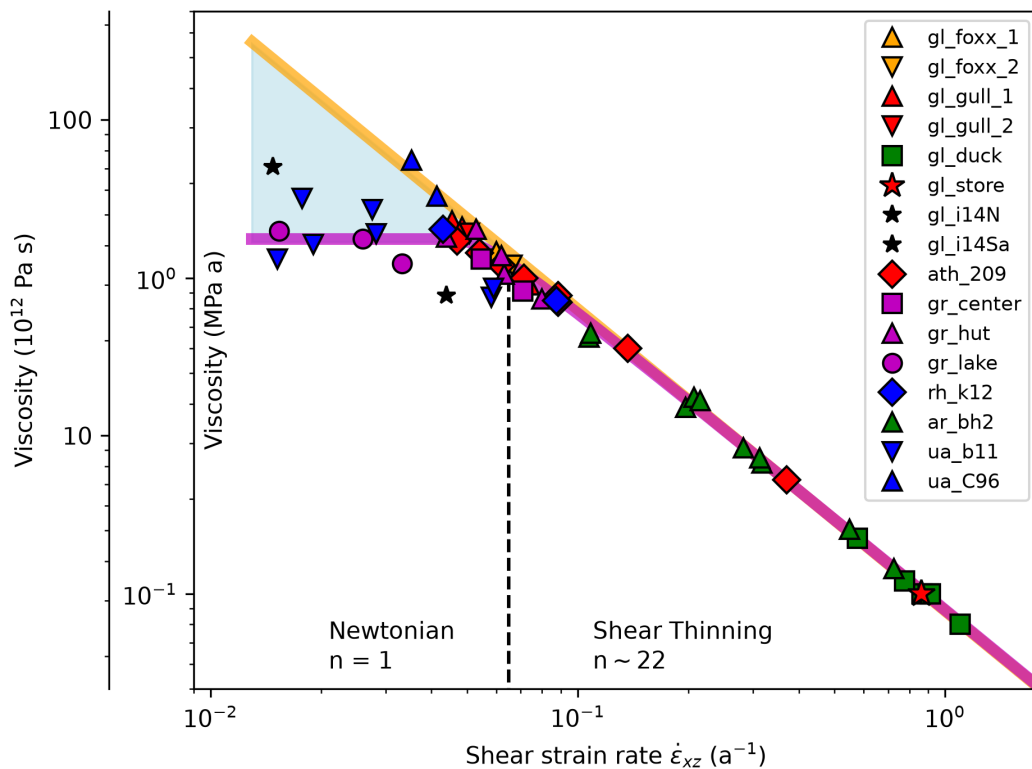


Fig. 5. The shear viscosity calculated from shear stresses and strain rates. The wide purple line shows the Ellis rheology. The blue area marks the region where different water contents lead to viscosity variations of a factor two. The orange line (pseudo-plastic) appears to be an upper limit for the viscosity.

borehole ua_C96 would align these points with the calculated curve. It is likely that along-flow and transverse stresses alter the conditions at the sensors, as do the influence of basal undulations which create stress shadows and concentrations. Given all these uncertainties it is encouraging to observe that one single flow law captures the essential features of the vertical ice deformation profiles.

Significance for glaciology

Most alpine-type glaciers are temperate, or feature layers of temperate ice close to the base where stresses are highest. Also polar ice masses often rest on temperate ice close to their base (e.g., MacGregor and others, 2016). In numerical models of such ice bodies the RTIG could be used as a drop-in replacement for Glen's flow law. Obviously, for high stresses the strong non-linearity of an Ellis fluid leads to higher fluxes than calculated with the power-law rheology of Glen's flow law. This effect would alter the emergent surface topography and likely reduce the need for tuning of parametrizations for basal motion and ice enhancement factors, which is currently often done to achieve better agreement between models and data.

We illustrate the importance of the rheological switch at $\sigma_t = 0.157$ MPa by a comparison between the Ellis/RTIG rheology and Glen's flow law in Figure 6. The RTIG rheology features a strong nonlinearity above the transition stress σ_t that imposes a soft upper bound on the ice thickness that a glacier can achieve. Figure 6A shows that at a fixed surface slope and velocity the ice thickness can reach higher values using Glen's flow. Also the shapes of the deformation profiles are sensitively dependent on the ice thickness (panels B and

C in Fig. 6). For an increase in ice thickness of only 4 m the velocity nearly doubles if the thickness is such that the shear stress is close to the transition stress.

The ice flux of most glaciers is limited by their finite accumulation area, meaning that a strong increase in ice velocity cannot be compensated by increased upstream ice flux, and consequently the glacier is thinning until balance is established. This has the effect that most temperate glaciers should be close to the pseudo-plastic limit. In view of these ideas, the clustering of data points between 0.13 and 0.15 MPa in Figure 3 is striking. Our interpretation is that these glaciers have relaxed to a geometry where shear stress near the base is close to the transition stress. Only glaciers with a very large accumulation area, such as Athabasca glacier or ice sheet outlet glaciers, have an effectively unlimited supply of ice that is able to sustain high ice velocities during extended periods of time. Indeed, all data points on the pseudo-plastic limit are from such settings. A normal mountain glacier can only temporarily, and under rare climatological circumstances, pass the transition stress.

The strong nonlinearity in the RTIG flow law could also provide an explanation for kinematic waves on glaciers, such as the one observed on Mer de Glace from 1891 to 1899 (Lliboutry and Reynaud, 1981) and the surge-like advance of Vernagtferner at the end of the Little Ice Age. Under even more extreme circumstances, the nonlinearity might also explain the occurrence of rapidly repeating glacier surges such as the ones observed in the Karakoram and Pamir mountain ranges (Paul, 2015), which to date have been attributed mostly to basal processes (Harrison and Post, 2003; Benn and others, 2019).

For models relying on a depth-integrated flux (shallow ice approximation) an analytic expression for the flow speed and the ice flux are needed. The flow of an Ellis fluid has been analyzed for internal flow and various geometries (Matsuhisa and Bird, 1965), for gravity-driven flow on an infinite plane (Chhabra and Richardson, 2008, p. 162), and for a 3D free-surface flow (Kheifets and Kieweg, 2013). The solution for an infinite plane-strain flow of thickness H , width W and slope α is (using the short notation $S = \rho g \sin \alpha$)

$$v = \dot{\epsilon}_t \frac{S}{\sigma_t} H^2 + \frac{2\dot{\epsilon}_t}{22} \left(\frac{S}{\sigma_t} \right)^{22} H^{23} \quad (18)$$

$$Q = \frac{\dot{\epsilon}_t}{3} \frac{S}{\sigma_t} W H^3 + \frac{2\dot{\epsilon}_t}{23} \left(\frac{S}{\sigma_t} \right)^{22} W H^{24}. \quad (19)$$

The first term in Equation (19) is the flux due to the linear viscous flow. It is clear that the strong nonlinearity of the second term leads to very high ice fluxes as soon as the transition stress σ_t is exceeded. Such a strongly nonlinear behaviour therefore controls the shape of glaciers. If the ice grows somewhat thicker than a reference steady state, the fast ice flux leads to rapid thinning of thick glacier parts, and is therefore a limit for ice thickness. The opposite effect is active when the thickness is less than the reference state. The ice flux slows considerably, which lets the ice thickness grow towards the reference state.

CONCLUSIONS

In most dynamically important parts of glaciers and polar ice streams the ice at the base is at its melting temperature and contains liquid water between crystal boundaries. We investigated the rheology of such ice using tilt sensor data from 16 boreholes. The emergent rheology shows a linear ($n = 1$) rheology for intermediate shear stresses. A deformation-rate dependent upper bound for ice viscosity was found with a strong nonlinearity of $n \sim 22$. The effect of the new RTIG rheology is a limit on ice thickness that can be achieved in those parts of glaciers and ice sheets where the ice accumulates water from dissipative heating during many decades.

The results of this study clearly show that the commonly assumed power-law rheology (Glen's flow law) is a poor approximation for the dynamics of temperate glacier ice, and thus for most fast-flowing portions of ice sheets and mountain glaciers. Obviously, such a simple relation cannot capture the wide variety of fabric, grain size and water content in ice sheets and glaciers. The hallmark signature of this shortcoming is the frequent use in the literature of an enhancement factor, that merely captures the misfit without much explanatory power.

The RTIG rheology has been derived and tested for deformation in vertical shear. It is desirable to extend it to multi-axial stress states, and to test such a relation with the available data from field and remote sensing studies. The resulting rheological parameters should be constrained by laboratory experiments and by in-situ deformation tests on glaciers. Especially the power of stress in the pseudo-plastic regime has a large uncertainty which ranges from $n \sim 11$ (Fig. 7B) to 25. Expressing the same relation in terms of strain rate, the power μ in Equation (13) ranges from 0.91 to 0.96 which is an uncertainty range that appears more acceptable.

At stresses below 0.03 MPa no rheology can be inferred from our data. Considering other borehole studies (Marshall and others, 2002; Helbing, 2006) it seems reasonable to assume $n = 1$, potentially with the same rheology. While experiments at low strain rates are difficult due to very long experiment duration, the rheology at low stress is barely relevant for large-scale ice flow, as it affects modeled flow velocities only marginally. Besides the RTIG also other alternative rheologies, such as the polynomial flow law (Colbeck and Evans, 1973), yield a better agreement with the borehole data than Glen's flow law (Fig. 8). It seems surprising that such relations have not found a more widespread adoption in the modeling community.

The new RTIG rheology facilitates more accurate modeling of the deformation of temperate ice which is essential for assessing ice sheet discharge to the ocean and therefore predictions of sea-level rise.

ACKNOWLEDGEMENTS

We thank the many people involved in the development and manufacturing of the borehole sensor systems and the participants of the drilling campaigns, especially Cornelius Senn, Thomi Wyder and Claudia Ryser. This research was funded by Swiss National Science Foundation grants 21-39582.93, 200021-127197, 200020-111892 and 200020-197015.

REFERENCES

- Adams CJC, Iverson NR, Helanow C, Zoet LK and Bate CE (2021) Softening of temperate ice by interstitial water. *Frontiers in Earth Science*, **9** (doi: 10.3389/feart.2021.702761)
- Aschwanden A, Fahnestock MA and Truffer M (2016) Complex Greenland outlet glacier flow captured. *Nature. Communications*, **7** (doi: 10.1038/ncomms10524)
- Bassis JN and Ultee L (2019) A thin film viscoplastic theory for calving glaciers: Toward a bound on the calving rate of glaciers. *Journal of Geophysical Research: Earth Surface*, **124**(8), 2036–2055 (doi: 10.1029/2019j005160)
- Bauder A (2001) Bestimmung der Massenbilanz von Gletschern mit Fernerkundungsmethoden und Fließmodellierungen: Eine Sensitivitätsstudie auf dem Unteraargletscher. Mitteilung 169, Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie der ETH Zürich, Diss. ETH 14110
- Behn MD, Goldsby DL and Hirth G (2021) The role of grain size evolution in the rheology of ice: Implications for reconciling laboratory creep data and the Glen flow law. *The Cryosphere*, **15**(9), 4589–4605 (doi: 10.5194/tc-15-4589-2021)
- Benn DI, Fowler AC, Hewitt I and Sevestre H (2019) A general theory of glacier surges. *Journal of Glaciology*, **65**(253), 701–716, ISSN 1727-5652 (doi: 10.1017/jog.2019.62)
- Boger DV (1977) Demonstration of upper and lower newtonian fluid behaviour in a pseudoplastic fluid. *Nature*, **265**(5590), 126–128, ISSN 1476-4687 (doi: 10.1038/265126a0)
- Bons PD and 6 others (2018) Greenland Ice Sheet: Higher nonlinearity of ice flow significantly reduces estimated basal motion. *Geophysical Research Letters*, **45**(13), 6542–6548, ISSN 1944-8007 (doi: 10.1029/2018gl078356)
- Bradford JH, Nichols J, Mikesell TD and Harper JT (2009) Continuous profiles of electromagnetic wave velocity and water content in glaciers: an example from Bench Glacier, Alaska, USA. *Annals of Glaciology*, **50**, 1–9 (doi: 10.3189/172756409789097540)
- Butkovich TR and Landauer JK (1960) Creep of ice at low stresses. *CRREL Res. Rep.*, **72**, 6
- Chandler D, Hubbard B, Hubbard A, Murray T and Rippin D (2008) Optimising ice flow law parameters using bore-

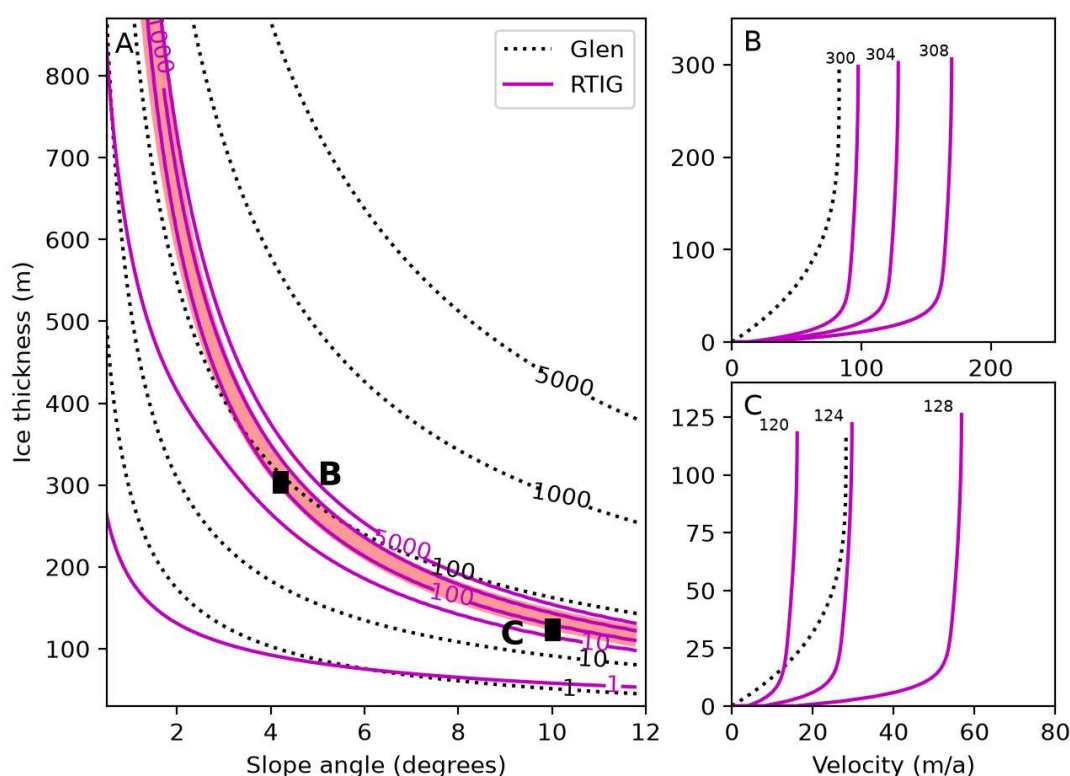


Fig. 6. Comparison of velocities calculated with Glen's flow law (dotted line) and RTIG (purple). A: Contours of surface velocity for a given slope angle and ice thickness. Velocities are indicated on the contours. Black squares locate the velocity profiles shown in the small panels. B,C: Vertical velocity profiles for surface slopes of 4.2° and 10° show a strong sensitivity with respect to the ice thickness, which is indicated next to the lines.

- hole deformation measurements and numerical modelling. *Geophysical Research Letters*, **35**(12), 2008GL033801 (doi: 10.1029/2008gl033801)
- Chhabra R and Richardson J (2008) *Non-Newtonian Fluid Behaviour*, 1–55. Elsevier, ISBN 9780750685320 (doi: 10.1016/b978-0-7506-8532-0.00001-9)
- Colbeck SC and Evans RJ (1973) A flow law for temperate glacier ice. *Journal of Glaciology*, **12**(64), 71–86, ISSN 1727-5652 (doi: 10.3189/s0022143000022711)
- Cuffey K and Paterson W (2010) *The Physics of Glaciers*. Elsevier, Burlington, MA, USA, ISBN 978-0-12-369461-4
- Dahl-Jensen D, Thorsteinsson T, Alley R and HShoji (1997) Flow properties of the ice from the Greenland Ice Core Project ice core: The reason for folds? *Journal of Geophysical Research*, **102**(C12), 26831–26840 (doi: 10.1029/97jc01266)
- Deubelbeiss Y, Kaus BJP, Connolly JAD and Caricchi L (2011) Potential causes for the non-Newtonian rheology of crystal-bearing magmas. *Geochemistry, Geophysics, Geosystems*, **12**(5), 2010GC003485 (doi: 10.1029/2010gc003485)
- Disbrow-Monz ME, Hudleston PJ and Prior DJ (2024) Multimaxima crystallographic fabrics (CPO) in warm, coarse-grained ice: New insights. *Journal of Structural Geology*, **182**, 105107, ISSN 0191-8141 (doi: 10.1016/j.jsg.2024.105107)
- Doyle SH and 7 others (2018) Physical conditions of fast glacier flow: 1. measurements from boreholes drilled to the bed of Store Glacier, West Greenland. *Journal of Geophysical Research: Earth Surface*, **123**(2), 324–348 (doi: 10.1002/2017j004529)
- Duval P (1977) The role of water content on the creep rate of polycrystalline ice. In *Isotopes and impurities in snow and ice*, 29–33, International Association of Hydrological Sciences, publication No. 118
- Fowler JR and Iverson NR (2023) The relationship between the permeability and liquid water content of polycrystalline temperate ice. *Journal of Glaciology*, **69**(278), 1–9 (doi: 10.1017/jog.2023.91)
- Frost H (2001) Mechanisms of crack nucleation in ice. *Engineering Fracture Mechanics*, **68**(17–18), 1823–1837 (doi: 10.1016/s0013-7944(01)00036-4)
- Gerrard J, Perutz M and Roch A (1952) Measurement of the velocity distribution along a vertical line through a glacier. *Proceedings of the Royal Society of London. Series A. Mathematical and Physical Sciences*, **213**(1115), 546–558, ISSN 2053-9169 (doi: 10.1098/rspa.1952.0144)
- Glen JW (1955) The creep of polycrystalline ice. *Proceedings of the Royal Society of London. Series A. Mathematical and Physical Sciences*, **228**(1175), 519–538, ISSN 2053-9169 (doi: 10.1098/rspa.1955.0066)
- Goldsby DL and Kohlstedt DL (2001) Superplastic deformation of ice: Experimental observations. *Journal of Geophysical Research: Solid Earth*, **106**(B6), 11017–11030, ISSN 0148-0227 (doi: 10.1029/2000jb900336)
- Gudmundsson GH (2007) Tides and the flow of Rutford Ice Stream, West Antarctica. *Journal of Geophysical Research*, **112**(F04007) (doi: 10.1029/2006JF000731)
- Gudmundsson GH, Bauder A, Lüthi M, Fischer UH and Funk M (1999) Estimating rates of basal motion and internal ice deformation from continuous tilt measurements. *Annals of Glaciology*, **28**, 247–252, ISSN 1727-5644 (doi: 10.3189/172756499781821751)
- Gusmeroli A, Murray T, Jansson P, Pettersson R, Aschwanden A and Booth AD (2010) Vertical distribution of water within the polythermal Storglaciären, Sweden. *Journal of Geophysical Research*, **115**(F04002) (doi: 10.1029/2009JF001539)

- Harrison WD and Post AS (2003) How much do we really know about glacier surging? *Annals of Glaciology*, **36**(1), 1–6
- Haseloff M, Hewitt IJ and Katz RF (2019) Englacial pore water localizes shear in temperate ice stream margins. *Journal of Geophysical Research: Earth Surface*, **124**(11), 2521–2541 (doi: 10.1029/2019jg005399)
- Helbing J (2006) Glacier dynamics of Unteraargletscher: Verifying theoretical concepts through flow modeling. Mitteilungen 196, Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie der ETH Zürich, Gloriastrasse 37-39, ETH Zürich, CH-8092 Zürich, pp. 214
- Hellmann S and 7 others (2021) Crystallographic analysis of temperate ice on Rhonegletscher, Swiss Alps. *The Cryosphere*, **15**(2), 677–694 (doi: 10.5194/tc-15-677-2021)
- Hoinkes HC (1969) Surges of the Vernagtferner in the ötztal Alps since 1599. *Canadian Journal of Earth Sciences*, **6**(4), 853–861 (doi: 10.1139/e69-086)
- Hubbard BP, Hubbard A, M Mader H, Tison JL, Grust K and Nienow PW (2003) Spatial variability in the water content and rheology of temperate glaciers: Glacier de Tsanfleuron, Switzerland. *Annals of Glaciology*, **37**, 1–6, ISSN 1727-5644 (doi: 10.3189/172756403781815474)
- Johnson SG (2012) *The NLOpt nonlinear-optimization package*. [Http://ab-initio.mit.edu/nlopt](http://ab-initio.mit.edu/nlopt)
- Joughin IR, Tulaczyk S and Engelhardt HF (2003) Basal melt beneath Whillans Ice Stream and Ice Streams A and C, West Antarctica. *Annals of Glaciology*, **36**, 257–262, ISSN 1727-5644 (doi: 10.3189/172756403781816130)
- Keller A and Blatter H (2012) Measurement of strain-rate components in a glacier with embedded inclinometers. *Journal of Glaciology*, **58**(210), 692–698 (doi: 10.3189/2012JoG11J234)
- Kheyfets VO and Kieweg SL (2013) Gravity-driven thin film flow of an ellis fluid. *Journal of Non-Newtonian Fluid Mechanics*, **202**, 88–98, ISSN 0377-0257 (doi: 10.1016/j.jnnfm.2013.09.010)
- Krabbendam M (2016) Sliding of temperate basal ice on a rough, hard bed: Creep mechanisms, pressure melting, and implications for ice streaming. *The Cryosphere*, **10**(5), 1915–1932 (doi: 10.5194/tc-10-1915-2016)
- Kuiper EJN, de Bresser JHP, Drury MR, Eichler J, Pennock GM and Weikusat I (2020) Using a composite flow law to model deformation in the neem deep ice core, greenland - part 2: the role of grain size and premelting on ice deformation at high homologous temperature. *The Cryosphere*, **14**(7), 2449–2467 (doi: 10.5194/tc-14-2449-2020)
- Law R, Christoffersen P, MacKie E, Cook S, Haseloff M and Gagliardini O (2023) Complex motion of Greenland Ice Sheet outlet glaciers with basal temperate ice. *Science Advances*, **9**(6), eabq5180 (doi: 10.1126/sciadv.abq5180)
- Leger TPM and 9 others (2025) A data-consistent model of the last glaciation in the Alps achieved with physics-driven AI. *Nature Communications*, **16**(1), 848 (doi: 10.1038/s41467-025-56168-3)
- Lilien DA, Ranganathan M and Shapero DR (2026) Effect of the flow-law exponent on ice-stream sensitivity to melt. *Journal of Geophysical Research: Earth Surface*, **131**(3), e2025JF008726 (doi: 10.1029/2025jg008726)
- Lliboutry L (1976) Physical processes in temperate glaciers. *Journal of Glaciology*, **16**(74), 151–158, ISSN 1727-5652 (doi: 10.3189/s002214300003149x)
- Lliboutry L and Reynaud L (1981) 'Global dynamics' of temperate vally glacier, Mer de Glace, and past velocity deduced from Forbes' bands. *Journal of Glaciology*, **27**(96), 207–226, ISSN 1727-5652 (doi: 10.3189/s0022143000015367)
- Lüthi MP, Funk M, Iken A, Gogineni S and Truffer M (2002) Mechanisms of fast flow in Jakobshavn Isbrae, Greenland; Part III: measurements of ice deformation, temperature and cross-borehole conductivity in boreholes to the bedrock. *Journal of Glaciology*, **48**(162), 369–385 (doi: 10.3189/172756502781831322)
- MacGregor JA and 11 others (2016) A synthesis of the basal thermal state of the Greenland Ice Sheet. *Journal of Geophysical Research: Earth Surface*, **121**(7), 1328–1350 (doi: 10.1002/2015jg003803)
- Maier N, Humphrey N, Harper J and Meierbachtol T (2019) Sliding dominates slow-flowing margin regions, Greenland Ice Sheet. *Science Advances*, **5**(7), eaaw5406 (doi: 10.1126/sciadv.aaw5406)
- Marshall HP, Harper JT, Pfeffer WT and Humphrey NF (2002) Depth-varying constitutive properties observed in an isothermal glacier. *Geophysical Research Letters*, **29**(23), ISSN 1944-8007 (doi: 10.1029/2002gl015412)
- Matsuhisa S and Bird RB (1965) Analytical and numerical solutions for laminar flow of the non-Newtonian Ellis fluid. *AIChE Journal*, **11**(4), 588–595, ISSN 1547-5905 (doi: 10.1002/aic.690110407)
- May T and Bassis J (2025) Estimating glacier ice thickness and yield strength using surface elevation and the perfect-plastic approximation. *Journal of Glaciology*, **71**, e128 (doi: 10.1017/jog.2025.10091)
- Meier MF (1960) Mode of flow of Saskatchewan Glacier, Alberta, Canada. *U. S. Geological Survey*, professional Paper 351
- Mercenier R, Lüthi M and Vieli A (2018) Calving relation for tidewater glaciers based on detailed stress field analysis. *The Cryosphere*, **12**(2), 721–739 (doi: 10.5194/tc-12-721-2018)
- Meyer CR, Bellamy J and Rempel AW (2024) Subtemperate regelation exhibits power-law premelting. *Proceedings of the Royal Society A: Mathematical, Physical and Engineering Sciences*, **480**(2290), 20240032 (doi: 10.1098/rspa.2024.0032)
- Millstein JD, Minchew BM and Pegler SS (2022) Ice viscosity is more sensitive to stress than commonly assumed. *Communications Earth & Environment*, **3**(1), ISSN 2662-4435 (doi: 10.1038/s43247-022-00385-x)
- Morgan V (1991) High-temperature ice creep tests. *Cold Regions Science and Technology*, **19**(3), 295–300 (doi: 10.1016/0165-232x(91)90044-h)
- Navarro F, Macheret YY and Benjumea B (2005) Application of radar and seismic methods for the investigation of temperate glaciers. *J. Appl. Geophys.*, **57**, 193211 (doi: 10.1016/j.jappgeo.2004.11.002)
- Nye JF (1953) The flow law of ice from measurements in glacier tunnels, laboratory experiments and the Jungfraufirn borehole experiment. *Proceedings of the Royal Society of London. Series A. Mathematical and Physical Sciences*, **219**(1139), 477–489, ISSN 2053-9169 (doi: 10.1098/rspa.1953.0161)
- Paterson WSB and Savage JC (1963) Measurements on athabasca glacier relating to the flow law of ice. *Journal of Geophysical Research*, **68**(15), 4537–4543, ISSN 0148-0227 (doi: 10.1029/jz068i015p04537)
- Paul F (2015) Revealing glacier flow and surge dynamics from animated satellite image sequences: examples from the Karakoram. *The Cryosphere*, **9**(6), 2201–2214 (doi: 10.5194/tc-9-2201-2015)
- Pettersson R, Jansson P and Blatter H (2004) Spatial variability in water content at the cold-temperate transition surface of the polythermal Storglaciären, Sweden. *Journal of Geophysical Research*, **109**(F2), F02009 (doi: 10.1029/2003JF000110)
- Ranganathan M and Minchew B (2024) A modified viscous flow law for natural glacier ice: Scaling from laboratories to ice sheets. *Proceedings of the National Academy of Sciences*, **121**(23), e2309788121 (doi: 10.1073/pnas.2309788121)
- Ranganathan M, Minchew B, Meyer CR and Peč M (2021) Recrystallization of ice enhances the creep and vulnerability to fracture of ice shelves. *Earth and Planetary Science Letters*, **576**, 117219 (doi: 10.1016/j.epsl.2021.117219)
- Raymond C (1969) *Flow in a transverse section of Athabasca Glacier, Alberta, Canada*. Ph.D. thesis, California institute of Technology, Pasadena, California

- Reeh N (1982) A plasticity theory approach to the steady-state shape of a three-dimensional ice sheet. *Journal of Glaciology*, **28**(100), 431–455, ISSN 1727-5652 (doi: 10.3189/s0022143000005049)
- Roldán-Blasco JP and 8 others (2025) Creep enhancement and sliding in a temperate, hard-bedded alpine glacier. *The Cryosphere*, **19**(1), 267–282 (doi: 10.5194/tc-19-267-2025)
- Rosenberg CL and Handy MR (2005) Experimental deformation of partially melted granite revisited: implications for the continental crust. *Journal of Metamorphic Geology*, **23**(1), 19–28, ISSN 1525-1314 (doi: 10.1111/j.1525-1314.2005.00555.x)
- Ryser C, Lüthi M, Blindow N, Suckro S, Funk M and Bauder A (2013) Cold ice in the ablation zone: its relation to glacier hydrology and ice water content. *Journal of Geophysical Research*, **118**(F02006), 693–705 (doi: 10.1029/2012JF002526)
- Ryser C and 7 others (2014a) Caterpillar-like ice motion in the ablation zone of the Greenland Ice Sheet. *J. Geophys. Res. Earth Surf.*, **119**, 2258–2271 (doi: 10.1002/2013JF003067)
- Ryser C and 7 others (2014b) Sustained high basal motion of the Greenland Ice Sheet revealed by borehole deformation. *Journal of Glaciology*, **60**(222), 647–660 (doi: 10.3189/2014JoG13J196)
- Savage J and Paterson W (1963) Borehole measurements in the Athabasca Glacier. *Journal of Geophysical Research*, **68**(15), 4521–4536 (doi: 10.1029/JZ068i015p04521)
- Schohn CM, Iverson NR, Zoet LK, Fowler JR and Morgan-Witts N (2025) Linear-viscous flow of temperate ice. *Science*, **387**(6730), 182–185, ISSN 1095-9203 (doi: 10.1126/science.adp7708)
- Smith GD and Morland LW (1981) Viscous relations for the steady creep of polycrystalline ice. *Cold Regions Science and Technology*, **5**, 141–150 (doi: 10.1016/0165-232x(81)90048-3)
- Steinemann S (1958) Experimentelle Untersuchungen zur Plastizität von Eis. Geotechnische Serie Nr. 10, Beiträge zur Geologie der Schweiz., Kommissionsverlag Kümmerli & Frey AG, Geographischer Verlag, Bern
- Thorsteinsson T, Waddington ED, Taylor KC, Alley RB and Blankenship DD (1999) Strain-rate enhancement at Dye 3, Greenland. *Journal of Glaciology*, **45**(150), 338–345 (doi: 10.3189/002214399793377185)
- Treverrow A, Budd WF, Jacka TH and Warner RC (2012) The tertiary creep of polycrystalline ice: Experimental evidence for stress-dependent levels of strain-rate enhancement. *Journal of Glaciology*, **58**(208), 301–314 (doi: 10.3189/2012jog11j149)
- Ultee L and Bassis J (2016) The future is Nye: an extension of the perfect plastic approximation to tidewater glaciers. *Journal of Glaciology*, **62**(236), 1143–1152, ISSN 1727-5652 (doi: 10.1017/jog.2016.108)
- Vallon M, Petit JR and Fabre B (1976) Study of an ice core to the bedrock in the accumulation zone of an alpine glacier. *Journal of Glaciology*, **17**(75), 13–28 (doi: 10.3189/S0022143000030677)
- Vasseur J, Wadsworth FB and Dingwell DB (2023) Shear thinning and brittle failure in crystal-bearing magmas arise from local non-Newtonian effects in the melt. *Earth and Planetary Science Letters*, **603**, 117988 (doi: 10.1016/j.epsl.2023.117988)
- Vialov S (1958) Regularities of glacial shields movement and the theory of plastic viscous flow. *International Association of Hydrological Sciences*
- Wadsworth FB and 7 others (2018) Combined effusive-explosive silicic volcanism straddles the multiphase viscous-to-brittle transition. *Nature Communications*, **9**(1), 4696 (doi: 10.1038/s41467-018-07187-w)
- Webb S and Dingwell D (1990) The onset of non-Newtonian rheology of silicate melts: A fiber elongation study. *Physics and Chemistry of Minerals*, **17**(2), ISSN 1432-2021 (doi: 10.1007/bf00199663)
- Wilson CJ, Peternell M, Piazzolo S and Luzin V (2014) Microstructure and fabric development in ice: Lessons learned from in situ experiments and implications for understanding rock evolution. *Journal of Structural Geology*, **61**(nil), 50–77 (doi: 10.1016/j.jsg.2013.05.006)

APPENDIX

Temperature correction factor

The temperature dependence of Glen's flow law can be described by a correction factor which is calculated according to Smith and Morland (1981) as

$$C = \begin{cases} 0.9316 \cdot \exp(6.4426 \Theta) + 0.0686 \cdot \exp(1.4410 \Theta) & T \geq -7.65^\circ \text{C} \\ 0.7242 \cdot \exp(11.9567 \Theta) + 0.3438 \cdot \exp(2.9494 \Theta) & T < -7.65^\circ \text{C} \end{cases}$$

$$\Theta = T/20 \quad (20)$$

This temperature correction is a continuous and smooth version of a double Arrhenius relation that also agrees with the temperature dependence of Cuffey and Paterson (2010).

Borehole tilt sensor data

Tables 2 and 3 list the data for all borehole sections and tilt sensors with temperate and near-temperate conditions. A database of all tilt sensor data is given in the supporting material.

Viscosity

Figure 7A further investigates the upper limit of viscosity implied by Equation (13). Plotting all data points from our data set, including those from cold ice, shows that the viscosity limit extends to much lower and higher strain rates than the original data set shown in Figure 5. The data from the laboratory experiments (Duval, 1977; Adams and others, 2021; Schohn and others, 2025) also approximately aligns with purple curve. Alternatively, a separate Ellis relation can be drawn which is shown with a blue line.

The same data points are shown in Figure 7B. The red line is a power law with $n = 11$ which also passes through the high-stress values, and is given by

$$\dot{\epsilon} = 5 \cdot 10^7 \sigma^{11}. \quad (21)$$

The blue curve shows the average viscosity determined by Raymond (1969) from boreholes on Athabasca glacier (his Figure 36).

Polynomial rheology

The Polynomial rheology of (Colbeck and Evans, 1973), which was derived from in-situ experiments in a glacier tunnel, also provides reasonable estimates for the ice deformation rates in many boreholes. Their relation from uniaxial tests (their equation 6) is given as

$$\dot{\epsilon} = 0.21\sigma + 0.14\sigma^3 + 0.055\sigma^5, \quad (22)$$

and is transformed to shear stress and shear strain rate. Using this flow law gives reasonable borehole deformations as is shown in Figure 8.

Borehole	Sensor	Slope (°)	Depth (m)	σ_{xz} (MPa)	L_{xz} (a ⁻¹)	P (W m ⁻³)	η (MPa a)	T (° C)	T_m (° C)	f_{SM} (1)
gl.foxx.1	B*	1.55	604.0	0.144	0.1364	0.62	1.06	-0.76	-0.39	0.794
gl.foxx.1	A*	1.55	606.0	0.145	0.1200	0.55	1.21	-0.49	-0.40	0.862
gl.foxx.2	D*	1.55	597.0	0.143	0.0863	0.39	1.65	-0.69	-0.39	0.811
gl.foxx.2	C*	1.55	606.0	0.145	0.1083	0.50	1.34	-0.74	-0.40	0.799
gl.foxx.2	B*	1.55	609.0	0.145	0.1327	0.61	1.10	-0.42	-0.40	0.880
gl.foxx.2	A*	1.55	612.0	0.146	nan	nan	nan	-0.37	-0.40	0.894
gl.gull.1	E*	1.30	676.0	0.135	0.0832	0.36	1.63	-0.60	-0.44	0.834
gl.gull.1	D*	1.30	690.0	0.138	0.0910	0.40	1.52	-0.49	-0.45	0.862
gl.gull.1	C*	1.30	699.0	0.140	0.1477	0.66	0.95	-0.42	-0.46	0.880
gl.gull.1	B*	1.30	702.0	0.141	nan	nan	nan	-0.47	-0.46	0.867
gl.gull.1	A*	1.30	705.0	0.141	0.0971	0.43	1.45	-0.39	-0.46	0.888
gl.gull.2	F*	1.30	667.0	0.134	0.0396	0.17	3.37	-0.83	-0.44	0.778
gl.gull.2	E*	1.30	687.0	0.138	0.0909	0.40	1.51	-0.56	-0.45	0.844
gl.gull.2	D*	1.30	697.0	0.140	nan	nan	nan	-0.54	-0.46	0.849
gl.gull.2	C*	1.30	702.0	0.141	nan	nan	nan	-0.47	-0.46	0.867
gl.gull.2	B*	1.30	705.0	0.141	nan	nan	nan	-0.38	-0.46	0.891
gl.gull.2	A*	1.30	707.0	0.142	0.1016	0.46	1.39	-0.50	-0.46	0.859
gl.duck	15*	1.40	795.5	0.172	1.0142	5.52	0.17	-0.73	-0.52	0.801
gl.duck	14*	1.40	820.6	0.177	1.8250	10.25	0.10	-0.60	-0.54	0.834
gl.duck	13*	1.40	827.6	0.179	1.7337	9.82	0.10	-0.58	-0.54	0.839
gl.duck	12*	1.40	828.5	0.179	1.5547	8.81	0.11	-0.59	-0.54	0.836
gl.duck	11*	1.40	829.4	0.179	1.1550	6.55	0.15	-0.59	-0.54	0.836
gl.duck	23*	1.40	830.0	0.179	1.7990	10.22	0.10	-0.59	-0.54	0.836
gl.duck	22*	1.40	831.9	0.180	1.8250	10.39	0.10	-0.59	-0.54	0.836
gl.duck	21*	1.40	832.8	0.180	2.2000	12.54	0.08	-0.58	-0.54	0.839
gl.store	A3	1.80	592.3	0.164	1.7250	8.99	0.10	-0.00	-0.39	1.000
gl.store	A1	1.80	601.2	0.167	0.1060	0.56	1.57	-0.71	-0.39	0.806
gl.i14N		0.80	628.0	0.077	0.0878	0.22	0.88	-0.44	-0.41	0.875
gl.i14Sa		0.65	655.0	0.066	0.0728	0.15	0.90	-0.63	-0.43	0.826
gl.i14Sa		0.65	665.0	0.067	0.0296	0.06	2.25	-0.38	-0.43	0.891
gl.i14Sa		0.65	675.0	0.068	-0.0345	-0.07	nan	-0.13	-0.44	0.961
ath.209		5.70	140.0	0.123	0.0940	0.37	1.31	0.00	-0.09	1.000
ath.209		5.70	147.0	0.130	0.1080	0.44	1.20	0.00	-0.10	1.000
ath.209		5.70	155.0	0.137	0.1240	0.54	1.10	0.00	-0.10	1.000
ath.209		5.70	162.0	0.143	0.1430	0.65	1.00	0.00	-0.11	1.000
ath.209		5.70	169.0	0.149	0.1770	0.84	0.84	0.00	-0.11	1.000
ath.209		5.70	176.0	0.155	0.1770	0.87	0.88	0.00	-0.11	1.000
ath.209		5.70	187.0	0.165	0.2740	1.43	0.60	0.00	-0.12	1.000
ath.209		5.70	192.0	0.169	0.7410	3.98	0.23	0.00	-0.12	1.000

Table 2. Tilt data for temperate glacier ice. Sensors with inclination time series that were analyzed for this publication are marked by a star. Time series that could not be fitted are indicated by “nan”.

Borehole	Sensor	Slope (°)	Depth (m)	σ_{xz} (MPa)	L_{xz} (a ⁻¹)	P (W m ⁻³)	η (MPa a)	T (° C)	T_m (° C)	f_{SM} (1)
gr_center	D*	2.40	285.9	0.106	0.0628	0.21	1.68	-0.33	-0.19	0.904
gr_center	C*	2.40	320.6	0.119	0.0914	0.34	1.30	-0.32	-0.21	0.907
gr_center	B*	2.40	340.3	0.126	0.1094	0.44	1.15	-0.28	-0.22	0.917
gr_center	A*	2.40	349.6	0.129	0.1416	0.58	0.91	-0.30	-0.23	0.913
gr_hut	F*	3.60	189.3	0.105	0.0802	0.27	1.31	-0.39	-0.12	0.888
gr_hut	E*	3.60	214.0	0.119	0.0879	0.33	1.35	-0.20	-0.14	0.943
gr_hut	D*	3.60	233.8	0.130	0.1260	0.52	1.03	-0.22	-0.15	0.937
gr_hut	C*	3.60	248.5	0.138	0.1600	0.70	0.86	-0.21	-0.16	0.939
gr_hut	B*	3.60	263.0	0.146	0.1240	0.57	1.18	-0.21	-0.17	0.937
gr_hut	A*	3.60	273.0	0.152	0.1060	0.51	1.43	-0.21	-0.18	0.937
gr_lake	F*	2.00	140.6	0.043	0.0308	0.04	1.41	-0.14	-0.09	0.960
gr_lake	E*	2.00	175.2	0.054	nan	nan	nan	-0.18	-0.11	0.945
gr_lake	D*	2.00	199.8	0.062	0.0775	0.15	0.79	-0.17	-0.13	0.948
gr_lake	C*	2.00	224.5	0.069	0.0520	0.11	1.33	-0.18	-0.15	0.945
gr_lake	B*	2.00	239.1	0.074	0.0665	0.16	1.11	-0.18	-0.16	0.947
gr_lake	A*	2.00	248.4	0.077	nan	nan	nan	-0.28	-0.16	0.918
rh_k12	C	9.70	79.0	0.119	0.0860	0.33	1.39	0.00	-0.05	1.000
rh_k12	B	9.70	96.0	0.145	0.1750	0.80	0.83	0.00	-0.06	1.000
ar_bh2	10	4.90	183.4	0.139	0.2150	0.95	0.65	0.00	-0.12	1.000
ar_bh2	9	4.90	193.5	0.146	0.2172	1.01	0.67	0.00	-0.13	1.000
ar_bh2	8	4.90	203.7	0.154	0.3928	1.92	0.39	0.00	-0.13	1.000
ar_bh2	7	4.90	213.9	0.162	0.5655	2.90	0.29	0.00	-0.14	1.000
ar_bh2	6	4.90	218.9	0.166	0.6340	3.33	0.26	0.00	-0.14	1.000
ar_bh2	5	4.90	224.0	0.170	0.6264	3.37	0.27	0.00	-0.15	1.000
ar_bh2	4	4.90	229.0	0.173	0.4151	2.28	0.42	0.00	-0.15	1.000
ar_bh2	3	4.90	231.1	0.175	0.4309	2.39	0.41	0.00	-0.15	1.000
ar_bh2	2	4.90	233.1	0.176	1.0996	6.15	0.16	0.00	-0.15	1.000
ar_bh2	1	4.90	234.2	0.177	1.4497	8.15	0.12	0.00	-0.15	1.000
ua_b11		3.50	64.9	0.035	0.0305	0.03	1.15	0.00	-0.04	1.000
ua_b11		3.50	90.2	0.049	0.0382	0.06	1.28	0.00	-0.06	1.000
ua_b11		3.50	117.7	0.064	0.0355	0.07	1.79	0.00	-0.08	1.000
ua_b11		3.50	144.9	0.078	0.0566	0.14	1.38	0.00	-0.09	1.000
ua_b11		3.50	168.6	0.091	0.0552	0.16	1.65	0.00	-0.11	1.000
ua_b11		3.50	187.5	0.101	0.1163	0.37	0.87	0.00	-0.12	1.000
ua_b11		3.50	203.1	0.110	0.1179	0.41	0.93	0.00	-0.13	1.000
ua_C96	A*	4.15	260.0	0.167	0.0705	0.37	2.36	0.00	-0.17	1.000
ua_C96	B*	4.15	235.0	0.151	0.0828	0.40	1.82	0.00	-0.15	1.000

Table 3. Tilt data for temperate glacier ice.

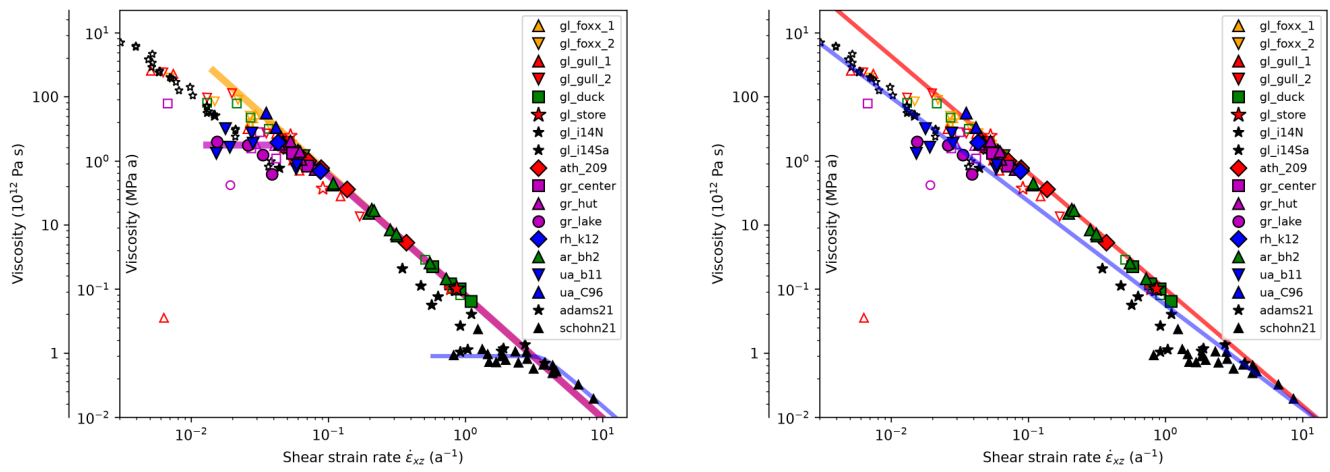


Fig. 7. Viscosity plot for all measurements. Sensors in cold ice are shown by empty symbols. In panel A an additional Ellis curve (blue) is shown for the laboratory measurements (black symbols). In panel B a power law with $n = 11$ is drawn in red, and the average viscosity from Raymond (1969) is shown in blue.

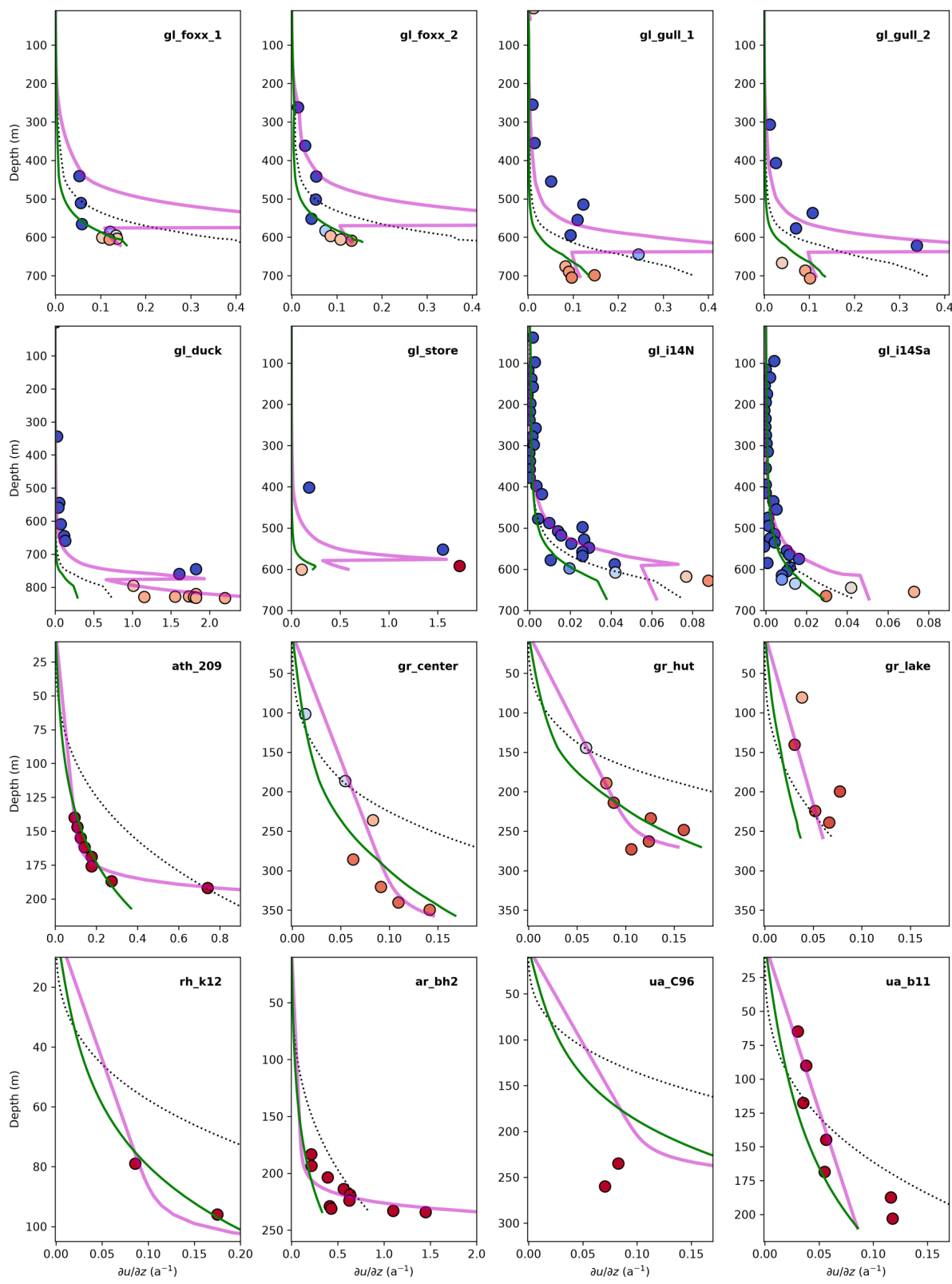


Fig. 8. Same as Figure 1 but with borehole deformations calculated from Equation (22) shown in green (Colbeck and Evans, 1973).