DEFORMATION STRUCTURES IN COLD BASAL ICE: 
INSIGHTS INTO SUBGLACIAL GLACIOTECTONIC PROCESSES

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ABSTRACT
A progressive simple shear regime in a cold-based glacier in the McMurdo dry valleys has produced three shear domains and several sliding interfaces within a 4 m-thick basal ice sequence. Although the deformation structures produced in each of the strain domains form distinctive associations, it appears that the principal control on the style of deformation is material strength. Simultaneous development of ductile and brittle deformation structures in adjacent strain domains suggests that heterogeneous deformation may produce complex structural associations in glacial deposits. These conclusions have important implications for studies that seek to reconstruct strain behaviour from subglacial deformation structures because co-existence of ductile and brittle deformation may be the result of a single deformation episode.

INTRODUCTION
The last two decades have seen a paradigm shift in glaciology and glacial geology as research on the deformation of sediment beneath glaciers has resulted in new views of the behaviour of glaciers (Boulton, 1986; Murray, 1997; Alley, 2000). These new views suggest that subglacial sediment deforms at least through part of its thickness thereby influencing large scale glacier behaviour as well as erosion, transportation and deposition processes. These processes have important implications for Quaternary geologists because during glaciations ice sheets in North America and Europe advanced over large areas of unconsolidated sediment.

Motivated by a need to understand and reconstruct the behaviour of glacier beds, a wealth of new information on structural interpretation of glacial landforms and sediments has been produced. Much of the work has focussed on meso- and micromorphological analysis structures in Quaternary sediments (e.g. Menzies, 2000; Van der Meer, 1986). However, despite the wealth of information, the significance and relevance of individual structures often remains vague and indeterminate (Menzies, 2000). Consequently Van der Wateren et al. (2000) have argued that it is critical to establish an unambiguous set of criteria that can be used to identify subglacial deformation. The approach outlined

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by Van der Wateren et al. (2000) involves understanding the kinematics of subglacial deformation, analysing structures that are left behind by former glaciers and evaluating the conditions which prevailed within the bed during deformation. Unfortunately the necessary first step of understanding the kinematics of subglacial deformation is problematic because the beds of glaciers are very difficult to access. The lack of information of subglacial kinematics has meant that most geomorphologists have focused on analysing structures in glacial deposits and inferring stress regimes beneath the glacier during deformation. Consequently, interpretations often lack a strong physical basis.

An alternative approach to this problem is to undertake direct studies of the kinematics and structures that have developed at the glacier-bed interface. This paper describes such an approach using direct measurement and observation of subglacial deformation processes and structures beneath a small cold-based glacier in Antarctica (Fig. 1).

![Figure 1: Location map of Suess Glacier.](image)

The objectives of the paper are to explore and understand the relationships between strain intensity, deformation processes and the formation of deformation structures. To do this we attempt to answer two questions:

1. Are distinctive assemblage of deformation structures associated with different shear regimes?
2. Are deformation structures likely to be a reliable way of reconstructing the stress field beneath a glacier?

METHODS

A 2 m × 1 m × 25 m tunnel (Fig. 2a) was excavated at the bed of Suess Glacier, a small glacier in the Taylor Valley, Antarctica with a basal temperature of −17 °C. At the end of the tunnel a 4.5 m high shaft was cut and instrumented with linear variable displacement...
transducers, plumb lines and thermocouples. The LVDTs and thermocouples were left in place for a year, the plumb lines were resurveyed annually for 4 years and the deformation of the tunnel was surveyed annually. Mechanical tests on the strength of basal ice and substrate were made in the tunnel and laboratory using a direct shear device (Fitzsimons et al., 1999, 2000). Deformation structures within the basal ice and at the apparent ice-substrate contact were made during the excavation and the entire suite of structures was mapped after the excavation was completed.

The terminology used in the analysis of deformation structures is fraught with difficulty because researchers frequently use different terms to describe the same feature or process (Maltman, 1994). In this paper we follow the terminology of Van der Wateren et al. (2000). The term ductile is used to denote permanent deformation without fracturing and rheology is used in a wide sense as pertaining to the mechanical behaviour of materials.

STRAIN DISTRIBUTION IN THE BASAL LAYER

Figure 3 shows a velocity profile through 4 m of basal ice together with a column that summarises the structure of the basal ice. Three patterns can be identified in the velocity profile. In the lower part of the profile (0–2 m) there is a near linear increase in velocity, between 2 m and 2.8 m there is minimal change in velocity and above 3 m there is a logarithmic increase. Comparison of the velocity profile with the basal ice stratigraphic column shows that the velocity variations can be directly associated with changes in the physical characteristics of the ice and the occurrence of debris-rich ice. The near-linear increase in velocity occurs in the basal stratified facies, the area of minimal increase in velocity occurs within the basal solid facies and the section of logarithmic increase in velocity occurs within the amber facies.

The distribution of strain shown by the velocity profile through the entire basal ice layer shows that strain is heterogeneous and that there is a strong partitioning into three recognisable domains. These domains together with the deformation structures associated with each are summarised below.
THE HIGH SHEAR DOMAIN
The high shear domain is confined to a layer of “amber” ice (sensu Holdsworth, 1974) that has a characteristic amber colour, low debris concentration and high solute concentration compared with the overlying clean glacier ice that has a meteoric origin. The amber ice layer is generally about 1 m thick although it varies from 1.1 m at the end of the tunnel to 0.2 m at the glacier margin. There are few visible deformation structures in the amber ice either in the field or in thin sections. Larger particles within the amber ice frequently have air-filled cavities on the upstream and downstream sides. The cavities are absent in particles less that about 20 mm in diameter. However, both the large and small particles displayed well-developed pressure shadows on their lee sides. The pressure shadows and cavities are wake effects produced by sheltered zones forming in the stoss and lee of strong objects held within a more deformable matrix. Some of these structures resemble sigma and delta structures observed in mylonites where they are associated with low and high shear strains respectively (cf. Passchier and Simpson, 1986).

At the boundary between the rapidly shearing amber ice and the underlying frozen sand and gravel numerous air-filled cavities are present, both where the frozen layer protruded into the amber ice and where the contact was relatively flat. This boundary is where dial gauges recorded movement from 0.93 to 5.65 mm.a$^{-1}$. These measurements were made using a 5 mm peg drilled and frozen into ice immediately above the layer and with the dial gauge bolted to the frozen sand. The measurements can be interpreted as sliding velocities or zones of high shear concentrated in a very thin layer. However, the co-incidence of the high strain rates with the presence of slickensides on the cavity roofs (Fig. 2b) clearly indicates the sliding has occurred at the interface. The slickensides have formed as ice sliding over the frozen sediment layer has moulded itself around the roughness elements on top of the layer and produce an imprint of the form roughness in the cavity. Several of the cavities rapidly degassed when they were punctured as the tunnel was excavated indicating that at least some of the cavities were above atmospheric pressure.

LOW SHEAR DOMAIN
The low shear domain consists of the layer of frozen sand and gravel that is apparently the product of erosion and entrainment of part of the glacier substrate (Fitzsimons et al., 1999). Samples taken from the frozen sand and gravel indicate that the debris concentrations range from 50 to 75% indicating that the sediment is very close to saturation, i.e. the only ice in the sediment occupies the pore spaces. The velocity profile through the basal ice shows that there is a very steep segment associated with the layer between 2.3 and 2.8 m above the bed (Fig. 3) suggesting that there is minimal deformation within the layer (Table 1). All the deformation structures observed within this domain are brittle.

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<th>Table 1: Basal ice characteristics.</th>
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<td>Strain gradient (mm.a$^{-1}$.mm$^{-1}$)</td>
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<td>-------------------------------------</td>
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<tr>
<td>High shear domain 0.159 0.9 0.01</td>
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<td>Moderate shear domain 0.03 1.28 17.81</td>
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<td>Low shear domain 0.003 2.53 55.5</td>
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The layer is characterised by numerous cracks and fissures (Fig. 2c) that are filled with a
mixture of air, and dry sand and gravel that is most likely to be the product of sublimation of the ice matrix adjacent to the crack or fissure. Most of the cracks and fissures are oriented transverse to the glacier flow direction (Fig. 4) and are near-vertical.

Figure 4: Orientation of tension cracks in frozen sand 18 m into the Suess Glacier tunnel.

They are interpreted as tension cracks formed when stresses exceeded the tensional strength of the materials. At the boundary with overlying ice the larger fissures were often partially occupied by ice intruding into the fissure.

MODERATE SHEAR DOMAIN

The moderate shear domain, located between 0 and 2.3 m above the glacier bed (Fig. 3), consists of a stratified ice-sediment mixture with debris concentrations ranging from 1 to 60%. The velocity gradient through this domain is $0.03 \text{ mm.a}^{-1}.\text{mm}^{-1}$ (Table 1) which is intermediate between the high and low shear domains. Within the domain there are several discontinuities that appear to be sliding interfaces or very thin zones of high shear (Fig. 2c).

A wide range of brittle and ductile deformation structures are associated with this domain including folds, riedel shears, cavities, slickensides and boudinage. Numerous brittle deformation structures were associated with layers and lenses of frozen sand. The most common brittle structures were faults and tension cracks associated with boudinage structures (Fig. 2f). Early-development boudinage consists of a series of small tension cracks in the more competent sediment layers. These cracks develop into small cavities develop which are eventually filled as the surrounding basal ice into the cavity (Fig. 2f). This form of boundinage has been described as competence-contrast boudinage by Hambrey
and Milnes (1975) because it results from a rheological contrast between the readily deformable glacier ice and more competent frozen sediment that has more resistance to shear (Table 1).

A wide range of fold types and sizes occur in this domain. The largest fold observed occurred 21 m into the tunnel. It was a recumbent fold with limbs about 3 m long and the hinge was oriented parallel to the ice-flow direction, i.e. it rested transverse to flow (Fig. 2d). The hinge of the fold is characterised by polyharmonic folds that form small, sheared parasitic folds (Fig. 2d). Geometrical analysis of such features suggests that the parasitic folds are probably formed by layer-shortening and folding of the stratified ice and sediment mixture, after which the initial symmetrical forms are modified into asymmetric forms by subsequent deformation (Ramsay and Huber, 1987). Several large granite boulders up to 1.5 m in diameter were encountered in the stratified basal ice. All of these boulders have stoss- and lee-side air-filled cavities that have formed as the more ductile sediment-bearing ice has flowed around them. Like the cavities found in the high shear domain described above, the roofs of the cavities contain well-developed slickensides. Cavities were also found at structural discontinuities where ice with a pervasive near-vertical foliation was overlain by ice with a near-horizontal foliation (Fig. 2e). These cavities were considerably smaller than those associated with boulders and the frozen sand and gravel layer. In other locations in the moderate shear domain there are numerous detached folds and some evidence for the formation of reidel shears and low-angle thrusts, both oblique to the bed.

The end of the tunnel revealed an apparent bed that consists of frozen sand and fine gravel. In two locations the thickness of the “bed” is at least 600 mm but, in others, ice layers and lenses were observed within this apparent bed material. Although more effort that was expended on defining the bed, identification of a simple ice-frozen sediment bed have proven elusive. The apparent contact between the stratified ice and the bed was instrumented with several LVDTs and no sliding was detected. In addition a segmented rod inserted into a 400 mm-deep hole drilled in the substrate recorded no detectable deformation when it was excavated after 1 year. Numerous air-filled cavities were also observed at this apparent bed.

**DISCUSSION AND CONCLUSIONS**

The basal velocity profile, together with tunnel deformation measurements in Suess Glacier, demonstrates that deformation in the lower 4.5 m of the glacier is characterised by progressive simple shear and that strain is heterogeneous. Four recognisable strain domains in the basal ice are associated with different assemblages of deformation structures. A superficial examination of the assemblages suggests that there is a systematic relationship between strain rate and the type of deformation (Fig. 3). However, the underlying control of the structural assemblage appears to be the strength of the ice and sediment mixture (Table 1). Material with a low viscosity (low shear strength) supports the highest strain rates and the material with the highest viscosity (high shear strength) supports the lowest strain rates (Table 1).

Structures observed in the basal zone demonstrate that brittle and ductile deformation structures have developed adjacent to each other simultaneously. It is likely therefore
that such structures are likely to co-exist in the geological record, as suggested by Van
der Wateren et al. (2000). Boulton and Hart (1991) also argued that there is a change
in deformation style and resultant structures as shear strain increases, and that zones of
deformation within sediment are related to similar zones of strain within ice sheets. How-
ever, they suggested that complex deformati
onal sequences are produced by superimpos-
that complex deformational sequences are produced by superimposition of these differing styles upon one another as the ice sheet advances and retreats.
The evidence from Suess Glacier described above demonstrates that a complex suite of
structures may be produced by a single ice advance if the deformation is heterogeneous.

Studies that seek to reconstruct deformation regimes from deformation structures are
hampered by having a very limited number of field observations on deforming sediments
beneath glaciers. Consequently our thinking about subglacial deformation is framed by
idealised velocity profiles (e.g. Fig. 1 in Boulton and Hart, 1991; Fig. 1 in Boulton, 1996
and Figs. 2 and 3 in Murray, 1997). The problem with using conceptual models built
on idealised velocity profiles is that they promote a simplistic view of the ice-sediment
interface as a clean boundary between ice and the glacier substrate. In Suess Glacier, the
composition and structure of the basal ice, together with the compound velocity profile,
demonstrate that the glacier bed is not a simple ice-substrate contact. In fact, our investig-
ation of the basal zone suggests that where a glacier rests on unconsolidated sediments
the glacier “bed” is a problematic concept because the more closely it is examined the
more difficult it becomes to define. The field observations support the idea of a trans-
ition zone in which strain is partially transmitted into the bed and partly accommodated
by strain within basal ice. The presence of sliding interfaces and narrow zones of high
shear mean that the glacier is partly uncoupled from its substrate - a condition not usually
associated with cold-based glaciers, although Echelmeyer and Wang (1987) recorded a
similar situation in a glacier bed at −4°C.

These considerations raise the question as to what extent are deformation structures likely
to be a reliable way of reconstructing the stress regime experienced by deposits that have
experienced glaciotectonic deformation? The strain patterns and associated structures we
have observed in Suess Glacier lead us to the conclusion that a great deal more investiga-
tion of deforming subglacial material needs to be undertaken, before deformation regimes
can be reliably reconstructed from structures in sediments. Although there are well de-
veloped classifications of meso- and micro-structures in glacial sediments (Van der Meer,
1996; Van der Wateren, 1994; Menzies, 2000) there are likely to be continuing problems
with this approach because of uncertainty concerning their origin and whether different
processes can result in similar features.

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