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Direct shear tests of materials from a cold glacier: implications for landform development

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Abstract

The mechanical properties of materials at the ice-bed interface exert a major control on glacier behaviour and define the nature and rate of subglacial erosion. This paper presents new data on the strength and behaviour of basal ice and substrate of the Suess Glacier, south Victoria Land, Antarctica which is a small alpine dry-based glacier that has a basal temperature of $-17^{\circ}C$. A tunnel excavated in the glacier revealed a substrate composed of frozen sand and gravel and a basal zone that was 3.8-m-thick. From bottom to top, the basal zone was composed of 1.8 m of stratified, complexly deformed layers of ice and debris-laden ice that was overlain by a 0.9-m-thick layer of frozen sediment and a 0.8-m-thick layer of discoloured ice that lies immediately beneath clean glacier ice. Direct shear tests were performed in the field on 36 samples using a strain rate of 0.85 mm/h and on replicate samples in a laboratory at strain rates of 0.08 and 0.01 mm/h.

The experiments performed in the field show that the average peak shear strength of substrate samples was 2.53 MPa, which is almost twice as strong as the average value for basal ice (1.28 MPa), and the glacier ice samples (1.39 MPa). The laboratory experiments show that the behaviour and peak strength of samples sheared at the higher strain rates are considerably greater than the peak and residual strengths measured during the lower strain rate tests. The direct shear tests suggest that the glacier substrate is unlikely to deform at the current temperature.

1. Introduction

Many geomorphological studies attempt to reconstruct thermal conditions at the beds of Pleistocene ice sheets using form-process inferences. Often, these studies regard dry-based glaciers as incapable or largely incapable of erosion. For example, Kleman and Borgström (1994) have argued that patches of periglacial surfaces adjacent to areas of subglacially sculpted landforms are indicative of patches of continuous frozen bed conditions. A later study by Kleman and Borgström (1996) attempted the reconstruction of parts of the Scandinavian ice sheet from geomorphological evidence. This study defined three geomorphological systems: a dry bed system in which the frozen bed condition leads to a hiatus in landform development without the formation of new landforms, a wet bed system in which basal sliding occurs and flow parallel lineations are continuously produced, destroyed or reoriented and a marginal meltwater system that consists of a spatially coherent system of linear features produced during deglaciation. The principal assumption of the model for reconstructing dry bed conditions are that basal sliding can only occur in wet bed conditions and that there is no net change to subglacial landforms if frozen bed conditions are present.

It is widely accepted that sliding is minimal in frozen beds because the bond strength between the ice and the substrate is very strong (Holdsworth and Bull, 1970) and liquid water is necessary for sliding to occur (Murray, 1997). However, theoretical work by Shreeve (1984), Fowler (1986) has concluded that sliding and regelation is possible at subfreezing temperatures. Subsequently, a study of the Urumqi Glacier No. 1 in China concluded that 60–80% of glacier motion was accomplished by deformation of a layer of ice-laden sediment several metres thick despite the fact that the temperature was around $-4^{\circ}C$ (Echelmeyer and Zhongxiang, 1987). More recently, Cuffey et al. (1999) have argued that

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thin films of liquid which facilitate sliding can persist at temperatures as low as $-17°C$. Echelmeyer and Zhongxiang's work is important becauses it suggests that it is possible to have significant bed coupling and deformation at temperatures below freezing and that rheology of frozen debris is more complex than previously realised. One approach to this problem is to examine the mechanical properties of the basal ice and the substrate in order to identify the likely conditions of deformation and entrainment. There have been very few laboratory experiments on subglacial materials and they have yielded contradictory results (Iverson et al., 1997). Kamb (1991) undertook a series of direct shear tests on sediment from beneath Ice Stream B which suggested that the material behaved in a plastic manner consistent with fine-grained sediment. However, Iverson et al. (1997) argued that direct shear tests may not be appropriate because the walls of the test cell constrain deformation to a thin zone and because the relatively small strains achievable by direct shear devices may not be great enough to achieve the residual strength, a criticism that can also be directed toward triaxial and simple-shear tests (Atkinson, 1993; Iverson et al., 1997). In contrast, Ho et al. (1996) have argued that triaxial and simple shear tests can be appropriately applied to fine-grained tills. Iverson et al. argue that the problem of low strains associated with direct shear, triaxial and simple shear tests can be overcome by using a rotary device. The device they used was sufficiently large to accommodate very large clasts and had transparent walls to allow the distribution of strain to be observed directly during a test. In the experiments described in this paper, we have opted to employ direct shear tests because we required a simple, robust, portable apparatus that we could use in the field and laboratory at temperatures around $-17°C$.

The objective of the research programme of which this paper is a part is to examine the role of cold-based glaciers in landscape development. In this paper, we do this by examining ice marginal landforms, reviewing the nature of debris entrained in basal ice and examining a series of direct shear tests with different strain rates to consider under what conditions the glacier bed is likely to deform. The paper describes how the direct shear tests were conducted, summarises the test results, presents an interpretation of the data and examines the implications of the data for landscape development.

2. Methods

Suess Glacier is a 5-km-long glacier that descends from 1350 m on the Asgard Range to about 50 m on the floor of the Taylor Valley. The moraine at the right margin of Suess Glacier was surveyed using an automatic level and cutting trenches to expose the sedimentary and glaciotectonic structures. A $2\,m \times 1\,m \times 25\,m$ tunnel was excavated at the bed of the glacier in 1996 using chainsaws to cut the ice and a demolition hammer to break-up the debris-bearing ice. The tunnel was extended in 1997 and a vertical shaft of 4.5 m high was cut at the end of the tunnel to expose the entire debris zone. At the back of the tunnel, thermocouples and alcohol thermometers showed that the basal ice was $-17°C$. Structures in the basal ice were mapped and samples were cut to determine debris concentrations and solute chemistry.

A series of direct shear tests were undertaken in the tunnel using a modified laboratory direct shear device (Fig. 1). The shear box was constructed on stainless steel plates that were drilled to accomodate a cylindrical sample with a 59 mm diameter. The device was capable of displacing the sample by about 5 mm which is a strain of approximately 8%.

Samples were cut from basal ice and the substrate using a diamond corer driven by an electric drill. The field experiments were conducted in ambient temperatures of $-14°C$ and displacement rates of 0.85 mm/h (7 m/a) were used in order to test a number of samples and gain an appreciation of the variability of peak

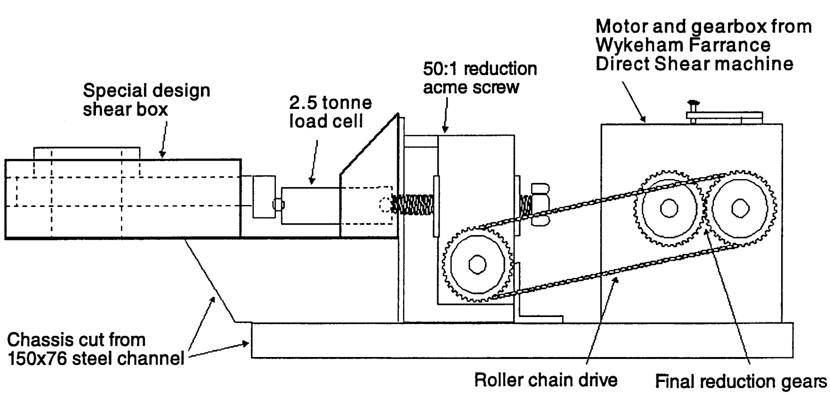

Fig. 1. Modified direct shear device used in field and laboratory experiments. In the field the normal load was applied using a pneumatic shock absorber and load cell and in the laboratory steel weights were hung from the sample.

strengths of the different materials. This displacement rate is more than an order of magnitude greater than ice velocities measured in the basal zone where the maximum velocity was 0.25 m/a 3.5 m above the bed. Because the displacement rates were considerably greater than the velocities experienced in the basal zone, a set of samples were also tested in a laboratory under the more controlled conditions of $-17°C$ using displacement rates of 0.08 and 0.01 mm/h which necessitated single experiments that ran for up to 21 days.

3. Marginal landforms and basal ice characteristics

Excavations in the moraine at the margin of Suess Glacier show that it consists of blocks of sand and gravel resting unconformably against each other (Fig. 2). Within the individual block, some sedimentary structures including planar bedding and cross bedding are well preserved.

An exposure of basal ice at the end of the tunnel shows a sequence of clean englacial ice resting on amber ice, a layer of frozen debris, a complex sequence of stratified ice and a frozen sand and gravel bed. Table 2 presents a summary of the main facies types, their debris concentration and the solute chemistry. Fitzsimons et al. (1999) described the basal ice characteristics and concluded that the blocks of frozen sediment have been eroded from the bed in many cases without disaggregation (Fig. 3a) and that many of the structures observable at the glacier margin are similar to subglacial structures. There are two distinct styles of deformation that are observable in the basal ice. Ductile structures are the most common and associated with basal ice with relatively low debris concentrations (Fig. 3b). Broken blocks of frozen sediment, often with gas-filled cavities, provide evidence that brittle deformation is widespread in materials with high debris concentrations (Figs. 3c and d).

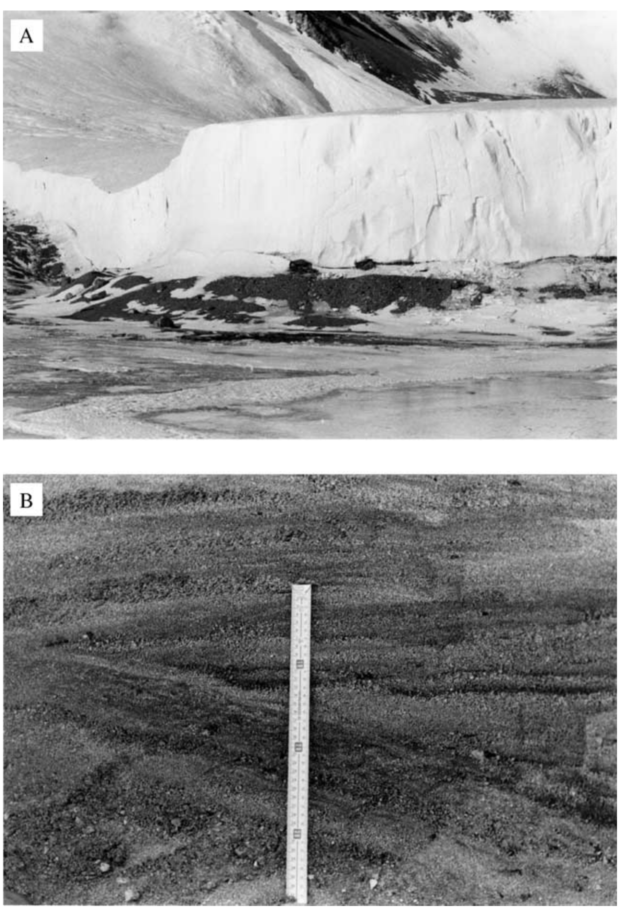

Fig. 2. Moraine at the margin of the Suess Glacier (A); exposure of the moraine showing the unconformable contact between two blocks of stratified sediment (B).

4. Direct shear tests

Tables 2 and 3 summarise the peak shear strength values for englacial, amber, stratified basal ice and the substrate for tests with three different displacement rates and Fig. 4 shows representative stress–strain curves of the different ice types. For each of the three ice materials, a steady state condition was reached by the end of each test with a constant shear stress being maintained at a constant rate of strain. For the englacial ice and basal ice samples tested at a high strain rate, the shear stress increased initially to a peak value of 1.39 and 1.28 MPa before falling back to a steady state value of 1.1 MPa. For the amber ice tested at a high strain rate, the shear stress increased steadily to reach the steady state value of 0.9 MPa. All of the ice samples reached a steady state condition without an initial peak when tested at the medium and low strain rates. Table 2 gives that the peak strength of englacial ice, amber ice and basal ice decreases as the strain rate decreases. Although there are only two stress–strain curves for the englacial ice, the strong peaks observed in the high strain rate tests is not present in experiments conducted at the low strain rate which suggests the sample has achieved its residual strength.

Fig. 4 shows that the amber ice attained a steady shear strength in all the tests but there is a significant and progressive decrease in the residual strength as the strain rate was lowered. The basal ice samples (basal diffused, basal laminated and basal clear facies, Table 1) show that the strong peak strength observed in the high strain rate tests is not present in the lower strain rate tests. In addition, the samples tested at intermediate (0.66 m/a) and low rates (0.01 m/a) are virtually indistinguishable because they achieve nearly identical residual strengths.

Although the substrate samples tested with the low and medium strain rates appear to have lower peak

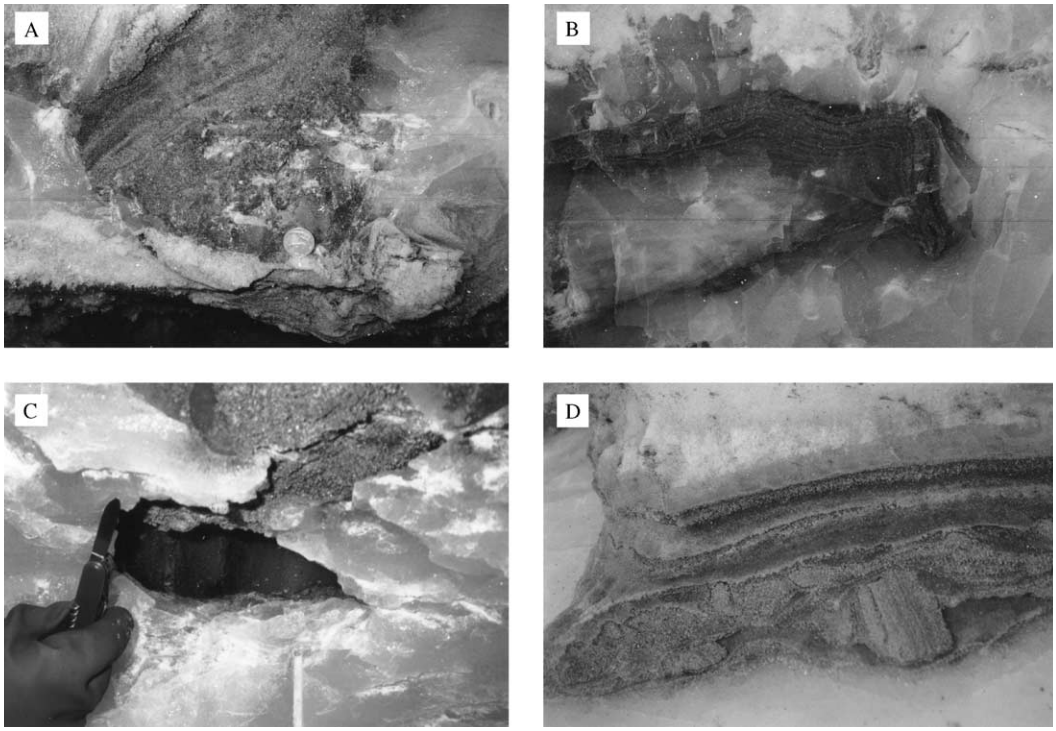

Fig. 3. Structures in basal ice: (A) sedimentary structures in the solid facies resting 2.6 m above the bed; (B) ductile structures in laminated basal ice; (C) broken pieces of solid facies ice with a cavity showing slickenslides at the roof of the cavity; (D) fragments of frozen sediment in basal ice.

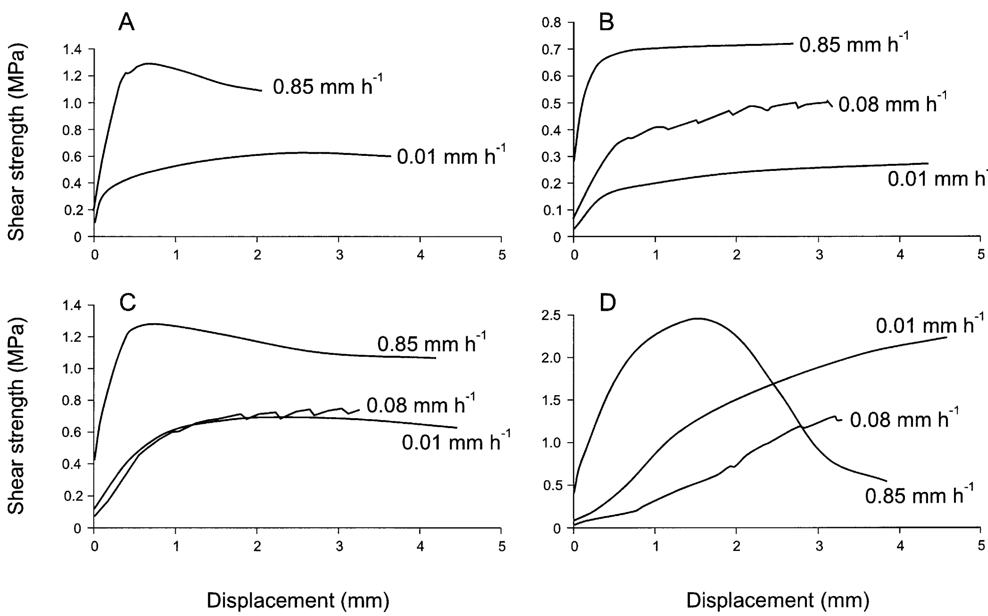

Fig. 4. Stress–strain curves for four different ice types and three displacement rates: (A) englacial ice; (B) amber ice; (C) basal stratificed ice; (D) substrate.

strengths (Table 2), the stress–strain curves show that the samples did not reach either a peak or steady state stress condition by the end of the tests. The test conducted at the low strain rate (0.1 m/a) reached a maximum shear stress of 2.16 MPa, which is close to the peak strength obtained in the high strain rate test (2.53 MPa), and was still increasing slowly at the end of the test. The test conducted at the medium strain rate (0.66 m/a) reached a maximum shear stress of only 1.26 MPa. Although this test was terminated

Table 1
Debris concentrations and total dissolved solids in the ice facies

Ice facies	Debris content (%)	Dissolved solids[a] (ppm)
Englacial diffused	0.06	3.79
Basal amber	0.87	6.23
Basal solid	96.00	
Basal diffused[b]	0.30	6.11
Basal laminated[b]		
Debris-bearing	73.99	14.05
Clean	13.26	8.33
Basal clear[b]	3.03	2.28

[a] Sodium, potassium, calcium, magnesium, and chlorides.
[b] Collectively called the basal stratified facies in the text.

Table 2
Peak shear strength values for four different types of ice

Material	Peak strength		
	Fast 7 m/a	Medium 0.66 m/a	Slow 0.1 m/a
Englacial	1.39	0.60	0.64
Amber	0.90	0.50	0.28
Stratified	1.28	0.76	0.71
Substrate	2.53	1.26	2.16

Table 3
Comparison of basal conditions in Meserve and Suess Glaciers

Meserve Glacier	Suess Glacier
Boulder/rock bed	Unconsolidated bed
Low debris concentrations	Low to very high debris concentrations
Structurally simple	Structurally complex
Basal ice ca. 1 m thick	Basal ice up to 3.5 m thick
No basal solid facies	Thick slabs and blocks of solid facies

prematurely, the shear stress was increasing at a lesser rate than for the low strain rate test. Most of the substrate samples that were tested at high strain rates were fractured along the shear plane when they were removed from the shear device. It seems likely that the strong peak on the high displacement tests records fracture of the substrate samples in the test cell.

5. Interpretation

However, several issues are raised by the descriptions and data reported above the issue that we focus on here is whether the tests provide an insight into bed deformation and entrainment processes. The results for the direct shear tests on the three different ice materials are consistent with the known general behaviour of ice, with the shear stress required to cause shearing decreasing with the rate of strain. The behaviour observed was ductile deformation.

The results for the substrate material are more difficult to understand. The stress–strain curve for the high strain rate test shown in Fig. 4 may be described as brittle, with a high peak strength followed by a rapid decrease in shear stress. Many of the samples were found to have fractured along the plane of shearing when they were removed from the apparatus. Others remained intact but may have fractured and subsequently healed. This style of deformation in the substrate samples is consistent with observations made in the tunnel which show that brittle deformation structures were always associated with frozen debris where the ice was confined to the pores in the sediment (e.g. Fig. 3c). The shape of the curves for the tests at both medium and low strain rates on the other hand may be described as ductile, although none of these tests had reached a steady state condition by the end of testing. It is possible that the medium and low strain rate tests may have exhibited brittle failure if strained far enough. Most likely, however, they would have continued to deform in a ductile manner eventually reaching a steady state shear stress, possibly the same value as the peak strength reached for the high strain rate test.

The result for the substrate material tested at a medium strain rate appears to be anomalous. Instead of being intermediate between the high strain rate and low strain rate test results it falls well below both. This result is difficult to explain but is probably due to variable particle size, packing and porosity in the material.

The results indicate that over a wide range of strain rates, the shear stress required to strain the substrate material at any given rate is substantially higher (by a factor of 2 or more) than for any of the ice samples. At the high strain rate (7 m/a) the substrate material exhibited brittle behaviour while the ice materials all remained ductile. At low strain rates, the substrate behaviour was not determined although it remained ductile within the range of deformation able to be applied. Further tests with greater total strains need to be performed on the substrate material to determine the steady state behaviour of the material.

A straightforward interpretation of the data is that the strength of the substrate is much greater than the ice that lies above it so it could be expected that under normal circumstances the bed will not deform significantly. However, the results necessitate consideration of the controls on the behaviour of mixtures of ice and debris. While the strength of ice is controlled by its visco-plastic behaviour, strength is also partly determined by the presence of solutes and solid detrital particles that are important constituents of basal ice. Hooke et al. (1972) provided some insights into the nature of deforming ice and debris mixtures with a series of unconfined compressive strength tests conducted at temperatures between −7.4°C and −9.4°C on ice with fine sand concentrations between 0% and 35% volume. Although the results at low debris concentrations are inconclusive, the creep rate of ice decreased exponentially as the debris concentration increased above 10%. Hooke et al. suggested that their observations were consistent with theories of dispersion hardening in metals where tangles of dislocations developing from or to particles impede primary glide dislocations. However, as the volume fraction of sand increases above 35% grain to grain contacts become an increasingly important component of strength. Goughnour and Andersland (1968) examined the problem from the slightly different perspective of the behaviour of frozen sand as the ice content increases above saturation. They found that as the volume of ice increases above saturation, individual grains become progressively detached from each other until a critical point is reached when almost all the particles are out of contact with each other and the strength of the material is almost entirely dependent on the strength of the ice. This point was called the critical ice content by Goughnour and Andersland (1968) who argued that it is characterised by a distinct break in slope in the curve of peak strength and ice content. Nickling and Bennett (1984) contributed to the problem of the behaviour of frozen materials by undertaking a series of shear box experiments that examined the relationship between shear strength, ice content and normal load. They found that peak shear strength appears to rise dramatically from 0% to 25% ice content (saturation) and decrease as ice content increases (Fig. 5). They argued that the initial sharp increase in strength is the result of increased cohesion produced by pore ice which cements particles together. Maximum peak strength is the point of saturation where strength is cohesion plus the product of internal friction and normal stress. As the ice volume increases above saturation particles that were in contact become separated thereby decreasing particle friction and strength becomes increasingly dependent on ice strength.

The results of the direct shear tests are generally consistent with previous studies because the strength of the frozen sand substrate is greater than basal and englacial ice. The low shear strength of the amber ice facies is also consistent with the findings of Jones and Glen (1969) and Nakamura and Jones (1970) who concluded that ice doped with some chemicals tends to soften and therefore increase creep rates. However, the stratified facies samples tested at the high strain rates are weaker than the clean englacial ice samples which contrasts with the results of Goughnour and Andersland (1968) and Hooke et al. (1972). At lower strain rates, the stratified ice samples are stronger than clean ice samples which suggests the relationship between debris concentration and strength is not linear. Such a conclusion has

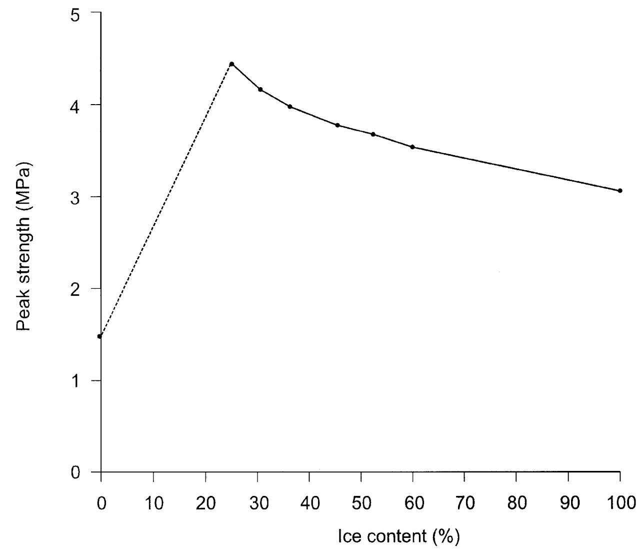

Fig. 5. Variation of peak shear strength with ice content. Tests done with a normal load of about 1 MPa based on data from Nickling and Bennett (1984).

been reached by Lawson (1996) who conducted a series of unconfined compressive strength tests that showed that between −5°C and 0°C ice that contained 5–20% debris was weaker than glacier ice but that at temperatures lower than −5°C, the debris-bearing ice was stronger than glacier ice. Although it appears that debris-bearing ice may be weaker than clean ice, variations in solute content may be equally important as the behaviour of the amber ice suggests. In the case of Suess Glacier, samples of debris-bearing basal ice have dissolved solid concentrations greater than the amber ice (Table 1). However, debris and solute concentrations are inter-related because particles are important sites for the attachment of ions which tend to disorb when ice is melted. Consequently, interpretation of the results of solute assays is difficult because of the uncertainty over whether the values are the product of disorbtion or solutes carried within the ice. For the reasons outlined above, it is very difficult to distinguish whether strength is primarily related to debris concentration or solute concentration.

One of the limitations of the direct shear tests is that they are conducted on very thin shear planes. This raises a problem because all the materials we have tested are anisotropic. Clean glacier ice consists of ice layers with different bubble concentrations, the stratified facies contains layers of variable debris concentrations and bubble concentration, and the substrate contains sedimentary structures that record variations in particle size and materials (e.g. fine algae layers and variations in ice concentrations). The anisotropy of materials is an important consideration because strain will be concentrated in materials that are more readily deformable. Two suggested possible sources of variation in strength of material from Suess Glacier are the presence of cavities and ice lenses/layers and thin mud layers (Fitzsimons et al., 1999). The effect of localised sublimation on the strength of frozen debris is described by the left of Fig. 5 where the peak strength reduces as the amount of ice in the sediment decreases below saturation. For this to occur, the cavity must develop within the substrate or at the ice–substrate contact. Dried sand accumulations in cavities in Suess Glacier and large sublimation ice crystals within cavities in Suess and Meserve glaciers (Holdsworth, 1974) provide evidence for the potential for this mechanism to occur. Data from this study and from Nickling and Bennett (1984) and Goughnour and Andersland (1968) also provide support for the possibility that if the substrate contains layers or lenses within which the ice concentrations are greater than saturation then the peak strength of the layers will be lower than that of the surrounding material (Fig. 5).

In a situations where the ice content was above saturation there is potential for the substrate to deform because greater strain will occur where the ice content is higher. The effect of variations in ice content is described by the right of the peak of the curve shown in Fig. 5 where the peak strength reduces as the amount of ice in the sediment increases above saturation. If segregated ice layers or lenses occur close to the ice–substrate interface, the direct shear tests suggest that the ice is likely to deform at a greater rate than the frozen sediment (Fig. 6). The outcome of this scenario is that the substrate between the segregated ice and the basal ice becomes displaced and the effective bed of the glacier descends into the substrate. This process describes a potential bed-coupling mechanism for a cold glacier resting on a sedimentary substrate. It is interesting to note that the only empirical study that has reported bed deformation at subfreezing temperatures involved "ice-laden drift" with ice concentrations between 30% and 100% (Echelmeyer and Zhongxiang, 1987). Given the controls on the behaviour of ice discussed above, the conclusions of Echelmeyer and Zhongxiang (1987) are not surprising because an ice-laden substrate would have a similar or lower strength than overlying glacier ice and therefore would be coupled to the glacier ice. Data from this study together with previous studies described above suggest that a layered substrate in which the ice concentrations vary will experience greater strain where volume fraction of ice rises above saturation. Material between glacier ice and the substrate that experience high strains could be described as entrained. The related problem of how the debris becomes elevated in the basal ice can only be addressed directly through examination of the nature of deformation within the basal zone which is the subject of further study. Although the interpretation proposed above is plausible, it remains speculative because the direct shear tests are not an appropriate tool for examining small-scale differences in the behaviour of anisotropic materials. Triaxial and simple shear tests may be a more appropriate tool for investigating this behaviour.

6. Implications for landform development

The question that arises from the data and interpretation is, what is the geomorphological impact of the interaction between a dry-based glacier and its substrate? A previous study of the moraines at the margin of Suess Glacier suggested that they formed as proglacial thrust block moraines at the glacier margin

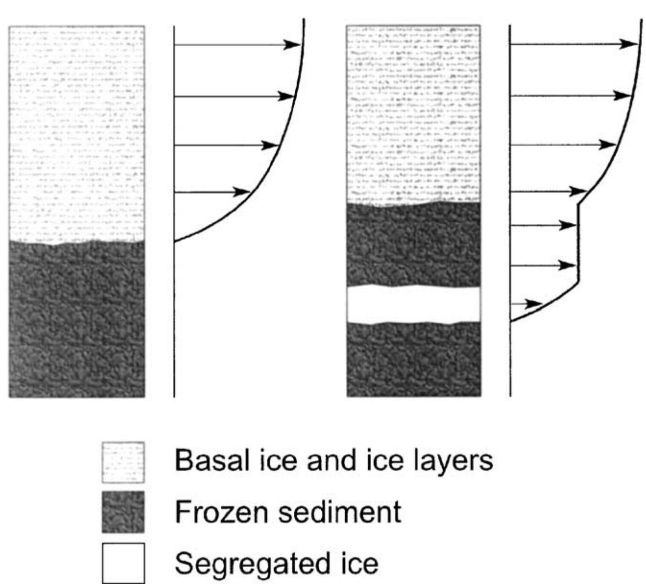

Fig. 6. Diagram showing the lowering of the effective bed into the substrate that contains ice layers or lenses.

as the glacier flowed into unfrozen sediments around an ice marginal lake (Fitzsimons, 1996). This conclusion was based on interpretation of the structure of the moraines which revealed blocks of sediments unconformably stacked against each other and interpretation of the structure of the basal zone from mapping around the ice margin. Despite the structural similarities to thrust block, moraines excavation of the tunnel has revealed a more complex origin.

Extrapolating the position of the main band of solid debris that rests about 2 m above the bed of the glacier to the glacier margin along well defined flow lines shows that the debris band is feeding the moraine (Fig. 7). A thrust block moraine *senso stricto* is a proglacial glaciotectonic feature formed as ice marginal sediments are thrust and stacked without being entrained by the ice (M. Sharp, pers. comm., 1996). The tunnel reveals that debris that feeds the moraine has been detached from the bed and entrained by the ice in a subglacial position and transported to the ice margin within basal ice. The continuity of proglacial and subglacial structures demonstrated by exposures in the tunnel shows that the moraines are extensions of subglacial materials and structures. Consequently, it seems inappropriate to use the term thrust block moraine. Rather than being thrust, it seems that the moraines accumulate at the ice margin as the debris is discharged from basal ice as it reaches the glacier terminus. Sedimentary structures in blocks of sediment that have been entrained, transported and deposited are preserved because the sediment does not experience pervasive deformation when it is entrained (Fig. 2) and because the climatic environment is conducive to passive removal of the surrounding ice by slow melting or sublimation which constitutes about 60% of ice loss (Lewis et al., 1998). A further climatic constraint on deformation of sedimentary structures during deposition is provided by the thin active layer in this environment (ca. 100–200 mm) which means that a relatively thick block that emerges from the ice margin could remain largely frozen during and after deposition.

The moraine at the left margin of Suess Glacier has a cross sectional area of 28 m^2. The composition and debris concentration observations of basal ice show that if the glacier were to melt in situ, there is sufficient debris to result in deposition of up to 0.6 m of debris. Assuming a stable ice margin, a steady velocity within the basal zone, and an unlimited supply of debris, the velocity and debris concentration measurements equate to a debris flux of 0.36 m^3/m of terminus ice margin. If this debris flux remained constant the moraine would take more than 750 years to form.

The moraines, together with basal ice observations, record erosion and evacuation of bed material to the ice margin which demonstrates that this glacier is or has been an effective agent of erosion. However, Suess Glacier may not be representative of the many small alpine glaciers in the McMurdo Dry Valleys. Relatively few of these glaciers rest on unconsolidated valley–floor sediments and are currently transporting much debris. Despite observations of modern cold glaciers that carry very little debris (e.g. Holdsworth, 1974) and geomorphological evidence that suggests cold glaciers can override sediments without significant modification (e.g. England, 1986), evidence from Suess Glacier demonstrates that under some circumstances cold glaciers that rest on unconsolidated sediments can erode their beds.

Given the evidence for subglacial erosion in some circumstances, absolute statements that conclude that cold-based glaciers are not capable of basal erosion appear misleading. The implications of this conclusion for geomorphologists who seek to reconstruct glaciological conditions under former glaciers using simplistic assumptions is that some reconstructions will be incorrect.

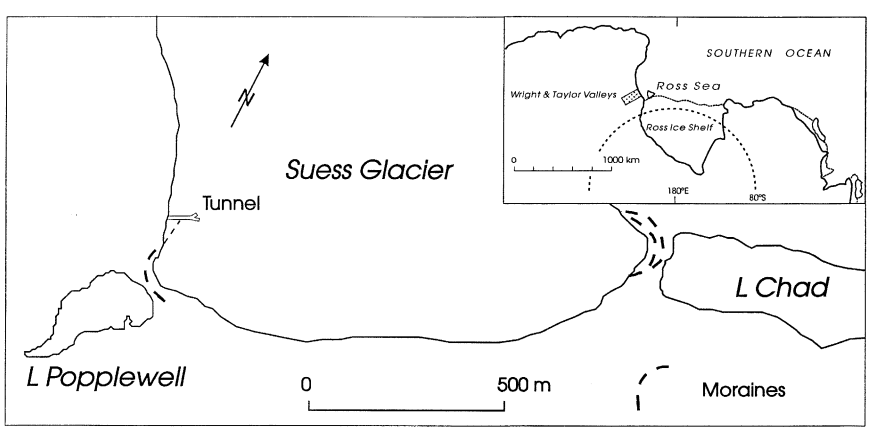

Fig. 7. Plan of the right margin of Suess Glacier showing the location of the solid facies debris band in the tunnel, flow direction of the glacier and position of the moraine.

7. Conclusions

The amber ice facies is significantly weaker than englacial ice, basal stratified ice and the substrate. The strength characteristics suggest that strain would be preferentially accommodated in this layer, which is consistent with previous interpretations of the behaviour of amber ice in other Dry Valleys glaciers.

Although the direct shear tests do not provide a clear explanation of erosion and entrainment processes, they suggest that the bed is not likely to deform unless it contains significant structural weaknesses that remain undetected by the tests.

The structure and composition of the basal ice demonstrate that large pieces of the substrate have been detached, entrained and lifted into the lower part of the glacier. Calculation of the basal debris flux and the volume of the moraines suggest that the processes are very slow and that the relatively small moraines would take more than 750 years to form.

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