The Structure and Strength of Basal Ice in the Suess Glacier, Antarctica

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Abstract

Direct-shear tests were conducted on clean glacier ice, several types of basal ice and frozen substrate material from the basal zone of the Suess Glacier, Taylor Valley, Antarctica. Different strain-rates approximating three orders of magnitude were employed in order to examine whether the rheology of basal material changes with a change in the rate of strain. The results of testing indicated that the peak shear-strength of the materials decreased as slower strain-rates were applied. High debris-concentration, solid facies and frozen substrate attained the greatest values of shear-strength. The lowest values were found in amber ice, which contains a significant solute concentration and finely dispersed sediment particles. High solute concentration along with low sediment content within the ice lattice appears to decrease the shear-strength of basal ice. Results from this study indicate that preferential deformation is likely to occur in basal ice above the saturation point (>25% ice) and that if frozen substrate and solid debris facies deform, it is likely to be at a much slower rate than basal ice.
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A recent reconnaissance of the basal zone in the Suess Glacier, Antarctica has revealed bands of solid facies that appear to have primary sedimentary structures indicative of the entrainment of proglacial lacustrine sediments (Fitzsimons et al., 1998). However, the Suess Glacier is a dry-based alpine glacier that are considered by many authors of glaciological literature to be incapable of any significant basal erosion, for example Boulton (1979); Chinn (1991); Kleman (1993). Such apriori evidence of basal erosion by a dry-based glacier may indicate an inconsistency in glacial literature and therefore requires closer consideration.

Basal erosion in temperate glaciers is accomplished by abrasion and glacial plucking along with the erosive presence of meltwater moving within and below the glacier. Polar or dry-based glaciers that are below the freezing point of water throughout their mass and considered to be completely frozen to their beds have therefore traditionally been excluded from these models for erosion. If basal sliding was possible in dry-based glaciers this could provide a mechanism for denudation of the landscape. A possible mechanism for erosion under dry-based conditions has been identified from by work by Shreve (1984), that offered theoretical evidence for basal sliding at sub-freezing temperatures. Recent examination by direct observation of the basal processes in the dry-based Urumqi Glacier No.1, China (-4°C), has revealed movement through the deformation of unconsolidated drift material and basal sliding at the ice/rock interface (Echelmeyer and Zhongxiang, 1987). Further empirical evidence of basal sliding under the Meserve Glacier, Antarctica (-17°C) has been measured by Cuffey et al. (1999).
If interaction between dry-based glaciers and their beds does take place, the mechanisms and rates by which it occurs require comprehensive understanding in order to facilitate greater accuracy in modelling the movement of these glaciers. Due to the weight of this empirical evidence, questions remain as to the magnitude of the interactive and erosive power of dry-based glaciers on their underlying substrates, as this may alter the nature of the more delicate landforms developed by these glaciers over time.

The mechanical properties of basal material place major controls on glacial effects with regard to landscape morphology (Fitzsimons et al. 1999). These controls affect the nature and rate of change wrought on the landscape, that occur through processes of erosion and deposition of sediment (Hambrey et al. 1999). Strength and rheology of material in the basal zone of glaciers has therefore been a focus of investigation for glaciologists in recent times. Several authors have examined the mechanical strength of basal ice and ice-cemented substrate under varying forms of structural stress (Iverson et al. 1997, Lawson 1996, Rist et al. 1996, Nickling and Bennett 1984). Little quantitative theory of glacial erosion is available, and development of the necessary and fundamental unifying concepts has been slow (Drewry, 1986; Iverson et al. 1994; Menzies, 1995). Several factors are responsible for the deficiencies in the current understanding of the complex nature of the erosive process. Among these is the incomplete knowledge of the behaviour of ice and rock material properties, ice dynamics, thermodynamics, friction and lubrication along with the physico-chemical behaviour of the basal zone (Drewry, 1986). Owing to these deficiencies, a need has been identified for direct observations and measurement of the processes operating at the ice/bed interface of cold dry-based glaciers (Echelmeyer and Zhongxiang, 1987).

In response to these gaps in the knowledge surrounding the basal processes of dry-based glaciers, this study examines the mechanical shear-strength associated with various types of basal ice and ice-cemented substrate material. Utilising a direct-shear device and employing strain rates approximating three orders of magnitude, sampled cores of frozen substrate, debris-laden basal ice, amber ice and cleaner polycrystalline glacier ice are subjected to uniaxial shear tests. This study seeks to establish whether the rheology of different types of basal material changes with variation in the strain rates.
Chapter 1. Introduction

Chapter 2 of this dissertation presents a theoretical review of the current literature of mechanisms and processes operating in the basal zone of dry-based polar glaciers. The theoretical review concludes with a synthesis of some of the outstanding questions surrounding the basal processes of dry-based glaciers as they pertain to this study. An overview of the McMurdo Dry Valleys, Antarctica used as the study area in this dissertation is considered in chapter 3, along with relevant information concerning the Suess Glacier. Chapter 4 contains results synthesised from data collected over the course of the study, followed by a discussion of the findings in chapter 5, and summary conclusions in chapter 6.
2

Glacial structure and processes.

2.1 Introduction.

The knowledge of processes that contribute to movement in the basal zone of dry-based glaciers with regard to aspects concerning glacier mechanics and ice dynamics is for the most part incomplete (Drewry, 1986; Menzies 1995). In this chapter the generic issues relating to the knowledge of the structure and movement of ice and debris in glaciers is reviewed. As an extension of this review the current understanding of dry-based glaciers in the McMurdo Dry Valleys, Antarctica is outlined. A synthesis of outstanding questions follows the reviews of dry-based glacial knowledge so as to provide a logical link to the objectives of this study.

2.2 Movement of ice in glaciers.

The movement of glacial ice may occur as the result of several different processes. Ice in glaciers may move through processes of internal deformation, often referred to as creep, or through folding and faulting within the glacier mass. Other forms of glacial movement may affect the body of the glacier, such as shifting along the bed by basal sliding, or through the deformation of an unconsolidated substrate underlying the glacier. Boulton (1996) proposed theoretical velocity profiles for glaciers of varying thermal regimes and substrates (Figure 2.1). An overview of some of the processes by which glaciers move is described below.
Figure 2.1 Theoretical horizontal velocity profiles for glaciers of varying thermal regimes and substrates. Source: Boulton (1996).
2.2.1 Internal deformation of glacier ice.

The internal deformation of ice occurs as the process of creep, or as large scale folding or faulting. Creep involves both the deformation of ice crystals (basal glide) and the mutual displacement of ice crystals relative to one another in response to the shear stresses placed upon them (Bennett and Glasser, 1997; Wang and Warner, 1998). Ice deforms as a visco-plastic material, the rate of which is temperature dependent coupled with a yield strength of approximately 1 bar or 100 KN.m\(^{-2}\) (Eyles, 1983). This form of behaviour is confined to an elastic response and the plastic (viscous) creep as illustrated by a stress-strain curve (Figure 2.2).

![Figure 2.2 Typical stress-strain curve for polycrystalline ice.](image)

Figure 2.2 Typical stress-strain curve for polycrystalline ice. Plastic creep begins once the yield strength is exceeded culminating in fracture of the crystal lattice once the tensile strength is surmounted. Source: Drewry (1986).

Figure 2.2 depicts a typical stress strain curve for ice and defines the relationship between its elastic and plastic behaviour under stress. Plastic creep begins at the yield strength, \(\sigma_y\), which signifies the termination of elastic creep. Fracturing occurs when the maximum tensile strength, \(\sigma_t\), of the ice crystal lattice is surpassed. Ice creep may exhibit three states, that of primary, secondary and tertiary (Figure 2.3).
The initial elastic strain of polycrystalline ice is followed successively by transient (primary) creep in which the strain rate decreases continuously until a minimum value known as steady-state (secondary) creep is reached. The strain rate increases after this (tertiary creep) and if monitored over a sufficient period of time, a steady value of creep is eventually reached (Paterson, 1981). Simulating typical glacial conditions, it may take several years to achieve steady-state creep at low stresses (Drewry, 1986).

The second form of internal deformation is through large scale folding and faulting. Faulting may be generated by the inability of the ice fabric to creep with sufficient speed in response to stresses placed upon it. This causes fracturing, shearing and folding to occur in the ice fabric. Folding occurs as a result of plastic deformation generating ductile behaviour in ice facies under constant strain (Bennett and Glasser, 1997).

Plastic deformation expressed as ice flow, is a response to gravitational attraction on both the ice fabric and from the overburden pressing down upon it. This is the driver of internal strain deformation (Eyles, 1983). The rate of creep may vary on a
continuum of scale depending on whether the glacier is experiencing accelerating (extensional) or decelerating (compressional) flow. Compressive flow often occurs where the glacial ice thickness starts to decrease near the margins in the ablation zone. Extending flow tends to occur as ice thickness increases, as often found in the accumulation zone. On a smaller scale, compressive flow may occur as a result of obstructions protruding from the bed or from changes in the slope of the bed where the down-slope angle becomes less acute. Extensional flow conversely occurs where the down-slope angle becomes more acute down-glacier (Bennett and Glasser, 1997).

Strain rate is not linearly related to stress for glacier ice, as it is for a viscous fluid. Rather, stress and strain are related in the form of a power law as described by Glen (1955).

Glen’s flow law as applied to isotropic polycrystalline ice, takes the form:

\[ \dot{\xi} = A \tau^n \]  

where, \( \dot{\xi} \) is strain rate
\( \tau \) is shear stress

and, \( A \) and \( n \) are constants.

\( A \) is a viscosity constant which is temperature and position dependent due to changes in the properties of ice with temperature and pressure. The value of \( A \) tends to decrease as temperature decreases. For example, the strain of a glacier at -22°C is therefore 1/10 that of a glacier at 0°C (Chinn, 1991). The value of \( n \) ranges from 2-4 (mean of ca. 3), with its exact value depending on the orientation of crystallographic axes of ice crystals with respect to the direction of applied shear stress (Eyles, 1983). This flow law is based on the assumption that ice consists of fine-grained randomly oriented crystals forming a statistically homogeneous and isotropic material. However, deeper ice may have highly anisotropic crystal structures that are a result of continuous deformation due to large strains involving crystal rotation, recrystallization and crystal growth (Wang and Warner, 1998).

Any naturally occurring body of ice deforms under the influence of gravity given enough time therefore very small amounts of deformation produced by low stresses
can be significant over many thousands of years. Stress-strain rates plotted from field data, for the flow law of ice in steady state glaciers, indicate that there is an effective yield stress of 1 bar that must be exceeded in order for normal flow to take place (Glen, 1953). Even though a significant amount of time is needed, this has implications for the deformation of both ice and debris-laden ice within glaciers.

2.2.2 Temperature dependence of creep in ice

Polar, dry-based, glacier ice tends to be more viscous than that of temperate glaciers because of its lower temperature. This is due to the changes in the rheology of ice with temperature and depth in a glacier. An increase in the overburden pressure decreases the viscous behaviour of ice and allows it to flow more freely. Hence, polycrystalline ice behaves more plastic than viscous at higher levels of stress (Eyles, 1983). Polycrystalline ice softens at temperatures above about -10°C so that the steady-state creep may be expressed in terms of temperatures, usually by an Arrhenius-type equation:

\[ \dot{\varepsilon} = B_0 \exp\left(-\frac{Q}{R_g T_a}\right)\sigma^n \]  

(equation 2)

where,

- \( \dot{\varepsilon} \) = strain rate in steady-state creep
- \( B_0 \) = constant, independent of temperature.
- \( Q \) = activation energy.
- \( R_g \) = gas content \((8.314 \times 10^3 \text{JK}^{-1}\text{kmol}^{-1})\).
- \( T_a \) = absolute temperature.
- \( \sigma^n \) = stress with creep exponent, \( n \), (a visco-elastic parameter).

This equation suggests that under conditions of constant stress the logarithm of the strain rate exhibit a linear dependence on absolute temperature. The temperature dependence of strain-rate in ice however, falls outside of the scope of this investigation since tests on sampled cores from the Suess Glacier were conducted at a constant temperature of \(-17.5^\circ\text{C}\), which is the ambient temperature measured at the base of the glacier. The rheology of basal ice types is examined by using different
strain-rates in order to determine comparative changes in the strength characteristics of the sampled materials in this case.

### 2.2.3 Basal Sliding.

Movement in glaciers may not be achieved by creep alone, the body of the glacier may move along its bed *en masse* by sliding at its base. Basal sliding of glaciers may occur as a result of two main processes: enhanced basal creep and regelation slip. Enhanced basal creep results from the deformation of basal ice around obstacles protruding from the bed of the glacier (Weertman, 1957). The basal ice pressure increases on the upstream side of the obstacle and tends to increase the rate of deformation at this point hence allowing more efficient movement of the ice around the obstruction. Basal pressure is related to the size of the object obstructing the flow, with the larger the object the greater the increase in basal pressure (Bennett and Glasser, 1997).

Regelation slip may occur as a result of ice near the pressure melting point coming in contact with smaller obstacles in the bed of the glacier. The increase in pressure on the upstream side of the obstacle may cause ice near the pressure melting point to melt and flow past the obstruction, as the melting point of ice decreases as pressure rises. The meltwater tends to refreeze downstream of the obstruction as it encounters an area of lower pressure forming regelation ice. In smaller objects the release of latent heat from the re-freezing of ice may pass through the object and aid in the melting process on the upstream side of the obstacle and in turn aid lubrication of the bed (*ibid.*). Basal sliding at sub-freezing temperatures down to -20°C has been shown to be a theoretical possibility and may account for movement in dry-based glaciers of up to 350m over 100,000 years (Shreve, 1984). Cuffey *et al.* (1999) have recently measured basal sliding at the Meserve Glacier, Antarctica, at temperatures of -17°C. It has been suggested that this sliding could be attributed to the presence of the high solute concentration of liquid-like interfacial films existing between the ice and the bedrock, that aids sliding (Cuffey *et al.* 1999).
Sliding may also occur in the absence of bumps on the glacier bed. This form of sliding often occurs as a result of a film of meltwater beneath temperate glaciers or within a network of cavities beneath the glacier. Due to the relative incompressibility of water, the film of meltwater tends to reduce the effective bed pressure by buoying up the glacier and reducing friction with the bed.

### 2.2.4 Weertman regelation.

Protuberances that obstruct glacial flow tend to create zones of higher pressure on their upstream (or stoss) sides. The pressure dependent melting point of ice may allow the ice on the upglacier side to melt and flow, or be squeezed around to the lee side of the obstacle where the pressure is lower. As the meltwater encounters a lower pressure zone it tends to refreeze and release the latent heat of fusion. This allows the transfer of heat to be conducted back through the obstacle and surrounding ice to the stoss side of the obstacle to aid further melting. The Weertman regelation mechanism is considered to be an important part of basal sliding in temperate glaciers where ice is near the pressure melting point and obstacles encountered by the ice are \(<1\text{m}\) in diameter (Weertman, 1957). Due to the basal temperature being well below the pressure melting point, basal sliding remains problematic for dry-based glaciers with regard to this mechanism as both the substrate and basal ice appear frozen together.

### 2.2.5 Deformation of subglacial material.

Subglacial deformation may occur when the shear stress of an unconsolidated substrate containing pore water or interstitial ice is exceeded by the movement of the glacier and thus it becomes part of the flowing ice mass. Through this process the effective bed of the glacier, that is the surface below which there is no forward motion, is located within the unconsolidated ice-laden debris beneath the glacier sole. The deformation of subglacial material may comprise a large proportion of the forward motion of a glacier (Boulton, 1979).
In a study of the Urumqi Glacier No.1, China, (-4°C), Echelmeyer and Zhongxiang (1987) observed directly the enhanced deformation of an ice-laden debris layer at the base of the glacier at sub-freezing temperatures. Brugman (1985, cited in Echelmeyer and Zhongxiang, 1987) also noted that an ice-laden subglacial debris layer deformed so easily that virtually all the observed surface motion of the Shoestring Glacier, Mount St. Helens, Washington, could be attributed to such deformation. In glaciers the rate of deformation, however, will be dependent on the mechanical strength of the frozen debris beneath it. The relation of this strength to that of overlying basal and englacial ice will determine the relative movement with regard to preferential deformation of each material.

2.3 Ice crystallography.

Basal ice is an accumulation of the mechanisms and processes of ice formation from various sources, such as regelation, refrozen surface meltwater and the overriding of pre-existing material. The characteristics of basal ice will be a combination of these processes coupled with the history of deformation within the basal region (Hubbard and Sharp, 1989). Due to concentrations of particulate matter found in the basal zone that may inhibit the growth of ice crystals (Baker, 1978), the ice fabric tends to be finer than that of the cleaner overlying polycrystalline ice (Echelmeyer and Zhongxiang, 1987). Orientation of ice crystals found in glaciers often shows very pronounced, tight single pole ‘e’ axis fabrics orthogonal to the maximum shear stress that may be the result of the growth of strain-free crystals via recrystallization that in turn eases gliding (Holdsworth, 1974; Drewry, 1986; Meyssonnier & Philip, 1996). Basal ice displays greater anisotropy than that of overlying polycrystalline ice which is most likely a result of continuous deformation due to large strains involving several processes such as crystal rotation, recrystallization and crystal growth (Wang and Warner, 1998). This transformation of ice has implications for the strength of basal ice, as discussed below.

Stress is transmitted within any ice mass through intergranular ice crystal contacts, and through bottom and side drag effects that translate as longitudinal push-pull
transient influences (van der Veen and Whillans, 1989). Grain aggregates of ice which are favourably oriented to deform by basal slip, progressively transfer strain (and hence load) to those less favourably oriented. To accommodate the inhomogeneous intracrystalline strain, ductile processes such as recrystallization and kinking effectively allow the deforming grains to accommodate each other without involving other intracrystalline slip systems or micro cracking (Wilson, 1986).

Microscopic examination of the crystalline structure associated with the preferential orientation of ice crystals along a shear plane through the use of thin sections illuminated by polarised light, would provide useful information as to the rheological properties of debris-laden basal ice. The degree of interference by debris particles with the boundaries of ice crystals within the crystal lattice as deformation occurs requires further investigation. Examination of the magnitude of particle to particle contacts and associated frictional heat produced may also prove informative with regard to the effects on the speed of recrystallization of ice grains within deforming frozen sediments. These avenues of research are however outside the scope of the current study.

2.3.1 Shear stresses within glaciers.

Shear stresses develop as the ice deforms near its margins and in the case of simple-shear, under the influence of its own weight (Menzies, 1995). Nye (1952) developed a two-dimensional model that showed that simple-shear stresses vary with depth and ice velocity from a value at the base to a value at the upper surface, when consideration is given to a parallel-sided ice slab. Along the direction of flow, the weight component of a slab of ice is balanced by the shear stress, $\tau$, so that:

$$\tau = \rho \cdot g \cdot h \sin \alpha$$

(equation 3)

where, $\alpha$ = approximate ground surface gradient
$\rho$ = density of glacier ice (0.917gm/cm$^3$)
$g$ = gravitational constant (9.8 m sec$^{-2}$)
$h$ = height of ice column.
2.4 Shear-strength characteristics of debris-laden basal ice.

Sediment begins to deform irrecoverably at some critical shear stress, $\tau_c$ that is given by the Coulomb failure criterion;

$$\tau_c = c + P_e \tan \phi$$

(equation 4)

where, $c$ = cohesion

$P_e$ = the effective normal stress

$\phi$ = the internal frictional angle.

For shear stress $\geq \tau_c$, most proposed expressions for the rate of shear strain, $\gamma$ of a subglacial till layer takes the form;

$$\gamma = \frac{k(\tau - \tau_c)^n}{P_e^m}$$

(equation 5)

where, $k$, $m$ and $n$ are material constants (Boulton and Hindmarsh, 1987).

The value of $n$ in this relation is considered to be critical as it determines the sensitivity of the sediment strain to temporal and spatial variations in the applied shear stress. Hence both the stability of glaciers on deformable beds and the distribution of motion in such beds depend on the value of $n$ (Iverson et al. 1997). The evaluation of material constants such as $n$ is however, outside the scope of this study.

Nickling and Bennett (1984) conducted simple shear-strength experiments on the mechanical behaviour of debris-laden ice manufactured in the laboratory. From this study they concluded that the strength of debris-laden ice exceeds that of clean polycrystalline ice and is possibly attributable to inhibition of dislocation motion by the physical presence of solid particles (Hooke et al. 1972). Strength of frozen soils below -1.0°C is considered by Nickling and Bennett (1984) to be primarily derived from cohesion resulting from ice-to-ice or rock-to-ice contacts. However, as the mineral grain concentration increases this strength is substituted or supplemented by rock-to-rock frictional component as grains come into contact with one another. This frictional contact could promote pressure-melting of ice and re-freezing in an area of
lower pressure with the re-formed crystals being generally smaller and having a preferred orientation to allow dissipation of planar stresses. Goughnour and Andersland (1968), have suggested that a critical point may be attained when ice saturation reaches a value where all grains are out of contact with each other resulting in strength characteristics that are almost totally dependent on the ice fabric rather than internal friction of the granular material.

These results are complimentary to those found from structural field observations by Echelmeyer and Zhongxiang (1987) and from other experiments on samples from ice-sheet cores that have indicated that debris-laden ice is weaker than contiguous clean ice (Lawson, 1996). The workers on debris-laden ice from ice-sheets however, have generally attributed this behaviour not directly to the presence of the debris but to a variation of ice crystal fabric and texture characteristics (Lawson, 1996). Temperature change also showed dramatic effects on the strength of debris-laden ice when subjected to uniaxial compression strength tests by Lawson (1996). At temperatures close to 0°C clean glacier ice displayed shear strengths 2.5 times the mean of debris-laden ice, however at temperatures below -5°C the strength of debris-laden ice became increasingly stronger than clean ice.

2.5 Cold, polar, dry-based glaciers.

Cold, dry-based or polar glaciers are so named because the ice is below the pressure melting point throughout the glacier and they are therefore generally frozen to their bed. Due to the inaccessibility of the sole and basal zone of most polar glaciers, authors of glacial literature tend to make theoretical generalisations about these glacial processes and how they interact with their sediments. For example Boulton (1970a), Eyles (1983a), Sudgen and John (1984), Drewry (1986), Chinn (1986, 1990), have made theoretical statements of motion and the possibility of erosion by dry-based glaciers.

A common generalisation made by many of these authors is that dry-based glaciers do not interact with their beds and hence are virtually incapable of erosion and that their deposits retain more characteristics of glacial transport than other glacial systems.
Holdsworth (1969) has identified four major rheological flow zones operating in the region of the ice-cliff of dry-based glaciers. These consist of a rigid outer zone, a semi-rigid zone, a semi-plastic zone and a plastic zone (Fig2.4).

Figure 2.4 Four rheological zones of a dry-based glacier as identified by Holdsworth (1969). Source: Chinn (1991).
Holdsworth regarded the ca. 20m thick, stiff, semi-rigid zone containing the upper surface of the glacier and the rigid ice-cliff as important to the movement of dry-based glaciers. This zone becomes grounded at the terminus excluding the more easily deformed plastic basal layer and becomes an obstruction to the thick upstream ice. Subsequent examination of the Urumqi Glacier No.1 (-5°C) by Echelmeyer & Zhongxiang (1987) has revealed that some dry-based glaciers may also move by other processes such as that of subglacial, ice-laden drift deformation and slow rates of basal sliding.

Examination of the basal zone of an alpine glacier from a tunnel excavated in the Suess Glacier, Taylor Valley, Antarctica revealed a basal zone extending upward some 4.2m from the bed. A velocity profile of the basal zone indicated movement throughout the entire zone, the lowest displacement of which was found within 1 meter of the bed. Linear velocity displacement transducers (LVDT’s) detected no movement, but some precision dial gauges did record small displacements at the bed of the glacier, as indicated later in this paper.

2.5.1 Velocities of dry-based glaciers.

Ice movement in glaciers occurs by processes of internal deformation, basal sliding or deformation of unconsolidated substrates (Boulton, 1979; Drewry, 1986; Echelmeyer and Zhongxiang, 1987; Bennett and Glasser, 1997). Small alpine glaciers such as those found in the McMurdo Dry Valleys, Antarctica exhibit slow rates of movement, with annual rates that are more typical of daily values of temperate valley glaciers (Chinn, 1991). Bull and Carnein (1970) found faster rates of movement in the steep trunk of the Meserve Glacier (ca 13°), with values of 3 m.a⁻¹ being recorded. Typical rates for dry-based glaciers however, reflect those mean rates of 0.1 - 0.5 m.a⁻¹ and 1.02 m.a⁻¹ that have been recorded for the Jeremy Sykes Glacier and Heidmall Glacier respectively (Chinn, 1981). From these mean flow rates, the time taken for ice to travel the full length of a 5 km glacier from headwall to terminus would be expected to be a minimum of 5000 years. The kinematic wave from a response to climate
change however, would only take about one third of this time to pass throughout the glacier (Chinn, 1991).

2.6 The formation of basal ice.

The cryostatic pressure of glacier ice coupled with movement near the glacial sole results in high shear stresses in close proximity to the substrate upon which the glacier rests. These shear stresses cause straining of the ice fabric across the bed and around obstacles in the path of glacial flow, as discussed earlier. Interaction with the mineral content of the substrate combined with stresses and strain causes distinctive structural, sedimentological and chemical features that characterise the basal ice layer of glaciers. As such, the basal zone is distinguished from the main body of ‘glacial ice’ that is a product of the firnification of snow at or near the upper surface of the ice masses (Paterson, 1971). In general, basal ice contains more debris, chemical solutes and has an anisotropic structure coupled with a lower gas content that aids in its distinction from “glacial ice” (Hubbard and Sharp, 1989). These characteristics that differentiate the basal zone of glaciers also lead to the postulation that the rheological behaviour of this portion of glaciers may be different to that of glacier ice or clean laboratory made polycrystalline ice.

2.6.1 Stratigraphy and sedimentology of the basal zone.

The stratification of sediments within the basal zone of glaciers are characterised by a variety of layers of sediments of different grain sizes, and lenses of clean and bubbly ice. Sediment layers may appear as solid plates, be stratified or seem dispersed throughout an ice matrix. The appearance of basal ice and debris is dependent on the mechanisms of entrainment and deposition within the glacier and the composition of the glacier bed. Where a glacier overrides unconsolidated sediment, the basal composition is usually more complex in character (Hubbard and Sharp, 1989). Identification of the processes forming basal ice may be aided by the examination of
associated ice and debris facies and through the characteristics of basal sedimentology and stratigraphy.

Lawson (1979) suggested one approach to the systematic study of the stratigraphy of ice sequences. Ice facies of the Matanuska Glacier, Alaska were defined according to the range in ice characteristics present in the glacier. In this study the ice facies were divided into three main zones including the basal zone, which was further, subdivided into dispersed and stratified facies. Subfacies consisting of solid, suspended and dispersed debris was identified within the stratified basal layer (Figure 2.5).

![Image of ice facies and associated debris zones in a vertical section of the Matanuska Glacier, Alaska. Three subfacies are identified within the stratified facies of the basal zone; solid, suspended and dispersed. Source: Lawson (1979).](image)

**Figure 2.5** Ice facies and associated debris zones in a vertical section of the Matanuska Glacier, Alaska. Three subfacies are identified within the stratified facies of the basal zone; solid, suspended and dispersed. Source: Lawson (1979).

A more detailed description of ice cores sampled from this basal sequence that were used in simple-shear strength tests are supplied later in this study.
2.6.2 Solute content of basal ice in Dry Valley glaciers.

Most studies investigating the chemical composition of basal ice have been concerned with temperate glaciers (Hooker, 1998). The majority of these studies have been involved in examining the mechanisms for the formation and refreezing of meltwater at the glacier sole, and identifying water sources involved in these interactions. Due to the importance placed on temperate glaciers and the associated chemical composition of meltwater and basal ice as it relates to movement, few studies have been conducted on dry-based glaciers as they are frozen to their underlying substrates. Recent studies by Hooker (1998) and Lambourne (1998) have discussed the solutes found in the basal ice of dry-based glaciers and the mechanisms by which the solute content of basal ice may be enriched.

From a review of work by Souchez et al. (1995) and Holdsworth (1974), the diffusion of solutes within the crystal fabric of ice has been identified as a possible mechanism of enrichment of ice layers within the basal zone of some dry-based glaciers of the McMurdo Dry Valleys, Antarctica. High solute concentrations were found in the amber ice facies of the Suess Glacier (Hooker, 1998; Lambourne, 1998), and in the Meserve Glacier (Holdsworth, 1974). The source of these solutes is attributed to the salt-rich substrate underlying these glaciers.

Three possible forms of solute diffusion have been identified in ice, these are volume diffusion, grain-boundary diffusion and glide-plane diffusion (Holdsworth, 1974). Volume diffusion is generated by a concentration gradient through the basal ice and operates on a vacancy mechanism, which is discussed later. Volume diffusion may operate at low temperatures, as demonstrated by the molecular diffusion of H\textsuperscript{3} and O\textsuperscript{18} in ice at temperatures as low as –35.9°C (Holdsworth, 1974). It is thought that the diffusion of solutes through ice is most efficient in old ice with high concentration gradients (Souchez et al. 1995).

Grain-boundary diffusion is thought to speed up the diffusion process, as implied by an experiment involving the Na concentration of water from between the grains of partially melted ice. The concentration of Na was found to be greater at the crystal
boundaries than that at the centre of the ice crystals (Holdsworth, 1974). A liquid-like layer has been postulated to exist at the grain boundaries of ice and foreign solids that may be enhanced by increased solute content. The presence of this liquid layer may enhance the diffusion of solutes through the crystal lattice (Cuffey et al. 1999). Raymond and Harrison (1975) have examined water veins at three-grain intersections after ice cores were retrieved from depths of up to 60m and inferred the percolation rate of 19cm.a\(^{-1}\) in fine grained ice (≈2mm), and 0.05cm.a\(^{-1}\) in coarse grained white ice. Defects in the crystal lattice such as holes or small foreign atoms occupying an oxygen atom position may facilitate the passage of diffusing entities such as solutes with small atomic radii (Holdsworth, 1974).

 Glide-plane diffusion is thought to occur near the base of glaciers where plastic deformation is greatest and the significant transportation of salts and debris is possible (Holdsworth, 1974). Evidence of plastic deformation can be seen in the basal ice near the bed of the Suess Glacier (Figure 2.6). It is within this decimetric bottom layer that ions from the bedrock can diffuse into the ice (Lliboutry, 1993).

**Figure 2.6** Ductile deformation of sediment within one meter of the bed in Suess Glacier.
2.7 Debris entrainment into basal ice of dry-based glaciers.

Bands of frozen sediment together with layers of debris laden ice and a well-developed layer of discoloured “amber ice” were observed within a tunnel excavated in the Suess Glacier, a dry-based alpine glacier in the Taylor Valley, south Victoria Land, Antarctica. A poor explanation of reasons for these observations has been obtained from the literature to date, as most authors still tend to believe that dry-based glaciers do not interact with their beds to such a great degree (Chinn 1991, 1987b; Kleman, 1994). The mechanisms of entrainment of sediment into dry-based glaciers remain poorly understood, but there remains a general consensus in the literature of the existence of two major mechanisms for debris entrainment at the sole of glaciers and large ice masses.

1. Pressure-melting regelation (Weertman, 1957, 1964) that is induced by pressure-melting fluctuations around bed protuberances and usually yields low debris/ice ratios exhibiting separated sequences of clear-ice laminae alternating with thin debris layers (Tison et al. 1993).

2. Freezing-on (Weertman, 1961), allowing the net accretion of ice at the glacier sole as a result of fluctuations in the position of the pressure-melting point (pmp.) isotherm with respect to the ice/bed interface. High yields of debris/ice ratios are associated with this mechanism where the ice/bed interface consists of soft unconsolidated substrate material (Tison et al. 1993).

Figure 2.7 depicts the pathways by which debris may be entrained into a glacier (Boulton, 1978). These pathways consist of supraglacial debris moving within and below the glacier along with debris that may be entrained and transported subglacially.
2.7.1 Robin’s heat pump effect.

A possible mechanism for the entrainment of debris has been proposed by Robin (1976), involving the melting of ice as it enters high pressure zones at the stoss side of obstacles, as described above (Section 2.2.4). Cold patches may occur on the lee side of these obstacles due to the “heat pump” effect that is postulated to be generated by the squeezing of meltwater along interangular veins. Some of these veins may close under pressure resulting in a loss of meltwater from the system. As lower pressures are encountered by the meltwater on the lee side of the obstacle the ice is unable to adjust rapidly to the new melting point as only a portion of the meltwater is present to supply the latent heat that would be released during refreezing. Unless there is free water available in the ice that can release this latent heat the ice will remain at a lower pressure melting point, causing the formation of a cold patch that may extend laterally...
for several meters (Souchez and Lorrain, 1991). The heat pump effect allows a mechanism by which water that has not been melted at the sole of the glacier to be frozen on to the glacier ice such as in the case of Weertman regulation. Debris may accompany this refrozen water thus allowing a mechanism for entrainment. The vertical extent of both Weertman regulation and the Robin heat pump effect are limited, with both considered only in the order of centimetres. Hence, other processes and field evidence are required to explain the greater vertical extent of the basal ice found in some polar glaciers (Hubbard and Sharp, 1989).

2.7.2 Entrainment of ice and debris through incorporation of apron ice.

Debris entrainment from the incorporation of the ice and debris aprons found at the margins of glaciers as proposed by Shaw (1977), is one method by which debris may be incorporated into the basal zone of polar glaciers (Figure 2.8). Shaw (1977) suggested that as dry-based glaciers and their substrates are frozen together, the most effective process of debris entrainment in arid environments would be frontal apron incorporation. Frontal aprons are accumulations of fallen ice blocks, supraglacial debris, wind blown snow and sediment and regelation ice from refreezing of supraglacial meltwater streams at the inflection of the glacier cliff-face and the ground (Shaw, 1977). Shaw’s model (Figure 2.8) suggests that the frontal apron is overridden and incorporated and then gradually attenuated by an advancing glacier. Shaw (1977b) suggests that this process results in distinctive foliation and debris content in the basal ice layers in dry based glaciers. Evans (1989) evaluated Shaw’s hypothesis for apron incorporation at the margins of Philips Inlet Glacier, Ellesmere Island in the Canadian High Arctic. Debris was present in the terminal ice cliffs as debris-poor folia expressing internal flow patterns, coupled with debris-rich bands of various thicknesses with a variety of grain sizes. Evans (1989) concluded that the recycling and reincorporating of fluvial and deltaic sediments, aprons and proglacially thrusted blocks near the ice face was a suitable alternative to debris entrainment by glacial plucking and larger scale freeze-on processes. Subsequent examination of the inner apron and basal ice layer of the Suess Glacier, Taylor Valley, Antarctica by
Larking (1997) revealed several inconsistencies associated with Shaw’s (1977) model. Larking (1997) concluded that Shaw’s (1977) apron entrainment model did not fully explain the entrainment process of debris within the Suess Glacier. These inconsistencies included the overtly dramatic attenuation of the apron ice within meters of the glacier margin coupled with large sediment blocks found at elevation further into the body of the glacier. Hooker (1998) substantiated these conclusions after examination of solute concentrations in both basal and apron ice. The high solute content of the basal ice layers in the Suess Glacier were distinct from those of the apron ice as the chemical signature of the apron ice was much closer to that of clean glacier ice.

2.7.3 Formation of shear and thrust-block moraines.

Chinn (1991) reports that it is the grounding of the rigid surface “crust” of ice around the margins of polar glaciers that may be responsible for the shear moraines and other forms taken by the termini of dry-based glaciers. “Shear moraines” may form in response to the pressure of upstream ice causing upwarping of the glacier sole and debris towards the surface. This response in the ice constitutes an attempt to override the obstacle caused by the stiff, grounded ice around the glacier margins (Chinn, 1991). Chinn’s (1991) observation was based on the rheological zones identified by Holdsworth (1969) who identified the stiffer ice crust at the margins of dry-based glaciers as compared to the more plastic behaviour of ice within the body of the glacier. Tison et al. (1993) suggest that the local stress field caused by the pressure of upstream ice against the buttressed rigid ice near the glacier margin causes mylonitization of the active glacier ice as it over rides the dead slow-moving ice near the margin. Even though shearing has been measured within glaciers and may account for uplift of sediment into the glacier (Echelmeyer and Zhongxiang, 1987), it does not account for how sediment initially became entrained, especially in the case of dry-based glaciers. The mechanism for entrainment of sediment blocks within the Suess Glacier remains unexplained with absolute certainty due to the frozen nature (-17.5°C) of the underlying sediment.
Figure 2.8 Debris entrainment by incorporation of the frontal apron. Source: Shaw (1977).
Three alternative models for the formation of thrust-block moraines at the margins of dry-based glaciers have been proposed by Fitzsimons (1996); block entrainment of sediment associated with frozen bed deformation, entrainment by overriding and accretion of marginal ice and debris-aprons, and transient wet-based conditions associated with glaciers flowing into lakes. Well preserved, primary sedimentary structures coupled with beds of algae have been observed within sediment blocks entrained by the Suess Glacier (Fitzsimons, 1996). Chinn (1987) has observed similar structures in the Victoria Lower Glacier, another dry-based glacier in the Victoria Valley, Antarctica. Boulton (1979) suggested that whereas un lithified unfrozen sediment deforms beneath the glacier rather than being incorporated within it, ice-cemented subglacial sediments can behave like bedrock because of their relative rigidity, and may be plucked and entrained englacially. This method of entrainment however would require the ice-cemented blocks to have a lower tensile strength than the overlying ice or the ice/sediment-block contact strength. Additionally the pressure melting point would need to be achieved within the block of sediment well below the ice/sediment-block interface so as to lower the tensile strength of the ice-cemented sediment. Chinn (1987) suggested a possible mechanism for these conditions after chemical and structural analysis of the ice along with radiocarbon dating of the algae. Chinn (1987) hypothesised, that during a Ross Sea advance small saline ponds and streams existed in the Victoria Valley at the glacier’s present location. The high saline content of these ponds may depress the melting point at the glacier sole sufficiently to permit basal erosion and entrainment of large blocks of lake sediment. Fitzsimons (1996) developed the ice/debris accretion model further by examining the structure and sedimentology of thrust-block moraines at the Wright Lower Glacier and Suess Glacier (Figure 2.9). Accretion of ice and debris were possible, it was suggested, at ice margins terminating in proglacial lakes due to relatively warm lake waters keeping the surrounding sediment unfrozen. The warm bottom waters were a consequence of heat conducted through the surface ice to the water below and where water circulation is weak, such as beneath a glacier, the heat flow from the water to the ice will continue until the water and ice equilibrate. At this time it is possible that the lake water debris may start to freeze to the glacier sole (Hooker, 1998).
Figure 2.9. Model for the accretion of ice and debris from proglacial lakes. Source: Fitzsimons (1996).
2.8 Previous mechanical strength tests of basal ice.

In unconfined uniaxial compression tests performed on cylindrical samples of sand/ice mixtures at $-9.1\,^\circ\text{C}$, Hooke et al. (1972) found the creep rates in samples with greater than 35% sand volumes to be lower than those of samples with less sand in the ice matrix. The results of their test suggested that the creep rate of ice with low sand concentrations was sensitive to factors such as ice crystal size, the distribution of sand particles relative to crystal boundaries and the concentration of chemical impurities in the ice.

Nickling and Bennett (1984) conducted simple-shear strength experiments on the mechanical behaviour of debris-laden ice, and from this study they concluded that the strength of ice-laden debris at saturation level (25%) exceeds that of clean polycrystalline ice at temperatures of $-1\,^\circ\text{C}$. The shear strength of frozen debris is thought to increase with decreasing temperatures but to decrease with decrease in strain-rate (Nickling and Bennett, 1984).

Lawson (1996) subjected cores sampled from the Taylor Glacier, Antarctica to uniaxial compression strength (UCS) tests in the field using high strain-rates to collect a large data set during a short field season. Lawson (1996) reported that at near-zero temperatures the UCS of clean ice was nearly 2.5 times as great as that of the mean UCS of debris-laden ice. At temperatures of $-5\,^\circ\text{C}$ and colder however, UCS of debris-laden ice became greater than that of clean ice and continued to gain strength systematically as temperatures became colder. Pressure melting was observed in the ice cores during the compression tests. Small amounts of pressure melting were still detectable at temperatures of $-25\,^\circ\text{C}$. Lawson (1996) suggests that the reversal in relative strength of debris-laden ice to that of clean ice with decreasing temperature may be linked to the changing pressure melting rate of the debris-laden ice. Lawson’s (1996) form of strength test may not be directly transferable to ice sheet modelling, in that unconfined uniaxial compression at rapid strain rates are unlikely to be found at the base of large ice masses. It does however, raise questions as to the temperature dependency of the rheological properties of debris-laden basal ice. The rate of strain applied to ice-laden debris may also affect the rheological properties associated with
both this material and that of clean ice and therefore also warrants further investigation.

The shear stress supported by sediment in direct-shear tests, which allow only small strains (up to ca.6-8%), may be sometimes as much as 75% larger than the residual strength (steady shear strength at a steady rate of strain) (Atkinson, 1993, cited in Iverson et al. 1997). Ho et al. (1996, cited in Iverson et al. 1997), have argued that visco-plastic parameters can still be extracted from such tests. Strain rates such as those used by Lawson (1996) are experimentally convenient but do not reflect those operating under normal ‘glaciological’ conditions (low strain rates in shear, confining conditions) (Lawson, 1996). Slower, and therefore more realistic strain rates are employed in this study, although temporal constraints do not allow absolutely realistic glacial strain rates to be used.

2.9 Research Objectives.

The basic ingredients of an ice-sheet model include prescriptions for viscous creep flow by internal deformation (Wang and Warner, 1998). However, controversy still surrounds the relative strengths and viscous behaviour of debris-laden and clean basal ice along with that of frozen substrates. This study investigates relative mechanical strength by employing a direct-shear device to strain cored samples of several types of basal ice and frozen substrate along a thin shear plane. Strain-rates approximating three different orders of magnitude are employed by this study to assess whether changes in the rheology of basal material occur at different rates of strain. The peak shear-strength attained by each material is compared to that of the other materials tested at each strain-rate. Experiments were conducted in a cold-room (-17.5°C) in an effort to mimic temperatures measured within the tunnel excavated into the basal zone of the Suess Glacier, Antarctica.
The objective of this dissertation is:

To examine the rheology of different types of basal ice and frozen sediment using a direct-shear device, and to investigate whether strength characteristics in these materials change at different strain-rates.

The hypotheses to be investigated are:

**H 1.** The rheology of ice-cemented substrate and basal ice will change under different rates of strain.

**H 2.** A mechanism by which substrate may yield at stresses that are the same or lower than that of basal ice will be identifiable from direct-shear tests at strain-rates approximating three different orders of magnitude.
3

Field Area and Research Methodology.

3.1 Field Area.

The McMurdo Dry Valleys are one of the largest ice-free regions on the Antarctic continent and lie along the western margin of the Ross Sea. The Taylor Valley (77°00′S, 162°E), is approximately 400 km² in area, running roughly east-west and bounded on the north by the Asgaard Range and on the south by the Kukri Hills (Fig.3.1). A mosaic of perennially ice-covered lakes, ephemeral streams, bare rocky ground, permafrost and glaciers characterise the valleys. Glaciers cover approximately 35% of the Taylor Valley with the predominant type consisting of small alpine glaciers that flow out of the bordering mountains (Lewis et al. 1998). All of the alpine glaciers in the valley are frozen to their beds hence meltwater runoff is restricted to glacier surfaces generating ephemeral proglacial streams, with an absence of englacial/subglacial hydraulic systems (Chinn, 1990).

3.1.1 Climate of the Dry Valleys.

The aridity of the Dry Valley region is attributed to a precipitation shadow formed by mountain ranges up to 2500m in height aligned in an east-west arrangement in the area. Snow bearing cyclonic systems that tend to track southward over the Ross Sea are intercepted and blocked from further inland penetration by these mountains, causing a marked precipitation deficit in the inland regions. The ground surface remains predominantly devoid of snow due to the extreme vapour gradient. The
atmosphere of the region is so dry that evaporation causes snow to sublimate soon after it settles on the ground. At higher elevations on the ranges heavier snowfalls are redistributed by strong winds to accumulate enough snow in sheltered places to form alpine glaciers. Temperatures in the valleys range from about 0°C during summer to colder than -50°C during winter, with a mean annual temperature below -20°C (Chinn, 1990). Occasionally, December and January air temperatures may reach +5°C (Holdsworth, 1974). Winds from the northeast and southwest are predominant in the valleys and are virtually continuous. In summer northeast winds dominate but there is considerable evidence that during winter strong westerly katabatic winds flow down into the valleys from the ice dome, causing erosion (Bull, 1966). The presence of darker bare rock influences the extremes in the temperature regime of the valleys by absorbing more heat during summer and losing more heat as a black body radiator during winter. The influence of bare rock may be seen when comparing temperatures at lower elevation sites such as Scott Base, where summer temperatures can be up to 5°C cooler in summer and 5°C warmer in winter, than that of the Dry Valleys (Lambourne, 1998).

3.1.2 Glaciers of the Dry Valleys.

The glaciers of the Dry Valleys have been classified into three types according to size and location (Chinn, 1990). 1. Outlet glaciers similar to the Taylor Glacier that drain from inland ice domes. 2. Piedmont glaciers such as the Wright Lower Glacier and Canada glacier, that cover most of the seaward margins of the Dry Valleys. 3. Alpine glaciers as typified by the Suess Glacier, that flow down the valley sides from cirques that accumulate snow, some of which may reach the valley floors.
Figure 3.1 McMurdо Dry Valleys area of Antarctica. The Suess Glacier is the focus of this study and is highlighted. Source: Larking (1997).
3.1.3 Suess Glacier.

The Suess Glacier is a cold, dry-based, alpine glacier descending from 1750 m in the Asgaard Range, with a flow of ca. 5km. Although most of the glacier rests on gneiss, the terminus reaches the valley floor and rests on unconsolidated sediments that are frozen to a depth of greater than 300m. Relatively smooth surfaced and free of surface moraine, Suess Glacier flows across the floor of the Taylor Valley where it terminates in an unsymmetrical lobate ice cliff (Figure 3.1). This lobate form is due to the interaction of the terminus with an intra-valley ridge on the opposite side of the Taylor Valley, causing the distinct termini, one up-valley and the other down-valley to form. A narrow passage persists between the ice terminus and a step scree slope on the southeast valley side, known as ‘The Defile’. The right (up-valley) terminal margin residing alongside Lake Popplewell is a steep sided ice cliff with some small moraines extending from the glacier apron. Calving of ice was observed from this margin during the study period. The left (down-valley) terminal margin bordered by Lake Chad consists of a series of ice cliffs coupled with numerous complex ice cored moraines (Fitzsimons, 1996). These moraines and ice-debris apron are a part of a well developed basal zone present in the lateral and terminal margins of the Suess Glacier, and extends up to 3.5m from the glacier bed (Figure 3.2). The debris bands are considered to consist of bed material including blocks of proglacial lacustrine deposits (Fitzsimons, 1996). A layer of amber ice as described by Holdsworth (1974), is present overlying the stratified basal debris bands. This layer continues up the profile to ca. 4.2m. Clean polycrystalline glacier ice is present in the remaining portion of the vertical profile.
Figure 3.2  The true right margin of the Suess Glacier as it flows across the Taylor Valley, Antarctica. A well developed basal ice layer and apron are evident near the edge of Lake Popplewell (foreground). The tunnel entrance is clearly visible in the basal zone of the glacier. Photo courtesy of R. Larking.
3.2 Methods.

3.2.1 Research strategy.

The relative inaccessibility of polar glaciers combined with the loss of daylight and extreme cold temperatures during winter months provides a limited field season available to researchers in Antarctica. These temporal constraints required that the methodology employed in this dissertation be conducted both in the field and in a laboratory in the Geography Department at the University of Otago, New Zealand.

The purpose of this study is motivated by an interest in how material can be eroded and entrained within a dry-based glacier, especially if they are considered to be permanently frozen to their beds. Evidence of erosion and entrainment of debris by a glacier in dry-based conditions is inferred from the debris layers observed within the basal zone of the Suess Glacier. Particularly interesting is the mechanical conditions surrounding activity within the basal zone of such glaciers as it relates to preferential movement of various types of basal ice and substrate material. Examination of the strength characteristics of the basal ice and substrate may provide clues that will help identify mechanisms of movement and entrainment at the bases of dry-based glaciers. As this is a reconnaissance-level study, samples were acquired from the basal zone of an actual glacier, rather than trying to reproduce basal material in a laboratory or conducting a theoretical or descriptive study.

The aim of the field component of this research was to gain access to an entire vertical sequence of the basal zone and substrate of the glacier bed within a dry-based glacier in order to take samples for experimental testing both in situ and in a laboratory. A tunnel excavated by Fitzsimons et al. in 1997 into the Suess Glacier provided access to obtain representative samples of a basal sequence and the underlying substrate for simple-shear stress test analysis.
Cores were subjected to direct-shear stress tests within the tunnel employing a strain rate of 0.80mm.hr\(^{-1}\). This rate of strain allowed most samples to be sheared in 3-4 hours under a normal load of 200 KPa (the value of the normal load equates to the normal load of ca.20m of ice that occurred at the tunnel site). Stress tests were conducted under laboratory conditions employing displacement rates of 0.08mm.hr\(^{-1}\) and 0.01 mm.hr\(^{-1}\) at the University of Otago, New Zealand one year later. The field, laboratory and analytical techniques employed to generate and synthesise data in this study are described in the following section.

### 3.3 Field Methods.

A tunnel measuring ca 25m long, 2m high and 1.5m wide was excavated in the basal zone of the true right side of the Suess Glacier, approximately 250 m from the terminus by Fitzsimons \textit{et al.} in 1997 (Figure 3.3). A 3.7m extension to the tunnel was added in 1998, oriented in an easterly direction and culminating in a vertical shaft ca. 4.5m high x 1.5m wide, exposing the full basal sequence.

![Figure 3.3](image)

**Figure 3.3** Excavated tunnel within the Suess Glacier, depicting orientation of the tunnel to the glacier margin and extension added in 1998. Source: Lambourne (1998).
Blocks were removed from the vertical sequence using tungsten-carbide tipped cutters bonded to the blade of a chainsaw. Cores measuring ca. 58 x 130mm were then removed from the blocks employing a slow speed diamond coring bit attached to an electric drill. Some cores were subjected to direct shear tests within the confines of the tunnel. The remainder of the cores were numbered and wrapped in air-tight polythene bags for removal and placed in an insulated wooden rock-box. Ensuring that the cores remained at temperatures close to that measured near the end of the tunnel (ca.-17°C) the cores were freighted to the Geography Department, University of Otago, New Zealand for analysis. The basal ice sequence was described by Lambourne (1998), based on a series of criteria outlined by Knight (1997). Physical characteristics of the cored samples were described based on criteria outlined in Hubbard et al. (1996).

Measurements of the velocity of the glacier bed were conducted using linear variable displacement transducers (LVDT) and precision dial gauges, capable of measuring displacements of 0.01 mm. Utilizing rock bolts, the dial gauges and LVDTs were anchored to the substrate and measured the movement in the basal zone from the displacement of wooden pegs drilled and frozen into the ice 5mm above the bed. Displacements of the LVDTs were recorded in a Campbell data logger every 2 days for 4 weeks and after 348 days.

Plumb lines were deployed in the vertical shaft at the end of the tunnel in order to characterise the form of the velocity profile in the lower portion of the glacier. A 4.2m line with a plumb bob was affixed above a brass target bolted to the substrate allowing displacements to be measured as the offsets of the line from a vertical row of wooden pegs. Holes were drilled into the vertical ice face and the wooden pegs were frozen into place approximately 200mm apart. The displacements were measured using digital callipers capable of measuring intervals of 0.01mm. The movement of the bed was monitored at four locations within the tunnel by drilling a 400mm-long hole and filling the hole with painted wooden dowel segments. After 348 days the hole was excavated and the position of the dowel segments were measured (Fitzsimons et al. 1998).
3.3.1 Field based direct-shear stress tests of ice cores.

Direct shear tests of the ice cores were conducted in the tunnel at a temperature of –10 ±2°C using a laboratory direct shear test apparatus modified for use in the field (Figure 3.4). Cylindrical core samples described above were frozen into place in the stainless steel plates of the shear box, and a normal load was applied via a pneumatic shock absorber, hand pump and a strain gauge. A normal load of 200 KPa was used to simulate the normal load of 20m of ice present above the site of the experiment (Fitzsimons et al. 1998). The cores rested atop a bench cut into the ice with the shearing device placed adjacent to it (Figure 3.5). Shear tests were conducted within the tunnel over 3-4 hours using a displacement rate of 0.80mm hr$^{-1}$. This displacement rate is approximately two orders of magnitude greater than the basal velocities that were subsequently measured using linear variable displacement transducers (LVDT) and precision dial gauges. Due to temporal constraints these strain rates were applied in order to facilitate the measurement of a representative number of cores within the limited field season. The shear strength was recorded as the output of a strain gauge that was logged every 61 seconds (ibid).

![Direct shear test apparatus modified for use in the field. Source: Fitzsimons et al. (1998).](image)

Figure 3.4 Direct shear test apparatus modified for use in the field. Source: Fitzsimons et al. (1998).
Figure 3.5  Field set-up of direct-shear device. Hydraulic pump is used to simulate normal load of ice above the sample site. Source: Photo courtesy of S.J. Fitzsimons (1998).
3.4 Laboratory Methods.

3.4.1 Laboratory based simple-shear stress tests.

A cold room was set up at the University of Otago, Geography Department, New Zealand in order to carry out simple-shear stress tests on basal ice and substrate cores sampled from the Suess Glacier, Antarctica. Laboratory methods were required for these tests as the field season in Antarctica is too short for slower strain rate tests to be completed in situ. The cold room temperature was set to $-17.5^\circ C$ to simulate conditions encountered at the end of the tunnel excavated in the glacier. A stable power supply (PP5017) and CR10 Data-Logger were set up outside the cold room to record the pressure (in newtons) exerted by the shear device on the ice cores through time. Measurements of stress exerted by the shear device were obtained from the output current of a load cell (Precision Transducer Ltd. ST1000) supported by bearings between the drive screw of the shearing device and the moving upper platen containing the core. The output voltage of the load cell was converted to newtons by software in the data-logger. Instantaneous, minimum, maximum and hour averaged pressure information was recorded every 61 seconds for displacement rates of 0.08 mm.hr$^{-1}$ and every 10 minutes for rates of 0.01 mm.hr$^{-1}$ on the data-logger.

Strain rates of 0.08 mm.hr$^{-1}$ and 0.01 mm.hr$^{-1}$ were applied to cores sampled from the vertical profile and bed exposed at the end of the tunnel. The cores were constrained by a normal load that was applied to the ice cores by employing a suspended weight apparatus hanging down from an alloy end cap with a countersunk groove in the centre and frozen to the top of the cores (Figure 3.6). The normal load was calculated to simulate the cryostatic pressure of the overburden of ca. 20 vertical meters of glacial ice present above the sample site. The chassis of the shearing device was extended so the cores could be constrained upon it. The face ends of the cores were ground flat using sandpaper to attain maximum contact with the end cap, suspended load and the chassis of the shearing device. Facing off the ends of the core helped to ensure even stress throughout the core from the constraining normal load by allowing the end of the core to sit flush against the chassis of the shear device with the top end
cap sitting flush on the core. The normal load was thus distributed over the entire upper and lower surfaces of the core.

### 3.5 Description of ice cores.

The cylindrical ice cores (ca. 130 x 59mm) were examined under fluorescent light in a freezer held at $-17.5^\circ$C. A physical description of the cores was conducted using the approaches described by Hubbard et. al. (1996) and Knight (1997). The physical properties of the cores were described in regard to colour of the ice, relative debris content and bubble content. Stratification of debris bands containing interstitial ice was described along with the relative angle of dip of both debris stratification and bubble elongation normal to the A axis of the cylindrical core. Laminations of debris present in the cores together with identification of regelation ice and larger particles such as pebbles were included in the description (Menzies 1995; Bennett and Glasser 1996). The degree of bubble content coupled with their morphology and spatial distribution was considered to be pertinent to the description.

![Figure 3.6](image)

**Figure 3.6** Laboratory set-up for direct-shear device. Weights are suspended from a cap on top of the core to simulate the normal load of ice at the sample site.
3.6 Direct-shear strains applied in laboratory.

Due to the recognition that varying strain-rates may have significant effects on peak strengths of frozen soils (Nickling and Bennett, 1984) the sampled basal ice cores were placed under strain-rates approximating three different orders of magnitude. Uniaxial strains of ca. 6 – 8% were applied to the ice cores along a direct-shear plane over 40 and 400 hours. Strain-rates of, 0.08mm.hr\(^{-1}\) and 0.01mm.hr\(^{-1}\) were applied to the cores. Cores that exhibited alternate clear ice and debris-laden layers were tested by shearing across the layered interface boundaries. The resultant stress-strain curves were plotted using Microsoft Excel to illustrate peak and residual strengths of each sample. The residual strength is attained when both the shear-stress and strain-rate remain at steady values.

3.7 Analysis of stress-strain curves.

Stress-strain curves generated for each of the basal materials tested were examined in relation to one another. A qualitative analysis of the varying types of basal ice and substrate material was undertaken. This analysis involved comparison of the peak strengths exhibited for each of the different strain rates applied to groups of similar samples. The analysis also included examination of the gradients of each curve as it approached peak shear strength and the subsequent values attained at the end of each strain. A summary of the peak strengths was constructed in order to simplify the comparison of the curves (Table 4.1).
4

Results

This chapter presents a synthesis of the results obtained from the stress tests conducted on cores sampled from the basal zone of the Suess Glacier, Antarctica. The chapter is divided into 3 sections. The first section contains a physical description of the basal sequence presented as a log of the basal material along with a description of cores representing clean glacial ice, clean basal ice, debris-laden stratified sediment and frozen substrate follows the log of the sequence. The second section contains amalgamated stress-strain curves for each type of basal material and a summary of the peak shear-stress attained for each rate of strain. The third section of this chapter illustrates the velocity profile recorded for the basal zone of the Suess Glacier from measurements taken after one year.

4.1 Description of basal sequence.

The basal ice sequence exposed by the excavation of a tunnel near the base of the Suess Glacier, Antarctica, in 1997 was described by Lambourne (1998), based on the approaches outlined by Lawson (1979) and Knight (1997). Thirty-one individual layers were recognised by Lambourne (1998) within the exposed sequence. A log of the sequence along with debris content is depicted by figure 4.1. Nine distinct facies were identified, eight of which were basal ice facies. The ninth facies represented the lowest extent of the clean glacier ice. Of these nine facies, five plus the bed substrate were selected for direct-shear stress tests. These consisted of; clean glacial ice, amber ice, stratified basal ice, a high debris-concentration sediment facies, clean basal ice and bed substrate. Photographs of representative cores from each of the identified basal ice types are illustrated in figure 4.2.
Figure 4.1 Log of the basal sequence exposed by a tunnel excavated at the base of the Suess Glacier indicating physical appearance and debris content. Source: Lambourne (1998).
Chapter 4. Results.

Figure 4.2 (a) Amber ice

Figure 4.2 (b) Frozen substrate material.

Figure 4.2 (c) Clean glacier ice.
Figure 4.2 (d) Clean basal ice.

Figure 4.2 (e) Stratified basal ice.

Figure 4.2 Representative cores of (a) amber ice (b) frozen substrate (c) clean glacier ice (d) clean basal ice (e) stratified basal ice.
A description of the physical properties of the cores of each different basal material is outlined below;

**Clean glacier ice.**

The ice sampled from within the clean glacier ice from heights greater than ca. 4m above the bed consisted of relatively clean, white, bubbly ice with few sediment particles (0.219g\(l^{-1}\), Lambourne, 1998). Bubbles were elongated and fairly evenly dispersed throughout the cores, dipping at ca. 10-15° to the length of the cores (Figure 4.2 (c)).

**Amber ice.**

Amber ice located below the clean glacier ice appeared translucent but discoloured, with a light amber hue (Figure 4.2 (a). Holdsworth (1974) first discovered amber ice during the comprehensive study of the Meserve Glacier, Antarctica. The term amber ice describes a layer of easily deformed, solute rich ice containing finely dispersed particles that give it an amber hue, and generally rests directly below the clean glacier ice (Holdsworth, 1974). Small elongated bubbles were evident throughout the cores, some exhibiting horizontal layering. Fine particles were dispersed evenly within the cores, along with small amounts of sand sized particles 0-2mm. The amber ice facies had low debris content (2.962g\(l^{-1}\), Lambourne, 1998).

**High debris-concentration sediment facies. (solid sediment facies).**

This sediment facies is essentially a solid facies, as described by Hubbard et al., (1996). These sediment bands consisted of a high debris concentration of fine and coarse sand cemented by interstitial ice (251.283g\(l^{-1}\), Lambourne, 1998). Bedding structures were found within the sediment bands by Fitzsimons (1998), together with fine layers of algae. No such structures were identified within the cores that were tested in this study.
Stratified sediment facies.

This facies consisted of layered bands of coarse and finer sandy sediment with interstitial ice, dipping at approximately $5^\circ$ to the length of the cores (Figure 4.2 (e)). Debris-poor, clear ice layers measuring between 1-5mm thick separated layers of sediment. Different concentrations of sediment exist within the stratified sediment facies that are often associated with the higher concentration debris facies. The average debris concentration was $45.019g.l^{-1}$ but the standard deviation was large at $39.089g.l^{-1}$ (Lambourne, 1998). Some elongated bubbles associated alongside the sediment layers were identified dipping parallel to the sediment layers.

Clean basal ice.

These cores consisted predominantly of debris-poor nearly clear ice ($0.290g.l^{-1}$, Lambourne, 1998). Some dispersed clots of elongated bubbles up to 1cm long, dipping between 45-60$^\circ$ to the long axis of the cores was observed (Figure 4.2 (d)). These layers were located in the bottom half of the sequence that is within 2 meters of the bed.

Substrate.

These cores consisted of a solid facies of coarse and fine, sandy sediment particles (Figure 4.2(b)). Debris concentration was high throughout the cores. Thin, clear, interstitial ice was identified within the pores of the sediment comprising the cores. Very small bubbles were detected within portions of the ice separating some sediment particles.

4.2 Stress-strain curves for basal cores.

Strain-rates approximating 3 orders of magnitude, $0.8mm.hr^{-1}$, $0.08mm.hr^{-1}$ and $0.01mm.hr^{-1}$ were implemented for strains between 6 and 8% of the core diameter.
Chapter 4. Results.

The shear stress was calculated from the electrical output of an ST1000 load cell (Precision Transducers Ltd.) converted by software programmed into the CR10 Datalogger to produce newtons equivalent. The shear stress is the pressure exerted along a simple shear plane by the shear device.

1 Pascal (Pa) is the unit of pressure produced by a 1 newton force applied over 1m$^2$.

$$1\text{MPa} = 10^6\text{ Pa}$$

$$\text{Shear stress} = \frac{\text{Force}}{\text{Area}}$$

$$\text{Area} = \frac{\pi}{4} \times \text{diameter}^2$$

$$\text{Units(MPa)} = \frac{\text{Newtons}}{\text{mm}^2}$$

The normal load applied to each of the cores was calculated from the weight of ice above the tunnel, (the weight of ice is 9 KN.m$^{-3}$);

ca. 20 m $\times$ 9 KN.m$^{-3}$ = 180 KN.m$^{-2}$ = (180 KPa).

Figure 4.3 illustrates the stress-strain curves generated for cores consisting of clean glacier ice found ca.4m and above from the bed. Peak shear-stress was obtained for the core strained at 0.8mm.hr$^{-1}$ and 0.01mm.hr$^{-1}$ but not for the core strained at 0.08mm.hr$^{-1}$. The core strained at 0.08mm.hr$^{-1}$ did not reach the peak shear strength within the displacement allowable by the shear device. The gradient displayed by the curve in this test leads to the belief that it would have obtained a higher value than the core strained at 0.01mm.hr$^{-1}$. The peak shear strength displayed by the clean glacier ice is obtained at the fastest strain-rate of 0.8mm.hr$^{-1}$. A small periodicity alternating in steps of ca. 0.2MPa is evident in the curve. A further periodicity is observable superimposed at ca. 1mm displacements along the curve in both the 0.8 and 0.01mm.hr$^{-1}$ strain-rates. This periodicity is also observable in some curves in the following figures although the displacements do not remain at 1mm intervals.
Figure 4.4 illustrates the stress-strain curves generated for the amber ice facies. Peak strengths exhibited by each of the cores tested decrease as the strain-rate decreases. The core strained at 0.8 mm.hr\(^{-1}\) appears to reach the residual strength of the material. The residual strength of the material is obtained when both shear-strength and strain-rate attain steady values. Peak strength can be seen to take longer to attain in the each of the curves utilising decreasing strain-rates.

Peak strengths for the high debris-concentration sediment facies are obtained at markedly different rates. Under high strain-rates the peak strength is smaller than that of the 0.08 and 0.01 mm.hr\(^{-1}\) strain-rates. Fracture of this core occurred ending the test after only 2 mm of displacement, as values of shear-strength decreased rapidly. Cores strained at 0.08 and 0.01 mm.hr\(^{-1}\) attain higher peak strengths and begin to display ductile deformation at the slower rates of strain. Again a periodicity can be observed in these curves. The peak strength of the slowest strain-rate is estimated as values exceeded the software capabilities for the CR10 data-logger. The software for the data-logger had been set to the sensitivity required for the predominantly smaller values measured in tests up to this point. Peak values for this solid sediment facies were much higher than any other facies measured. A periodicity is also discernible in two of these curves. In the core strained at 0.01 mm.hr\(^{-1}\) a small pebble was found to be present intersecting the shear plane. Fractures tangential to the pebble were observed, with small cavities beginning to form on the lee side of the pebble as the core strained over the pebble. This cavitation was observed near to the completion of the test.

Stratified sediment of lower debris concentration depicted in figure 4.6 shows a greater value of peak strength at greater strain-rates. The peak strength of the cores that were sheared across the sediment/clean ice boundaries are lower than those of the solid sediment facies (Figure 4.5). Lower strain-rates used in these tests exhibited similar peak shear strengths with a periodicity in evidence in both of the stress-strain curves.

Tests on clean basal ice shear tests also follow the trend of lower peak strength with slower rates of strain (Figure 4.7). Again the periodicity displayed in some of the other cores tested is noticeable in these curves. Both the finer saw-tooth changes
approximating 0.2MPa and the longer wavelength deviations at approximately 0.5mm displacements are discernible.

The substrate shows high peak values of shear strength with the highest attained at the greatest rate of shear (Figure 4.8). Values for this core can be seen to drop dramatically as fracture occurs along the shear plane. Peak shear-strength was not obtained for the core sheared at 0.08mm.hr$^{-1}$. Time constraints imposed upon the period available for conducting direct-shear testing did not allow this test to be repeated. Interestingly the bed substrate samples displays markedly lower peak values than those of the solid sediment facies.

### Figure 4.3

Stress-strain curves of clean glacier ice at various strain-rates. The highest peak shear-strength is attained at the highest rate of strain. The medium strain-rate test does not attain a discernible peak before the maximum displacement is reached.
Figure 4.4 Stress-strain curves for the amber ice under various strain-rates. Peak shear strength can be seen to decrease as strain-rate decreases.

Figure 4.5 Stress-strain curves of solid sediment facies under various strain-rates. Brittle fracture occurs at the high rate of strain. Ductile deformation is apparent at slower rates of strain. The medium strain-rate required estimation due to insensitivity of software supplying data-logger.
Figure 4.6 Stress-strain curves of stratified sediment facies under various strain-rates. Highest shear strain is displayed at the highest rate of strain. Smaller variations in peak shear strength are evident over each of the slower rates of strain.

Figure 4.7 Stress-strain curves of clean basal ice under various strain-rates. Peak shear strength decreases with decrease in strain-rate.
Figure 4.8 Stress-strain curves of bed samples under various strain-rates. Brittle fracture occurs at the high rate of strain as shown by the rapid decline in shear strength values. Peak strength is not attained for medium rate of strain as test was ended prematurely.

Table 4.1 summarises the results from shear tests. Highest values for peak strength was achieved by the solid sediment facies with the substrate samples collected from the bed attaining the next highest values for each rate of strain. Clean glacier ice and the stratified sediment facies displayed intermediate peak strength values. The clean basal ice and the amber ice facies exhibited the lowest peak strengths.
Table 4.1 Peak shear-strengths displayed by different types of basal and substrate material at various strain-rates.

<table>
<thead>
<tr>
<th>Strain-rate</th>
<th>0.8mm hr⁻¹</th>
<th>0.08mm hr⁻¹</th>
<th>0.01mm hr⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shear-strength Peak (MPa)</td>
<td>Peak (MPa)</td>
<td>Peak (MPa)</td>
<td>Peak (MPa)</td>
</tr>
<tr>
<td>Types of Basal Ice</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clean glacier ice</td>
<td>1.01</td>
<td>0.90</td>
<td>0.71</td>
</tr>
<tr>
<td>Amber ice</td>
<td>0.90</td>
<td>0.51</td>
<td>0.28</td>
</tr>
<tr>
<td>Solid sediment facies</td>
<td>2.29</td>
<td>3.47</td>
<td>2.61</td>
</tr>
<tr>
<td>Low debris-conc. Stratified facies</td>
<td>0.85</td>
<td>0.74</td>
<td>0.70</td>
</tr>
<tr>
<td>Clean basal ice</td>
<td>1.24</td>
<td>0.60*</td>
<td>0.64</td>
</tr>
<tr>
<td>Substrate</td>
<td>2.56</td>
<td>1.26*</td>
<td>2.10</td>
</tr>
</tbody>
</table>

* Denote samples for which peak strength was not obtained or was estimated.

4.3 Velocity profile of the basal zone of Suess Glacier.

A velocity profile of the basal zone of the Suess Glacier was constructed from displacements of wooden dowels placed alongside a plumbline one year earlier by Fitzsimons (1998) (Figure 4.9). The dowels were placed on one side of the exposed vertical sequence approximating the direction of flow in the glacier. Displacements were measured from a brass target anchored to the bed and from displacements of the dowels to the plumbline in order to calculate velocities. Highest velocity occurred in the amber ice facies 3 - 4m above the bed and directly below the clean glacial ice above. The lowest velocity is measured near the bed. An area of lower velocity in movement can be observed approximately 1m above the bed coinciding with a solid
Figure 4.9 Plumpline and vertically placed wooden dowels anchored to the vertical sequence exposed in the tunnel excavated in the base of the Suess Glacier, Antarctica. Measurements of displacement from a brass target anchored to the bed and from wooden dowels allow velocity profile to be calculated.
sediment facies (Figure 4.10). At 2.5m above the bed the velocity accelerates markedly coinciding with the transition of stratified sediment facies to that of the amber ice. For greater detail of the physical characteristics of the exposed vertical sequence refer to figure 4.1.

**Figure 4.10** Horizontal velocity profile within the basal zone of the Suess Glacier, Taylor Valley, Antarctica. Source: Fitzsimons (1998).
Chapter 5. Discussion.

5

Discussion

This chapter is divided into three sections. The first section examines the peak shear strengths of several types of basal ice, clean glacier ice and substrate material from a dry-based polar glacier when strained along a thin plane by a direct shear device. Three strain-rates are employed approximating three different orders of magnitude. Various strain-rates are used to investigate whether the rheology of different types of basal material changes from high experimentally convenient, to lower more glaciologically representative strain-rates. The second section discusses controls on the strength of basal ice, clean glacier ice and substrate with regard to their chemical and physical properties. Further inferences are drawn from the chemical and physical properties associated with each type of material tested and suggest possible explanations for some of the observations. The third section discusses the possible implications of the behaviour of dry-based glaciers along with suggested avenues for further research.

5.1 Peak shear strengths of basal material.

Peak shear strengths were ascertained for clean glacier ice, amber ice, a high debris-concentration, stratified sediment facies (solid sediment facies), a low debris-concentration, stratified facies (stratified facies), clean basal ice and substrate of the Suess Glacier. The peak strengths were deduced from stress-strain curves generated for each of the materials tested (Table 4.1). The nomenclature used for the direct shear tests that were performed on sampled cores of basal material are; 0.8 mm.hr\(^{-1}\) (high), 0.08mm.hr\(^{-1}\) (medium), and 0.01mm.hr\(^{-1}\) (slow), for strains of 6-8% (Figure 4.2).
Chapter 5. Discussion.

The results of both field and laboratory testing revealed that the solid sediment facies attained the highest peak shear strength at the medium strain-rate (3.47 MPa), followed by (2.61 MPa) at the slow strain-rate. The next highest peak strength was recorded by the frozen substrate from the glacier bed (2.56 MPa), at the high strain-rate. Under the highest rate of strain clear comparison of peak strengths were complicated by fracture along the shear plane in the solid sediment facies and the substrate material, while malleability of the other materials facilitated creep behaviour.

At the slower rates of strain the solid sediment facies exhibited the highest peak shear strengths (3.47 MPa) and (2.61 MPa), respectively. The substrate attained a peak shear strength of (2.1 MPa) at the slowest strain-rate with the medium rate test being ended prematurely before a peak strength could be ascertained. Interestingly this peak is lower than any attained by the solid sediment facies.

Clean glacier ice showed the highest peak strength of the remaining materials tested, at the high rate of strain (1.24 MPa), followed by amber ice (0.90 MPa), the stratified facies (0.85 MPa) and the clean basal ice (0.8 MPa). At the medium and slow strain-rates the stratified facies recorded peaks of 0.74 MPa and 0.70 MPa respectively, which although very similar are lower than the peak recorded for the high strain-rate. Clean basal ice and amber ice recorded the lowest peaks at the two slower rates of strain, with clean basal ice peaking at 0.71 MPa, and amber ice peaking at 0.28 MPa. Interestingly the clean glacier ice peaks between these last two values at the lowest rate of strain (0.64 MPa) reflecting the difference in peak strength exhibited by ice with changes in the rate of strain.

One feature of the tests conducted using the direct shear device over the three rates of strain, is that in virtually every set of tests involving each type of material tested, the peak strength of the material decreases with decreasing rates of strain. The gradient of the curves to peak strength exhibited for the majority of the tests of each material also tends to increase with increase in the strain-rate. Initial observations of the results of the shear tests appear to support results found by other workers investigating the strength of frozen materials, in that frozen sediments have greater shear strengths than ice. Unfortunately residual strengths were not reached for most of the tests, as
the direct shear device was unable to strain cores sufficiently to achieve a steady reading. The residual strength of a material is attained when the rate of strain and the shear strength reach steady value. Some factors that may offer at least partial explanation for the results observed in this study are outlined in the following section.

5.1.1 Periodicity in the stress-strain curves.

The stress-strain curves from the tests contain two distinct of periodicities within the overall strain of test cores. The first order of periodicity involves small steps displaying rises and falls of approximately 0.2MPa at relatively evenly spaced intervals along the curve. At first this was thought to be a consequence of flex within the load cell. The load cell is designed to vary the output of current passed through a stress-sensitive piece of metal to a data-logger. With increase or decrease in stress exerted by the shear device, the metal strip responds by varying resistance to the input current. This is accomplished by a small amount of flex that the load cell is designed to accommodate in order to vary the resistance across the piece of metal within the cell. The small steps were thought to be a consequence of pressure and then relaxation of the stress in the load cell as the ice moved. Further investigation of the literature found that other workers had also observed this periodicity involved with measurement of the creep of ice. Baker (1978), while investigating the influence of ice-crystal size on creep observed that the curves of total strain vs. time were not always smooth, but often contained numerous 1-2mm steps representing $3.1 \times 10^{-3}$ to $6.2 \times 10^{-3}$mm of strain. It is not known whether the steps are due to sticking in a seemingly well lubricated system or actual jerky displacements in the samples. In contrast, Parameswaran (1975, cited by Baker, 1978) observed similar strain steps in ice and attributed them to a stress-induced ordered region moving with dislocations through the ice lattice.

The second order of periodicity involved larger steps in the curve, recognisable at intervals of approximately 0.5mm displacement. These steps are observable to some extent in nearly every curve produced, and have been attributed tentatively to some aspect associated with the mechanism of the shear device. From the nature of the
steps it is probably some mismatch of spacing within the gearing causing a sticky spot every rotation of the gearing. The rationale behind this is that different periods of time are taken for the shear device to accomplish 1mm of displacement, so if this periodicity is associated with stress rather than strain then it should be discernible at varying frequencies.

5.2 Controls on the strength of ice and frozen debris.

The mechanical strength of basal ice, and frozen debris may be affected by several factors that may operate independently or in combination to produce changes in rheology of these materials. These factors include the chemical composition of the ice and the distribution of impurities throughout the basal zone. The physical structure of facies regarding size of ice and debris grains coupled with the relative content of ice and sediment in each of the materials also have implications for structural strength. Some of the controls on the strength of basal material are outlined in the following section.

5.2.1 The strength and structure of ice.

The strength of ice is affected by its temperature, grain size and orientation, structural anisotropy and by the sediment particles and other inclusions within it. As ice deforms, the textural and structural changes that occur may influence its subsequent behaviour significantly (Theakstone, 1979). The formation of ice crystals is dependent on pressure and temperature along with the chemical composition of the water being frozen. The firnification of snow accounts for the majority of glacier ice, which displays a coarser polycrystalline fabric than that of basal ice (Hubbard and Sharp, 1989). Butkovitch and Landauer (1958, 1960) found that for randomly oriented polycrystalline ice, in general, the samples with large crystals tended to deform more rapidly than those with smaller crystals. Their studies employing low stresses, showed that the deformation rate of ice with an average crystal size of 10 to 20mm was about four to five times as great as that of ice with an average grain size of
about 3mm (Baker, 1978). As a result of stress, grain boundaries in ice will be subjected to localised shear forces or shear couples. Shear couples are associated with bending adjacent to the boundary and the operation of recovery processes in the mantle of the ice crystal grain (Wilson and Russel-Head, 1982). From experiments conducted on laboratory manufactured ice Wilson (1986) found that these movements of grains along their boundaries, combined with lateral stress, leads to the formation of subgrains and eventually high angle boundaries that immediately undergo migration recrystallization. The newly nucleated grains can also undergo further deformation (Figure 5.1). Once recrystallization nuclei are established in the ice aggregate, subsequent grain-growth results in a rapid intensification of the preferred orientation pattern. Preferred-orientation development is associated with grain reduction mechanisms, which involve solid state processes of dynamic recovery and recrystallization. The deformation and recrystallization structures, grain-growth features and preferred orientations are developed more rapidly in the finer–grained ice aggregate (Wilson and Russel-Head, 1982). Smaller ice crystals formed with preferred orientations may have significant implications for strength and movement of basal ice.

Figure 5.1 Sketch of recrystallization sequence: (a) the shape of the original grain C (b) grain boundary adjustments and development of undulose extinction (c) relict core C with new grains nucleated and replacing pre-existing grain in region of greatest lattice rotation. (d) subsequent growth of new grain existing at stage C at expense of deformed grain C. Source: Wilson (1986).

Grain Boundaries are obstacles to dislocation motion and resist penetration by moving (Armstrong, 1970 cited in Baker, 1978). Hence, dislocations have a tendency to pile-
up at crystal boundaries. These pile-ups act as stress concentrators and can cause plastic flow to begin in the neighbouring grain due to deformation from the transference of stress.

The studies by Butkovitch and Landauer (1958, 1960) outlined above, were conducted on ice crystals prepared in the laboratory from de-ionised water and therefore contained few impurities. The basal ice used in the present study however, was found to contain significant amounts of solutes. High solute concentration in the amber ice, clean basal ice and the stratified sediment facies was derived at least in part from the underlying substrate and meteoric origin (Lambourne, 1998; Hooker, 1998). Some solutes have been found to decrease the mechanical strength of ice (Jones and Glen, 1968). Figure 5.2 shows a polarised light image of the contrasting crystal structure of the amber ice facies underlying the clean glacier ice. A small layer of debris separates the two facies.

5.2.2 The effect of chemical impurities and inclusions in ice.

The driving force for crystal growth is surface tension (Smith, 1948, cited in Baker, 1978), therefore inclusions and solutes may play an important role in determining the final grain-size achieved during crystalline growth, due to their effects on surface tension. Because an inclusion will tend to become attached to a grain boundary, rather than be included in the crystal, a rigid and insoluble inclusion will effectively anchor the boundary locally (Van Vlack, 1965, cited in Baker, 1978). If a moving crystal boundary encounters such an inclusion it will locally cling to it causing an indentation in the surface. Before a boundary can move beyond a dispersed particle, the total boundary area must be increased and the radius of the curvature reversed locally. If a sufficient number of inclusions are present, grain boundaries can be effectively pinned and crystal growth is thus limited (Baker, 1978).

A liquid like layer (as discussed in greater detail in the next section) has been found to exist at the grain boundaries of ice crystals and foreign solids (Gilpin, 1979, in Cuffey et al. 1999). The intersections of these layers at crystal boundaries may act as veins
such as those measured in cores from Blue Glacier. These veins had a measured average diameter of 25\(\mu\)m (Raymond and Harrison, 1975, in Paterson, 1994). It is thought that solutes may interrupt crystal structure, or cause larger formation of the liquid layer as the concentration of solutes is greater between the crystal grains (Holdsworth, 1974; Cuffey et al. 1999). Increase in solutes tends to depress the melting point and induce this film growth (Beaglehole and Wilson, 1994). This mechanism may aid the movement of solutes through the crystal lattice causing implications for strength of the crystal fabric. The highest rate of strain measured in the vertical profile exposed in the basal zone of the Suess Glacier was found in the amber ice facies (Fitzsimons et al. 1999, forthcoming). Solute content of the amber ice was also high, coupled with a low but present fine-debris content that was well dispersed throughout the facies.

In studies concerning the strength of ice crystals doped with chemical impurities, Jones and Glen (1968) found that certain chemicals, such as hydrofluoric acid (HF), affect the behaviour of ice crystals, causing them to weaken significantly. From their studies, they suggested that it is possible that the mechanism responsible for the dielectric behaviour that causes reorientation of point defects in the lattice is also responsible for the mechanical softness of HF-doped crystals. Other chemicals however, were found to harden the structure of ice crystals (Jones and Glen, 1968). Solid state diffusion of solutes through the crystal lattice suggested by Holdsworth (1974) supports earlier work by Jones and Glen (1968), who suggest that when some chemicals are dissolved in ice, it is thought to that they enter the ice lattice substitutionally (Gränicher, 1963, in Jones and Glen, 1968). By doing so, each HF molecule, since it contains only one hydrogen atom, introduces an L defect on one of its neighbouring bonds, compromising the strength of the crystal lattice (Jones and Glen, 1968). It has also been suggested that the weakness of Pleistocene ice in comparison to Holocene ice on the Barnes Ice Cap, may be due to the effect of microparticles and chemical impurities that are known to be present in this ice (Hooke, 1998).
Figure 5.2 Photo of ice crystal fabric as viewed through polarised and transmitted light. (a) shows the amber ice/clean glacier ice interface separated by a fine layer of debris, shown by arrow (b) the relative crystal sizes of the amber ice with the smaller crystals at bottom, and the larger crystals of the clean glacier ice (top).
Photographs courtesy of S.J. Fitzsimons.
5.2.3 Interfacial films.

Interfacial films exist due to the reduction of the chemical potential of water very close to the surface of a foreign solid (Gilpin, 1976 in Cuffey et al., 1999). The chemical potential is lowered due to the solubility of parts of the mineral that is in close contact with the liquid like layer close to the ice crystal. This depresses the melting point of ice as an inverse function of the film thickness. As the thickness of the layer increases the solute concentration is decreased. The impurity concentration in grain boundaries and veins in polycrystalline ice is several orders of magnitude higher than that of the crystal lattice (Holdsworth, 1970; Alley, et al. 1986). As ice freezes from water, such as in regelation cycles, it rejects solutes at a very high efficiency (Gross, et al. 1975). Interfacial films will generally intersect grain boundaries, allowing solutes to exchange within the vein network between ice crystals. Chemical exchanges will take place over areas with differing solute concentrations, moving in order to establish equilibrium throughout the film. Because of this, and the refreezing accompanying regelation, the solute content of the interfacial films will be generally several orders of magnitude higher than the bulk concentration of the adjacent ice, and this makes for generally thicker films (Holdsworth, 1974; Cuffey et al. 1999). The presence of interfacial films between ice crystals and foreign solids may also have implications for the movement of ice in the basal zone of dry-based glaciers. Such films have been postulated to exist at temperatures well below 0°C, especially where solute content is high (Cuffey et al. 1999). Again, the high solute content existing within the basal facies of the glacier surveyed in this study may have significant implications for movement within these facies and provide a means for regelation at temperatures below 0°C.

5.2.4 Structural defects in the ice fabric.

From flex tests involving fracture strength experiments in ice, Gagnon and Gammon (1995) have found the flexural strength of glacier ice is dependent on bubble content. They suggested that intragranular air bubbles increase ice strength by accommodating greater strain and reducing stress concentrations through dissipation. Grain boundary
bubbles may act to weaken the ice by taking up space that would otherwise be filled by ice. Increasing grain boundary recrystallization and migration with depth tends to decrease the proportion of intergranular bubbles and may help to explain the trend for increasing resistance to fracture through the firn and meteoric ice. Similarly, the concentration of inclusions at grain boundaries in marine ice may help explain its apparent weakness as explained above (Rist et al. 1996). Data obtained from the present study also indicate that the peak strength of clean glacier ice is greater than that of clean basal ice, which exhibits less bubble content than that of the overlying glacier ice tested (Table 4.1). Clean basal ice has been shown to contain a greater solute content (Lambourne, 1998), so it may be possible that the solute content of ice is a greater controlling factor on structural strength than that of the inclusion of bubbles.

5.2.5 Strength of frozen debris and substrate.

The physical characteristics of material in the basal zone of dry-based glaciers include debris-laden stratified ice and high debris-concentration, solid sediment facies. The content of ice within the sediment of such facies affects the mechanical strength associated with the facies through the relationship of contacts between particles or between the contact strength of ice and foreign solids. The strength of frozen fine-grained materials, under constant normal load, strain-rate, and temperature increases with increasing ice content until reaching a threshold after which the strength decreases to some limit (Goughnor and Andersland, 1968). This threshold corresponds to the point where the ice content of the sample is just sufficient to fill the available void space. As ice content increases beyond this threshold, a point is reached when the strength of the frozen sample becomes more dependent on the strength of the ice than the strength of the soil matrix. This point, termed the critical ice content, is indicated by the break of slope on the plot of peak strength against ice content (Goughnor and Andersland, 1968).

The changes in shear strength with increasing ice content are attributed directly to changes in internal friction and the cohesive effects of pore ice. Grain-bridging
(Figure 5.3) has been suggested as a mechanism for increasing the internal friction and hence apparent strength of debris-laden ice above the critical threshold (Hooke 1998).

![Diagram](image)

**Figure 5.3** (a) A grain bridge, formed by nearly coaxial alignment of several grains in a deforming granular medium. The bridge may fail by (b) fracture of a grain or (c) slip between grains. (d) Stresses at contacts between grains are reduced when additional particles occupy pore space. Heavy arrows show shear stress applied to material, $\tau$, and component of this stress along grain bridge, $\sigma$. Source: Hooke (1998) modified from Hooke and Iverson (1995).

The continuation of creep in frozen granular materials depends on two predominant factors, changing cohesion and the changing angle of internal friction. As the ice content is reduced in areas of stress due to dynamic recrystallization within pore spaces, particles are placed in closer contact. Thus internal friction is increased and molecular cohesion is increased. If the strengthening process (particle rearrangement and bond creation) exceeds the weakening process (ice reorientation and bond creation), then the creep of the frozen granular material is reduced (Nickling and Bennett, 1984). From shear tests performed on coarse granular material Nickling and Bennett (1984) reported that the stress strain curves of the 31 tests indicated that samples with higher ice contents tended to reach peak strength with less displacement during shear. Moreover, the difference between peak strength and residual strength was the lowest for pure polycrystalline ice and highest for ice saturated samples. The results of the present study show that the peak strength for both frozen substrate and
solid sediment facies were far in excess of any of the other materials tested by direct shear. The difference between peak strength and residual strength could not be ascertained due to the small strains attainable by the direct shear device.

When a granular material accumulates gradually, it compacts under its own weight. Materials that accumulate in this fashion are called ‘normally consolidated’. An ‘over-consolidated’ material is formed by removal of part of the overburden previously resting upon the material or added stress to the material (Hooke, 1998). The debris of the solid sediment facies exposed in the Suess Glacier would therefore be considered over-consolidated if it has been subjected to stress from movement in the basal zone that exceeds that of just the normal stress from the overburden of glacier ice alone. Depending on the granulometry (size distribution of particles) of a material, it may either compact slightly or dilate (move apart) during movement. Dilation occurs in consolidated materials when they begin to deform, in order for particle grains to move past one another (Hooke, 1998). The movement of particles past one another causes friction, if there is particle to particle contact (Figure 5.4). With friction comes the generation of heat at the particle boundaries. The stresses caused by shifting debris grains may promote pressure melting of the ice and refreezing in areas of lower pressure (Nickling and Bennett, 1984). Lawson (1996) employing rapid strain-rates in uniaxial compression tests observed pressure melting at temperatures down to -25°C, even though the pressures were somewhat unrealistic in terms of normal glacial movement. Nevertheless, if sufficient pressure were to occur at localised points in the basal fabric, such as between particle points, pressure melting may be expected to occur in glaciers. The transfer of water in this manner provides the opportunity for granular adjustment or creep in granular materials (Nickling and Bennett, 1984). The reformed crystals left in the wake of pressure melting are generally smaller and have a preferred orientation to allow dissipation of planar stresses. These processes may cause a local weakening of the material that is being deformed.
Temperature and strain-rate can also have a significant effect on the peak strength of frozen soils (Tsytovich, 1973 cited in Nickling and Bennett, 1984). These variables are considered in the sections below.

5.2.6 Temperature effects on ice.

The effects of temperature have been discussed by several researchers in work on the structural strength of ice (Wilson, 1982; Lawson, 1996; Wang and Warner, 1998). The strain rate of warmer ice has been shown to be greater than that of colder ice with explanation of this behaviour in part, being attributed to grain-boundary sliding due to the increase of liquid water at the grain boundaries (Paterson, 1994). Controls on crystal growth have also been attributed to temperature. At colder temperatures for example, $-10^\circ$ C (Wilson, 1982a), and $-115^\circ$ C (Durham et al. 1983), the grain size observed is appreciably finer than at $-1^\circ$ C, even when samples are subjected to comparable stresses. This observation has also been recorded by other workers (Baker, 1978), and suggests that temperature may be critical in controlling grain size in the deformed state (Wilson, 1986). It generally agreed that one of the major factors
that influence the extent of grain growth in ice is temperature, with marked grain size increases occurring at higher temperatures.

High temperature and relatively high fluctuating stresses initiate dynamic recrystallization, with fast grain boundary migration, and the formation of a multi-maxima fabric which adapts to the local state of stress (Meysonnier and Philip, 1996). The process of intragranular nucleation has been termed migration recrystallization in minerals (Tungatt and Humphreys, 1981, in Wilson, 1986). One of the most important process occurring in the plastic deformation of ice at high temperatures are the phenomena of recrystallization and the development of a non-random fabric (Wilson, 1986). According to Duval and Castelnau (1995) this “migration–recrystallization” regime begins when the temperature rises to reach -12°C. Therefore, temperature dependence may place some of the greatest controls on movement of ice in the basal zone of dry-based glaciers. This may be coupled with the presence of solutes that also depress the melting point of ice and allow the presence of liquid near the sliding interface.

In uniaxial compression tests (UCS) both Lawson (1996) and Wang and Warner (1998) have found that differences in temperature have changed the behaviour of ice and ice-laden sediment. Wang and Warner (1998) report that for temperature–dependent ice rheology, the shear strain overtakes the compressive strain at strains that are often in excess of 50%. Wang and Warner (1998) observed that enhancement develops over a relatively narrow transition zone, indicating that even a conservative estimate of the shear needed to establish a fully developed compatible anisotropic crystal fabric produces striking modifications to strain–rate profiles. Lawson (1996) found that the maximum UCS for debris-laden basal ice was smaller than that clean glacier ice at field temperatures of approximately -5°C. This changed with decrease in temperature, until at -25°C the maximum UCS of debris-laden ice was significantly greater than that of clean glacier ice. The present study, employing direct shear tests has found that the peak shear strength of debris-laden ice and frozen substrate is approximately twice that of glean glacier ice at -17°C (Table 4.1).
5.2.7. Varying the rates of strain on frozen ice and debris.

Strain history of the ice is a contributing factor to the size and orientation of ice crystals. In temperate glaciers, this is a result of the processes operating in the basal zone of glaciers where the flow of ice over protuberances in the bed or other obstacles may give rise to pressure melting, regelation and crystal reorientation (Hubbard and Sharp, 1989). In dry-based glaciers this may be predominantly due to internal deformation of ice within the glacier. Deformation tests on polycrystalline ice with a developed crystal-orientation fabric have shown that creep rate is enhanced when the preferred orientation fabric is compatible with the stress configuration (Wang and Warner, 1998). Smaller crystals produced by dynamic recrystallization that are more favourably oriented for glide in the direction of the stress, causes increased strain-rate in tertiary creep. This process is also referred to as mylonitization, and occurs as the buttressed larger crystals of slow moving ice are seemingly crushed by faster moving ice that causes a compressive stress field within the slower moving ice. Hence the process of mylonitization of larger ice crystals may cause easier formation of shear planes within the ice fabric (Figure 5.5, Tison et al. 1993).

![Figure 5.5 Sketch showing movement of faster ice over slower moving, buttressed ice near a glacier margin. Mylonitization of ice occurs near the principal shear plane. Source: Tison et al. (1993).](image-url)
The difficulty of appealing to shear planes as agents of entrainment is that, for two dimensional flow, the down glacier ice must be stationary whilst the sediment layer is incorporated, which would require very strange patterns of movement to incorporate a multi–layered debris sequence. However, this difficulty is overcome somewhat when considering the flow of ice over a realistic three-dimensional bed. When a glacier flows over a hummock, which is not elongated transverse to flow, the ice moving around the flanks of the hummock often flows into the lee-side position more rapidly than that moving over the summit. Thus, sediment masses torn from the summit, or transported over the summit of a hummock, may rise above the glacier bed by flow of flanking ice beneath it. (Boulton, 1979).

Multiplication of dislocations and formation of micro-cracks may also contribute to weakening of the crystal fabric (Paterson, 1981). These factors may cause greater movement in finer grained ice fabrics with high solute contents, even though larger ice crystals have been shown to be somewhat weaker than smaller ice crystals (Baker, 1978). The amber ice facies tested in this study contained fine grains of ice, <1mm (Figure 5.2), a high solute content and finely dispersed sediment. At the lowest rate of strain the amber ice facies had the lowest peak strength (0.28MPa), however at the highest rate of strain the stratified sediment facies exhibited the lowest peak strength (0.854MPa) compared with 0.90MPa recorded for the amber ice. Considering the strain history of large strains associated with the amber ice facies (Figure 4.9), it would be expected to exhibit the lowest peak strength characteristics. The stratified sediment facies also displayed the least variability between the three orders of strain-rate, (0.854, 0.74, and 0.70MPa from high to slow respectively). The major physical difference between these two facies is their relative debris contents. The effects of strain on the two facies bear consideration with regard to distribution of strain throughout the facies.

In an effort to identify a mechanism for abrasion in dry-based glaciers, Drewry (1986) illustrated that the theoretical velocity profile of ice movement within the basal zone would cause an angular velocity of force to work on a clast or particle within the moving ice (Figure 5.6). This will cause rotation of clasts and could also help to rotate and compact sediment particles into the upper and lower sides of solid sediment
facies. This may be accomplished if ice was to move by ductile deformation along either side of the sediment facies, at a greater rate than the movement of the sediment facies itself. The velocity profile recorded for the Suess Glacier depicts greater strains above and below the solid sediment facies (Figure 4.9).

Figure 5.6 Rotation and net forward motion of a particle embedded in the basal layer of a dry-based glacier frozen to bedrock. The vertical velocity profile is shown (left) with zero forward movement at the bed, but with infinite horizontal flow, due to creep, increasing with distance above the bed (hence a shear zone). The clast experiences angular velocity ($\omega_c$) and forward velocity ($U_p$). Source: Drewry (1986).

Strain localisation may be due to the mechanical segregation of small grains, which support small forces, rotate and therefore acted as low-friction bearings between large particles. A second possibility is that grain bridges that develop in granular materials distribute shear strain. The maximum decrease in the internal friction angle may occur with dilation due to particles moving apart, and each shearing episode may therefore cause renewed dilation regardless of the total shear strain that the till has
experienced. The consequent dilatant hardening should cause the locus of strain to shift elsewhere (Schulsen et al. 1997). In the shear tests conducted in the present study, using slower rates of strain caused the peak shear strength recorded in each of the materials to decrease. This is in general agreement with other work that have employed various strain-rates on deforming till (Nickling and Bennett, 1984). The slower rates of strain may allow recrystallization to take place more easily and hence facilitate enhanced ductile deformation.

In uniaxial compressive shear tests performed on plate-like columns of salt-water ice, Schulsen et al. (1997) found that at lower deformation rates, cracks in the ice tended to nucleate but did not propagate, correspondingly the material exhibited macroscopically ductile behaviour. At higher rates, the cracks tended to grow and interact; macroscopic faults eventually developed and brittle failure ensued. This behaviour was found in the substrate and solid sediment facies tested in the present study, which both exhibited brittle failure at the high rate of strain and ductile deformation at the lower rates. The remainder of the tested materials however, exhibited ductile behaviour at the same high rate of strain that caused brittle failure in the substrate and solid sediment facies. Schulsen et al. (1997) observed that fewer across-column cracks formed at higher strain rates than at lower strain rates, and suggested that presumably, they were stabilised at the lower strain rate by crack-tip creep. Schulsen et al. (1997) also observed that cracks were preceded by clear development of the decohesive zone on the grain boundaries. These tests however, employed unconstrained uniaxial compression whereas the tests conducted in this study utilised direct shear. Caution should be observed before drawing any direct inferences from this work, other than that similar behaviour was found in some materials that were in themselves dissimilar.

In their study on the compression of saline ice columns, Rist et al. (1996) suggested that, although loading is compressive, crack nucleation still occurs as a result of tensile mismatch of stresses at grain boundaries and that the crack size is in the order of grain diameter. Under uniaxial compression, cracks tend to form with their long axes in the direction of maximum compression, parallel to the specimen axis and the stress at which these new cracks start to form can be used to infer the effective tensile-fracture toughness $K_Q$. (Rist et al. 1996). In the present study small cracks were
observed in some of the substrate and solid sediment facies cores, especially where larger particles of sediment were pushed along the shear plane. These cracks formed tangential to the shear plane and were probably due to the laterally unconstrained nature of the cores. This lack of lateral constraint was caused by a gap that was due to the diameter of the core measuring less than the diameter of the holes in the moving and stationary platens of the direct shear device. Schulsen et al. (1997) attributed such cracks to grain boundary sliding, suggesting that this assumes that the boundaries act like a smaller order of cracks. They suggested that upon sliding, they concentrate stress at microstructural impediments, such as ledges or steps. When the stress there is high enough, the tensile component nucleates cracks, which then tend to run across columns, traversing the grain (presumably) where the stress is more highly concentrated. A relatively irregular array of impediments would then account for the more or less uniformly spaced cracks within the sets. Cracks along the shear plane of the solid sediment facies core tested in the present study were observed to be uniformly spaced also. Rist et al. (1996) suggests that it is difficult to make direct comparisons to other work because of the strong dependence on ice type, porosity, impurity content and loading regime.

5.3 Implications for movement in dry-based glaciers.

The implications for glacial movement from results obtained in this study are complicated by the interaction of several factors whose cause and effect relationship remain clouded by uncertainty. These factors include; ice crystal size and orientation, solute content, debris content, size, and orientation, temperature and the previous strain history associated with each type of basal material.

On one hand, the solute content of basal ice alters overall strength characteristics associated with several of the ice and debris-laden ice facies identified in this and other studies (eg. Jones and Glen, 1968; Holdsworth, 1970). On the other hand, work by Goughnour and Andersland (1968), Hooke et al. (1972), Nickling and Bennett, (1984) and Iverson et al. (1998) have indicated the strong dependence of debris content within ice, as being a major, non-linear, controlling factor on the strength
associated with frozen material. The possibility of a non-linear relationship between temperature the strength on frozen materials has been implied by Wilson (1982), Lawson (1996) and Wang and Warner (1998). Rist et al. (1996), Schulsen et al. (1997) and Hambrey et al. (1999) have all discussed the implications of rates of strain on the fabric of ice and frozen debris. Each of these factors places some control on the size, orientation and subsequent movement of ice crystals in the fabric that either makes up or consolidates ice and debris facies found within the basal zone of dry-based glaciers. However, several of these variables appear to be interrelated thus the cause and effect relationship between the rate of strain and the strength of the material being tested cannot be defined without first defining each of these parameters.

Clearly the mechanical strength properties of ice in the basal zone of a glacier cannot be specified without first characterising such other properties as chemical impurity or particle inclusion content, texture and fabric. The flow law at any part of a glacier therefore must be a function of location, defined by the local crystallographic, fabric and stress situation hence the behaviour of ice close to the bed of a glacier should not be expected to conform to a simple flow law (Theakstone, 1979).

One simple observation that may be made from this study is that the substrate and solid facies has greater shear strength than those of the clean basal ice and the overlying amber ice and clean glacier ice. In most cases this has been recorded as greater by a factor of two or more. Therefore, it could be expected that preferential deformation would probably take place in the clean basal ice, amber ice and clean glacier ice. This deformation will be due to strain concentrations that can be expected in the basal zone, such as those in evidence in the velocity profile recorded for the Suess Glacier (Figure 4.9). It should be noted here that the relative mixtures of ice and debris, solute content and temperature are important factors controlling the shear strength, as mentioned above. Another source of variation in the strength of material from the Suess Glacier is the presence of cavities and which represent localised structural weaknesses. Ice layers or lenses have been observed in the substrate along with thin mud layers that may also compromise the structural strength of this material (Fitzsimons et al. 1999). So with this in mind, it is possible that although preferential movement may take place in the facies comprised predominantly of ice, that debris-laden ice (above saturation) and unconsolidated substrate (containing lenses of ice)
may also deform, albeit at a slower rate. The mechanism described here for deformation of the substrate of a dry-based glacier is similar to that described by Echelmeyer and Zhongxiang (1987) that described the deformation of ice-laden drift (ice concentration between 30-100%), at subfreezing temperatures (-5°C).

Measurements made of a solid band of sediment with comparison to well defined flow lines in the Suess Glacier show that the sediment band is feeding the moraine that has developed on the glacier margin (Fitzsimons et al. 1999). This moraine was previously thought to be a thrust block moraine, derived from thrusting and stacking of blocks of proglacial lacustrine sediments (Fitzsimons, 1996). Nevertheless, the moraine constitutes evidence for the evacuation and transportation of unconsolidated sediment from the bed of the glacier. The method of transportation is that of the increased strain of clean and stratified basal ice on either side of the sediment band coupled with ductile deformation parting the stratified basal ice above and below the band. This mechanism may cause a form of plug flow of the sediment band within the basal zone until it reaches the glacial margin where it is deposited in the moraines described by Fitzsimons (1996).

It is evident from these observations that in some cases dry-based glaciers can effectively erode unconsolidated substrates on which they rest. The direct shear tests used in this study do not offer any evidence for the mechanism by which this sediment is entrained but when combined with the velocity profile (Figure 4.9) they do offer an insight into the method by which they may be transported within the basal zone.

5.4 Main problems and shortcomings in this research.

The objective of this research was to examine the strength characteristics of several different types of basal material grouped into broad categories. These categories were clean glacier ice, amber ice, low debris-concentration stratified facies (stratified facies), high debris-concentration stratified sediment facies (solid sediment facies) and substrate material. From the information presented by pertinent literature, it appears that these broad categories do not sufficiently parameterise the materials
tested in order to make inferences as to why each exhibited the strength characteristics observed. A cause and effect relationship requires that the solute and debris content coupled with ice structure and orientation of the crystalline fabric be accurately deduced for each of the cores so that the effects of the variation of strain-rate has more meaning. The inability of the direct shear device to strain cores over sufficient distances so as to ascertain the residual strengths of each material at each of the strain-rates further inhibited the inference of meaningful relationships between peak and residual strengths of each of the cores tested.

Another limitation of the direct shear tests used in this study is that they are conducted on thin shear planes that do not account for the variation (anisotropy) of the layers of material that comprise the basal zone of dry-based glaciers. Hence structural differences such as bubble content or chemical and debris content are not incorporated in the tests. Part of this research was conducted as a field study, utilising the excavated tunnel as an onsite laboratory. As can be appreciated, the tunnel environment is not 100% laboratory conditions, as temperature (-10°C) which can affect the properties of ice could not be held constant. The thermal condition set-up in the laboratory was held near a constant –17.5°C. Therefore the data from this study could not be held as unequivocal.

Due to time constraints and the length of time taken to process and test cores at each of the strain-rates, not many cores were tested within the time period of the study. This reduces the ability to account for sample variability. Hence, it is hard to make solid inferences from the study, as the physical and chemical characteristics of each of the cores are not identical. With the length of time taken to process the number of cores necessary to make meaningful inferences, there is a greater likelihood of mechanical breakdown or disturbance of experiments by power failure or unintentional tampering. Each of these problems were encountered over the course of this study, causing temperature and strain fluctuations in ice cores, hence rendering some cores useless. Such setbacks meant that fewer cores could be tested due to the limited time and material available for testing. For example, we ran out of englacial cores.
Software that was designed to convert power input from the load cell to the CR10 Data-logger into a newtons equivalent was unable to record the complete range of sensitivity required for testing the stronger high debris-concentration sediments. Once a value of 6999 newtons was reached, subsequent higher values were not recorded. Hence, one of the peak strength readings for this study needed to be estimated. Uncertainty remains as to whether the crystal structure of the cores change from having been unconstrained for more than one year in a freezer at Otago University and then placed under a constraining load once again. This may have an effect on the structural strength of the cores when subsequently tested.

5.5 Avenues for further investigation through questions arising from this study.

Due to the related nature of many factors involved in this study, several questions have arisen that may form the basis for future investigation of basal material.

- At what strain-rate does the brittle fracture of debris-laden ice change to ductile deformation, and is this related to the relative debris content?
- Changes in velocity throughout the vertical extent of facies in the basal zone cause angular velocity and hence rotation in sediment grains. Can this cause packing of grains in solid facies, as once a grain rotates it may be able to move into a better place to lock between other grains on both the upper and lower sides of a solid sediment facies?
- Can grain-bridging cause pressure-melting leading to lubrication of small shear-planes from increases in water pressure and thus allow regelation to occur within stratified facies, in turn aiding ductile deformation?
- What is the form of the relationship between temperature and the strength of ice, and at what temperature does debris-laden ice become stronger than clean ice?
- Clean and stratified basal ice has been associated with high solute contents that may compromise their strength. Can this enable greater velocities of ductile flow above and below sediment parcels, causing cavitation on the lee side of the parcel and contribute to plug flow movement of the parcel?
The objectives of this study were to examine the peak strength characteristics of basal ice and substrate material that have been subjected to different strain-rates and to observe any changes in rheology that occur in the materials at each of the strain-rates. The conclusions drawn from the results and interpretations of direct-shear tests on basal material in this study in regard to the hypotheses stated in section 2.9 are:

- The peak shear-strengths of the materials tend to decrease as strain-rate decreases. This may be related to the greater time available for the mylonitization and subsequent recrystallization of larger ice crystals. These findings are consistent with previous work involving shear tests on ice and frozen soils.

- The peak shear-strength of the solid facies and frozen substrate material is greater than that of the other basal materials tested at –17.5°C. It is likely that this may attributed to an increase in friction between particles and/or grain bridging between sediment particles. The greater shear-strength found in solid facies and frozen substrate is coupled with the increased viscosity of pore ice and contact strength of ice/sediment at low temperatures.

- The lowest peak strengths were attained by basal ice that has the highest solute content and/or dispersed fine sediment that contain interstitial ice above the saturation point (25%). High solute concentration along with low sediment content within the ice lattice appears to decrease the peak shear-strength of basal ice.
• Differences in the strength of basal material will cause preferential motion within the clean basal ice and amber ice facies within the basal zone of dry-based glaciers. This is one of the driving mechanisms behind ductile folding and deformation of sediment.

• Many factors controlling the overall strength of basal ice act in combination with each other. This study has shown that many questions remain to be answered concerning the cause and effect relationship between composition, temperature and strength of basal ice and frozen sediment.

• This form of investigation does not appear to provide clear evidence for the mechanisms involved in the erosion and entrainment of sediment identified from observations beneath a dry-based glacier.
References.


