3. The development and effect of microscopic deformational structures

3.1 Introduction

Because glaciologists are interested in very large scale models they tend to generalise proglacial and subglacial sediments as homogenous continua. In other subject areas, however, small scale structures in the sediment have been shown to have a profound effect on the dynamics of the material as a whole. Such structures form through the sediments' grains varying in size and orientation. Two properties are considered to alter with the development of grain orientation in sediment; the sediment's drainage and its strength. In continuum models these two have been related via water pressure (which is often controlled by drainage) for example in the classic Mohr-Coulomb equation for sediment strength,

$$t^{*} = c + P_e \tan f \qquad \textit{...Equation 3.1}$$

where t^{*} is the maximum shear stress supported by the sediment (or 'yield strength'), c is the cohesion of the sediment, P_e is the effective pressure on the sediment, and f is the angle of internal friction for the sediment. In glaciology P_e is usually the ice pressure minus the pore fluid pressure in the sediment. Cohesion is the deviatoric stress that the sediment supports at zero effective pressure. This is often visualised as the force necessary to overcome the electrostatic attraction of sediment grains. The internal friction of the sediment is calculated from the rate at which the yield strength of the sediment changes with effective pressure (for a review see, Price and Cosgrove, 1990). It is often visualised as being determined by the force needed to mechanically move grains up and over each other (Hooke and Iverson, 1995).

Creep prior to the sediment yielding is usually ignored. Following yield, strain rate is usually modelled through equations relating it to the shear stress and effective pressure. Therefore, because they are part of the effective pressure, the pore-fluid pressure and ice overburden pressure affect the yield strength of the sediment *and* its post-yield strain.

The pore fluid pressure of homogeneous sediment is dependent on the ability of the sediment to drain. For homogeneous sediments drainage is usually modelled in two dimensional terms by Darcy's law,

$$\frac{Q}{A} = K\frac{\mathrm{d}H}{\mathrm{d}l} \qquad \textit{...Equation 3.2}$$

where Q is the discharge, A is the drainage area transverse to flow, $\mathrm{d}H/\mathrm{d}l$ is the fluid pressure gradient, and K is a material-specific constant known as the hydraulic conductivity. Hydraulic conductivity varies from 10^{-8} to 10^{-16} for clay sediments. Discharge may also be represented in terms of permeability (k), which is related to hydraulic conductivity by

$$K = \frac{krg}{h} \qquad \textit{...Equation 3.3}$$

where r is the density of water, g is the gravitational acceleration, and h is the viscosity of water. The permeability is chiefly dependent on the size of pores in the sediment. In turn, the pore size is partially determined by the effective pressure on the sediment, so that drainage reduces the pore size and reduces the permeability. Pore space is often represented in terms of the ratio between pore volume and the volume of the grains (the 'void ratio'). Mesri and Olson (1971) showed that permeability is a log-linear function of void ratio from ratios of 0.7 to 4.0. Permeability is likely to also be affected by mineralogy, grain size, cementation, water chemistry, and particle packing (Brown *et al*., in press). However, Mesri and Olson's work gives some indication of the considered importance of small porosity changes to fluid flow and, subsequently, the effective pressure acting on the sediment and the sediment's strain response.

Changes in permeability may also occur through changes in the path length or 'tortuosity' of the sediment (Arch, 1988), and changes in the electrostatic response of sediment grains (Mesri and Olson, 1971). The tortuosity is defined as the true path length across a chosen region divided by the length across the region in the absence of sediment grains. This will be dependant on the length-to-width ratio of the sediment grains and their alignment. Arch and Maltman (Arch, 1988; Arch and Maltman, 1990) give the following variant of the Kozeny-

Carman model of permeability for representing the control of permeability by tortuosity, pore shape, and size,

$$k = \frac{Cm^2}{BT^2} \qquad \text{...Equation 3.4}$$

where k is permeability (m^2), B is a shape factor, as is C which relates to flow at an angle to pores treated as pipes of a shape determined by B, m is the ratio of the pipe volume to its surface area in contact with the fluid, and T is the tortuosity.

This chapter examines work on the development of microstructures and their effect on the ideal models outlined above. Much of this information will be used in a glacial context in future chapters, however, a number of novel conclusions on the causes and effects of glacial microstructures are presented at the end of this review. The treatment of the structures will progress from depositional forms, to simple consolidation structures, through bodies intruding into sediments, to the complexities of shear deformational features. Some of these features are presented in Figure 3.1.

The geometrically simplest alteration a sediment can undergo is a bulk reduction of its pore space. This is commonly caused by electro-chemical processes or consolidation of the material under stress.

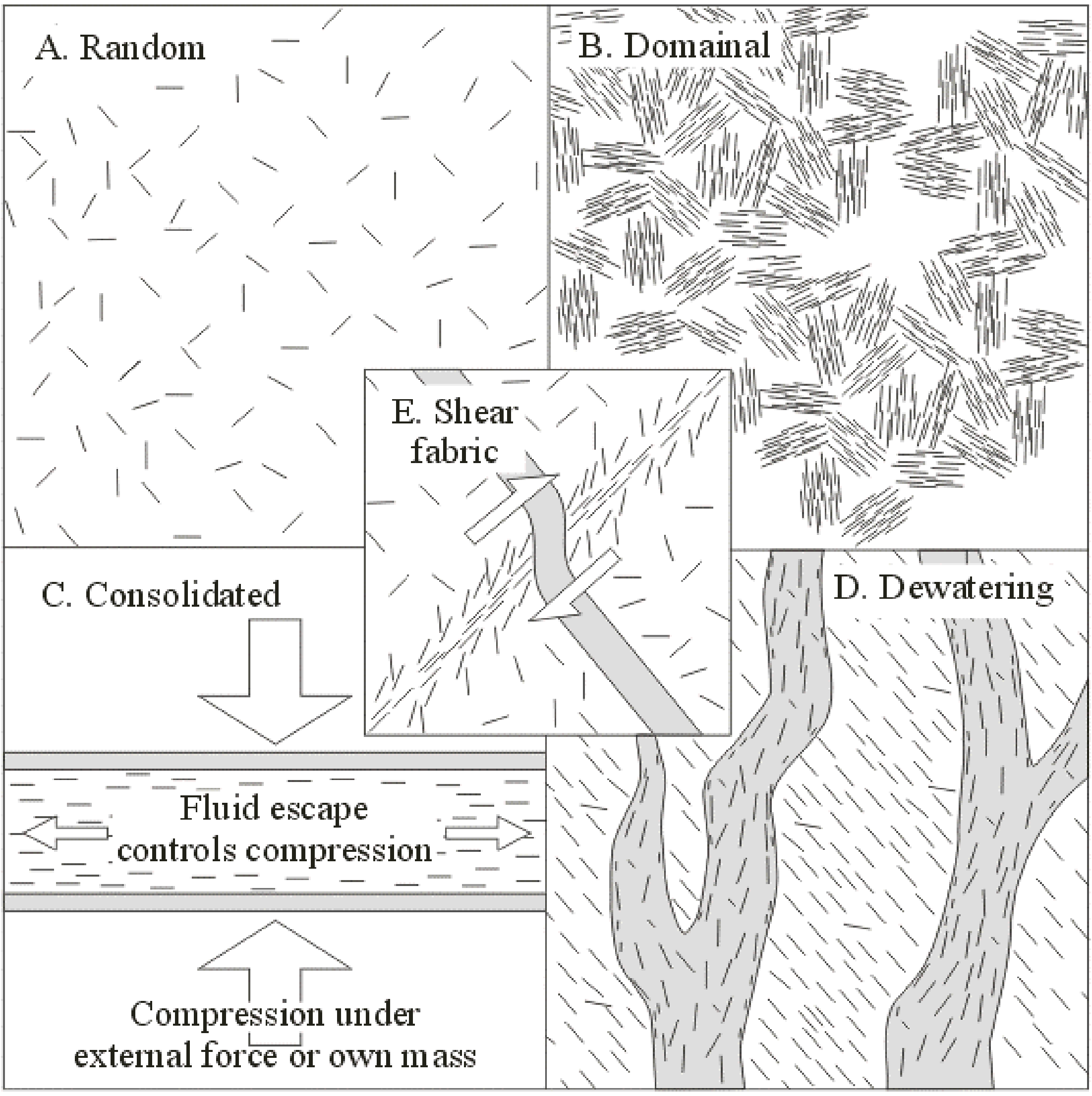

Figure 3.1 Some common soft sediment fabrics and microstructures: a) random orientation b) domains c) omnisepic fabric d) dewatering fabric e) shear fabric.

3.2 Depositional forms - The electro-chemical reduction of porosity

Unsatisfied charges on the surface of clays mean that structures can form through the attraction and repulsion of grains when sediments settle out of fluid suspension. The chemistry of fluid flowing through clays alters their permeability by changing the fluid viscosity, the clay structure and distribution, and thickness of water layers bound to the grains. Lower electrolyte concentrations lead to greater grain dispersion but the effect will be strongly dependent on the clay type (Mesri and Olson, 1971). Thus, in the glacial situation, the permeability may be dependent in part on the fluid's source and how far it has moved through the sediment already picking up solutes.

The classic electro-chemical structure is the domain (*Figure 3.1b*). These are randomly oriented patches, some 50 μm to 1 mm square, of aligned particles (though note the term is also used in reference to broader areas, such as shear zones by some authors, for example, Morgenstern and Tchalenko, 1967; M^cConnachie, 1974). As domains form through electrostatic attraction the porosity will be lower than in clays with random grain alignments. This reduction would suggest a decrease in the sample permeability. However, the tortuosity of randomly aligned domains will be less than a sediment which is totally random. Large pores may also develop between the aggregates (Mesri and Olson, 1971). Flow is generally increased through domainal sediments by the latter two processes, but this is dependent on the effective pressure on the sediment (Mesri and Olson, 1971). That domainal structures are rare in glacial sediments suggests that the pore fluid chemistry is not suitable, or the features are destroyed after formation.

Substantial subglacial solution of grains can occur (May, 1980). The characteristic embayment of grains affected by this process has been seen in tills from the Yorkshire coast, the micromorphology of which is examined in Chapter Seven. Possible glacial iron deposits found in Criccieth in Wales are examined in Chapter Four. It seems possible that the solution and redeposition of minerals may also significantly alter the permeability of glacial sediments.

3.3 Consolidation

Consolidation is the reduction of a sediment's porosity under its own mass or external forces. It is usually taken to refer to gravitational coaxial strain, that is, a vertical length change. Rates of consolidation are modelled using the stress and the rate at which fluid is expelled from the reducing pore space; the 'diffusivity'. The diffusivity is dependent on the permeability, the friction, and cohesion of the sediment as well as the pore fluid (see Rieke and Chilingarian, 1974, and Murray, in press, for a comprehensive review). The porosity of sediments is often in equilibrium with the effective pressure they are under. Such sediments are said to be normally consolidated. Where more porosity has been retained than is normal, the sediments are said to be underconsolidated for the present pressure or the maximum they have experienced *in situ*. (the latter is also known as overpressuring). When there is less porosity than expected for the present situation or a past maximum pressure the sediments are known as overconsolidated.

Under loading particles become strongly aligned perpendicular to the compressive stress (*Figure 3.1c*) (Delage and Lefebvre, 1984; Dewhurst *et al*., in press). However, alignment is in the direction of the maximum strain where this direction differs from that of maximum stress because of heterogeneous deformation (Baker *et al*., 1993). Alignment is strongly dependant on the material (Dewhurst *et al*., in press). Where domains form under progressive loading, the sediment first develops grain clusters which break down between 200-500 kPa to be replaced by domains and smaller pores (Dewhurst *et al*., in press; Brown *et al*., in press). Most porosity is then lost through the reduction of large inter-aggregate pores despite the internal deformation of the domains (Delage and Lefebvre, 1984). Domain deformation may be constrained by loading thresholds (M^{c}Connachie, 1974). Inter-domainal pores may elongate before they collapse (Delage and Lefebvre, 1984) potentially leading to increased horizontal permeability without a change in vertical permeability.

Alignment of grains and pores under consolidation produces an anisotropic permeability in sediments (Al-Tabbaa and Wood, 1987; Znidarcic and Aiban, 1988; Dewhurst *et al*., in press; Brown *et al*., in press). For kaolinite and calcium montmorillonite Brown *et al*. found that the horizontal drainage was greater than the vertical (horizontal to vertical ratios of 1.7 and 8 at void ratios of 0.7 and 1.6). This result matches Dewhurst *et al*.'s similar findings. The effect was attributed to increases in the vertical tortuosity (perpendicular to fabric). However, these ratios are not as great as those expected from the Kozeny-Carmen model (*Equation 3.4*). It was suggested that domains, shears caused by uneven consolidation, and other heterogeneities, reduce the effect of the alignment. The permeability response of glacial material might be expected to fall between that of Brown *et al*.'s (in press) mixed clay material and their silt-clay mix. The ratio between horizontal and vertical permeability would thus range between five and zero if only gravitational consolidation took place.

It should be noted that glacial clays have been reported which have suffered no consolidation. Meltout till might maintain 'fossil' porosities of 40%+ from ice filled ‘pores’ in the basal ice (Ronnert and Mickelson, 1992). In addition, the consolidation alignment of fabrics is thought strongly dependent on the depositional fabric. Lateral pressures can also cause consolidation. The widespread glaciotectonic alignment of particles under horizontal compression was reported by Kluiving *et al*. (1991).

Consolidation may be represented by more complex features than horizontal alignment. Even sediments that have undergone no obvious tectonic event can contain deformational features formed in the process of consolidation. Though Paul and Eyles (1990) pointed out the process on the visible scale, this fact has been too often ignored in glacial micromorphology. Many of the classic signs of tectonic deformation (for example shears, *Figure 3.1e*) can be found in material that has suffered only local vertical stresses (Arch *et al.*, 1988; Brown *et al*., in press). Such features are produced by differential compaction. This strain difference results from varying diffusivity in heterogeneous sediments (Athy, 1930), fluid pressure gradients set up by drainage in homogeneous material (Roscoe, 1970; Al-Tabbaa and Muir Wood, 1991), and sub-sediment topography (Athy, 1930). In the first two cases, the material will eventually appear to have consolidated evenly on a large scale.

Consolidation may also produce dewatering fabrics (Cowan, 1982) (*Figure 3.1d*) when the force from expelled fluids exceeds the sediment strength and the force needed to transport the particles (Lowe, 1975; Paul and Eyles, 1990).

3.4 Localised bodies of altered porosity

The intrusion of exotic sediments, localised sorting, or comminution will alter the porosity and drainage potential of sediments (Athy, 1930; Barron, 1948; Åmark, 1986; Boulton and Dobbie, 1993; McCabe and Dardis, 1994). Structures showing such variation include dewatering paths (fine sediments winnowed out and deposited elsewhere), load-failure structures between two beds (sediments move up or down into another bed), brittle fracture fills, drag-folds, channel fills, and soft clasts which are strung out by deformation. The effect of such structures will depend on the drainage potential of the material around them and the areas they link with. Often it is implicit in such structures that they link areas of different permeability. For example, gravity structures (Lowe, 1975; Allen, 1977), stop dropping into sediments when the resistance of the surrounding material becomes too great. Such resistance is intimately tied to sediment porosity and permeability through the fluid pressure and friction. The drainage effects on the bulk sediment containing the structures will control the ability of the external material to deform both locally and at a considerable distance (for quantified examples see, Athy, 1930, Barron, 1948, Al-Tabbaa and Muir Wood, 1991).

An example of these drainage processes may be the clay brecciation seen around sand intraclasts by Menzies (1990). While Menzies attributes this to the drawing of water onto freezing fronts on the clast surfaces, the clasts could equally maintain a low pressure in the impermeable till for long enough for the brecciation to form (implied in Menzies and Maltman, 1992). Heterogeneities in drainage may provide the initial stiffness responsible for drumlins (Boulton, 1987; M[c]Cabe and Dardis, 1994) or even tougher "sticky-spots" controlling glacier velocity (Kamb, 1991; Fischer, 1992).

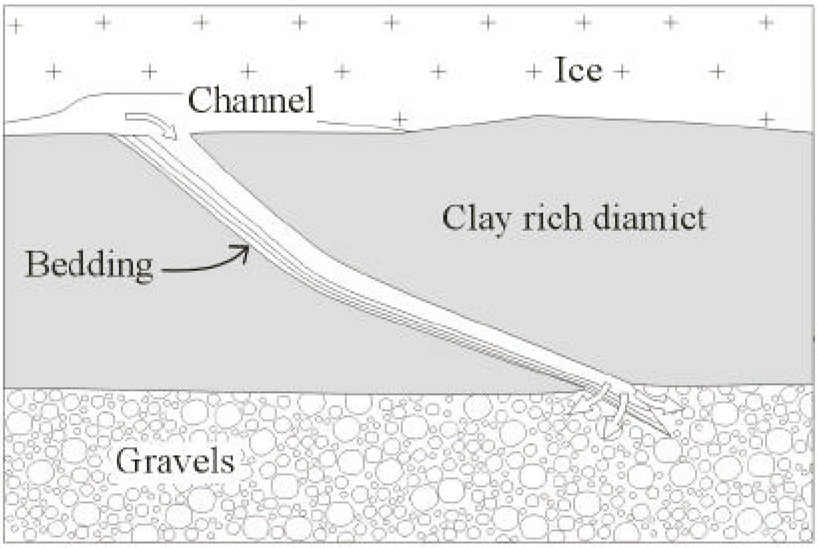

Figure 3.2 Diagrammatic representation of one of Åmark's (1986) clastic dykes. The dykes link the glacial bed to highly permeable gravels through an aquitard of till.

As an example of this, Åmark (1986) found clastic dykes linking the ice interface to a lower gravel unit through a clay-rich diamict bed (*Figure 3.2*). The bedding inside the dykes suggests the fluid depositing the beds drained easily into the gravels. As the dykes filled with material that is coarser than the clay-rich bed it seems likely the dykes continued to act as drainage conduits after they filled. Boulton and Hindmarsh (1987) have argued that one would need a thickness of 1.5 million metres of till at the terminus of a warm bedded glacier to evacuate its meltwater by throughflow. However, one would need only 150m of sandy till (as in the dykes) and only 15 cm of coarse gravel such as that under the diamict at Åmark's site. Thus, the dykes (up to 2.5m across) control the flow between the top of an aquitard and the highly permeable gravels. This drainage may explain the low strain after their formation. A similar situation could be envisioned for the coarse material filling a large-scale pipe recorded by Kluiving (1994). This was interpreted as having been implaced into a frozen fine-grained body. Paré *et al*. (1984) found till with a hydraulic conductivity anisotropy of 10 to 20 (horizontal to vertical ratio) at the base of dams in Northern Québec. This anisotropy was attributed to sand body orientation. The orientation of glacial sand bodies seen at the microscale has also been shown to control larger-scale flow anisotropies (Barron, 1948; Nyborg, 1989).

All this makes for a problematic reconstruction of the hydrology of any palaeoglacier in three dimensions and across time, however, it should not be viewed as impossible. M^cCabe and Dardis (1994) have shown the qualitative possibilities for such work, and researchers in soil science and civil engineering tackling similar problems have devised methodologies that simplify the problems from a quantitative viewpoint. Unique solutions are possible for individual features and stress fields (for example, Barron, 1948; Al-Tabbaa and Muir Wood, 1991), however, generalisations of features can successfully represent their effects (Nyborg, 1989; Paré *et al*., 1984). Estimates of bulk hydraulic properties can be made from field samples (Bouma *et al*., 1989; Ahuja *et al*., 1989), and these approaches may make possible the interpolation of till hydraulic characteristics from borehole data, micromorphology, and outcrop samples.

3.5 Shear zones

It is becoming increasingly accepted that many of the fabrics found in glaciogenic sediments are formed in pure and simple shear situations (Chapter Two). Measurements of subglacial deformation imply that simple shear, and possibly pure shear, geometries exist subglacially (Boulton and Hindmarsh, 1987; Blake, 1992; Porter, 1997). It is a possibility that the shear fabrics found in glaciogenic sediments are due to such subglacial deformation. Alternative causes are proglacial flow, low strain because of the meltout from basal or buried ice, or heterogeneous consolidation shearing. Large-scale shear structures can have a controlling influence on the dynamics of sediment bodies (Moore, 1989; Maltman *et al*., 1993b). Such macroscale shear *zones* may be formed from a mass of microscopic shears (*Figure 3.3*) (Maltman *et al*, 1993a). Glacial studies have shown macroscale shear features in the proglacial zone (Hart, 1989; Hart *et al.*, 1990; Thomas, 1984; van der Wateren, 1986;

Kluiving, 1994), and in frozen subglacial sediments (Echelmeyer and Zhongxiang, 1987). However work on the development of subglacial shears has largely been hypothetical (Harrison, 1957; Evenson, 1971; Menzies, 1986; Boulton and Hindmarsh, 1987).

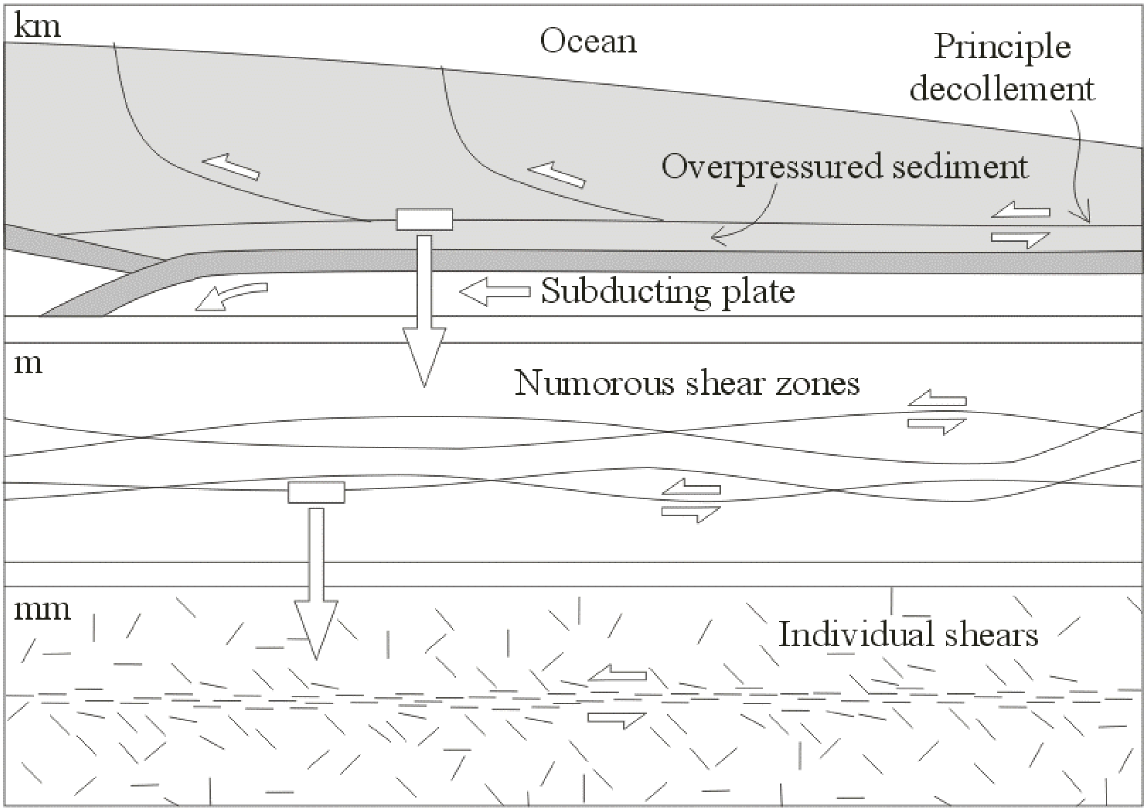

Figure 3.3 Diagram of the structure of the Nankai Prism, East of Japan, showing the décollement at different scales. After Maltman et al., 1993a/b.

The next sections examine the development of shear zones, and the effect they have on the hydrology and strength of the sediments they form in.

3.5.1 The development of shear zones

Shears are thought to develop at localised stress concentrations or material heterogeneities when the ability of material to maintain higher and higher stresses through compaction is exceeded (White *et al*., 1980). Early work showed that shear in soft sediments produces a complex morphology of structures (Tchalenko, 1968; 1970; Morgenstern and Tchalenko, 1967). This shear zone morphology develops in the following order (*Figure 3.4*).

1) The material undergoes particle realignment and strain before yield. This appears pervasive, but is partly due to microscopic shearing (Maltman, 1987; Arch, 1988) (*Figure 3.4a*).

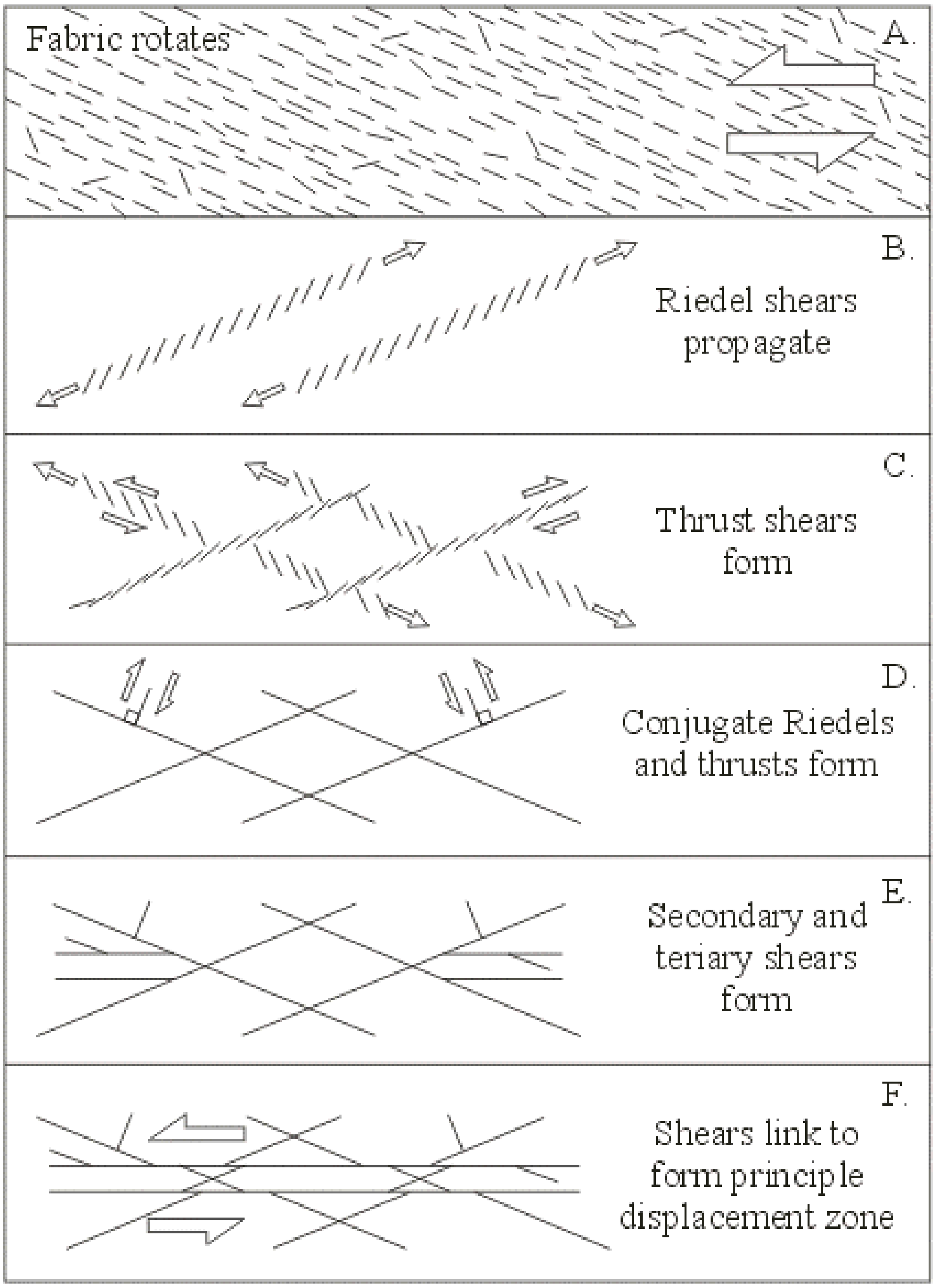

Figure 3.4 The development of shear zones (after Morgenstern and Tchalenko, 1967; Moore et al., 1986, p.42; Rutter et al., 1986; Tchalenko, 1968; 1970; Arch, 1988).

2) At the yield strength of the sediment (i.e. when the maximum stress is supported) yield at weak spots or stress concentrations transfers stress to nearby areas causing catastrophic propagation (Byerlee *et al*., 1978). Propagation may be further aided by the pumping of fluid from collapsing areas of the shear into areas which then dilate (Arch, 1988). 'Riedel' shears form at an acute angle (a) to the major principle stress (*Figure 3.4b*), modelled by

$$a = 45° - \frac{f}{2} \qquad \textit{...Equation 3.5}$$

where f is the internal friction of the sediment. Atkinson and Richardson (1987) show there is a maximum a of 45° from the applied stress, which is also the figure when there is no bulk drainage from the sample

3) Riedel shears represent only small strain, being arranged at an orientation not conducive to large amounts of sliding in the direction of shear. As the material softens with strain, 'Thrust' shears form, mirroring the Riedel shears (with respect to a line perpendicular to the major principle strain direction) (*Figure 3.4c*) and conjugate Riedels form (*Figure 3.4d*). Conjugate Riedels form at higher pore water pressures in unconstrained simple shear experiments (Tchalenko, 1970). When their internal fabric is sub-perpendicular to their shear direction, the conjugate Riedels form structures known as kink bands (*Figure 3.5*) (Tchalenko, 1970).

4) The residual strength of the material is the stress supported as the material undergoes steady state infinite strain. This state occurs when the shears form a long 'Principle Displacement Zone' (PDZ) through the development of new shears in the Riedel and Thrust angles to the old shears (*Figure 3.4e*). The PDZs form parallel to the shear strain direction in simple shear experiments (*Figure 3.4f*), and along the Riedel shears in triaxial experiments. These shears may then undergo infinite strain without change.

The PDZs often bound lenses of undeformed material in cataclastic zones (Rutter *et al*., 1986). A similar situation may occur in glaciogenic sediments (Seifert, in 1954, quoted in translation by Harrison, 1957). Thus it is essential that low strain thin sections are not taken as indicative of bulk finite strain without the boundaries and any resistant layers in outcrops being sampled. Often the lens fabric reorientates normal to the major principle stress (*Figure 3.5*). The grains rotate until the shear stress on the grains is zero (Tchalenko, 1968). Such reorientation can represent considerable strain (Morgenstern and Tchalenko, 1967; Tchalenko, 1968).

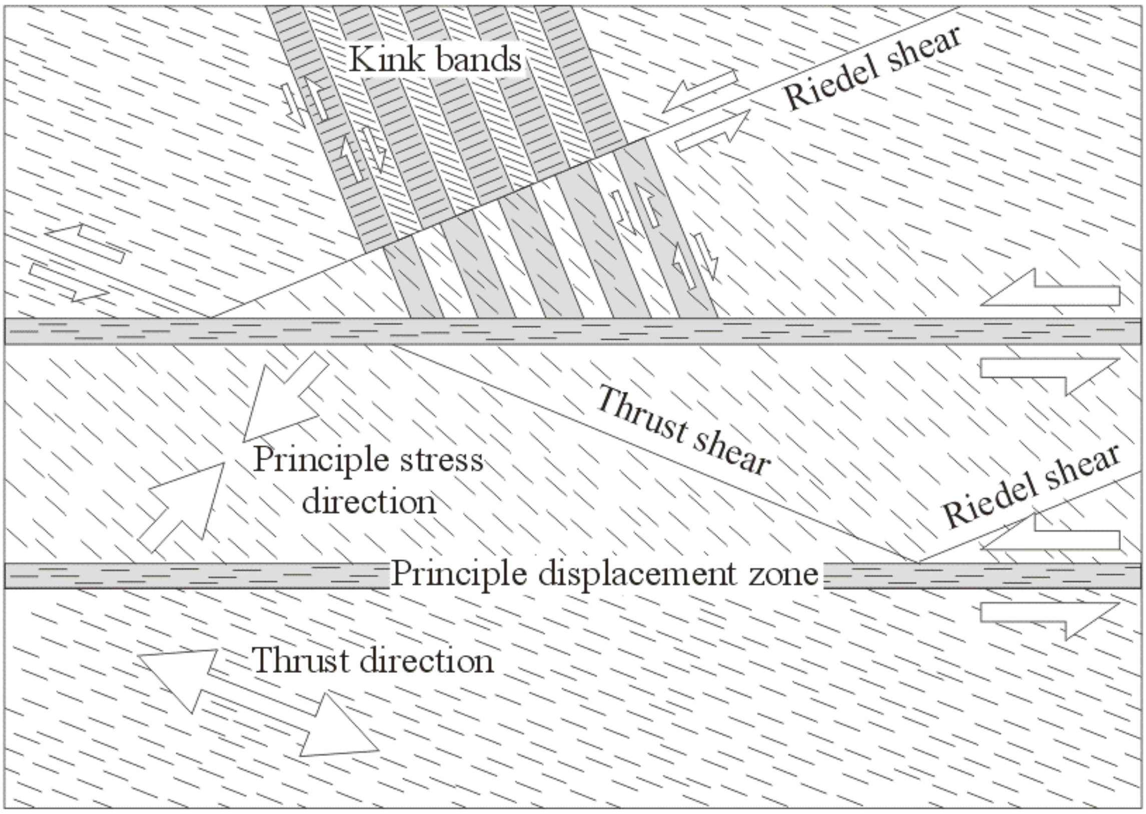

Figure 3.5 The orientation of fabrics in kink bands and lenses bound by Principle Displacement Shears (after Tchalenko, 1968; 1970).

A mix of primary fabric and a single PDZ could both represent infinite shear and give information about the final principle stress direction. Van der Meer (1993) has described broad zones of fault bounded lenses in glaciogenic deposits. However, increased shear may lead to more shears, rather than a development of individual PDZs (Maltman, 1987). It should be noted that local reorientation around shears occurs prior to yield. In *overconsolidated* samples (contrast with Rutter *et al*. (1986), above) the fabric in the area where the shear zone will develop rotates to *a* prior to peak strength (Tchalenko, 1968); when not overconsolidated the fabric is in the Thrust shear direction.

Interaction of shears may lead to strain hardening, and Riedel and Thrust shears that have developed in the absence of a PDZ (bimasepic or lattisepic fabrics) may imply this stiffening has occurred (Rutter *et al*., 1986). It has been suggested that shear zone thickening compensates for local hardening on the shears (Rutter *et al*., 1986). However, it is impossible to tell from geometrically constrained lab tests whether such widening of a lattisepic fabric fully compensates for hardening when the thickness of the fabric zone is not limited.

Kink bands are rare in triaxial tests, and have only been found in samples with water contents greater than 30% (Maltman *et al*., 1993a; 1993b). However, such features have been found in nature in conjugate sets (Maltman *et al*., 1993b), and in the absence of other shear forms (Maltman, 1977; Maltman *et al*., 1993a). In the latter the unusual breadth of the features suggests they may be associated with conditions in which shear stress is not released through the normal shear mechanism. Byrne *et al.* (1993) showed that kink bands can narrow and become shears, suggesting they are low strain forms of normal shears. Maltman *et al*. (1993a) suggest kinking arises from compression associated with the volumetric problem of moving sediment along shears while there is no movement occurring at the propagating tips. Kink bands can be both extensional and compressive features dependent on their throw. Whatever their cause, it would appear that kink bands represent strain prior to the sediment yielding *locally*. It may, however, be that sufficient stress could be supported by the residual strength of a Principle Displacement Zone to cause such behaviour in its periphery. Kink band formation can only be taken to represent pre-yield creep if there are no PDZs.

Repeated, close spaced kink banding forms ‘crenulations’ parallel to the principle displacement shear. Massive crenulation in glaciogenic sediments has been attributed to regional compression by Kluiving *et al*. (1991). In thin sections kinking can be so intense that there is the potential for misinterpreting the stress field and strain response responsible for it.

Factors affecting formation

It has been noted that shear *zones* may contain several sets of each geometry of shear outlined above. The orientation, number, and length of different shears is controlled by the skeletal and hydraulic properties of the sediment under deformation, and the strain. Thus, the geometry of the shear zones can give information about the sediment and the conditions under which deformation occurs. The general form (Riedels, Thrusts etc.) is not controlled by the material as it appears in a wide variety of geologies. This overall form must relate to strain/stress application (Logan *et al*., 1992), or the development of material anisotropy (Rutter *et al*., 1986). The response of shear geometry to changing several important conditions is given in Table 3.1.

Condition	Response	Authors
Effective pressure increases.	Work hardening of sediment.	Feeser, 1988
	The increased *deviatoric* stress resulting from increased loading leads to acceleration of the evolution of shears, the material reaching the residual strength faster.	Simamoto and Logan, 1981; Maltman, 1987; Logan *et al.*, 1992; Parry, presented by Roscoe, 1970
	Riedels are shorter, more numerous, and inclined at higher angles to the Principle Displacement zone.	Byerlee *et al.*, 1978; Logan *et al.*, 1992
	Higher angle between the confining stress and shear, some dry specimens having an angle greater than the theoretical maximum of 45 degrees	Arch *et al.*, 1988
	Principle Displacement shears are less frequent and longer (cf. Riedel shears, above). Strain concentrates in a few Principle Displacement shears.	Maltman, 1987; Arch *et al.*, 1988.
	For water contents <25% shear zones are narrow, and with >25% they become more complex.	Arch *et al.*, 1988; Logan *et al.*, 1992
Increased depth of material available to shear.	Displacement needed to produce the same shear zone geometry increases. Because shear geometry is related to the shear stress supported at any time, it follows that a higher shear stress is needed to deform thicker zones over an equal displacement to thin ones	Logan *et al.*, 1992
Change in material.	In unimodal sediments the form is roughly constant whatever the material.	Maltman, 1987; Logan *et al.*, 1992
A fabric exists prior to deformation (for example, because of depositional processes or due to changes in glacial thermal and stress regimes as the glaciers advance).	The reorientation fabrics in mylonites theoretically induces considerable hardening	White *et al.*, 1980
	If the initial fabric is close enough (<15°) to the expected shear angle, shears develop along the primary fabric because the high shear stress on grain boundaries at an angle close to that of the shear direction preferentially rotates these grains. At higher angles shears become pervasive or anastomosing.	Arch, 1988; Arch *et al.*, 1988; Platt and Vissers, 1980; Stephenson *et al.*, in press

Table 3.1 The response of sediment shearing and shear geometry to changing effective pressure, material thickness, constitution and fabric.

The relationship between shear geometry and effective pressure will relate to the number of stress concentrations on the shears from which subsequent shears propagate, the ease of realignment under different porosities (Maltman, 1987), and the ease of fluid movement within the shears, which will control the ease of propagation (Arch, 1988). The conflicting signatures of PDZs (get longer and less numerous with increased effective pressures) and Riedels (get shorter and more numerous) plainly demands a great deal more experimental work, particularly as one may become the other. However, the geometry of shears can broadly be used to qualitatively estimate the effective pressure.

The effect of potential shear zone depth is important in till, where shear zone size may be nominally unlimited. The only controls on shear zone width are fluid diffusivity and the stress-strain relationship of the sediment, both of which are controlled in part by shear zone development. There are, thus, two alternatives for thick glacial tills; either it takes a great deal more stress to produce less and less strain as the initial thickness of the sediment grows, or the deforming thickness reaches a threshold after which it is internally controlled. Thick tills may develop Riedels without PDZs up to considerable displacements, maintaining a high strength. However, PDZs do develop on a larger scale in other unconstrained environments (Tchalenko, 1970).

The relationship between shear form and material is also complex under glaciers, where the sediments are often multimodal. It was suggested in Chapter Two that bimasepic fabrics may be shear zones constrained by larger grains, and such constraints could act at an outcrop scale. This hypothesis is backed up by Rutter *et al*. (1986) who found that in natural and experimental fault gouges, Riedel and conjugate Riedel development was constrained to areas between clasts rather than expelling them. The shears developed preferentially next to clasts because of stress or strain heterogeneities associated with their surfaces. It is possible that clast expulsion from the shears will also occur. Alternatively, soft clasts may break up through slip on Riedel and Thrust shears (Rutter *et al*., 1986), giving the strung-out ‘tails’ of material seen in glacial sediments. This explanation suggests that pervasive flow is not necessary to form these features.

In much of the above, it can be seen that the water pressure is an important factor in determining the form of shears and, therefore, their strength and effects. Thus it is worth examining in greater detail the ways fluid in sediments and deformation interact. Despite the fact that many sediments act as though pervasively deforming up to failure at ~8 to 12% strain (Arch *et al*., 1988), true pervasive deformation is rare, and shears form rapidly in sediments with pore water volumes of up to 60% (Maltman, 1987). Equally, even in so-called 'undrained' lab tests samples are locally drained by shear zones (Atkinson and Richardson, 1987), and their strength is controlled by this process. Thus, an examination of water movement in sediments is largely an examination of fluid interactions with shear zones. Following such a study, the effect of shear development on strength will be reviewed.

3.5.2 Hydrological effect

There are two viewpoints from which the interaction of shears and fluid flow can be examined. Shears can be considered as static objects which affect permeability, or more rarely, they can be considered as evolving structures interacting with, and being affected by, fluid flow. In both cases the shears will affect the consolidation of surrounding material, and this will affect the shear's development (Moore, 1989).

Shears as static structures

Large-scale shears may act as dewatering paths (Moore, 1989; Maltman *et al*., 1992; Maltman, 1988; Byrne *et al*., 1993). Dilatant shears will have a higher porosity than the surrounding material, however, increases in permeability have also been seen in clay materials in which the platy grains in the dilatant areas may be expected to collapse and porosity be reduced. Arch and Maltman (1990) quote shearing as reducing porosity from 30% to 10%. They found that the permeability of samples increased as shear zones developed, if the fluid flow and shear were aligned, and decreased if they were perpendicular (Arch, 1988; Arch and Maltman, 1990). As porosity should have been the same in each case tortuosity was invoked to account for the variation (*Equation 3.4*).

It has already been noted that Equation 3.4 overestimates the effect of tortuosity by an order of magnitude for consolidation structures because of heterogeneities along the structures formed by alignment (Dewhurst *et al*., in press). Brown *et al*. (in press) compared the permeability of consolidated kaolinite to consolidated kaolinite sheared in a ring shear rig.

They show that shear zone behaviour can be reasonably predicted with the equation, although only in the vertical direction.

Brown *et al*. (in press) found the greatest permeability was horizontally through vertically consolidated samples. Horizontal drainage for the material that was both consolidated and sheared was identical to the vertical drainage in a sample that had simply been consolidated. The two samples had the same overall void ratio. For the authors these results indicate particle alignment cannot increase permeability, though greater tortuosity may *decrease* it. However, they tested bulk samples, not individual shears, and it should be remembered that there is a logarithmic relationship between void ratio and permeability. Shear specific effects like tortuosity will be swamped by bulk effects.

The lowest permeability found by Brown *et al*. was the vertical drainage through the sheared sample, with a horizontal to vertical ratio of 25 at a void ratio of 0.85. This is in line with Equation 3.4. Brown *et al*. suggest that the tests reflect porosity decreases under shear. The situation is further complicated by shears being preferential areas for the flow of fluidized sediments, forming sediment dykes (Talbot and Brunn, 1987; Maltman, 1988; Menzies, 1990).

Long horizontal pores may form in consolidating material as the grains align (Delage and Lefebvre, 1984; Dewhurst *et al*., in press). The fact that, despite lab work, natural shears *do* appear to act as enhanced fluid flow paths led Dewhurst *et al*. (unpub.) to suggest that these pores were causing enhanced horizontal permeability (B.Clennell, pers. comm., 1996). Such pores may only form when there is an increase in the pore fluid pressure after grain alignment (Delage and Lefebvre, 1984). It is this propensity to open under reduced effective pressure which may be responsible for hydraulic anisotropy (B.Clennell, pers. comm., 1996). In traditional fracture mechanics it is presumed that there is a grading between tension cracks to shears (Price and Cosgrove, 1990). With the possible localisation of fluids in shears subglacially there is no reason why this should not be so. Freeze-thaw cycles can rapidly lead to horizontal macrochannels (mm+) in unsheared soils (Sole-Benet *et al*., 1964), and may form similarly under glaciers, closing as the effective pressure increases. Other features larger than the grain-grain scale that will effect permeability are slickenlines and broad undulations parallel to the movement direction (seen by Bryne *et al*., 1993; Maltman, 1987; van der Meer, 1993) which will affect the tortuosity of shears.

There are a number of conditions which significantly alter the permeability response of the sediment to shear strain. The nature of the material itself has an effect. For example, Bryne *et al*. (1993) found that in a sand/kaolinite mix permeability increased with strain (unlike kaolinite on its own), particularly at yield, the permeability then dropping with greater strain. This response is very similar to mixed sediments reported on in this thesis (Chapter Six). If we presume a mixed grain size distribution causes shear zone disruption it seems hard to explain why the permeability increases with greater pervasive movement. Large grains may preserve dilated areas in the manner suggested by Murray (1990).

The second major control on the permeability response is the sediment consolidation history. Consolidation itself introduces a hydraulic anisotropy (above). However, shear is much more efficient. Brown *et al*. (in press) found that it took twice as high an effective pressure to consolidate samples down to a 0.85 void ratio as it did to shear them to this ratio. Consolidation affects the development of shears. Stephenson *et al*. (in press) found that under-consolidated and normally-consolidated sediments decrease permeability with strain because the fabric compresses, whereas over-consolidated sediments increase in permeability because they dilate to strain. Some lightly consolidated sediments vary in a complex fashion during deformation, while decreasing in permeability over all. Such complexity develops as the peak strength is reached and shears develop. Stephenson *et al*. (in press) worked in terms of 'dynamic' permeability, which is the effective permeability measured during continuous deformation and therefore includes fluid movements forced by deformation. This viewpoint provides a more realistic understanding of active systems, such as those likely under deforming-bed glaciers. Therefore, the next section examines the dynamic considerations which such authors have revealed.

Shears as dynamic structures

When an individual shear zone forms, particles must realign. In normal and overconsolidated sediments this realignment will include some element of dilation as the grains must move up and over their neighbours. In underconsolidated and some pre-aligned normally consolidated sediments this is unnecessary and alignment occurs by compaction in the shear zone (Stephenson *et al*., in press). In saturated sediments at low stresses fluid is drawn into the shear zones as they form. If these zones then collapse this fluid is expelled (Stephenson *et al*., in press).

Such fluid movements will affect the porosity of nearby sediments (Boulton and Hindmarsh, 1987; Moore, 1989; Brown *et al*., in press), dilation possibly draining the surrounding material and inducing brittle behaviour. This is recognisable in *anisotropic* hard rocks by the development of reverse drag (*Figure 3.6*) (Platt and Vissers, 1980). In the case of glaciogenic sediments this process may be expected below a depth determined by the overburden and the fluid pressure. The latter is presently thought to decrease with depth in homogeneous sediments (Boulton and Hindmarsh, 1987; Clarke and Murray, 1991). The increase of such reverse drag with depth would be a strong indicator that such a fluid pressure regime existed in a sediment body.

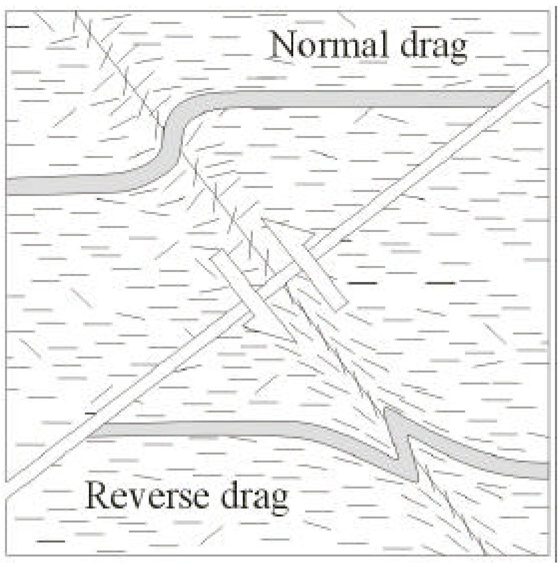

Figure 3.6 Reverse and normal shear in anisotropic rocks. Solid shaded band is a marker horizon that has no effect on material properties of rock.

3.5.3 Strength effects

The form and, therefore, strength of shears will be determined in part by the consolidation history of the sediment. The strength of overconsolidated soils is shear controlled. The shears dilate and soften, allowing further softening in these areas, dilating until they reach a stable porosity and strength (the 'critical state') (Atkinson and Richardson, 1987). Underconsolidated soils deform by compression so shears are less likely (Atkinson and Richardson, 1987), however, at some point there is a balance between the two processes. For steady state deformation the critical state is, ideally, the same for a soil tested in normal and overconsolidated states, that is, both have the same final porosity (Roscoe, 1970). The residual strength is unaffected by the initial consolidation state, though the overconsolidated sample will have supported a yield stress greater than this around yield. In reality this behaviour is only found in truly undrained situations, and shears cause local drainage (Atkinson and Richardson, 1987). Thus, we cannot say that normally consolidated glacial sediments and any overridden material that is pre-consolidated will act similarly given enough time.

Particle alignment is a crucial factor in bringing a sediment to its residual strength (Early, 1964, quoted without source by Tchalenko, 1968; Tchalenko, 1968; Maltman, 1987). However, what is equally important is the form the alignment takes. Changes in the rheology and drainage potential of shear zones are likely to be strongly dependent on the shears' overall form. The overall form of such features controls the angle the aligned grains make to the direction of the stress application and the fluid throughflow. It also controls the interconnectivity of such zones. One can only compare the response of similar geometries. Because of this fact, response to varied stress can only be compared after the development of Principle Displacement Zones (Logan *et al*., 1992). Once this point has been reached, the residual strength of the material may be constant (Brown *et al*., in press), although at high strains PDZs develop new Riedels and conjugate Riedels between them (Logan *et al*., 1992). This suggests that a balance may be reached between the residual strength of the sediment and continued shear zone formation.

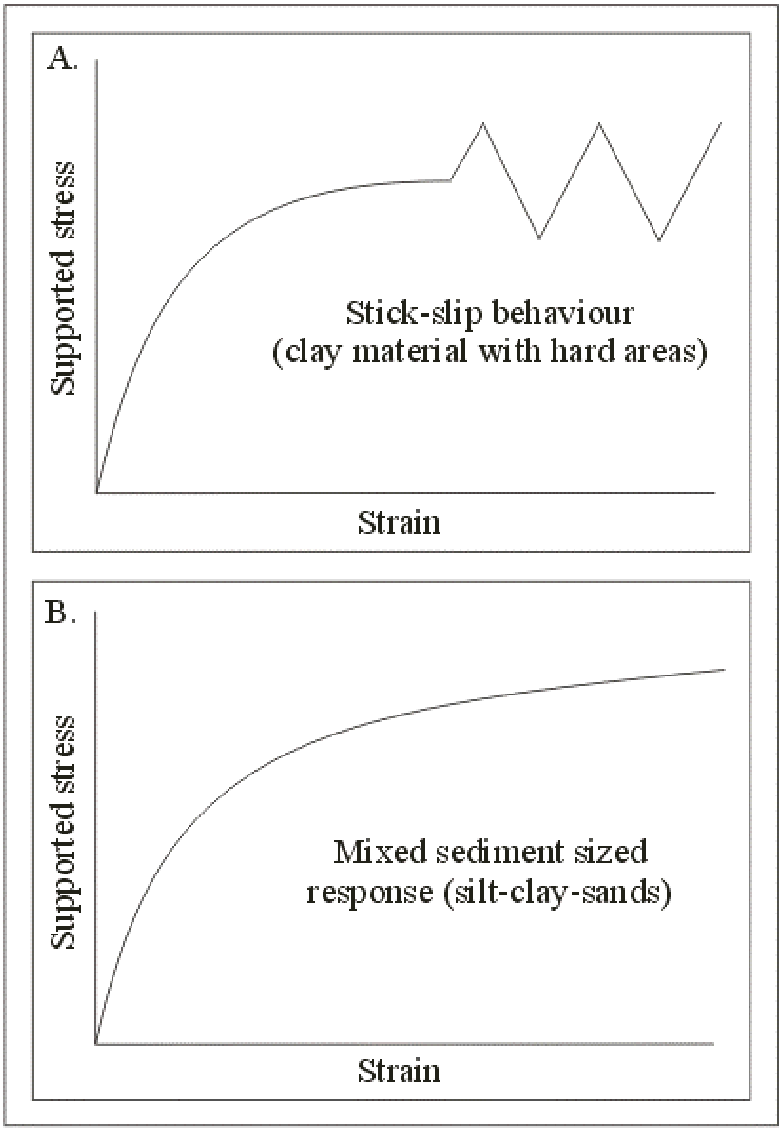

Figure 3.7 Idealised stress strain relationships: a) stick-slip behaviour b) response of mixed silt-clay-sand sediments (after Maltman, 1987, figure 1C).

In laboratory tests, both saw-cut and gouge-filled rock faults (normally consolidated) can develop stick-slip strain after a few percent of smooth movement (~3%-5.5%, Logan *et al*., 1992). During this behaviour supported stress oscillates between high and low values (*Figure 3.7a*). The stress variation amplitude reduces with increased gouge thickness. There is no accepted explanation for this effect. It may be that lenses in the faults are strain hardening while the shears soften (Logan *et al*., 1992), however, this would still necessitate an unexplained synchronisation of shear development. Gouge material includes fragments on the scale of the shear zone width, therefore, such stick-slip behaviour could be grain fracture events (Summers and Byerlee, 1977). This suggestion is backed up by the fact that soft materials do not display such behaviour (Simamoto and Logan, 1981). Such behaviour only develops in soft gouges at confinement pressures greater than 50 MPa (Logan *et al*., 1992), which is far higher than expected for glacial tills. However Simamoto and Logan (1981) found two important features of the behaviour:

a) it may be the presence of material compacted to rock strength in the gouge that is causing stick-slip.
b) the thicker the gouge, the more confining pressure is necessary to initiate stick-slip in fine grained material.

Both suggest that the interaction of hard areas and shears may be responsible for stick-slip behaviour, and that the confining pressure effect largely reflects the pressure necessary to produce hard patches. Such behaviour can then be imagined to occur at much lower stresses in till, where the hard areas are already present in the form of clasts. Indeed, if such a response was seen in tills under low pressures, this would be additional evidence for the effect being caused by hard patches. Such a response is tentatively outlined in Chapter Six.

Slickenlines or broad undulations parallel to shears are common in soft sediments. Slickenlines will strongly influence the friction across the shear zone as unorganised material is pushed out of the way by particles to make the grooves. Slickenlines will be highly significant in the heterogeneous diamicts under ice. In the situation where undulations develop they increase the surface area of the shear, presumably increasing the total frictional resistance (though other effects may lead to overall strain softening).

3.5.4 Disruption of shear under glaciers

Permeability and strength effects will combine to give a complex rheological response to the sediment under glaciers. The nearest analogue may be the movement of oceanic accretionary prisms (*Figure 3.3a*). The shear zone at the base of the actively moving area of the Nankai prism (East of Japan) is thought to determine the dynamics of much of the material above. This décollement affects the hydrology of the prism, raising the fluid pressure below the shear zone. The problem in the Nankai prism (one reflective of shears in general) is that the décollement is both dense enough to prevent fluid flow from below yet is weak enough that it deforms and no stress is transferred across it, that is, it is a 'dynamic seal'. One possibility is that particle alignment allows weakness along the shear zone, while preventing the flow of fluid across the zone. However, there is evidence that the shear zones act as drainage paths (Moore, 1989; Maltman *et al*., 1992). There is some controversy about the effect of particle alignment along shear zones (see above), and this has cast doubt on tortuosity as an explanation allowing both deformation and overpressuring. The present hypothesis is a cyclic change from overpressuring to movement (Bryne *et al*., 1993; Moore, 1989; Maltman *et al*., 1993b) controlled by hydrofracturing as the overpressuring below builds. In the case of glacial sediments under ice the overpressuring will probably be above the décollement, but otherwise the same problems hold if movement is on a shear zone. Equally impermeable shears may form early in the deformational history, raising the pore pressure and strain rate in the material above (Murray and Dowdeswell, 1992).

The Nankai sediments may decrease in porosity with each overpressuring cycle, increasing the chance of overpressuring (Bryne *et al*., 1993). Steady-state views of shears may also be poor for glacial sediments. In opposition to this process, healing processes could reset fault zones in 'slack' periods of movement (Logan *et al*., 1992); glacial effective pressure/water system changes for example (Menzies and Maltman, 1992). Equally subglacial shear zones may be disrupted by clasts, particularly if they move down through soft till during deformation (Clark, 1991).

Even clasts with neutral buoyancy could cause disruption if trapped and rotating in wide shear zones, leading to an increased residual strength. However, the effect of reducing drainage along disturbed shear zones would be softening of the material. Sediments of a wide size distribution have broader, anastomosing, shears (Maltman, 1987) and strain harden, displaying no distinct yield strength (*Figure 3.7b*) (Maltman, 1987, figure 1C; Chapter Six). Shear zones may expel clasts, however, in thick tills it would take a great deal of shear for this to occur. Rutter *et al*. (1986) have shown how shear stress is supported on clasts in clay-bearing gouges as their concentrations increase and the possibility of expulsion is reduced. Under their (high) pressures the grains fracture, rather than exiting the shear zones easily. Strings of material extending from soft clasts in till show that such clasts supported a force sufficient to cause them to yield, even though they were in a ductile matrix. It may be possible to quantify till residual strength from such clasts.

Clast comminution may also radically change strain patterns (Aydin, 1978; Logan *et al*., 1992). Shear could provide a mechanism for comminution in till (Boulton *et al.*, 1974; Hooke and Iverson, 1995; Iverson et al., 1996). Fractured grains are found in till, though this may be due to freeze-thaw processes. It is likely that tills become finer grained down-glacier and over time, causing more fluid to be retained and the till to become softer, with the possibility that the ice becomes thinner than expected on a hard bed. Laboratory tests show that the difference in properties between remixed comminuted material and harder surfaces leads to shears initially being concentrated at the interface between the two, then moving deeper into the sediment during work hardening (Logan *et al*., 1992; Maltman, 1987). This may explain the layered bed described from under Breiðamurkurjökull (Boulton and Hindmarsh, 1987; Benn, 1995), where there is a layer of pervasive mixing, dilatancy and high strain next to the ice, and more distinct shears deeper in the sequence.

3.6 Conclusions

The principle aim of this review has been to outline the work on the development and effect of a number of microstructures. These microstructures have been located in glacial sediments, but little significance has been attributed to them in glaciology. This review has been undertaken so that it may be referenced during the rest of this study, however, it also allows us to draw a number of novel conclusions in two areas. The first is using micromorphology to recognise glacial processes. Secondly, we can also hypothesise a number of processes that may occur glacially, but need further testing. To conclude this chapter, these conclusions are drawn out in the next section.

3.6.1 Test criteria for glacial processes

Initial deposition processes. The rarity of large domains in glaciogenic sediments so far thin sectioned suggests that the conditions necessary for their formation (low effective stresses and/or, the correct pore fluid solutes) did not occur after the structures that *are* seen were deposited. Sediments that have undergone no obvious tectonic event may display significant strain markers formed in the process of consolidation.

Pre-yield Creep. Kink banding without Principle Displacement Zones may represent pre-yield creep. Kinking can be so intense that it is mistakenly interpreted as forming under a uniaxial compression stress field, therefore it is essential the presence or absence of larger scale compressive features is included in any interpretation.

Bed deformation. Large-scale shears form in a sufficiently varied set of stress and hydrological conditions that it seems likely that they could form subglacially. The pervasive/discrete shear layering of some subglacial sediments may result from the progression of shears away from the interface between two differing materials. Infinite strain shears often bound lenses of undeformed material. Thus, low strain thin sections should not be taken as indicative of bulk finite strain without the boundaries and any resistant layers in outcrops being sampled. A non-horizontal primary fabric with Principle Displacement Zones could represent infinite shear and give information about the principle stress direction.

Deformation with a high residual strength. Thicker material requires a greater absolute shear distance to induce the same post-yield features and residual strength. In the absence of numerous Principle Displacement Zones it is possible that a high proportion of shears in the Riedel orientation represents the early deformation of a thick till deposit with a high residual strength. Thus, such microstructures would not seem to be implicit of steady-state glacial conditions. Bimasepic fabrics may develop when shear zones are constrained between, and initiated by, clasts. Soft clasts may break up through slip on Riedel and Thrust shears giving 'tails' without 'pervasive' flow.

Hydraulic conditions in deforming till. If developing shears in glacial sediments drain the surrounding material sufficiently it is possible that they may induce brittle behaviour in the local sediment recognisable through the presence of reverse drag features. A depth-dependant variation in these structures would be strong proof of a hydraulic gradient in glacial sediments. At high confining pressures Riedel shears are short, numerous, and at a high angle to the Principle Displacement Zone. Large numbers of anastomosing Principle Displacement shears may represent low effective pressures, with water contents above >25%.

3.6.2 Potential glacial situations

Changes in till permeability. Till permeability is likely to be dependent on the origin of the fluid flowing through it and the sediment. Such variation will be enhanced by the solution and redeposition of till constituents. The reduction of large inter-aggregate pores is the main change in porosity in aggregated sediments. This is likely to occur early through shear in proglacial or subglacial material. If glacial consolidation occurs, the ratio between horizontal and vertical permeability would range between five and zero. Small heterogeneities in sediment bodies may have a large drainage effect as well. On a larger scale, it is implicit in the formation of injection and gravity instability features that they bridge between two differing sets of drainage potential. The production of microchannels and slickensides during subglacial shear may alter permeability.

Sediment deformation processes. Stick-slip motion has been found in very thin cataclastic laboratory specimens of material similar to glacial sediments, and clast interactions may be responsible for such behaviour. Permeability and strength effects may combine to give complex rheological responses under glaciers, with dynamic seals developing across shear zones, or shears simply inhibiting the depth of deformation. Equally possible is that effective pressure/water system changes might ‘reset’ the sediment, clearing features. Similar resetting could be completed by clasts falling through soft sediments or caught in shear zones. The presence of ‘tails’ from soft clasts shows that they supported stresses up to their yield strength, even though they were imbedded in a ductile medium. Comminution of such material will affect glacial deformation. It is likely that the stress field changes considerably, both temporally and spatially during a glacial period. It is possible that the necessary reorientation of shears induces some shear hardening.

The rest of this thesis will examine a series of thin sections from various sites around the British Isles and laboratory tests. The information in this chapter will be used in the next two chapters to reconstruct the history of one site in North Wales. At this site there are examples of sediment that has responded to surrounding conditions passively, but also material that shows signs of having controlled the movement of the glacier over it. The former case is explored in the first of the next two chapters.