5. The Ploughing and Lodgement of Clasts: a field based case study and theoretical model

5.1 Introduction

Subglacial till is assumed to form by the deformation of pre-existing deposits, meltout of sediment from the ice, and the lodgement of sediment trapped in the base of the ice that is ploughing through the top of the till. The mechanism by which ploughing material becomes lodged is only poorly understood, chiefly because the rheology of till is uncertain. The present qualitative hypothesis for lodgement is that it occurs by the development of a prow of sediment in front of the ploughing object or by collision with other clasts (Boulton *et al*., 1974). There is evidence to suggest that prows *do* develop (Clark and Hansel, 1989; and references therein). However, as subglacial effective pressures are usually considered low, and till easily deformed, it is not clear how prow development acts to lodge clasts, particularly where clast density is low and collisions are therefore infrequent.

Present quantitative models of lodgement assume it occurs once the force on a clast drops below that needed to cause Mohr-Coulomb failure in the down stream sediment (*Equation 3.1*) (Brown *et al*., 1987). Such models do not provide a *steady state* method for the development of lodgement till, and assume the rheology of the sediment to be perfectly plastic. It has been shown that such models tend to lead to a pervasively deforming bed (Alley, 1989), indeed, it is implicit in some models that the stress necessary to cause deformation is transferred by ploughing (Alley, 1989). Deforming beds are often considered to have a low friction at the ice-sediment interface (implicit in Alley, 1989), and to be weakened by the low effective pressures acting on them (Boulton and Hindmarsh, 1987). Under such conditions basal frictional melting will be reduced, and it is unclear how a thick deforming till bed builds up. Under such models till thickness is limited to the height of the largest clast in it which can remain fixed relative to the ice and cause the regelation deposition of fine material. Such thin layers are at odds with the great till plains of the Northern Hemisphere.

This chapter will focus on a suite of micromorphological structures associated with the ploughing and lodgement of a clast, with the processes these structures suggest being built into a quantitative model of clast lodgement. The first half of this chapter will examine the microstructural evidence of ploughing and provide a suite of process-related features that can be used for the determination of lodgement in specific circumstances. The second half of the chapter puts forward a model of the processes to examine their implications, investigate their relative importance, and quantify the subglacial conditions necessary for their action.

The clast in question has lodged into the up ice ('stoss') side of a shallow, Devensian, subglacial ridge or hummock located near the town of Criccieth on the Lleyn Peninsula, Wales (*Figure 5.1*). The outcrop scale sedimentology of the feature has been discussed in detail in Chapter Four, however, an extensive discussion of some aspects of the lithofacies containing the clast (lithofacies E, see last chapter) has been left until this chapter because the techniques used in the following analysis are both considerably different from those used in the last chapter, and better served by being presented with the data they are developed from because of their complexity and length. The sediments discussed in this chapter are also not representative of the sediment of lithofacies E as a whole. A summary of the environmental interpretations discussed in the last chapter is repeated in Table 5.1. The following section indicates the stratigraphic position of thin section samples within the sediments. The sedimentology allows an estimation of the post-glacial alteration of any small-scale structures. These alterations need to be assessed before one can investigate glacial effects. The following section therefore also details such an assessment.

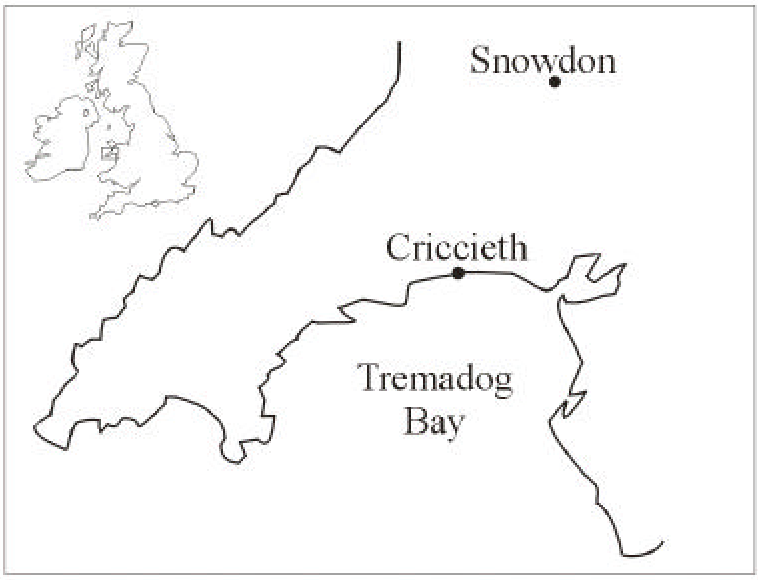


Figure 5.1 The Lleyn Peninsula, North Wales (inset of the UK).

5.2 The samples and their possible post-glacial disturbance

The exposure scale sedimentology of the area is given in Chapter Four, and is repeated in Table 5.1 for convenience.

Lithofacies	**Possible origins**	**Reasoning and caveats**
A	Periglacial loess	A micromorphological study would probably not confirm this, the less permeable materials below having encouraged subsurface flow illuviation and particle reorientation.
B	Periglacial solifluction material	Evidence of mass movements and small (~1m wide) stream deposition. Fills frost wedge (3, Figure 4.3) and is involuted at surface.
C	Periglacial cryoturbation of lithofacies B and D.	Random fabric with increased angular slate content. Frost wedge at 3 of fig.4.3 suggests depth for periglacial action. West of Castle Rock, the boundary with D has clast concentration suggesting fluidization of C or a fluvial event.
D	Glaciogenic sediment flow (Boulton, 1977; Grant, 1990) or lodgement till (Saunders, 1968), or flows under lake/marine conditions (Matley, 1936).	Clast orientation eigenvectors cover the first two possibilities (Grant, 1990). See Chapter Four for further discussion.
E	Subglacial till (Grant, 1990; this chapter)	It seems unlikely given the exposure of the site that this formed in a pre-Dimlington stadial (pre-δ^{18}O Stage 2). See Chapter Four and below for further discussion.

Table 5.1 Sedimentology of Criccieth bay middle deposit (see Chapter Four for further details).

The thin sections were taken from an area where the cliff trends sub-parallel to the lithofacies E clast orientation of 135-315 (Grant, 1990) (Area 5 on Figure 4.3). The samples were taken from the boundary between lithofacies E and lithofacies D, immediately up-ice of a large boulder (shown in black on the figure). At this point the blue grey lithofacies E alternates with sand lenses (~1 x ~5 cm) and a light yellow diamict in thin (<5 cm) bands. This alternation occurs in a zone approximately 50 cm thick (*Figure 5.3*) which becomes a wave-resistant band (30 cm thick) in the up-ice direction (*Figure 5.2*). This band runs approximately 20 cm below the surface of the blue grey diamict and extends some 15 m up-ice, disappearing under the beach with the blue grey diamict. A land-slip made extensive mapping to the west impossible, but immediately west of the boulder homogeneous blue till leaves the top of the erratic and drops to the west. This suggests the interbedded sediments do not extend to the west of the boulder.

Figure 5.3 Interbedded sample area from the Criccieth deposit. Material is a mix of yellow and blue silts and clays, interbedded with short (~50 mm) sand lenses. Object in centre is a food tin lid ~7 cm.

Figure 5.2 Sampled feature. Note the blue diamict at beach level and the large boulder within the blue diamict to the left of the picture. Material was sampled from interbedded material 1m right of the boulder, and from the blue diamict just below.

Thin section samples were taken from the interbedded material and from lithofacies E. On the microscale, it will be shown that this deposit represents the trace of the boulder as it ploughed. Given the environmental attribution outlined in Chapter Four, it is possible to indicate the potential postglacial alteration of the sectioned material.

The possible *postglacial* disturbances to the samples include periglacial cryoturbation, deep-water alteration and mass movements, and compression (Chapter Four). There is evidence at Criccieth that lithofacies E is partly a glacitectonized laminated mud (*D.Evans, pers.comm., 1997,* see also Chapter Four). However, the latter three processes can probably be discounted for the *postglacial* period here, as the samples are taken from an area of lithofacies E that includes interbedding which could only be formed subglacially (see analysis below) and which has not been disturbed. Cryoturbation can also be discounted as the uncryoturbated lithofacies D overlies the sampled area. Microscale periglacial changes that have been reported elsewhere (Chapter Two) are not seen here. It seems likely that such microstructures as *are* found could not have formed if the ground was frozen as they require free water. They are, therefore, extremely unlikely in a sediment mass undergoing permanent permafrost conditions.

5.3 The microstructure

5.3.1 Thin section description

Two thin sections were prepared from the interbedded zone immediately up-ice of the large boulder, from the west-east vertical plane (samples 3a/b). Two thin sections were also later prepared from lithofacies E to test the model suggested here. These have also been discussed in terms of the general sedimentology of lithofacies E in Chapter Four. The sections 3a and 3b show that the alternating banding seen in outcrop was a result of the interbedding of three sedimentary types.

1) Sands (100 point-count average length of ~0.3 mm) of schistose material. This is found in bands with clean pores constituting 45% of the sediment (*Figure 5.4*).
2) Silts in unimodal, graded, and melange units of various sizes (including some clays) and internal fabrics. These silts form the blue and yellow bands seen in hand-samples (*Figure 5.5*).

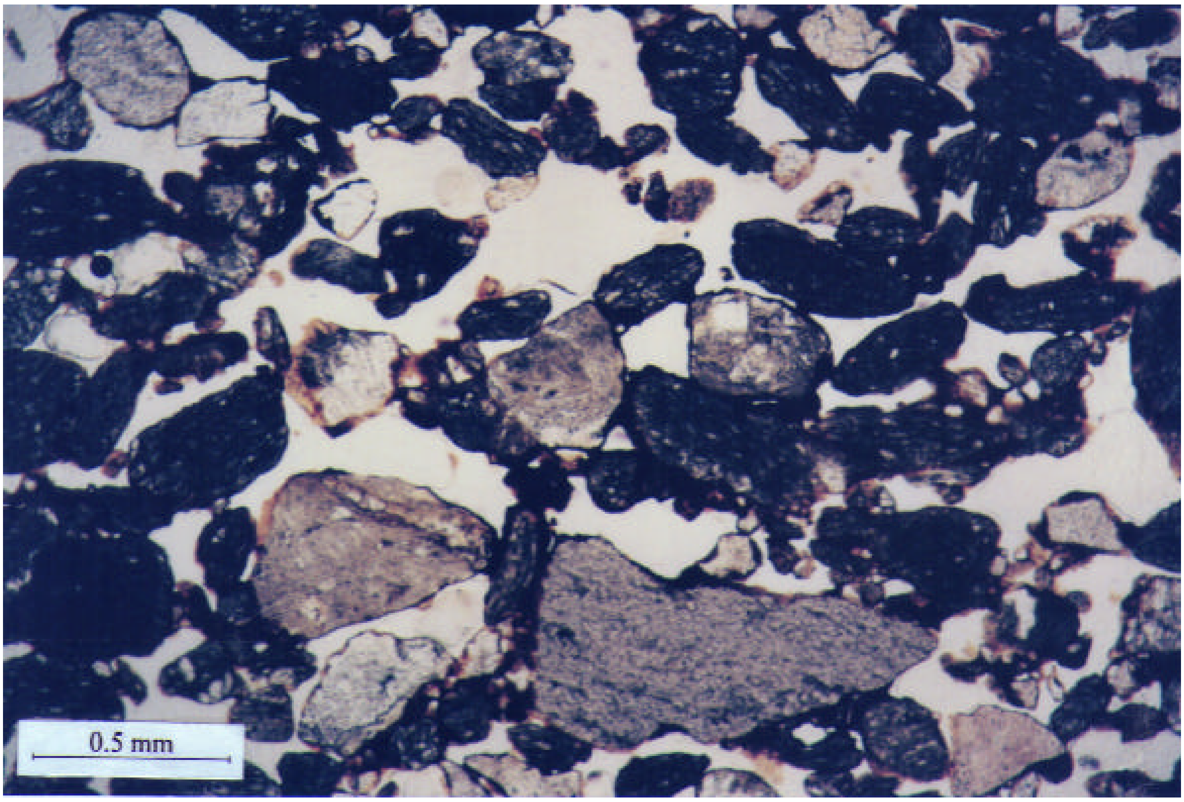


Figure 5.4 Photomicrograph of a sand layer from the base of section 3a. Unpolarized light. West-east vertical plane.

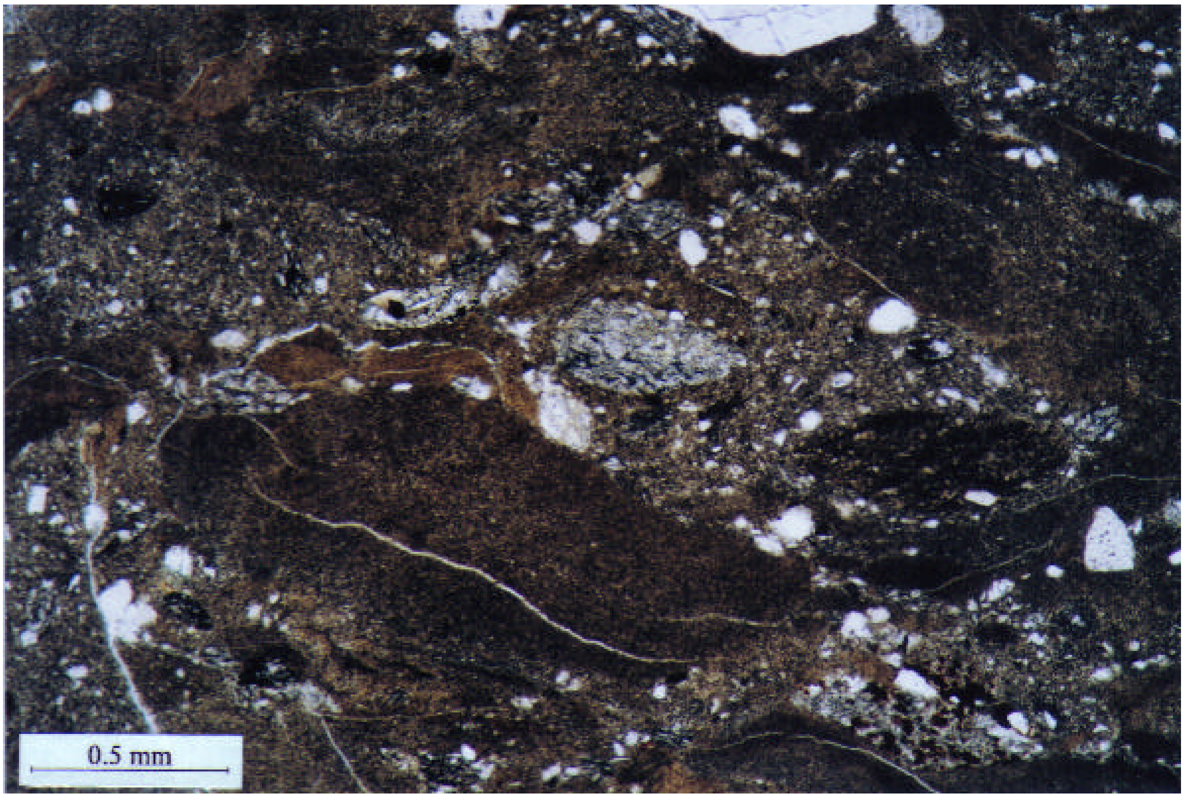


Figure 5.5 Photomicrograph of a melange of silts of various sizes mixed with larger quartz grains. From section 3a; unpolarized light conditions, west-east vertical plane.

3) Quartz grains (~0.1 mm) in separate masses and as units between/in other layers (*Figure 5.6*).

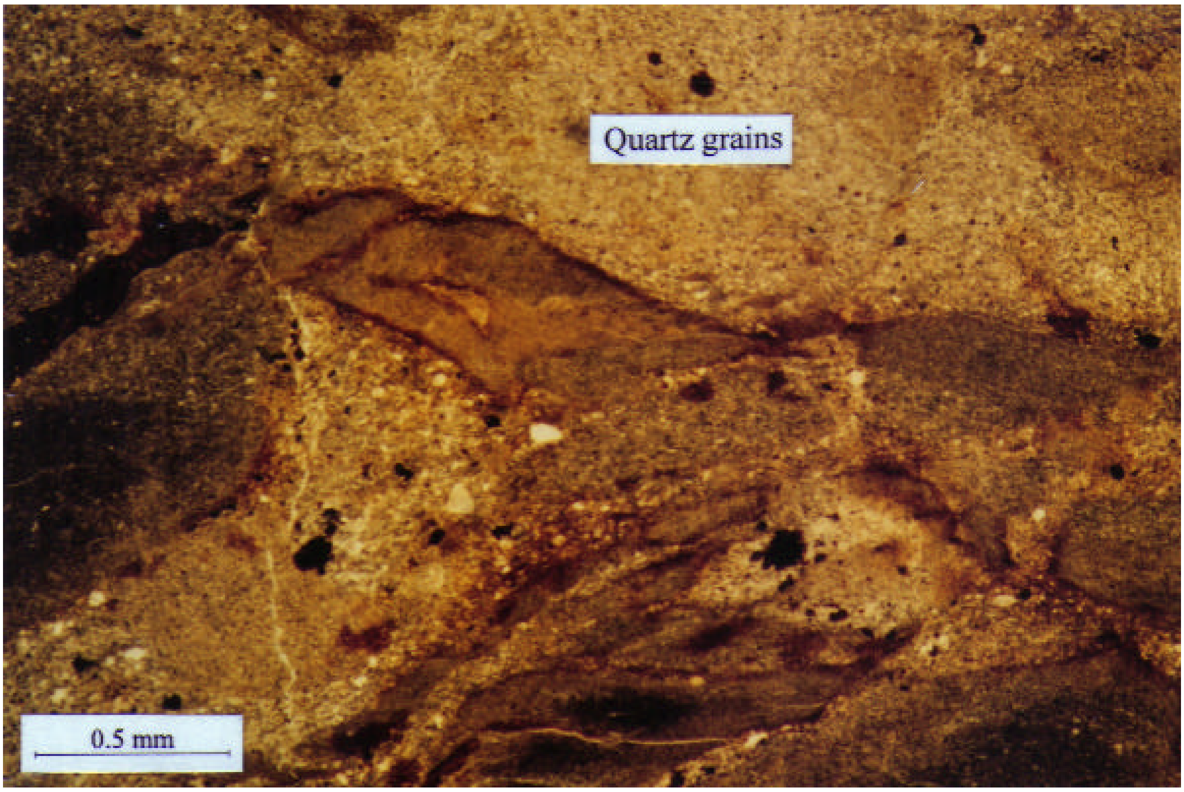


Figure 5.6 Photomicrograph of a quartz grain unit from section 3a (lower third of photo). Unpolarized light. West-east vertical plane.

The boundaries between the different bodies are discontinuities but are not formed by post-depositional shear; instead they are deep undulations. There *are* shear fabrics extending from some boundary irregularities at the top of these undulations, however, in most areas the material moves into the topography of lower sediment bodies. In places the boundaries undulate in response to the intrusion of discrete sediment units from above, suggesting deposition by displacement in semi-coherent flows (*Figure 5.7a*).

The sands were deposited in bands with little or no silts in the pore spaces (*Figure 5.4*). The sands are also weakly aligned horizontally. In one place the sands are overridden by silts, the two units being mixed together in a 4.3 mm zone indicating that the clean pores are probably due to washing of water rather than protection by ice fills during deposition. In places in this mixed-zone, slugs of silt are armoured with sand grains indicating the edge of flow units (*Figure 5.7b*). While it is impossible to determine the true flow direction of these features, it is apparent from the associated extensional features (*Figure 5.7b*) that they did not move east-west.

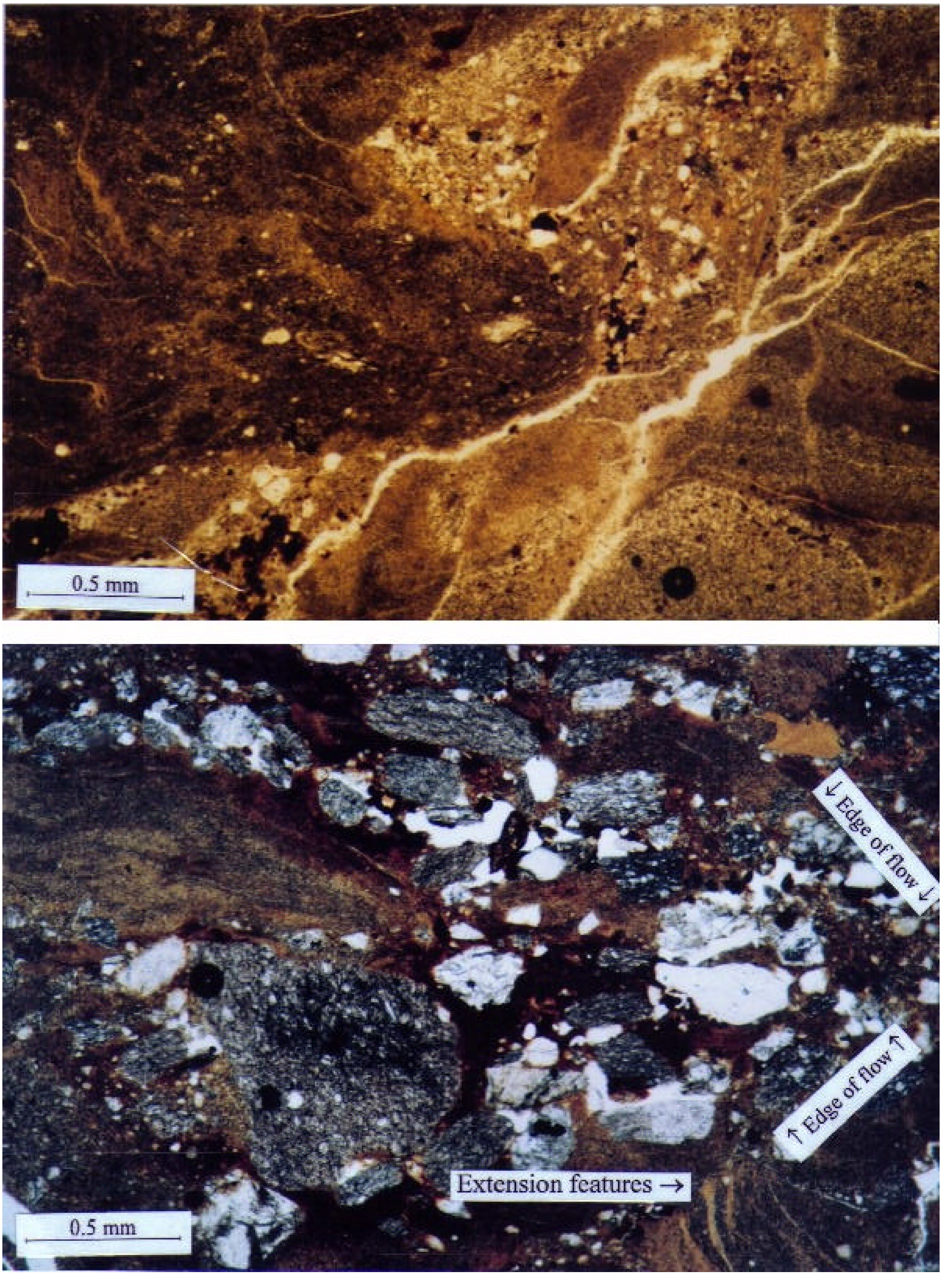


Figure 5.7 A. Photomicrograph of a silt unit that has moved into a quartz rich unit from section 3a. Unpolarized light, west-east vertical plane. B. Photomicrograph of a silt slug nose, armoured with sand grains, from section 3a; unpolarized light conditions, west-east vertical plane. Tension cracks, filled with light yellow, fine, material can be seen defining the lower forward edge of the nose.

The quartz grains were found throughout all the other sediment types except the sands, suggesting they may have winnowed out of the sands in a water body. The quartz grains are especially prevalent at the boundaries of sediment bodies indicating fallout deposition between mass movement events (*Figure 5.8*). However, the quartz grains also form units that moved with the surrounding mass movements and filled dewatering veins.

The silts can be subdivided into three types of bodies.

a) 'Typical' melanges consisting of blocks of coherent sediment within a groundmass of deformed material (*Figure 5.5*). Neither the blocks nor the 'matrix' are of one specific silt size. The rotated and varying fabrics in these bodies suggest the material did not suffer glacial consolidation after deposition (*Figure 5.5*). However, the strong internal fabrics of many of the blocks suggest overburden or shear alignment of individual grains *prior* to the melange formation. It is possible that the strong pre-deposition fabric was resistant to reorientation, and that these pre-orientated areas protected the weaker matrix. Thus, prior consolidation is certain, but post-depositional conditions suitable for consolidation cannot be ruled out.

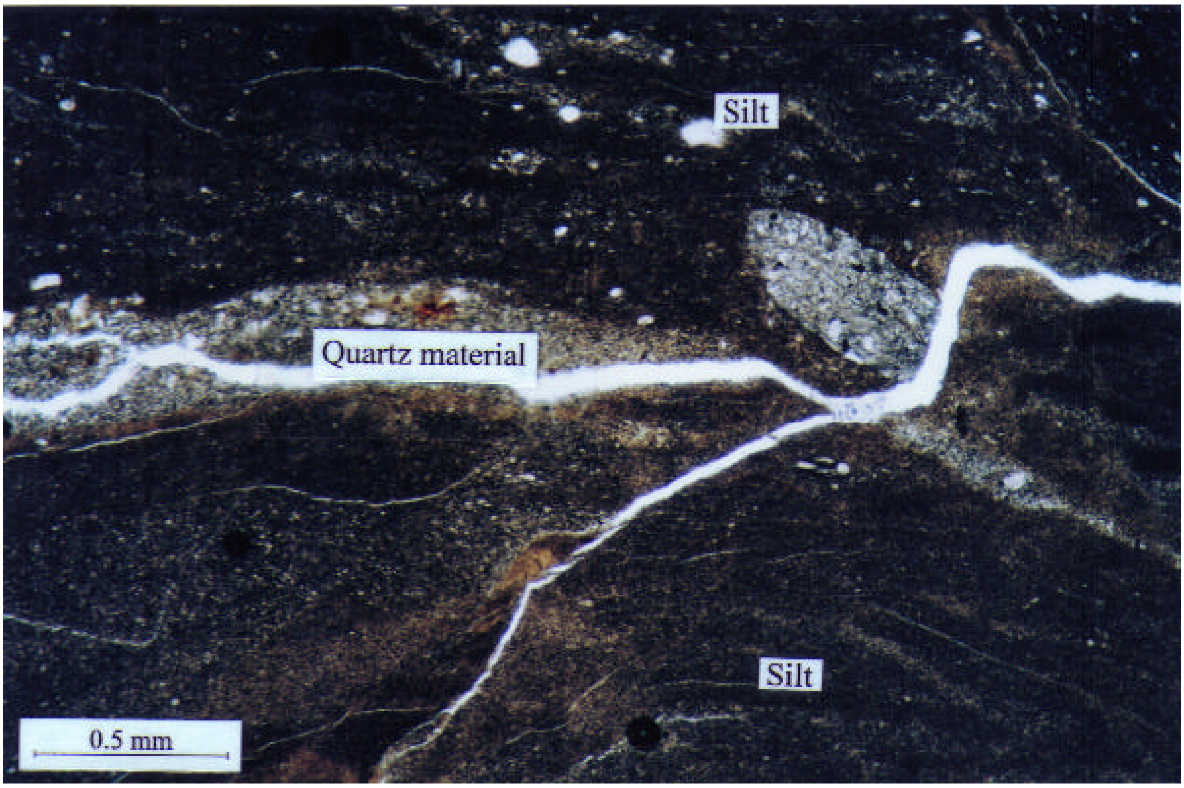


Figure 5.8 Photomicrograph of two silt slugs (blue in hand specimen) with an attenuated quartz layer between them, from section 3a; unpolarized light conditions, west-east vertical plane

b) Flow bodies with a unidirectional internal fabric. Between these flow bodies are, coarse grained quartz units which have been extended by shear by the moving units (*Figure 5.8*).

c) Graded units in overlapping lenses. These units have sharp boundaries between the coarse material of one lens and the fine material of the one above (*Figure 5.9*). The fine material has a single fabric orientation which is also weakly present in the coarse material.

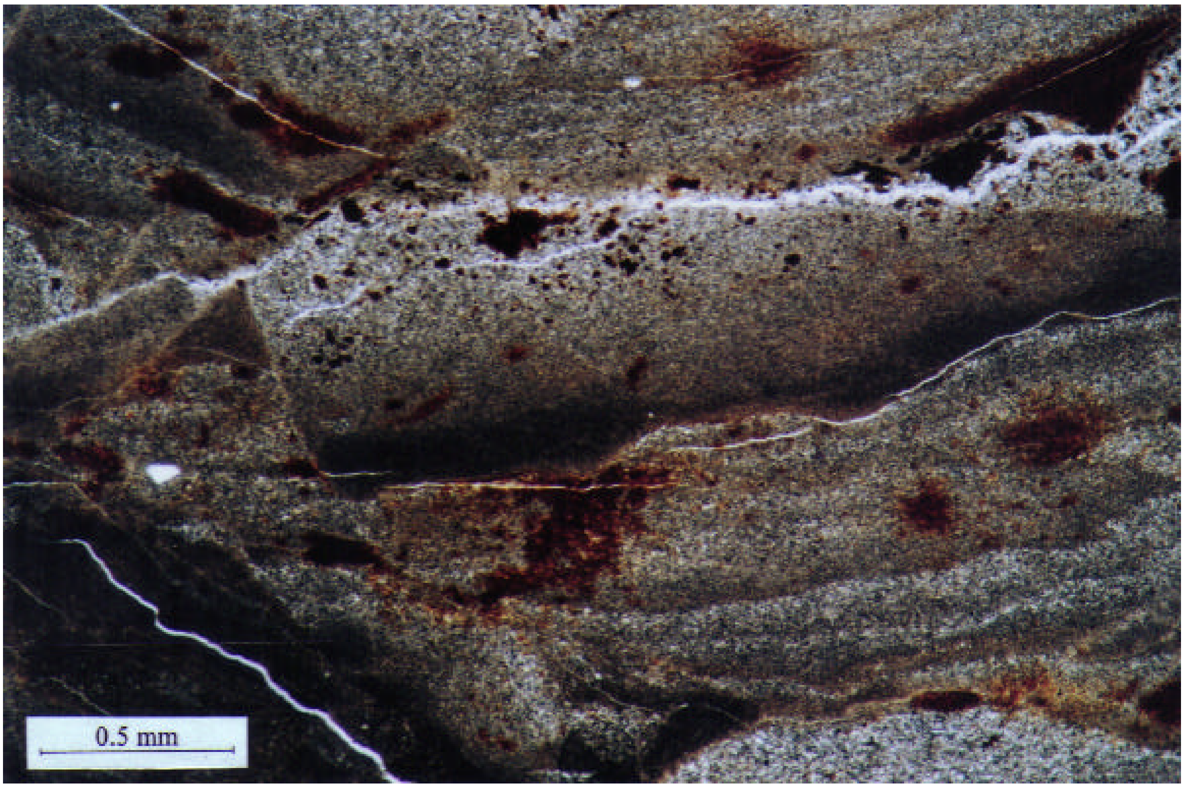


Figure 5.9 Photomicrograph of graded units from section 3a; unpolarized light conditions, west-east vertical plane.

5.3.2 Discussion of microstructural study

The fact that the larger grain sizes are *above* the fine silts in each graded silt 'unit' suggests these are not water-sorted features, most of which have fine material above the coarse, particularly in glacial environments where seasonal stratification leads to the long suspension of fines in the water column. While it is possible to suggest that the units were inverted in transport or the larger grains were expelled as the bodies were emplaced, a progression can be found which supports a deformational origin for these features. Figure 5.10a shows low strain fracturing of a coarse unit (fine-grained sand quartz unit) into lenses along listric faults. In the more developed features, shears can be seen to bound the lense-like areas (*Figure*

5.10b). This faulting and imbrication process causes the observed juxtaposition of fine and coarse material in lenses.

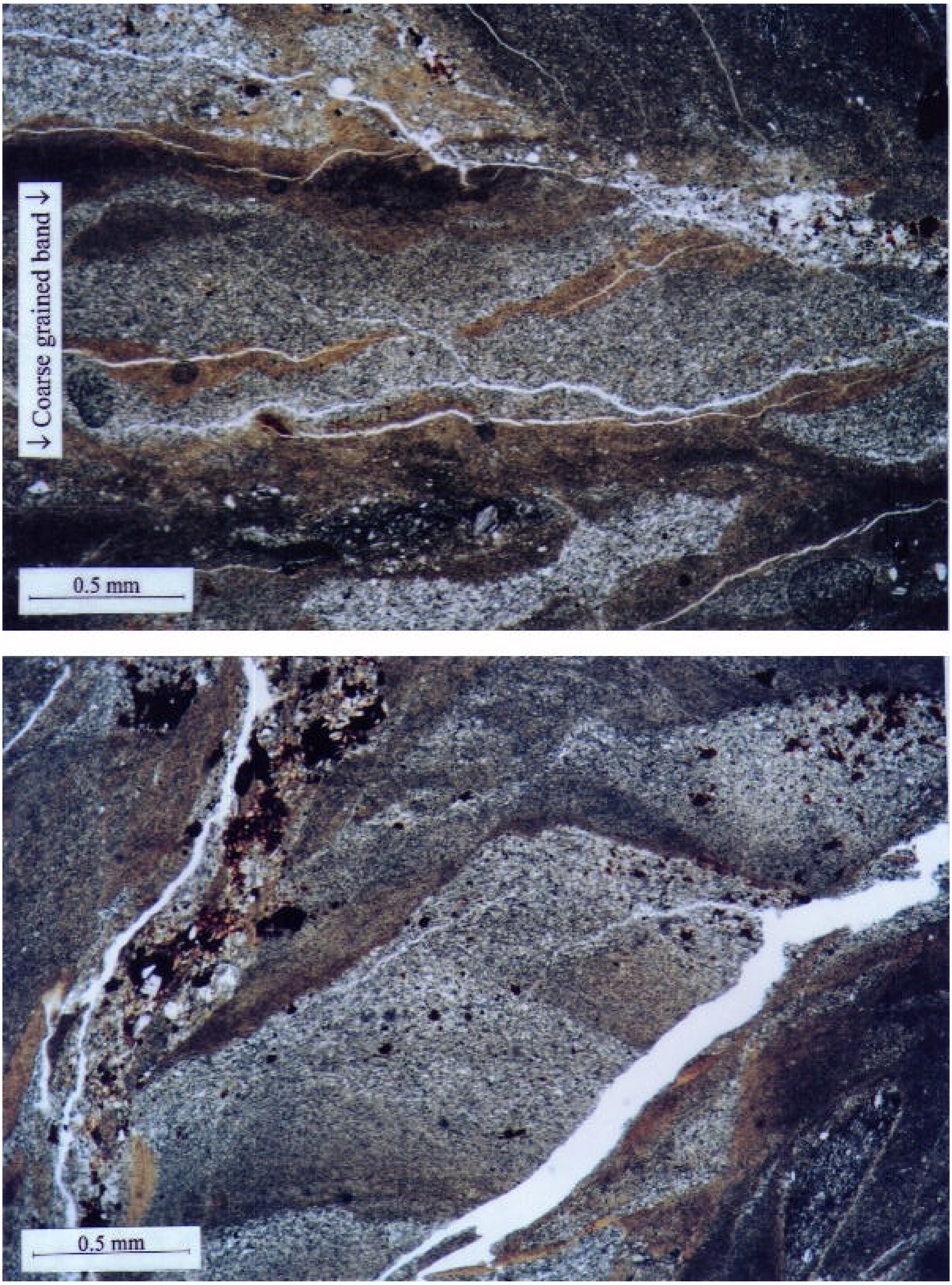


Figure 5.10 A. Photomicrograph of a coarse band of material breaking up. B. Photomicrograph of the same band as in (A), only here showing greater deformation. Listric shears have developed and the band has been compressed. The listric shears give a lens-like appearance to the coarse material. Both photos are taken under unpolarized light in west-east vertical planes.

It can be suggested that free water movement through the coarse units as the listric areas dilate and collapse allows sharp faults to form. In the post-faulting situation, the juxtaposed fine sediments restrict the fluid flow. This may lead to a reduced effective pressure close to shearing areas, and more pervasive movement. This pervasive movement would provide the mixing necessary to give the units' 'graded' appearance.

There are a number of other important features of the sediment that require explanation:

1) melanges of silt blocks with prior developed fabrics;

2) clean areas of sands;

3) quartz grain bands that may have suffered winnowing;

4) the movement of coherent sediment bodies into each other by small scale displacements.

There are a two possible situations accounting for these features.

1) Lodging clast gouging, and lodgement induced melt-out (*Figure 5.11*).

a) The boulder immediately in the down stream ice direction from the sediments described above (see *Figure 5.2*) gouges through pre-existing diamict.

b) Diamict moving from in front of the clast is joined by diamict flows from the sides of the gouge. These processes go some way to forming the 20 cm of blue-grey diamict above the 15 m + of harder material extending up-ice from the boulder.

c) The clast starts to lodge as material piles up in front.

d) The pressure of the ice on the clast causes pressure melting and water fills some of the gouge behind the clast. Meltout debris intermittently mixes with the reworked lower diamict. Smaller quartz grains rain out more slowly.

e) Lodgement of the till continues. As the large boulder is covered, normal ice flow can resume, with all the above processes acting, but on smaller particles. The particle size means that this later sediment is less protected and, thus, more mixed. This gives a homogeneous till which then suffered shear consolidation.

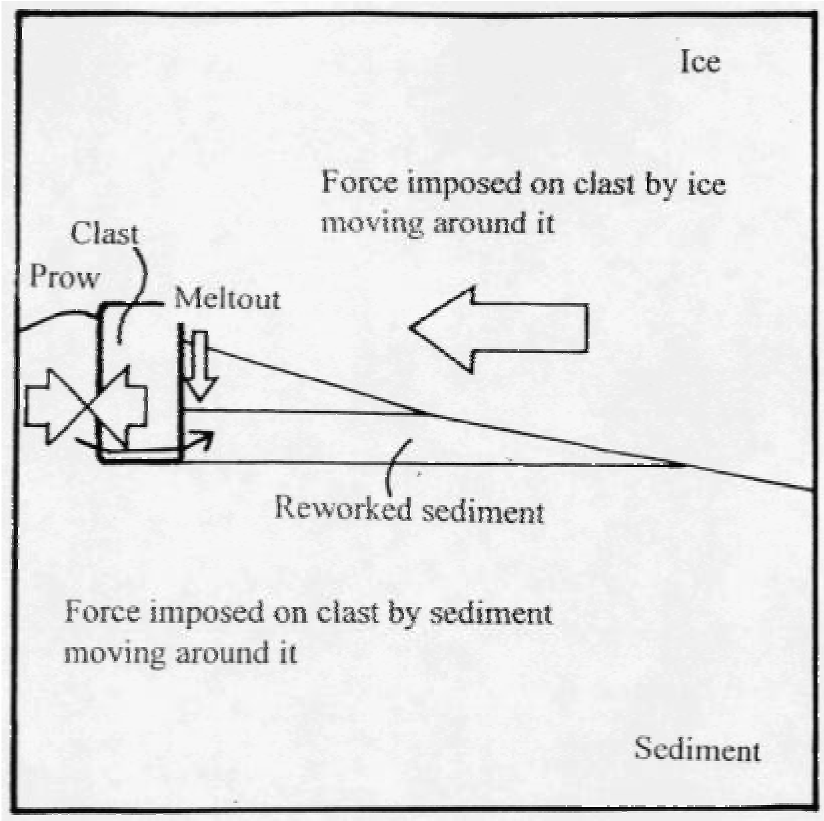


Figure 5.11 Processes involved in the lodgement model.

2) Pressure pumping of sediments.

a) Water pressure variations in distant till pump material from the upper diamict down shear zones and into the lower diamict or the ice (for mechanism see Talbot and von Brunn, 1987).

b) This mix of basal till and ice melts. High discharge meltwater moves preferentially through the diamct concentrating sands and washing them. Quartz grains represent till bodies with fines removed where escape could not occur through sands. These unimodal units may also have been deposited as such from the ice.

It is the macroscale form of the deposits which allows the attribution of the microscale features to process (1) with greater probability than process (2). The amount of meltout material in (1) would be expected to increase as the clast became lodged. The resistant layer seen in Figure 5.2 widens up and becomes more sandy towards the clast, so that the geometry strongly supports hypothesis (1). It is possible that the resistance of the layer is partly caused by consolidation under the lower part of the ploughing clast. The ice pushed the clast west and lodged it against a massive till prow, with meltout thickening the mixed zone only in the very last stages. It should be noted that the micromorphology of the sediments indicates a 'warm' (unfrozen) bed at the sample point, however, the inclusion of the silts with a prior fabric, and the general form of the feature on a macroscale indicate that it did not form through simple 'meltout' *alone*.

The ploughing hypothesis allows a prediction of the nature of lithofacies E under the resistant layer. The material is expected to show;

1) a generally homogeneous mix of material from clay to sand sizes;
2) strong shear fabrics.

Under the glacier pumping hypothesis we would expect to still see some injection features and areas of fabric disruption which may develop prior to throughflow winnowing. We may alternatively see a very heterogeneous meltout sediment under this model.

Two orthogonal, vertical samples (2a/b) were taken to compare with these predictions. They showed homogeneous material, though largely without sand-sized material, and a very strong shear fabric (*Figure 5.12a/b*) with no evidence of sediment intrusion. The evidence for lodgement is therefore good, though the sand sized material must probably be regarded as exotic.

It seems likely that the extreme size of the boulder in question meant that sufficient sediments were produced to survive further, limited, overriding. The low post-depositional reworking suggests the deforming layer at this time was thin and/or the deposition was towards the end of the glaciation. Model evidence (below) suggests the till was weak, which may indicate a thin deforming layer due to very low effective pressures and/or the low clast concentration of the original till. The material examined above may also have been protected by the introduced sand-sized material (which has a higher angle of friction). It is equally possible that the unaligned silt and clay material seen in the thin sections of the interbedded material was more resistant to shear. Indeed, it is interesting to speculate that there is a feedback here which could mildly enhance the further lodgement of clasts.

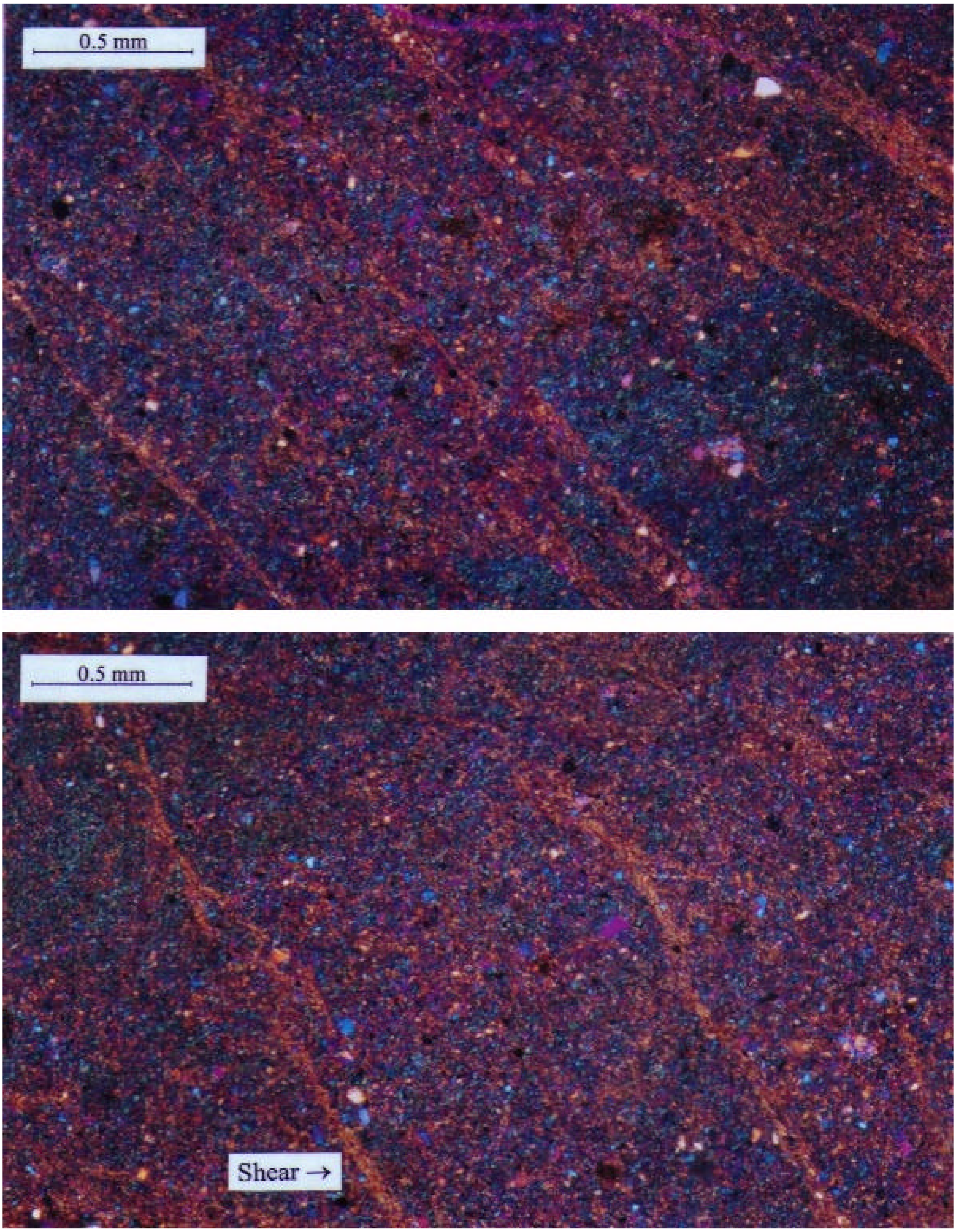


Figure 5.12a/b Sections from lithofacies E below the interbedded material (Sample 2a). Cross-polarized light with tint plate. West-east vertical plane.

The strong macroscale clast fabric makes it unlikely that simple meltout without reworking was a significant factor in the deposition of the whole of the two diamicts (Grant, 1990, see also Boulton, 1977). While similar areas of interbedded material *are* found elsewhere on the ice-upstream side of the outcrop, they are rare and discontinuous. This picture is in line with explanation (1) though localised simple meltout may also have produced some of these interbedded sediments (see Boulton, 1977). It is suggested that the microstructures seen here *are* associable with simple meltout, with the important exception of the pre-aligned slugs of silts and clays. These indicate the reworking of material that has already been deposited. It is possible such alignment occurred in the sediment while trapped in the ice. While shear occurs in sediment trapped in ice (for example, Echelmeyer and Zhongxiang, 1987), we have no knowledge of such sediment's microscale alignment. It seems likely that in the case of stagnant meltout, the sediment will lack cohesion and mix to form more homogeneous deposits (Paul and Eyles, 1990). A fluvial origin for the material can be dismissed as such a situation is unlikely to produce cohesive sediments horizontally juxtaposed with clean sand bodies and substantially winnowed material. Each sediment type suggests a differing fluid flow rate and sediment concentration. One must also remember that this feature is inextricably included in the lower diamict.

If the ploughing boulder interpretation (*Figure 5.11*) is taken as correct, it may greatly elucidate the process of clast lodgement. The inferences that can be drawn from the work so far, and avenues of further research, are outlined below.

5.4 Modelling

5.4.1 The qualitative model

There are four important stages to the clast lodgement process implied by the till flow and pressure meltout-induced microstructures discussed above. The potential contribution to the process from other meltout is discussed in the model assumptions (below).

1) The clast initially moves with the ice. As there is no velocity difference between the two, there is no longitudinal stress imposed on the clast by the ice. Thus, there is no horizontal net pressure gradient across the clast and no pressure melting ('regelation') or pressure-enhanced creep associated with the boulder (contrast with Weertman, 1957). Provided

there is some decoupling between the ice and the lithofacies E (which is anticlinal in section - *Figure 5.2*) the ice creep over the anticlinal hummock will probably be at a maximum at the hummock's top in the same manner as the 'streaming' effect around obstacles seen by Boulton (1979) (uncertainty is introduced by our ignorance of the three dimensional form). The thinning of any debris-rich ice with velocity increases the chance of vertical clast-clast collisions. These collisions may push clasts up into the ice and down into the stoss side of the soft till hummock. That is, clasts will plough horizontally into the forms because of the enhanced creep around the obstacle. This hypothesis of horizontal clast release is within the constraint on the release of clasts found by Iverson and Semmens (1995). Iverson and Semmens showed that meltout of clasts from ice will only occur if there is clast-clast contact bridging a determinable thickness of basal ice. This constraint needs to be satisfied in some manner before the model can be used as a flat-bed analogy (below, see also Philip, 1980; Iverson, 1990; Hallet, 1979; 1981; and Shoemaker, 1988, for the arguments leading to the development of this constraint).

2) The clast moves partially into the sediment. The difference between the flow of the till around the ploughing clast and the ice velocity will cause a stress to be imposed on the clast. This stress will result in regelation and ice creep around the clast. These processes produce the microstructures seen above. There is a problem here, in that we would expect till to be much softer than the ice and lodgement to be unlikely. Several factors may increase the till resistance. These are discussed further below.

3) As the till resistance increases the clast will start to lodge and the stress on the clast will increase. The till flow into the gouge behind the clast will be combined with meltwater and meltout material produced by regelation. This is in line with the regelation production of diamict matrix suggested by Kemmis (1981). At this point it is still considered that the ice-gouge interface could have a geometry such that the ice is fully in contact with the back of the clast having partly filled the gouge. However, the sediments filling the gouge will restrict the up-ice contact area between the ice and clast. This will increase the stress at this contact, in turn resulting in an acceleration of the pressure meltout. At the same time the clast-till contact area will be increasing, reducing the force per unit area on the yielding sediment in front, making lodgement more likely.

4) Meltout will continue to accelerate until meltout till and slumped out material from the gouge edges fills the gouge. At some point the melt and creep around the clast will equal the ice velocity, or the meltout material will cover the clast. At this point lodgement will have occurred. Normal ice flow will then take over. Lodgement of the till will continue, with all the above processes acting to produce the same mix of reworked matrix and meltout sediment, but on other, in this case much smaller, particles. With more numerous smaller particles, there is a greater chance of another small clast obliterating all the sedimentary signatures of the processes. Thus, the sediment is more mixed and homogeneous. This material is then shear consolidated during the remaining glaciation.

As mentioned above, it is difficult to explain lodgement in till which is assumed softer than the ice. Stress acting on the till will partly result in the flow of the till from in front of the clast into the gouged area behind. Under the present understanding of tills it is expected that this flow will be accentuated by the compressive stress in front of the clast temporarily increasing the water pressure in the till. Pore pressure in front of the clast will increase until the sediment yields pervasively or on shears. This increased pore fluid-induced weakening will continue until a balance is reached between clast advance and Darcian flow or discrete dewatering, with the fluid pressure reaching a steady state. It is possible that this weakening is catastrophic, however, we know that clasts do lodge, suggesting that processes that are more powerful are acting to increase resistance. Four processes are suggested as potential ways of increasing the effective till viscosity locally to the clast, all are feasible under our present understanding of the subglacial environment.

a) To maintain pressure induced regelation, melt must be balanced by freezing somewhere else to provide the energy for melting (Weertman, 1957). Freezing may be occurring in the sediment around the clast, the material becoming stiffer and increasing the clast's effective size in the till. By comparison with similar sediments we can reasonably expect the till to have a residual strength, that is, even when deforming, there will be some stress transferred across the clast, from the till, to the ice. This stress will initiate the regelation cycle. Freezing of water in the sediment rather than on to the ice (as in traditional regelation) would be dictated by the geometry of ploughing into a slope, and for the development of a prow in front of the boulder.

b) The residual strength of the till may cause a cavity to open down-ice of the clast as the clast slows in the sediment. The excess pore fluid pressure may be released into this cavity, with the excess meltwater from up-ice of the clast. Cavities downstream of clasts have been suggested by Boulton *et al.*, (1974) and have been seen by Boulton (1979). However, there is an observational bias towards marginal ice regions where the movement of ice into such features would be slow. The rate of ice creep closure and regelation refreezing in such a cavity would have to be less than the creep due to the fluid pressure increase keeping it open. The opening of such a cavity seems unlikely to initiate slowing when it is considered that a process has to be found that produces considerable slowing in order to develop a cavity. Thus, while cavitation may play a late-lodgement function, regelation refreezing in the till seems more likely to cause early clast deceleration. The relative importance of the two will relate to clast size.

c) A cavity opens in front of the clast, as above, and the soft till moves up into the space. This causes a flute as suggested by Boulton (1976) and the cavity re-forms further and further upstream, the flute acting as an extension of the clast. Thus, the area of dewatering and softened sediment is moved away from the clast, and stiff till forms just in front of the boulder. The same problem holds for initiating such a cavity as in (b). Prow size may provide detailed information on the relative till-ice rheologies, however, this chapter does not deal with this interesting avenue.

d) The chief mechanism of lodgement involves changes in the contact areas of ice and till due simply to the geometry of ploughing into a slope, or the development of a prow on a flat bed. This development will affect the stress transferred from the ice to the deforming till down-ice of the clast. The difference between this situation and that of the flute in (c) is that this geometrical change does not need to be initiated by the clast slowing relative to the ice, though this may well be a result of the change.

In this lodgement model, the clast slowly decouples from the ice, undergoes a transition period, and then becomes coupled to the till. This process of transitional decoupling may be a major process inducing deformation in the sediment and slowing the ice. It has been assumed in the above that the sediment has no effect on ice velocity. This may be the case for the small deposit at Criccieth, however, under more general conditions the process will feedback to link the till and ice velocities. It will be the ease of ploughing (the length of the decoupling transition) which determines the effect sediment has on reducing the ice-bed friction and increasing the velocity of the ice with respect to hard beds. It seems wise to construct a quantitative model of the proposed processes to test and explore their complex implications. A quantitative model will also allow us to determine whether these implications are consistent with our present knowledge of the subglacial environment.

5.4.2 The quantitative model

Initial caveat

Ideally we should model the till rheology as non-linear. Such an attempt is far from trivial. If the till were shown to be of a low effective viscosity in relation to the ice, the sediment could be modelled as deforming to release all the stress after a residual stress. This would remove the difficulties of non-linear modelling, while allowing the till to impose some stress on the clast, that is, giving the till a residual strength.

The effective viscosity of ice flowing around a clast can be calculated using the equations quoted by Weertman (1957; 1964; 1979). In Weertman's model the ice flow around an object of equal height and width transverse to flow and a given length is equal to the sum of the creep and regelation around it.

To set up the regelation component we assume there is a force difference (s) across a boulder of width and depth (w) and height (h) resisting ice flow. The stress is split between an upstream, compressive, component and a downstream, extensional, component, each with a value of $s\ /\ 2wh$. The stress difference across the boulder can be converted into an equivalent change in the melting point of the ice, $C\ (s\ /\ wh)$ where C is a constant equal to $7.42\text{x}10^{-5}$ K kPa^{-1} (Weertman, 1957). This causes ice to melt upstream and refreeze downstream. Following Weertman's derivation we gain the total volume of ice melted per unit time (m) by a longitudinal stress as given by

$$m = \frac{a\,s\,C h_i K}{H r} \qquad \text{...Equation 5.1}$$

giving the melt rate per unit area of clast ($\dot{m}$) as

$$\dot{m} = \frac{a s CK}{H r w}$$ *...Equation 5.2*

where h_i is the height of clast out of sediment, K is the conductivity of the rock (taken as 0.1194229 J °C s^{-1} cm^{-1} following Weertman, 1957), H is the heat of fusion of ice, r is the density of ice, a is a constant dependent on loss of heat into the ice rather than the rock.

To set up the ice creep component we assume the stress causes the creep of ice around the resisting boulder at a rate ($\dot{c}$) determined by a variant of Glen's flow law, the original form of which is,

$$\dot{c} = B s^{n}$$ *...Equation 5.3*

where B is taken (for 0°C and 50-130 kPa stress range) as 5.5 x $10^{-15}s^{-1}$ kPa^{-3} (table 3.2 of Paterson, 1981) and n is 3 or 4.

Weertman suggests the creep of the ice around the obstacle/boulder per unit area per unit time $\dot{c}$ is given by the variant of Glen's flow law,

$$\dot{c} = \frac{2}{9} B d s^{n}$$ *...Equation 5.4*

where d is Weertman's 'distance of action', taken here as one clast width in line with Weertman. The constant arises from the uniaxial nature of the force application (see p.32 of Paterson, 1981, for the assumptions involved).

The total velocity of ice past the clast (v) is thus the creep and melt rate combined. The equation for total velocity can then be used to calculate the effective viscosity (h_e) from

$$h_e = \frac{F}{v}$$ *...Equation 5.5*

This value can be compared with the effective viscosity for a sphere moving through a non-regelating Newtonian fluid (the till), calculated using Stoke's equation,

$$F = 12 p h_t R v$$ *...Equation 5.6*

where h_t is the viscosity of the till. Figure 5.13 shows the till viscosities at which movement around, and force on, a clast moving through till is equal to that around and on a clast of the same size moving through ice. This rather complicated set of values is taken to represent the till viscosities at which non-linear till rheology effects will start to become very apparent during lodgement. It is considered that if the viscosities of tills found in the field fall higher than these values the present model is a less suitable model of till behaviour. Plainly this comparison has no absolute mathematical basis as the two models are unrelatable for reasonable values.

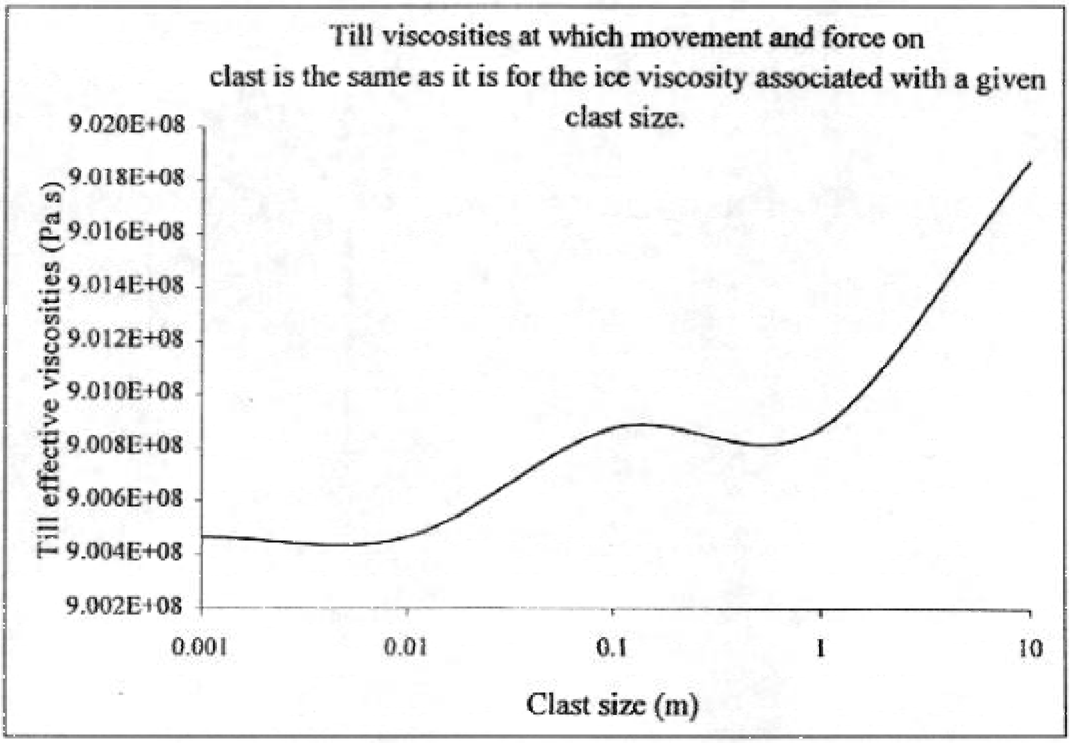


Figure 5.13 Till viscosities at which deformation is equal to ice deformation dependant on clast size for various ice viscosities.

Till viscosities measured in the field range from extremes of 10^8 to 10^{11} (Paterson, 1994). Thus, it can be seen that the model is a poor representation of reality for all but the weakest, least viscose tills. It is not a simple matter to calculate the stress supported on the clast in a non-linear viscose material. It must be accepted that the model presented here is the simpler end member of a continuum also dependant on the influence of sediment in the basal ice on the ice rheology. However, the model results suggest that in *this* case, the till has a low residual strength, and its low clast density may make it act in a manner closer to that of the model (these ideas will be enlarged upon in later chapters).

Assumptions

A number of important assumptions will be made in the model of ploughing into a slope that can be utilised to produce a flat bed analogue at a later date.

The forces on the clast are simplified here such that the clast is driven horizontally into a slope. There are two components of this simplification that may need to be altered in order to produce a model corresponding to other real situations. Firstly, the local slope may be different, indeed, may be horizontal. Secondly, the clast may be driven into the slope at a different angle. Taking the local slope first; it should be remembered that the primary process suggested for the lodgement of clasts is the resistance offered by a sediment prow. Even if the slope is horizontal, it has been seen that prows can develop (Clark and Hansel, 1989). The development of such a prow is determined by the sediment hydraulic diffusivity, internal friction, the clast size, and the stress acting on the sediment.

As a simplification it is presumed that the strength of the material is affected little by being pushed into a prow. Triaxial experiments presented in Chapter Six suggest work-hardening in sheared till at low total strains. The limited plough-meter evidence from under modern glaciers (Porter *et al*., 1997) shows force increasing as objects trapped in the ice move through sediment. However, this is usually interpreted as ice velocity changes rather than the development of resistance. It seems wise at this stage to take the simplest model and assume there is no hardening dependent on large strains. This assumption is easily remedied later as it can be handled with a variable in the calculation for the residual strength of the till. The stress build-up is therefore controlled by the changing contact area of till to clast including the prow effect, and may be calculated geometrically. The prow can reasonably be approximated as a slope angle local to the clast, provided the material is not allowed to flow into the gouge behind the clast. Thus, even on a horizontal bed, ploughed into horizontally, there will be a local change in contact heights that can be visualised as a slope.

Moving on to the direction the clast ploughs in; this angle is determined by the velocity parallel to the local sediment slope surface and the velocity normal to the slope. Slope-parallel movement is the enhanced creep rate minus the gravitational pull of the clast in the opposing direction. Slope-normal movement is the melt rate local to the slope, plus the gravitational movement in this direction, minus the regelation of the ice downwards around the clast normal to the slope (Hallet, 1979; 1981; Shoemaker, 1988; Iverson and Semmens, 1995). The resultant velocity can be assumed to be horizontal in the case considered here for the simple reason that there is empirical evidence for this situation. Iverson and Semmens (1995) suggest that thick sequences of clast supported sediments are necessary for clast expulsion. This criterion is satisfied at Criccieth because of the size of the clast, and the greater clast-clast contact caused by the enhanced flow of the ice over the mound, as discussed above. Any evidence for a slope-wide rapid melt rate will have been mixed into the till by later deformation.

A more general approach to the model would be to first calculate the direction of the resultant movement and include this in the slope calculation, along with regional slope and prow slope. It should be noted that the regelation that is calculated in the model should alter the driving force direction. Therefore, the entrance angle would vary as the model proceeded if it were not assumed to be constant on the basis of the ploughing trace seen in the field (*Figure 5.2*). Indeed, assuming a horizontal resultant velocity assumes an increase in the average melt rate over the slope to counteract this effect. This is the only non-steady-state assumption of this model.

Ideally, a more general solution is needed, showing the effect of many clasts on bed stresses. Kamb and Echelmeyer (1986) have produced a rigorous determination of hard-bed shear stress by including longitudinal stress and decoupling length terms. One way of getting from this investigatory *single* clast model to a more general solution may be to relate the decoupling lengths determined here to those of Kamb and Echelmeyer.

Model components

The release of stress caused by ice flow against a lodging clast is separated into the component removed by the ice moving around the clast, and that removed by the flow of till from in front of the clast.

Taking Weertman's (1957; 1964; 1979) model, which works reasonably for a rectangular prism, the flow of ice around the boulder is modelled as due to regelation and creep (*Equation 5.2* and *5.4*).

The till is assumed to yield to relieve all the stress above a residual stress supported by the material. This residual stress is transferred across the clast to cause regelation and enhanced creep in the ice. The movement of the ice past the clast is then totalled and removed from the ice velocity to give the resultant movement rate of the clast through the till. The shear stress that is traditionally used to drive regelation and creep is included by using the ice velocity. The residual strength of till cannot be related well to the Mohr-Coulomb equation, therefore empirical estimates are used when necessary (see below).

The till is assumed incompressible in the model, and the effective pressure is maintained as a constant. There is no freezing in the sediment. The empirical evidence suggests some of these factors *are* varying, however, they are not well understood. The compressibility and effective pressure are related, but not enough is known about the rheology of tills to calculate the effect yet. The same problem holds for the effect of freezing in the sediment. Therefore, these processes are assumed equal and opposite (which they will be for one particular clast size) and neglected here. When more information is available, these processes can be included in the calculation for the residual strength. The forces acting in the model are distributed over areas calculated geometrically.

Because the gouge fills with material, the height of the clast is not a measure of h_i the contact height between the clast and the ice. This is calculated assuming

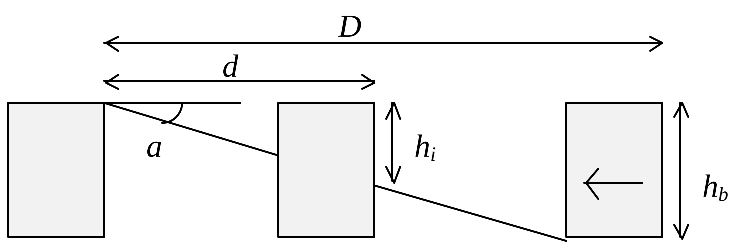
$$h_i = h_b - f\left(\frac{Md}{w}\right) - T \qquad \text{...Equation 5.7}$$

where h_b is the total clast height, w is the clast width, M is the total meltwater volume, d is the volume of debris in the ice divided by the fluid equivalent volume of the ice containing it, and T is the till inflow height, which is the average height of the till flowing around the clast. The value of d is taken as 0.2 for the ice at Criccieth. Values of d in modern glaciers range from 0.0002 to 0.74 (Kirkbride, 1995). A value of 0.2 is slightly above average, and is based on Boulton's (1977) contention that lithofacies D at Criccieth is stagnant meltout material. The method of calculating the debris height is difficult, however, it is suggested that a reasonable model is to proceed thus,

a) distribute the sediment as a ramp extending from the clast, where the angle of the slope is equal to the angle of internal friction of the sediment. Calculate the average height of this ramp,
b) calculate the area of the last increment's meltout that is covered by the base of the ramp and the volume of the last increment's meltout in that area,
c) add the overlapped material to the average over the new increment's base area.

The sediment is largely clay sized material with some silts. There are only low levels of sand sized clasts (above). Thus the angle of internal friction for the sediment is taken as 22°.

The clast-till contact height can be calculated from simple geometric considerations based on the following lengths;



Where d is the distance to complete burial (h_i tan$a$), D' is the total distance to burial, a is the slope angle, and h_i and h_b are the upstream height of the boulder in the ice and the absolute clast height respectively.

The model has two implicit states,

$$\dot{e} = v_i \, , F_c = 0$$
$$\dot{e} = 0 \, , F_c \geq F_v \qquad \text{...Equation 5.8}$$

where $\dot{e}$ is the strain rate in the till i.e. represents movement of the clast, v_i is the ice velocity minus the flow of ice around the clast, F_c is the force acting on the clast due to the residual strength of the till, and F_v is the viscous drag force of the ice around the clast when it is unmoving. The force terms will vary with clast exposure. Thus, F_c can be compared with F_v at each stage of a model run and the point at which they are equal will be that at which there is no clast movement through the till. The Weertman (1957) equations are used to calculate F_c. These equations give a cubic solution for force in terms of total ice velocity. Over the clast widths used in this study (below) the discriminant of this solution suggests there is only one real and positive root as required by the physical basis of the model (the other root being complex). The roots can therefore be gained using the Tartaglia / Cardano method (Daintith and Nelson, 1989).

Data used in model runs

Two sets of model runs were completed. In one set the ice velocity and residual strength were given and the lodgement characteristics calculated, in the other the residual strength was calculated for a number of velocities on the basis of the ploughing length at Criccieth. The ice velocity, when controlled, was taken as 20 ma^{-1}. Although there is no evidence that this velocity is suitable, there is no evidence for any other velocity, and it is both mathematically convenient and reasonably representative for a glacier. The bed slope local to the ploughing boulder is 5°. The residual strengths for the till, when required, are set to vary over three orders of magnitude from 0.5 kPa to 50 kPa, which covers field estimates (Table 8.1 of Paterson, 1994). Initial runs used clast sizes ranging over three orders of magnitude. Cubes of side 0.01m and 0.11m (the transition size assumed in all abrasion studies, following work quoted by Hallet, 1979; 1981) were used, and a boulder 1m height by 1.75m width and depth - the size of the clast discussed above at Criccieth. The code for the models can be found in Appendix B.

5.4.3 Results

Figure 5.14 shows the distance ploughed before lodgement by different sized clasts over time through tills of various residual strengths. The fact that in one case (0.5 kPa residual strength till) the model clast the same size as the Criccieth clast (1.75 m x 1.0 m) ploughs a similar distance (11.08 m) to that seen in the field (~10 to 11 m) strengthens our interpretation. This is further backed by the height of meltout associated with this clast size (*Figure 5.15*), which is close to the field estimates of ~0.5 m just before lodgement (note that the model does not include meltout after lodgement, or frictional/geothermal meltout, so the model results are a minimum estimate). The force supported by the clasts over time is given in Figure 5.16.

If the ice velocity and till strength are allowed to vary in steps of 10 m a^{-1} and 1 kPa respectively over 0 to 2000 m a^{-1} and 0 to 50 kPa, we can estimate the till strength for various velocities at the Criccieth site by searching for those clasts that plough the same distance as that suspected to have occurred in Criccieth (between 10 and 11 m). The potential residual strengths are given in Figure 5.17. In reality these values will form a continuous field of potential strengths, surrounded by a strength-velocity field which could *not* account for the feature at Criccieth. The total meltout fill for these cases is given in Figure 5.18 and the maximum force supported by these clasts is given in Figure 5.19. The thin sections suggest that lodgement is partly occurring because of the till flowing around the clast. This is allowed to fill the gouge in the model, pushing the ice off the clast back. This material would not flow into the clast gouge in the case where the local slope is being used as an analogue of prow development. If the model is run without this inflow to simulate this material having no effect, the potential residual strengths for the till and ice velocities are more limited (*Figure 5.20*), the meltout volumes are higher (*Figure 5.21*), and the supported forces for any given velocity are larger (*Figure 5.22*).

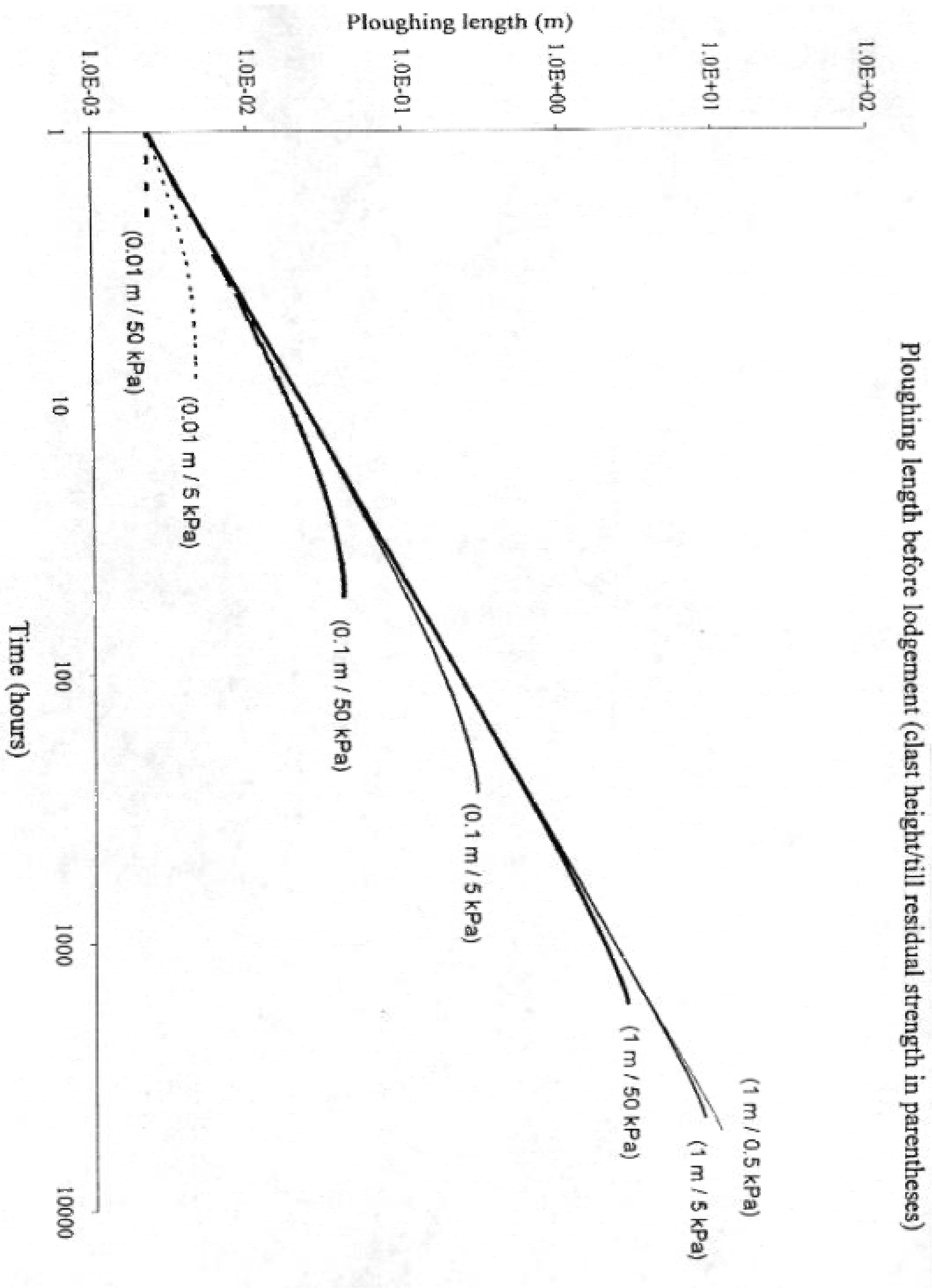


Figure 5.14 Distance travelled over time by different sized clasts through tills of various residual strengths up to lodgement when the model terminates (infill behind the ploughing clast by regelation and slumping allowed).

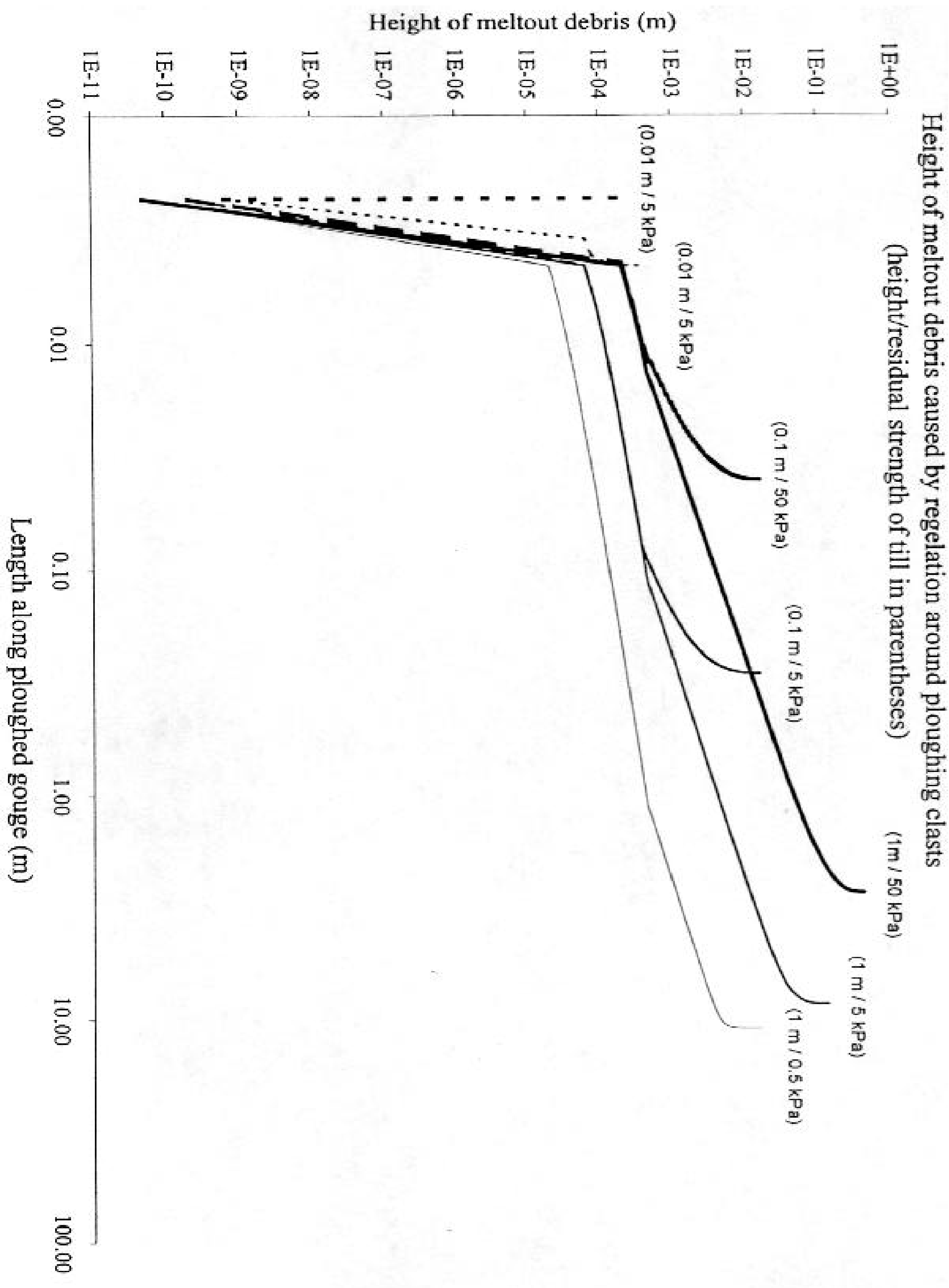


Figure 5.15 Meltout produced over the ploughed distance by the ploughing of different sized clasts through tills of various residual strengths (infill behind the clasts also allowed by slumping of the trench walls).Note that meltout production is halted in the model when lodgement occurs and frictional/geothermal meltout is not included, making these values minimum estimates.

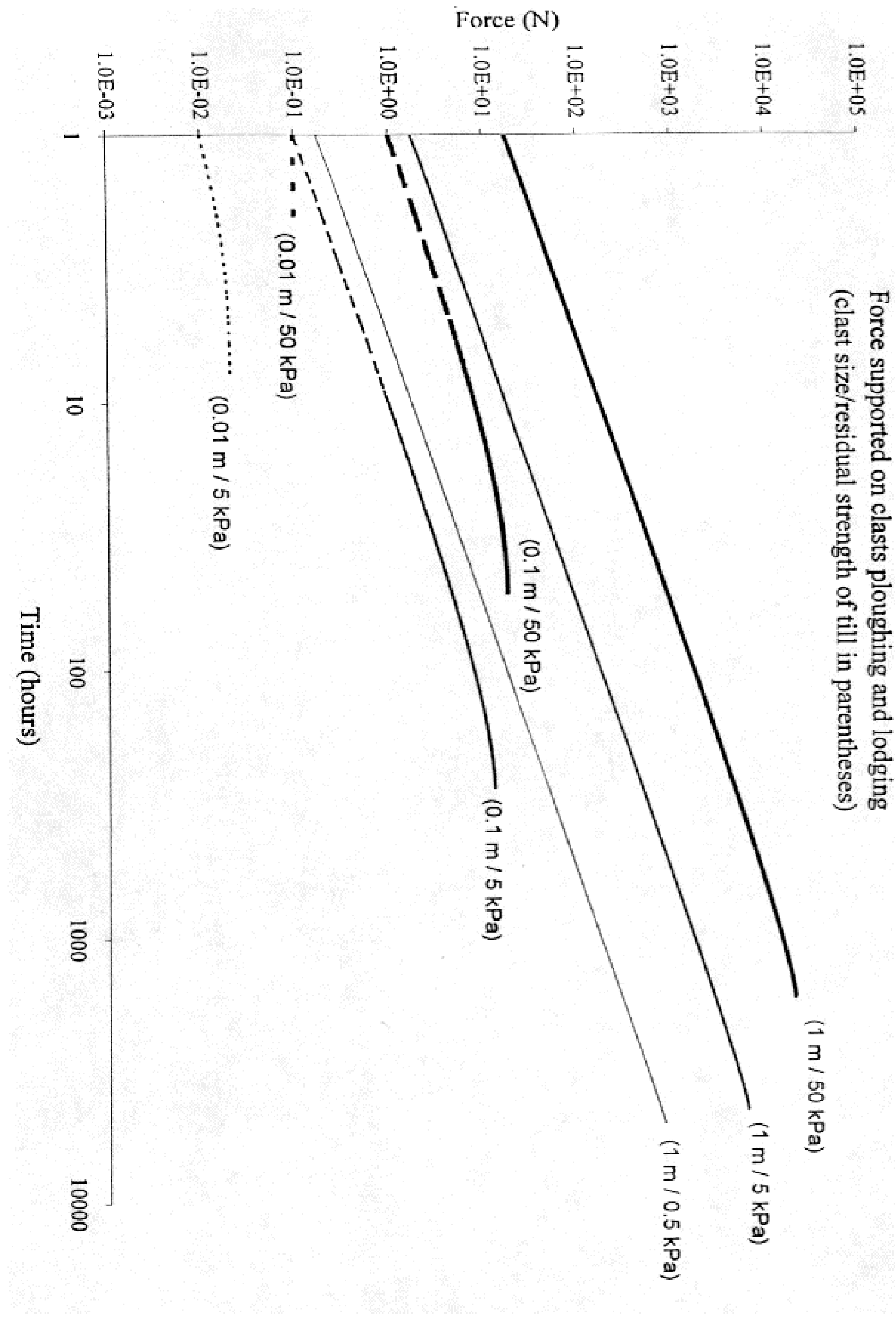


Figure 5.16 Force resisting ice flow over time up until lodgement, caused by clasts of various sizes ploughing through tills of various residual strengths (infilling by meltout and slumping of the trench walls allowed). The model terminates at lodgement.

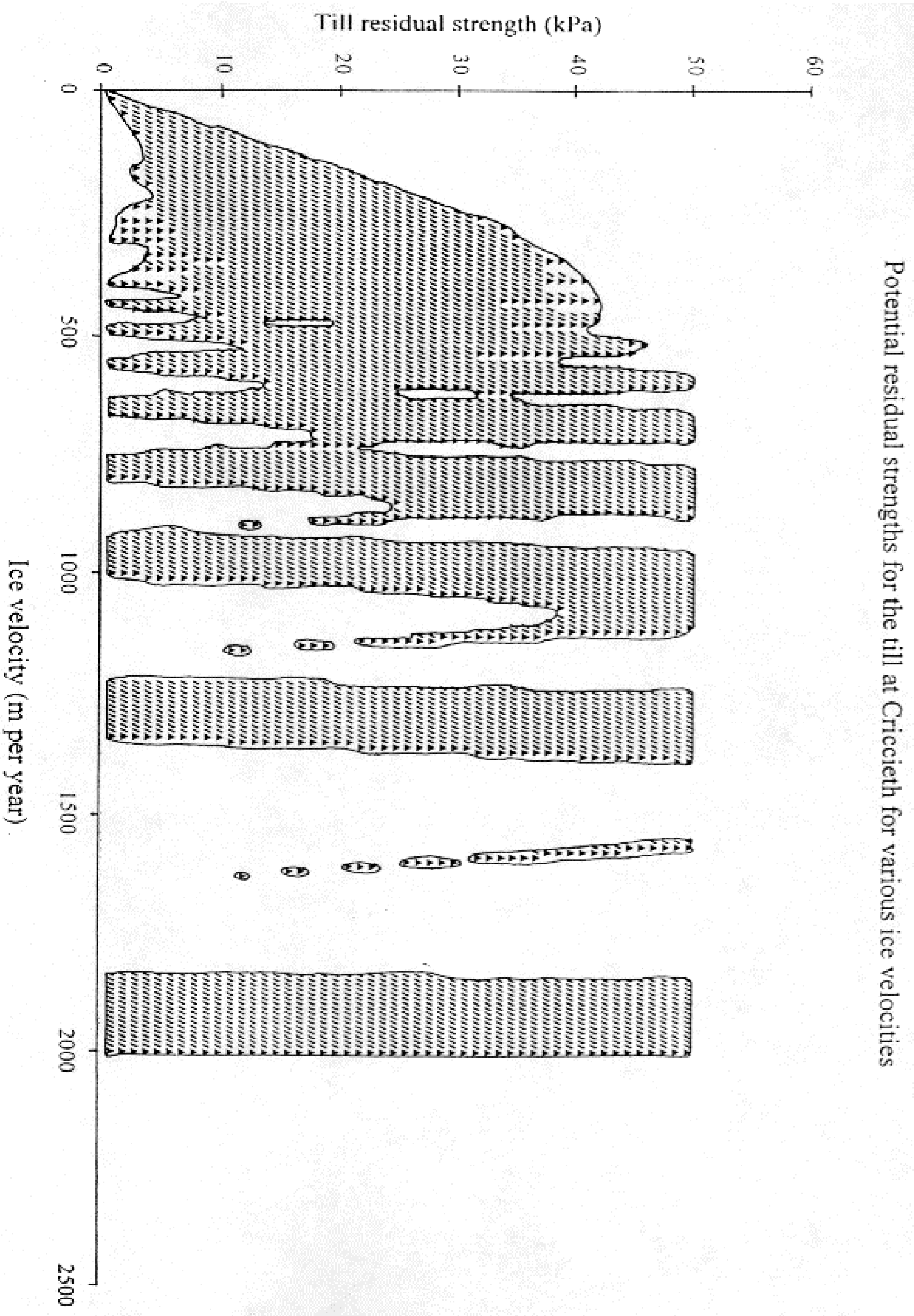


Figure 5.17 Potential residual strengths of the lithofacies E till at Criccieth for various ice velocities based on the length of the ploughing trace (infilling behind the ploughing clast by regelation and trench slumping allowed).The till strength was allowed to vary between 1 and 50 kPa in 1 kPa steps, the ice velocity was allowed to vary between 1 and 2000 m a^{-1} in steps of 10 m a^{-1}. Ploughing lengths were then compared with the length seen in the field and only those conditions producing this length accepted as likely.

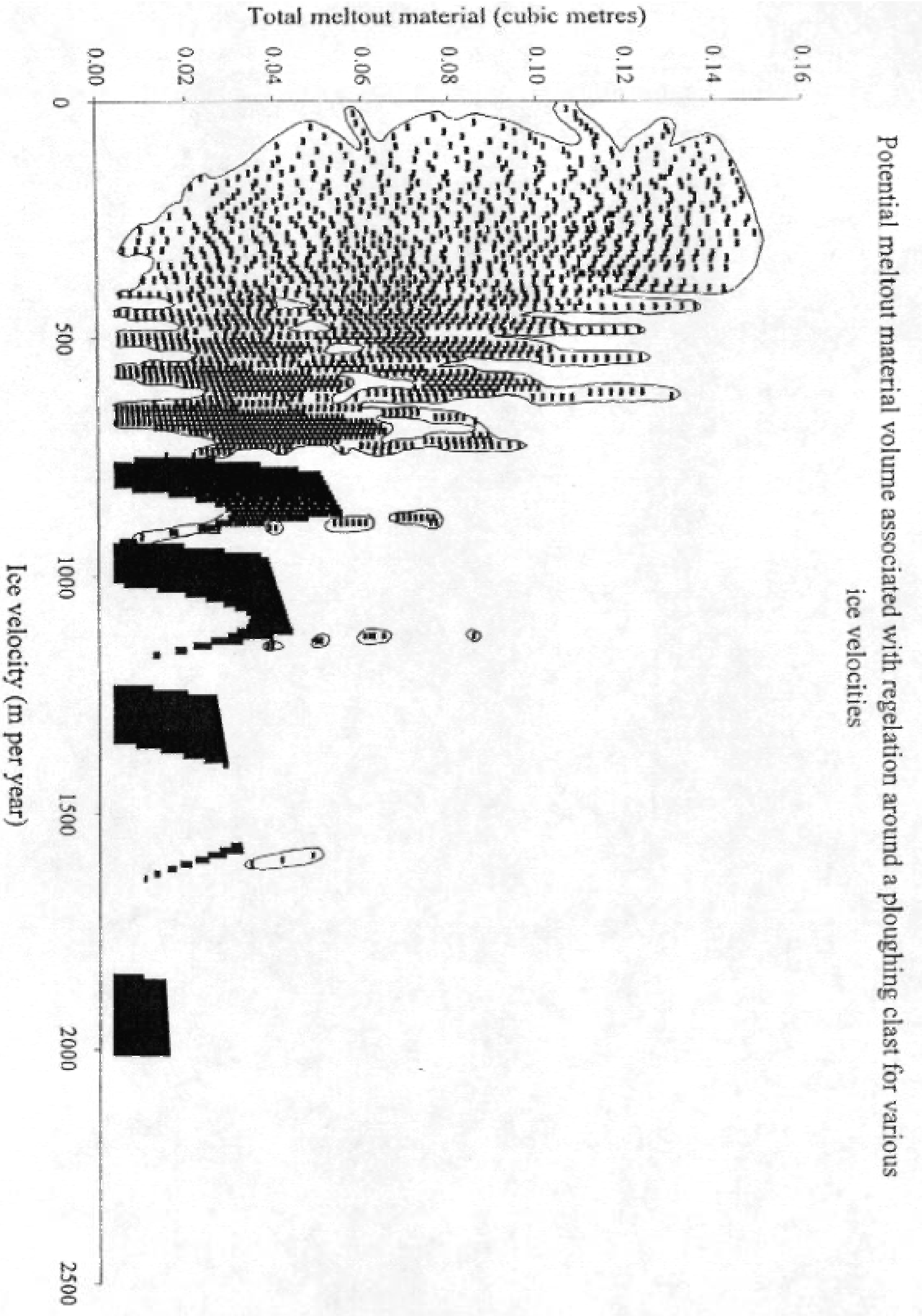


Figure 5.18 Meltout material associated with the potential residual strength estimations for various ice velocities based on the length of the ploughing trace (infilling behind the ploughing clast also allowed by trench slumping).

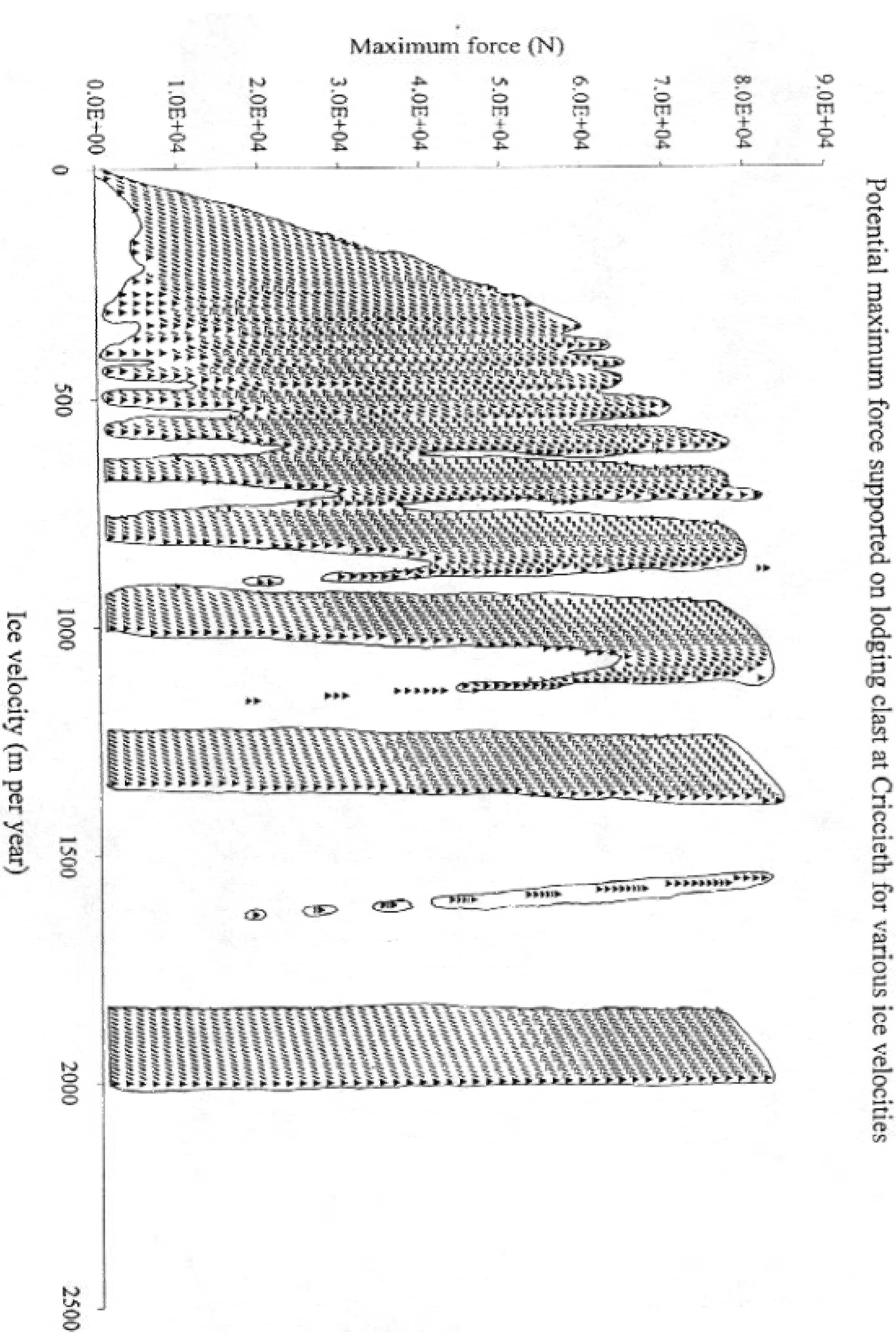


Figure 5.19 Maximum force resisting ice flow associated with the residual strength estimates for various ice velocities based on the length of the ploughing trace at Criccieth (infilling behind the ploughing clasts by meltout and slumping allowed).

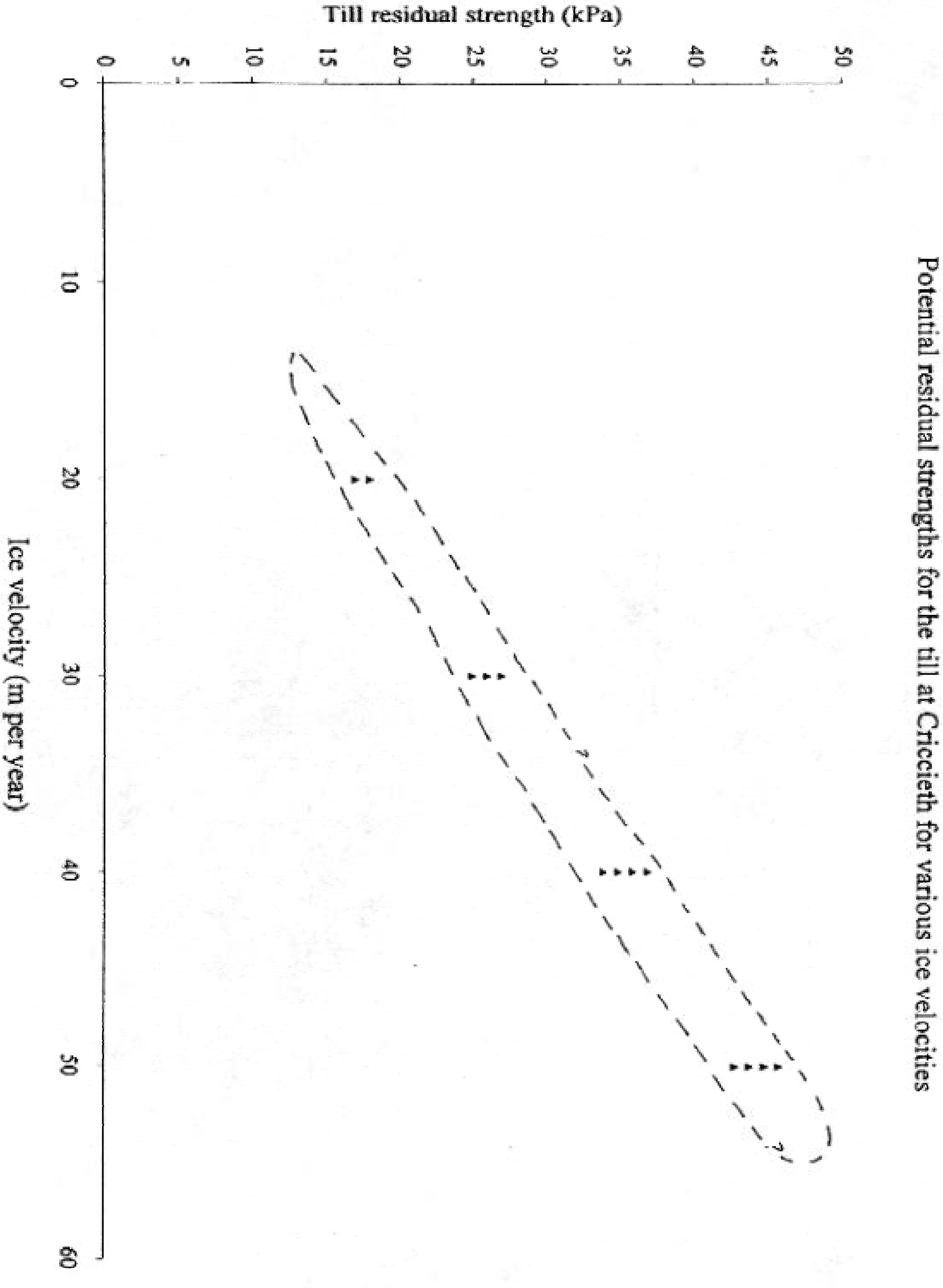


Figure 5.20 Potential residual strengths for lithofacies E at Criccieth for various ice velocities based on the length of the ploughing trace (infilling behind the clast not allowed).Note that the residual strength was allowed to vary between 1 and 50 kPa in steps of 1 kPa, and the ice velocity was allowed to vary between 1 and 2000 m a^{-1} in steps of 10 m a^{-1}, thus the values are not continuous. In reality the values would fall in a continuous field of potential strength-velocity combinations. The values seen here are the only ones found reasonable in the quoted ranges based on the criteria that the ploughing length should be that seen in the field.

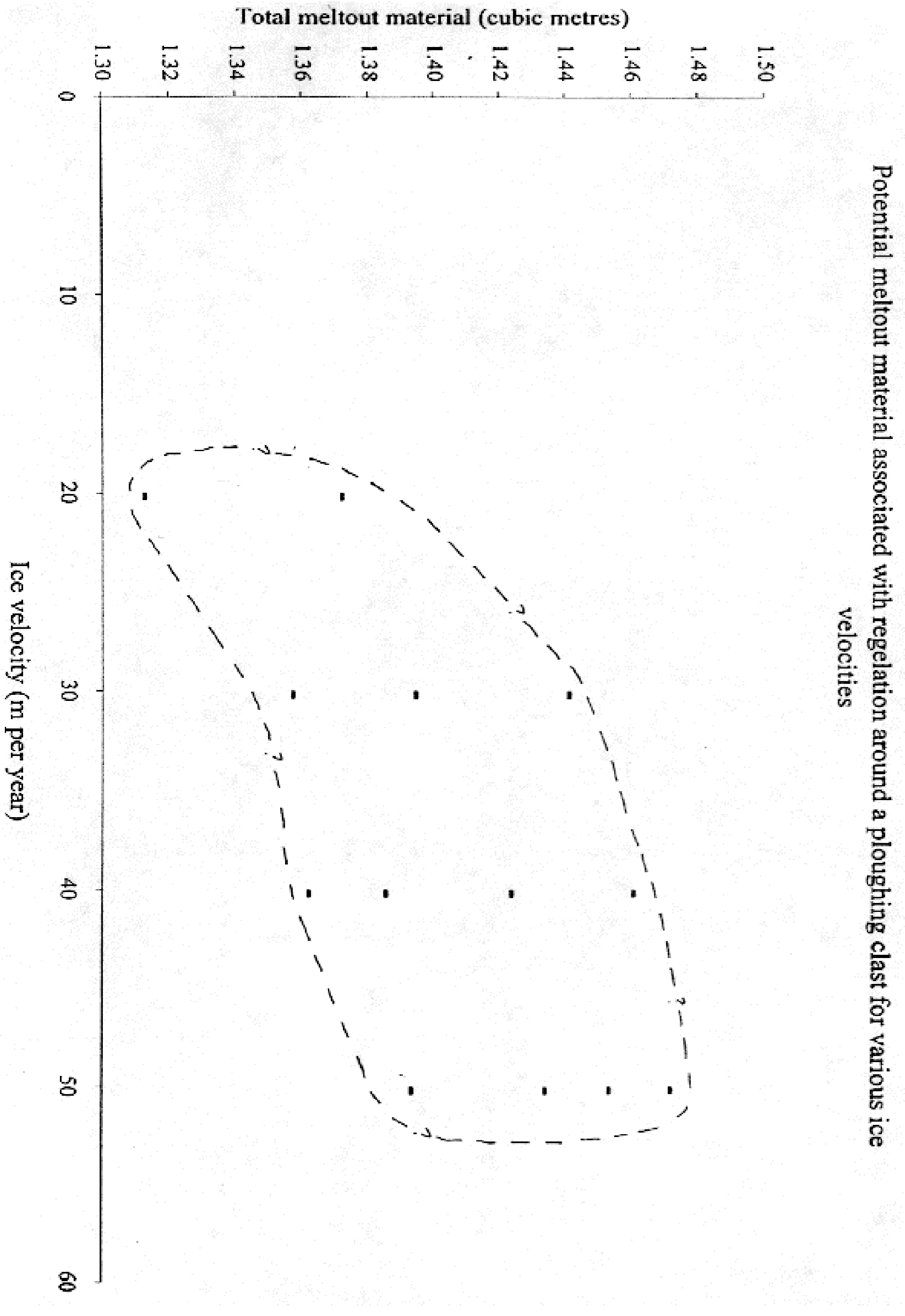


Figure 5.21 Meltout associated with the estimates of till residual strength for various ice velocities (infilling not allowed).

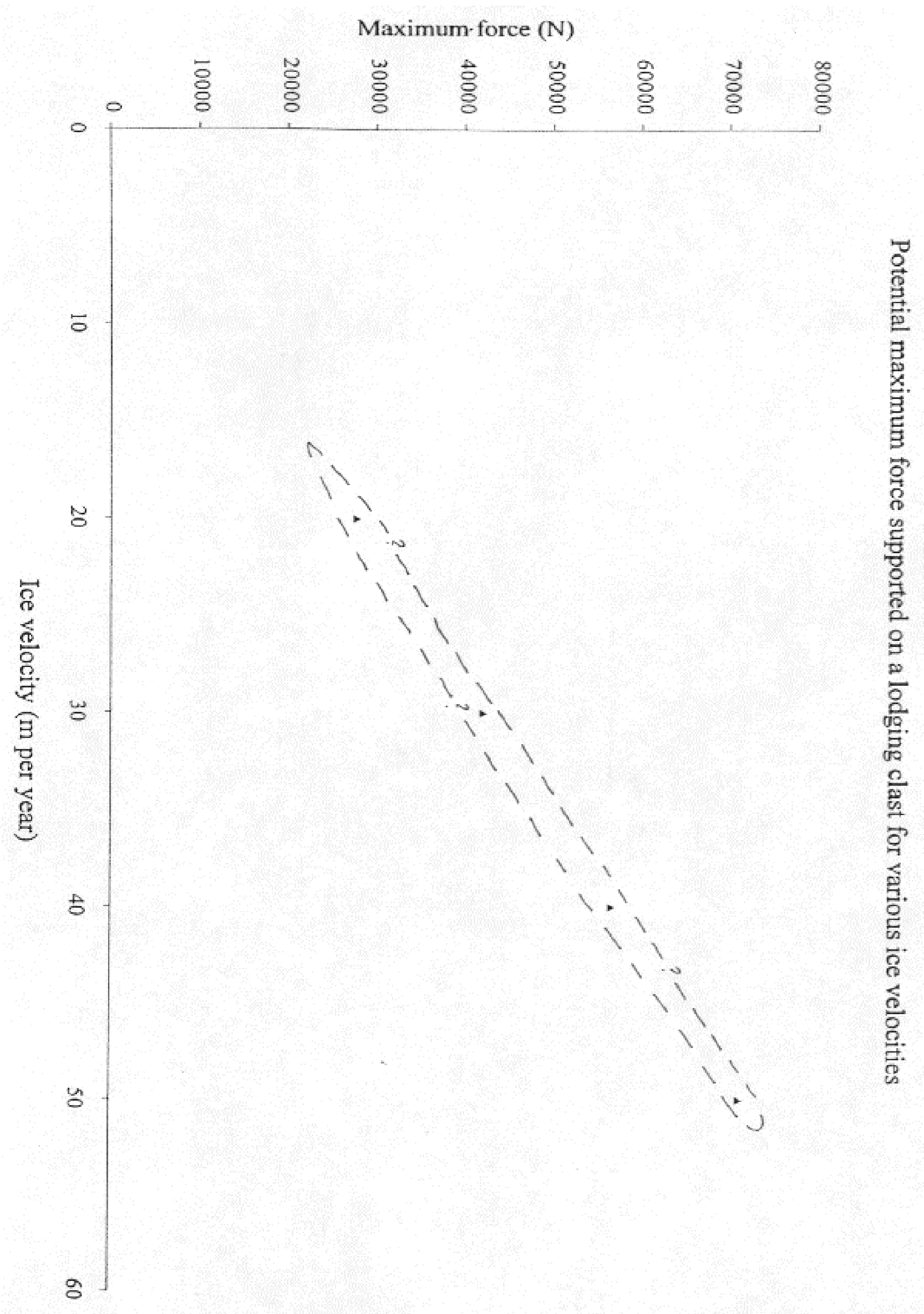


Figure 5.22 Maximum force supported on a ploughing clast for various potential till residual strengths associated with various potential ice velocities (infilling behind ploughing clast not allowed).

5.4.4 Discussion of the model results

The model which allows inflow of till from in front of the boulder gives convincing residual strengths for the till at Criccieth for ice velocities of 0-200 ma^{-1} (*Figure 5.17*), though does not give minimum values at higher velocities. This is because till inflow into the ploughed gouge will inevitably swamp the clast if allowed. In reality the micromorphology suggests periods of slack water in the gouge during which winnowed material could settle. This points to less till inflow than the model produces. Thus, we might expect the residual strength to fall between the values associated with till inflow (*Figure 5.17*), and those produced when the inflow is turned off (*Figure 5.20*). This would put the residual strength of the till between 1 and 10 kPa for ~20 m a^{-1} ice velocity. This matches other estimates well, though is on the lower end of observations (Paterson, 1994).

Such a low residual strength suggests a low effective pressure / high pore water pressure. This matches both the fluid-depositional nature of the micromorphology and the subsequent low deformation, which also possibly indicates a thin deformation layer. Free surface water nearby may explain the sand concentrations in the micromorphology and their apparent absence deeper in lithofacies E. Conditions at Criccieth may well have matched those under Ice Stream B today, where low effective pressures are associated with a thin deforming layer (M.Jackson, *pers.comm.*, 1996) and low residual strength (Kamb, 1991) (though the deposit is considered too small to have effected the ice velocity at Criccieth). The low residual strength of the material does not match the high strength of till from Yorkshire examined in Chapter Six. However, the Criccieth material is clay-rich, with few sand sized particles in the till as a whole (*Figure 5.12a*). This absence of sands will have allowed greater shear development (for confirmation of this shearing see *Figure 5.12b*), and easier deformation. This theme of a sediment's strength being related to its size distribution will be enlarged upon in the next two chapters.

One future improvement to the model will be to limit the potential residual strengths at Criccieth on the basis of the proportion of meltout material produced as well as the ploughing length. This may give an even greater accuracy to the potential strengths, and limit the potential ice velocities further.

5.5 Conclusions

1) One situation microscopic melanges may form in is in the ploughing trace of boulders (*Figure 5.23*). In the case of normal meltout there is no evidence to suggest that material should be preconsolidated to the extent seen here prior to deposition. Normal meltout material will lack cohesion and form more homogeneous deposits. In fluvial situations, large amounts of cohesive sediments would not be found on a microscopic scale horizontally juxtaposed with clean sand bodies and substantially winnowed material.

2) Until alternatives are found, the preconsolidated melanges should *sceptically* be taken as indicative of the ploughing of a boulder. The processes forming such melanges do not necessarily result in lodgement, therefore it is possible that these processes are responsible even if there are no clasts in the sediments. The melanges may be taken as indicative of a stress transfer between ice and till. It is likely that such diamicts will be found only at the top of till units, or in rare areas of intra-unit protection where they cannot be destroyed. In this case, lodgement was into the upstream side of a hummock as the last till deposition was occurring, and the preservation may partly have been a result of the particle size of the melt out.

3) The mechanism of ploughing discussed here can provide both the clasts and the matrix of diamictic tills. It is possible that the random alignment of the melanges in the till would slightly stiffen the deposit to make lodgement more likely, and facilitate the till deposition. The evidence for ploughing and lodgement discussed in this chapter strengthens the argument put forward in Chapter Four that lithofacies E was deformed subglacially, and the lodgement model explains how once laminated sediments contain large pebbles without the need to invoke ice rafting.

4) The model of the suggested processes (*Figure 5.23*) gives low residual strengths for the till (1 to 10 kPa) for reasonable velocities, and these match values from tills under present day glaciers (Paterson, 1994). The low values found here may reflect the fine grained nature of the till, and suggest a high pore fluid pressure environment matching the suggested deposition of the micromorphology through water.

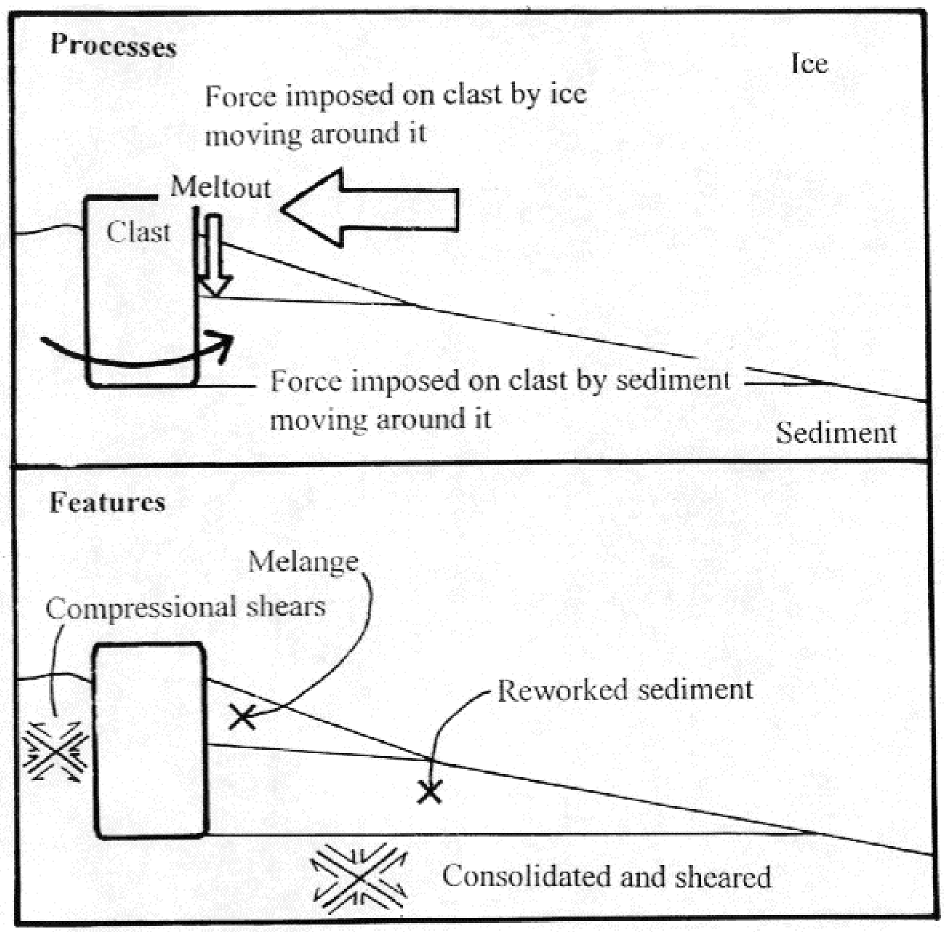


Figure 5.23 Processes and features associated with the ploughing and lodgement of clasts