DISTRIBUTION, CHARACTERISTICS, AND GENESIS OF JOINTS IN FINE-GRAINED TILL AND LACUSTRINE SEDIMENT, EASTERN AND NORTHWESTERN WISCONSIN

by

Douglas Edward Connell

A thesis submitted in partial fulfillment of the
requirements for the degree of

MASTER OF SCIENCE
(Geology)

at the

UNIVERSITY OF WISCONSIN--MADISON

1984

ABSTRACT

Joints in fine-grained sediment are caused by several processes, acting alone or in combination through time. Joints can be formed during shear or compaction beneath glacial ice, during unloading associated with deglaciation, erosion, or release of confining stress, from stress associated with differential isostatic rebound, or from shrinkage associated with physical and chemical weathering. The polygenetic nature of jointing and the superposition of processes through time makes determination of joint genesis difficult.

Joint orientation, spacing, maximum length, and asperity were measured at 22 locations in ten till units and associated lacustrine sediment in northern and eastern Wisconsin. Measurements were made in clay to silt loam, uniform, relatively unweathered basal till, at or near the water table, in natural exposures and excavations.

Preferred joint orientation was evaluated using PATCH (Mahtab and others, 1972) in combination with other computer plotting and data analysis programs. PATCH was tested for its ability to reject spurious clusters of joint poles and calculate mean joint orientation and dispersion. From an analysis of synthetic and

field data, PATCH was found to require supplementation with other methods to properly determine joint set orientation.

At all locations near-vertical joints show statistically significant preferred orientation, commonly perpendicular to and conjugate around the direction of ice flow as inferred from microfabric. Low-angle joints, although in places dipping up ice, generally have randomly oriented strike.

Although dessication may be responsible for opening joints, stresses during and just after till deposition seem to control the orientation of the subsequent failure plane. Joints do not appear to be the result of isostatic uplift or shrinkage following chemical weathering, although these processes may modify joint characteristics.

Sites that contain large numbers of joints perpendicular to ice flow were probably areas of compressive flow during or after till deposition. Most of the joints may have formed in a subglacial environment, at or near the ice margin, under conditions of elevated pore pressure.

Joints in fine-grained sediment can cause elevated permeability and lowered strength and slope stabilty compared to unjointed material. Significant anisotropy of strength and permeability can also occur. The resulting complexity makes accurate site characterization and simulation difficult and expensive.

Data from the Wisconsin study were combined with information from other studies in order to develop a data base for prediction of joints in fine-grained sediment. In till and lacustrine sediment, they are associated with fine-grained sediment (greater than 50% silt and clay), illite or smectite-dominated clay mineralogy, and are found primarily in the near-surface environment. The sediment is usually classified as high or low plasticity clay when using the Unified Soil Classification System. Joints have been observed in sediment with angles of internal friction ranging from 20 to 37 degrees, and cohesion values of 0 to 30 kPa. With increasing clay content and plastic limit, fracture length increases and the number of fractures decreases. It appears that joint length is controlled in part by clay content, and joint spacing is related to both clay content and till thickness.

Joint orientation, as it can be related to ice flow-direction, slope failure, regional tectonism, or the propagation of joints from underlying bedrock, is typically variable from site to site over large areas, but is fairly consistent in a limited area.

PREFACE

This thesis consists of four chapters. Two of these, Chapter 2: Measurement and Analysis of Jointing Characteristics in Fine-grained Glacial Soils by Peter J. Bosscher and Douglas E. Connell, and Chapter 3: Distribution, Characteristics, and Genesis of Fractures in Fine-grained Till and Lacustrine Sediment in Wisconsin, are intended for separate publication in refereed journals. Because they were written for publication, they include some of the material in Chapters 1 and 4. Chapter 2 is directed towards the geotechnical engineering community and so uses their vocabulary. Chapter 3 is intended for publication in a geology journal.

ACKNOWLEDGEMENTS

This thesis would not have been possible without the help of many people. My thanks go to:

Paul Huebner of the Wisconsin Department of Natural Resources, Fred Madison of the Wisconsin Geological and Natural History Survey, and Gary Stanke of the Soil Conservation Service for their assistance in locating suitable study sites.

The many landowners who allowed me to work on their property, particularily the Koenen family, and the staff of the Outagamie County Landfill.

Rick Stoll, my field assistant during the first summer of field work, Mark Johnson, and the rest of the Till Commandos, for their help in the field and with the lab methods.

Leon Fullmer and Ardith Hansel of the Illinois Geological Survey, and Al Schneider of the University of Wisconsin at Parkside, for their skepticism and alternative explanations of joint genesis.

Tim Holst of the University of Minnesota at Duluth, John Fischer of the United States Geological Survey, and Beverly Herzog of the Illinois Geological Survey, for sharing the results of their published and unpublished research.

Herb Wang and Mary Anderson, for their encouragement as members of my guidance and evaluation committee, and to Ken Potter, for his critical review of Chapter 2.

Argonne National Laboratory, represented by J.H. Kittel, the University of Wisconsin Graduate School, Mobil Oil Corporation, and the Wisconsin Alumni Research Foundation for funding this research.

Peter Bosscher, whose patience and good humor during the writing of Chapter 2 was remarkable, for his help and assistance on the ideas expressed in this thesis.

Dave Mickelson, for his interest in seeing this research carried out, for his support, skepticism, and encouragement, and for making my work at Wisconsin fulfilling and rewarding.

Joan Balogh, without whom this thesis would not have been finished, for her love, assistance, and attention to details.

Finally, to my parents, for providing me with the opportunities to complete my Masters degree, and for their help and encouragement throughout my education.

Table of Contents

List of Illustrations

Chapter 1:

Chapter 2:

Chapter 3:

Chapter 4:

Chapter 1:

Genesis and characteristics of fractures in fine-grained till and clay: Literature review

INTRODUCTION

The influence of discontinuities on the strength and permeability of rock and fine-grained sediment has become a topic of increasing concern because of the role these materials play in aquifer protection and waste disposal. This increased interest has led to research on discontinuities in a variety of geologic materials including till and glaciolacustrine clay. The goal of this paper is to summarize previous research on the genesis of discontinuities in fine textured sediment, particularily till.

DEFINITIONS

Discontinuities in unlithified sediment have been described by workers from a variety of disciplines, leading to somewhat inconsistent terminology. The terms fissure (Terzaghi, 1936), fracture (Cawsey, 1977), joint (Williams, 1975) and joint plane (Brewer, 1964) have all been used to describe planar discontinuities in unlithified sediment where there is with no evidence of movement and without restriction as to the specific mechanism of formation. Fracture is usually employed as a more general term than joint (e.g. as in the Bibliography of Geology), but has been defined as a discontinuity caused by mechanical failure (Bates and Jackson, 1980) while the term joint is free of any genetic interpretation (Bates and Jackson, 1980). By this scheme, joint should be used as a descriptive term, and fracture used only if its added genetic implication is appropriate.

As many of the processes responsible for creating discontinuities in fine-grained sediment may involve mechanical failure, in this paper the terms fracture and joint are used interchangeably to describe a planar or near-planar discontinuity of open or closed aperature and little or no apparant displacement along the plane, without regard to genesis. Although its usage is well established in the geotechnical literature, the term fissure is not used here, because in standard geological usage fissure denotes separation between the faces of the feature (Bates and Jackson, 1980). The term fissure has been used by engineers since the early work of Karl Terzaghi to describe discontinuities of limited extent, up to 1 square meter in area (McGown and others, 1975).

GENESIS

In his discussion on the characteristics of joints, Price (1966) states that "...they are perhaps the most difficult of all structures to analyze". Joints in fine grained sediment generally have a complex history, because different processes work to create joints alone or in combination through time (Grisak and others, 1976; Boulton and Paul, 1976). Like other geologic phenomena, similar structures may be formed by a variety of different processes, making the interpretation of fracture genesis difficult.

The following discussion is divided into two parts. The first reviews the processes responsible for fracture genesis and modification; the second describes the signatures that these processes leave behind, signatures which allow inference about their relative contribution to the final fracture form.

Fractures result from either tensile or shear stress during deposition, deformation, or weathering (Chandler, 1973). In the glacial environment (Figure 1), jointing can be found in till deposited beneath the ice (lodgement or melt-out till) and in supra- and pro-glacial sediment flows. Fracture formation due to deformation can take place in the glacial environment during consolidation, from stress associated with the rebound of the earth's

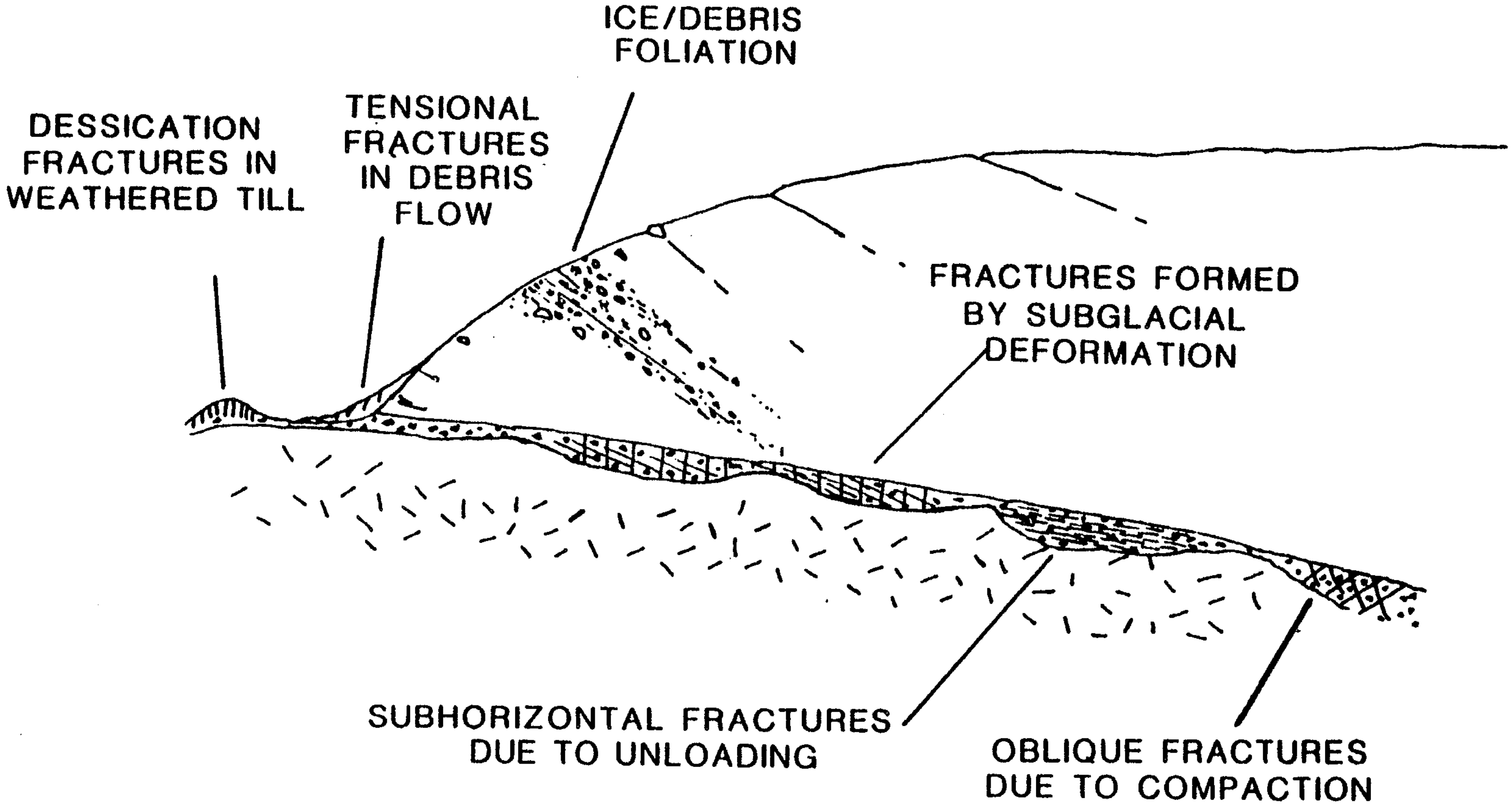


Figure 1: Fracture genesis in the glacial environment

(after McGown and Derbyshire, 1977)

surface during deglaciation, or from the stresses conducted from active ice into till or other material at the glacier bed. During and after deglaciation, secondary processes can lead to the exploitation of existing fractures or the creation of new ones. These secondary processes may include volume changes resulting from chemical change, dessication, and freezing, or stress relief associated with erosion of overlying or adjacent sediment or melting-out of buried ice.

FORMATION DURING GLACIATION

Subglacial Deformation

Several authors have reported joints in till described as lodgment till (Boulton, 1970; McGown and others, 1974, 1975a; Kruger and Marcussen, 1976; Kruger, 1979; Derbyshire and Jones, 1980; Eyles and Sladen, 1981; Schwan and Ritzema, 1982), which suggests a connection between the conditions that exist during deposition of lodgement till and the formation of fractures. Although by strict definition "lodgement" refers to till deposited grain-by-grain at the base of a moving glacier (as per Dreimanis, 1981), it has also been applied to other types of subglacial till.

The association of joints with subglacially deposited till has led to speculation that the joints formed due to subglacial stress (for example McGown and others, 1974; Boulton, 1970, Mickelson and others, 1979). Lodgement till is commonly thought to be deposited beneath temperate glaciers where meltwater promotes sliding at the glacier base (for example Boulton, 1972). The subglacial environment of temperate glaciers may be conducive for the formation of fractures because the large stresses and fluctuating pore pressure can lead to complex cycles of consolidation and deformation of the glacier bed (Boulton and others, 1974).

If jointing is associated only with lodgement till, as strictly defined, then it follows that joints created by deformation of sediment beneath a glacier form under wet-based conditions. However, as the term lodgement is sometimes more loosely applied, or used without explanation, the possibility that jointing has been observed in sediment formed beneath the base of a polar glacier cannot be discounted.

In contrast to temperate glaciers, polar glaciers characteristically have frozen beds. In this setting, shear is usually restricted to the ice itself because clean ice is typically weaker than the frozen ice/bed contact or frozen sediment (Boulton, 1972). Thus, it is usually assumed that deformation of the sediment beneath polar glaciers is restricted to the base of the permafrost layer (Mathews and Mackay, 1960). However, as Mathews and Mackay point out, plastic clay layers may deform under frozen conditions. If ice makes up a large portion of the frozen sediment, the strength of the clay/ice mixture may decrease to that of clean ice (Muller, 1947).

The possibility of joints forming beneath either a wet-based or polar glacier has to be examined in the light of current knowledge concerning conditions beneath glaciers. This knowledge is incomplete, because few direct measurements are available for the stress conditions and pore pressures beneath glacial ice (Paterson, 1981). This lack of knowledge is unfortunate because genesis of joints stemming from deformation depend on the conditions of subglacial stress.

The mechanical properties of glacial ice are commonly thought to limit the magnitude of basal shear stress that can be transmitted to the glacier bed. Glacial ice is commonly assumed to be unable to withstand shear stress in excess of 100-150 KPa without deforming (Paterson, 1981). This is relatively small when compared to the strength of most glacial sediment.

Because the basal shear component of stress is relatively small, the direction of the normal stress vectors (tensional or compressive) may influence the deformation of the glacier bed. Creation of horizontal maximum compressive stress can occur under compressive flow regimes (Nye, 1952) leading to passive earth pressures in the underlying till. These pressures may be large enough to cause deformation and failure, as demonstrated by the association of compressive flow regimes with glaciotectonic deformation (Banham, 1975; Shaw, 1979). Compressive flow regimes can occur in a variety of situations, such as where "...a glacier advances up a [concave] slope directed towards the ice surface" (Banham, 1975), or where the velocity decreases because of ablation or spreading of the ice front.

Nye (1952) has stated that during compressive flow, planes of maximum shear at the base of the ice are oriented either vertical and perpendicular to the direction of flow, or parallel to the ice surface (Figure 2). Fractures with similar orientation have been found beneath modern glaciers. For example, Boulton (1970) found slickensided horizontal fractures in till beneath the Nordenskioldbreen, Svalbard, as well as a set of unstriated near-vertical conjugate fractures arrayed at up to 65 degrees around the direction of ice flow. Similar

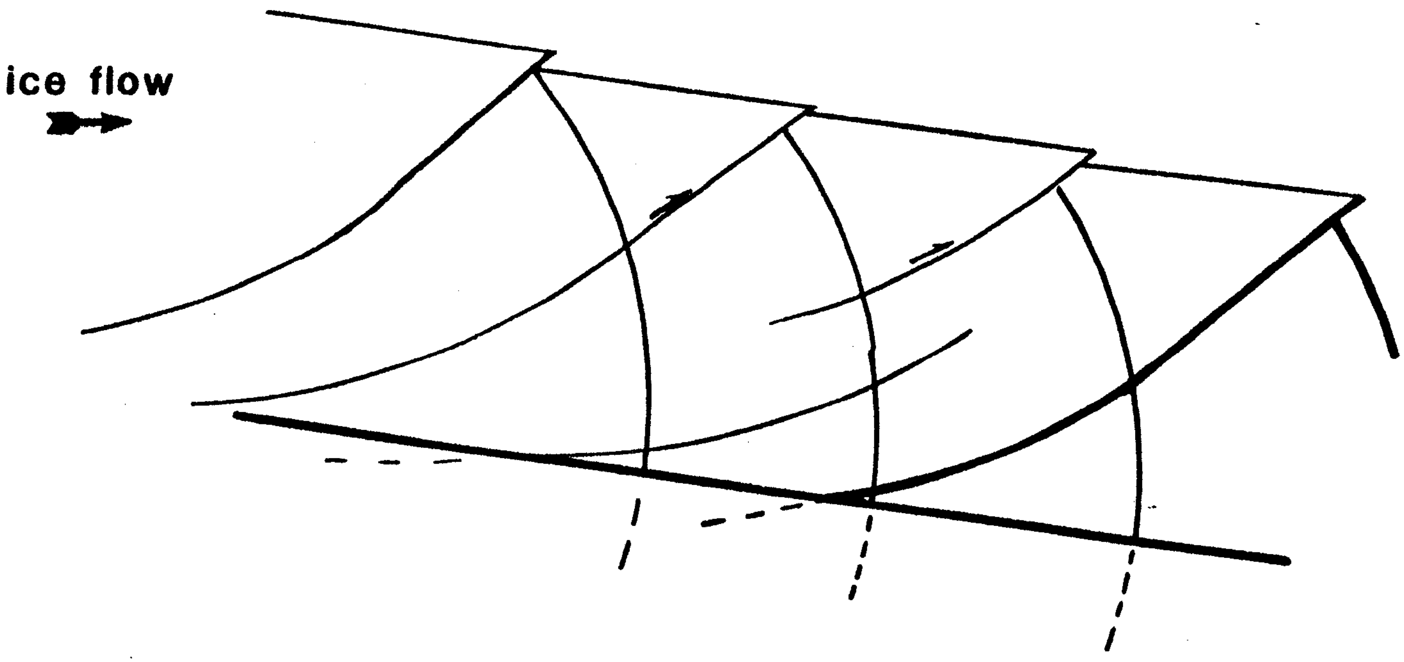


Figure 2: Orientation of maximum shear stresses and possible faults under compressive flow. (after Nye, 1952)

conjugate fractures have been described in Pleistocene deposits by McGown and others (1975) in till from west-central Scotland, by Kazi and Knill (1973) in overridden lacustrine clay, by Mickelson and others (1981) in till from south central Sweden, and by Kupsch (1955) in jointed boulders within a drumlinized kame in Saskatchawan. All of the authors cited above attribute the conjugate pattern to stress beneath the ice, although the type of stress is not clearly stated.

In contrast to the horizontal maximum compressive stress associated with compressive flow, under extending or steady-state flow the maximum compressive stress is typically vertical. This vertical compressive stress can cause compaction and deformation under some conditions.

LaFluer (1980) presents two mechanisms for till deformation during compaction. The first (shown in Figure 3a), occurs during gradual loading and compression of the subglacial sediment. Stress can be built up if lateral failure is inhibited by high lateral earth pressures. During subsequent rapid unloading, the residual stress that built up during consolidation can then be released through shear. Shear would take place along planes dipping at angles of $45 - \phi'/2$ if the built-up lateral earth pressure exceeds the strength of the till at lower confining pressure. The second mechanism involves rapid compression from a rapid application of load or change in pore pressure. Shear could then take place along planes oriented at angles of 45 + $\phi'/2$ to the horizontal (Figure 3b), similar to those shown by Pusch (1973, figure 2). LaFluer (1980) observed shears with both of these orientations in thin sections made from unoxidized, jointed Lavery Till from the West Valley site in New York (Appendix 7), and speculated that both processes had occured.

The magnitude of the vertical normal stress controls the deformation of the glacier bed by affecting the strength of the sediment. By Mohr-Coloumb failure, till or other sediment beneath a glacier will fail if the shear stress (τ) at the glacier base exceeds the strength of the sediment, as shown in Equation 1.

$$\tau >= c' + (\sigma - u) \tan(\phi') \quad \text{(Equation 1)}$$

where τ = basal shear stress (approximately 100-150 kN m-2)

c' = drained sediment cohesion

σ = normal vertical stress (equal to the density times the height of the ice)

u = pore pressure beneath the ice

ϕ' = angle of internal friction of the sediment

(after Menzies, 1979)

Figure 3

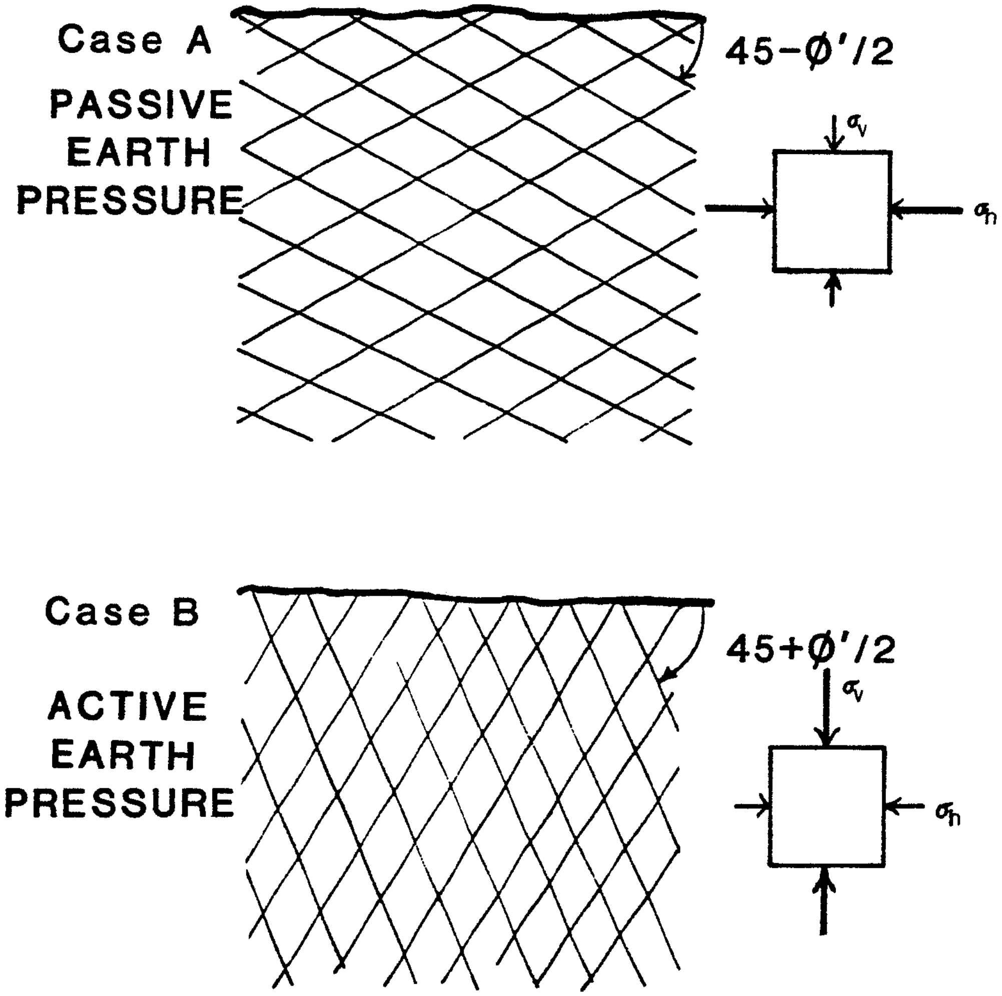


Figure 3: Cross section of joint pattern resulting from compaction. After La Fleur (1980).

Although the basal shear stress is constrained by the strength of the ice itself, the strength of the underlying till is controlled by the effective stress of the ice and by the mechnical properties of the till (c' and ϕ'). These properties may not be constant, as discussed below.

The orientation and magnitude of the effective principal stresses beneath a glacier are controlled by the surface slope, height and density of the overlying ice, the flow regime under which the ice is moving, and the pore pressure beneath the ice. Total stress due to the weight of the overlying ice can be estimated from the product of the height and density of the ice, while basal shear can be calculated from a simple model of ice stress known as Glen's law (Equation 2).

$$\tau = (\rho)(g)(h)\sin(\alpha) \quad \text{(Equation 2)}$$

where τ = shear stress beneath the ice

ρ = density of the ice column

g = acceleration due to gravity

h = depth beneath the ice surface

α = angle of the ice surface from the horizontal

(from Paterson, 1981)

Boulton and others (1974) found that in a model of the stress conditions beneath the Breidamerkurjokull, Iceland, the basal shear exceeded the strength of the till beneath the glacier only in a narrow zone at the ice margin where the effective stress was low. However, as pore pressure is significant in controlling the effective stress, any increase in pore pressure resulting from advance of the glacier over sediment of low permeability can lower the effective stress, allowing deformation of the glacier bed (Boulton and Paul, 1976). Pore pressure is also dependent on the basal water content, which is controlled by the balance of melting at the ice/sediment interface and the escape of the meltwater from beneath the ice (Muller, 1977). As pore pressures can fluctuate and so modify effective stresses, underlying till can be subjected to a complex sequence of loading and unloading without major changes in ice thickness (Boulton and Paul, 1976).

The shear strength of clay till or other fine-grained sediment cannot be considered a constant factor, because it is influenced by the water or ice content, grain properties, and rate of shear (Mathews and Mackay, 1960). Among glacial sediment types, shear strength is lowest in fine-grained and unfrozen material (Muller, 1947, in Mathews and Mackay, 1960) and decreases with increasing water content (Terzaghi and Peck, 1967, in Banham, 1975). Compressive and shearing strength of frozen clay is higher than unfrozen clay at low ice contents, but as pointed out above, decreases and approaches that of clean ice as the pores are progressively filled with ice (Muller, 1947, in Banham, 1975). Given the fact that deformation is more likely under conditions of high pore pressure, a high water content is also likely and may act to decrease the strength of fine-grained material at the glacier bed.

The discussion above describes the conditions controlling Mohr-Coloumb failure. Triaxial tests on normally-to-slightly overconsolidated, reconstituted silty clay suggest that Mohr-Coloumb (elastoplastic) failure may take place during undrained compression (Hight and others, 1979). Geologists have used the term brittle to describe this type of behavior and differentiate it from ductile failure. During brittle failure, deformation takes place by breaking along discrete discontinuites (Hobbs, Means, and Williams, 1976). Ductile failure represents the other end member, where deformation takes place throughout the deforming mass (Hobbs and others, 1976). Mead (1925) suggests that brittle failure occurs in granular materials at maximum packing density, if they cannot fail by dilation; or in cohesive material at the point when the available pore water cannot fill the expanding voids.

Clayey glacial sediment can behave in this manner if it is overconsolidated, and Mead goes on to state that "...the faulting and jointing in fine-grained glacial material so commonly observed... ...exemplify this property of granular aggregates".

It is unclear whether joints in till result from Mohr-Coloumb failure, as the indicators of failure mode are subtle and easily masked by later modification. No authors have reported finding evidence of shear, such as slickensides or feather fractures, on vertical joint surfaces, although they have been observed on near-horizontal fractures. In contrast to Mohr-Coloumb failure, deformation during secondary creep may take place at loadings less than those causing failure; this may create planes of weakness which control subsequent joint formation and orientation (Boulton and others, 1974).

The depth to which subglacial stresses can cause deformation is uncertain; MacClintock and Dreimanis (1964) found in one instance that reorientation of pebble fabric by overriding ice took place to depths of 10.7 m, while minor deformation without clast reorientation extended to depths of 19.8 to 21.3 m.

In summary, joints have been observed to form in the subglacial environment of a modern glacier, and similar joints have been found in Pleistocene till and other sediment overriden by ice. Formation resulting from subglacial deformation is probably restricted to a zone near the ice margin where the ice is thin, and to areas of elevated pore pressure and low effective stress. Deformation of fine-grained sediment may take place under either frozen or unfrozen conditions but is more probable at high ice or water contents. Joints may be formed by either Mohr-Coloumb failure or may be initiated during secondary creep.

Formation of Fissility

Muller (1977) has stressed dewatering at the glacier base during consolidation as an important component in the formation of narrowly spaced near-horizontal joints, sometimes termed fissility. Water films at the sediment surface may cause a reorientation of the grain fabric that may control subsequent failure. Others have proposed that a reduction in the effective stress because of a rise in pore pressure or a decrease in ice thickness may cause the development of fissility (Boulton and Paul, 1976). Near-horizontal joints have been observed in meltout till from polar glaciers where the till has inherited the original ice/debris foliation as the ice was removed (Shaw, 1977a). Formation of segregation ice lenses may also form near-horizontal joints, and is discussed below.

DEGLACIATION

The unstable nature of fresh tills may lead to deformation shortly after deposition. Slumping and flow of supraglacial debris can occur on both polar and temperate glaciers, leading to vertical fractures forming on the surface and at the nose of sediment flows (Figure 1) (McGown and Radwan, 1977).

Differential settling can create stress on a till mass during and after deglaciation. Differential settling can result from melting of buried ice (Selsing, 1981), or from deformation or diapirism of underlying strata during consolidation (Banham, 1975). This settling can lead to the failure of unstable slopes in the proglacial environment (Boulton and Paul, 1976).

Isostatic Rebound

Stress associated with isostatic rebound has been proposed as the cause of joints in till and glaciolacustrine sediment thought to parallel joints in the underlying bedrock (Grisak and Cherry, 1975). Differential stress from rebound during and after deglaciation may also explain the propagation of large-scale bedrock lineaments through overlying Pleistocene deposits (Barton, 1962).

Observations on propagation of bedrock joints into overlying glacial deposits do not provide a conclusive answer as to the importance of isostatic uplift. Burford and Dixon (1977) described systematic fractures in glaciolacustrine clay and clay till that have orientations similar to bedrock joints in Summit County, Ohio. They suggested that similarities in joint pattern may be the result of differential isostatic rebound, earth tides, or recent tectonic stress. Meltwater channels in thin till evidently controlled by the orientation of bedrock jointing in north-central New York State led Jordan and Andrews (1978) to conclude that joint propagation into the till was a result of rapid isostatic rebound. However, other studies have shown that jointed glacial sediment does not always reflect the jointing of the underlying bedrock. Babcock (1977) found no relationship between joints in glaciolacustrine silt and clayey silt and the bedrock jointing in central Alberta. Westgate (1976) pointed out that in eastern Alberta, jointing was consistent both with that of the underlying bedrock and with the direction of ice flow.

The magnitude of the stress that is associated with differential rebound is unknown. Loading by glacier ice can have significant effects on the stress conditions at depth in bedrock, as illustrated by the high remnant stresses in deep bedrock boreholes resulting from the crustal depression of central Sweden (Carlsson and Olsson, 1982).

Mörner (1970) among others has argued that rebound and its associated stresses are at a maximum during deglaciation; if this is generally the case, then stress due to uplift and the stress due to the weight or motion of the ice above may act in concert, making differentiation between the processes difficult.

SYNAERESIS

Joints in sediment not directly affected by glacial ice may form by processes taking place during or after deposition. One such process is synaeresis, or "the mutual attraction of clay particles to form closely knit aggregates with fissures between" (Mitchell, 1976). Synaeresis, which differs from dessication in that it can take place subaqueously, may be responsible for formation of fractures in fine-grained glaciolacustrine deposits or other waterlaid sediment. This process has been thought to create joints in normally consolidated clays at moisture contents above the shrinkage limit (Skempton and Northey, 1952). Vertical synaeresis cracks have been observed to form at random orientations in suspensions of flocculating clay and can form and then close during sedimentation, remaining closed until unloading or dessication reopens them (White, 1961).

POST-DEPOSITIONAL CHANGES

Weathering

Once deposited, till and glaciolacustrine sediment are altered by chemical and physical weathering processes acting in the near-surface environment. Weathering can act to both create new fractures and exploit pre-existing zones of weakness. Fractures can be formed because of volume changes resulting from physical processes such as dessication and freeze/thaw or from chemical changes such as leaching or alteration of clay minerals. Weathering does not seem to alter the orientation of existing fractures (McGown and others, 1974), but rather increases the number of fractures per unit volume (Fookes and Denness, 1969).

Drying causes loss of water and unequal shrinkage (Baver and others, 1972), leading to tensional failure as the water content decreases to the shrinkage limit (Sleeman, 1963). Shrinkage takes place as water is removed from the double layer of the clay minerals, leading to a more compact arrangment of the clay (Baver and others, 1972). Upon subsequent wetting, unequal swelling can cause shear in the soil, leading to the formation of shear planes at low angles to the surface (Birkland, 1974). Drying evidently produced vertical dessication cracks in the till of the Taylor Glacier, Antarctica (Shaw, 1977b). Both flow and lodgement tills have been observed to dry out rapidly after exposure, with the formation of "a close network of sub-vertical joints with a polygonal surface pattern" (Boulton and Paul, 1976). Donovan and Foster (1972) found dessication cracks with both polygonal and rectangular plan in Devonian sandstones in northeast Scotland.

Fracture spacing resulting from dessication is controlled by clay content and mineralogy, and increases with depth because of differential water loss (Baver and others, 1972). Dessication fracture surfaces can be coated with stress cutans (oriented clay particles) from repeated shrink and swell, or may be slickensided (Bullock and Murphy, 1976).

As much as fifty percent of the shrinkage resulting from dessication can take place above the wilting point of vegetation, making vegetation an important factor influencing dessication (White, 1966). For example, dessication can be enhanced by vegetation transpiring water through their root systems from significant depths (Barber, 1958 in Mitchell, 1976). Roots were thought to have accentuated differential drying and formation of a fractured overconsolidated crust of variable thickness in thick clay till in southwest Ontario (Adams, 1970).

freezing and thawing

Water freezing in sediment pores can impart large stresses to the sediment. These stresses are created by suction from the attraction of water to the freezing front (Chamberlain, 1981) and the pressure of the expanding ice lenses (Baver and others, 1972). The response of unlithified sediment to freezing and thawing is complex, and depends on the water content, grain size distribution, and mineralogy (Czeratzki and Frese, 1958a, 1958b; Bisal and Nielsen, 1967).

Formation of segregation ice lenses in till can create fractures by causing a rearrangment of the fracture surface during the growth of the ice lens. Subhorizontal joints have been observed to form in till after one season of exposure, and have typical spacings of about 0.5 - 5.0 cm (Boulton and Paul, 1976). Vertical ice wedges which form under permafrost conditions may also create fractures, fractures which grade into ice wedge casts if they are filled with sediment.

Chemical Alteration

Chemical weathering can affect fracture formation by causing a loss of volume by leaching, or by increasing the shrink/swell potential by altering clay mineralogy. Leaching of carbonate from originally carbonate-rich sediment allows greater contraction and expansion during moisture fluctuations because of the increased activity of clay minerals (Rimmer and Greenland, 1976). Volume loss caused by removal of carbonates may be responsible for the well developed joints found in the intensely leached pre-Illinoian till of eastern Iowa (G. Hallberg, personal communication, 1984) (Figure 4), as well as jointing in the weathered horizons of Devensian till in eastern Great Britain (Penny and Catt, 1967). Weathering changes in the mineralogy of clay sized minerals can cause a loss of

Figure 4: Oxidized vertical joints in pre-Illinoian till in eastern Iowa. Photograph by George Hallberg, Iowa Geological Survey.

volume or increase the swell potential (Quigly and Ogunbadejo, 1976). Typical weathering schemes of clay minerals are shown in Figure 5.

In the presence of high silica and magnesium concentrations, illite, mica, and chlorite tend to be altered into clay minerals that have higher swell potential. The swelling potential of smectite is influenced by the interlayer cation population and by the balance of aluminum and iron in the octahedral structure (Post, 1981). Swelling potential, as measured directly by linear shrinkage and indirectly by the plastic limit, was found to be proportional to the magnitude of fractures formed by dessication (Anderson and others, 1982).

Loss of volume can occur if sodium montmorillonite (marine) clays are exposed to a more dilute pore water, or if the sodium on the interlayer exchange sites is replaced by calcium. While this has been long recognized for the formation of sensitive clays (Terzaghi and Peck, 1967) its influence on joint formation is unknown.

Recent studies have pointed out that high pore fluid concentrations of organic solvents with low dielectric constants such as benzene, xylene, and carbon tetrachloride may cause dehydration and slight shrinkage of clays, a process that may be important in clay lined waste disposal facilities (Green, Lee and Jones, 1981;

Figure 5: Formation of clay minerals

Parent mineral			Shrinkage Potential
Alkali Feldspars	$\xrightarrow[\text{low silica concentrations}]{\text{acidic conditions}}$	Kaolinite	low
Micas, Alkali Feldspars	$\xrightarrow[\text{high silica concentrations}]{\text{alkaline conditions}}$	Illite	low
Illite, Mica	$\xrightarrow[\text{replacement by Mg+2}]{\text{leaching of K+}}$	Vermiculite	high
Low K Silicate, Chlorites	$\xrightarrow[\text{high silica concentrations}]{\text{alkaline conditions}}$	Smectites	high

after Quigly and Ogunbadejo, 1976; Quigly, 1975; Bohn, McNeal and O'Conner, 1978.

Figure 6: Schematic of clay/water system

High water content

clay mineral (face) <== H2O <== cation ==> H2O ==> (face) clay mineral

Low water content

clay mineral (edge) <==> clay mineral (face)

Green and others, 1983).

Stress Relief

Erosion of overburden or lateral removal of material can lead to the formation of joints oriented perpendicular to the direction of unloading. Non-planar fractures were found to form in clay parallel to an exposure in a few months, and formed a girdle perpendicular to the slope surface over a period of years (Fookes and Denness, 1969). The proportion of non-planar to planar fractures was found to increase with increasing exposure age.

Since the work of Terzaghi (1936), it has been observed that overconsolidated fine-grained sediment is typically fractured. In some cases, the fracturing may result from expansion that accomponies unloading. Overconsolidation can occur through the formation of segregation ice lenses (Boulton and Paul, 1976), through release of overburden pressures by erosion, consolidation under glacial ice (Harrison, 1957), or through fluctuations of the water table (Suderman and Kim, 1970). Tills deposited in the subglacial environment are commonly overconsolidated as a result of the large effective stress beneath thick ice, but may escape compaction if frozen during deposition (Mickelson and others, 1979) or if high pore pressures limit the effective stress that the glacier can exert on its bed.

Preservation

Fractures can persist in unlithified sediment through annual and long-term fluctuations in temperature and moisture. Fractures at depth remain at a relatively constant moisture content, and so can control the orientation of fractures in overlying active layers in sections undergoing periodic dessication (White, 1966).

dilation

Dilation, the rearrangement of particles on the shear surface caused by shear, can influence subsequent fracture orientation because it permanently alters the fracture surface. The surface is altered as skeletal grains are forced away from the shear plane during shear. This plane, due to its lower strength, may then localize stress and control the formation of future fractures, even if the original fracture is no longer open (White, 1966). The process is restricted to dense granular sediment and overconsolidated clay (White, 1966; Mead, 1925). Dilation may also be dependent on clay mineralogy. Smectite's water layer acts to prevent grain to grain contact, and thus limits the subsequent "rebound" caused by electrostatic repulsion upon shear (Nelson, 1973). Dilation's effects were noted by Boulton and Paul (1976), who found that shear failure took place preferentially along planes where dilation (during secondary creep) had previously occurred.

aggregation

Aggregation, because of dessication, synaeresis, or cementation by binding agents, can be an important factor in maintaining undisturbed blocks of material between joints. Aggregation caused by dessication or synaeresis occurs when water is removed from the clay/water system (Figure 6)(Russell, 1934). With the removal of water, electrostatic forces can bind the clay particles "edge to face" and create stable structures of clay minerals (Baver and others, 1972).

Inorganic colloids from the hydrolysis of minerals in the till can act as significant binding agents in weathered horizons (Gustafsson, 1950; Frei, 1950). Lab studies have found that both hydrous aluminum and iron oxides are able to increase the durability of soil aggregates (Krishna Marti and Richards, 1974). In northern temperate regions, iron compounds are more prevalent, and may act as a crude cement when precipitated in amorphous form (Blackmore, 1973). Organic binding agents, such as polysacchrides, may also act to stabilize aggregates in the solum (Low, 1972).

Weathering products can preserve blocks of unjointed material by coating joint faces. Coatings commonly observed on joint surfaces in till include hydrous iron oxides and manganese oxides, clay skins, and coatings of silt grains (Chandler, 1973). Loss of material from the joint plane through dissolution or mechanical removal of fines may also act to preserve joints (Sleeman, 1963). Other alterations of the joint face, such as precipitation of secondary carbonate and reduction or oxidation of the till matrix have been observed (Chapter 3), but their influence on joint persistence is uncertain.

RELATIONS BETWEEN FRACTURE FORM AND GENESIS

Identification of fracture genesis is important in any attempt to interpret fracture characteristics or apply limited site investigations to a broader area. The preceeding review of fracture genesis is summarized below and in Table 1 in an effort to catalog the fracture style associated with various modes of formation.

The important fracture characteristics for interpreting fracture genesis include the change in fracture length and intensity with depth, coatings or weathering products and the nature of the fracture surface, and fracture orientation. This information must be combined with other data such as ice flow direction, deformation of underlying material, and the lithology of the fractured sediment. It is not always an easy task to assign a genetic history to every joint because joints formed by different processes may have similar characteristics.

CHANGES WITH DEPTH

Change in fracture characteristics with depth can be used to distinquish between fractures resulting from weathering processes in the near-surface environment from those formed during deformation or deposition, which form fractures at greater depth. Increase of fracture spacing with depth is common in fractures formed by dessication (White, 1966). Formation of jointing due to segregation ice lenses is restricted to the depth of winter freezing or permafrost (Boulton and Paul, 1976).

JOINT COATINGS AND ALTERATION

Joint coatings and alteration are important because they allow relative dating of fractures of various types, and are indicative of the importance of weathering processes. For example, joints caused by lateral stress release in the wall of an excavation are usually uncoated and so can be distinguished from older joints formed by other processes. Oxidation along joints shows the limit of chemical weathering processes in the till, and the general position of the water table. This can then be used to evaluate the importance of dessication in joint formation.

ASSOCIATED DEFORMATION

Deformation of associated sediment is additional evidence for a deformational history of joint formation. Jointing caused by tectonism of unlithified sediment is commonly associated with regional folding or faulting (Sleeman, 1963). Differential settlement or diapirism in underlying sediment may form joints that are parallel to normal faults and associated with deformed stratified sediments (Selsing, 1981; Banham, 1975). Joints resulting from glaciotectonic deformation and subglacial shear are typically associated with low angle reverse faults and recumbent folds.

LITHOLOGY

The lithology of fractured sediment may suggest the importance of certain processes in fracture genesis. Differentiation between subglacial till, supraglacial till, or sediment flows, and between till or glaciolacustrine sediment, is an important initial step in studies of fracture formation because it may then be possible to rule out classes of processes, such as subglacial deformation. Lithology is also a controlling factor in fractures resulting from dessication and chemical change; the strength of the relationship between physical properties and fracture characteristics can be used to evaluate the importance of weathering processes in fracture formation. For example, joint magnitude was found to increase with increasing plastic limit in a clay embankment (Anderson and others, 1982). Changes in carbonate content or clay mineralogy with depth may also point to weathering processes in fracture genesis.

ORIENTATION

Fracture orientation can be used in several ways to distinguish between joints of different origin. Near-horizontal joints have been reported to form from inheritance of prexisting ice/debris foliation, from shearing at the base of the ice, from the formation of segregation ice lenses, from dewatering, and from unloading. Vertical fractures are thought to have formed through dessication, tectonics, isostatic rebound, subglacial deformation, propagation of jointing from underlying bedrock, and shrinkage from removal of soluble components. Fractures with intermediate dips have been interpreted to result from lateral stress release, differential settlement, dessication, and compaction.

The consistent orientation of vertical fractures can be used to distinguish between fractures formed by deformation or stress release, which generally have preferred orientation, from fractures formed by weathering or syneresis, which typically exhibit random orientation unless controlled by other factors.

Fractures in unlithified sediment resulting from tectonic activity or glacial deformation typically have a consistent orientation relative to the direction of the deforming stress. Change in fracture orientation areally may be related to fracture genesis; fractures formed by processes operating on a large scale, such as isostatic rebound, tectonism, or earth tides, should have consistent orientation over a wide area if they have not been affected by preexisting planes of weakness or upward propagation of bedrock jointing. This is in contrast to fractures formed by overriding ice. Joints formed by subglacial deformation may have joint orientations consistent with ice flow direction, and so may change over smaller distances in association with changes in ice flow.

The relationship between fracture orientation in till and the orientation of joints in the underlying bedrock can be used as evidence of tectonics or isostatic rebound, but the absence of agreement between joint patterns in the till and bedrock should not be used as evidence to discount these processes, because the dissimilar materials may behave differently under stress.

JOINT SURFACE CHARACTERISTICS

Many researchers have attempted to use joint surface characteristics as indicators of joint genesis. Price (1966) has stated that shear joints in lithified sediment commonly are smooth, planar, and unaffected by lithologic changes, while tensional joints tend to be more irregular and influenced by lithologic breaks. Feather or plumate structures have been observed on the surface of joints in rock which has failed under brittle shear (Roberts, 1961). However, their absence is not diagnostic, because shearing can also smooth joint faces (Hodgson, 1961). In unlithified sediment, this smoothing may be caused by dilation or reorientation of clay particles during plastic flow (White, 1966). Joint surfaces in unlithified sediment that undergo repeated opening and closing because of dessiciation often show preferred orientation of clay particles and formation of slickensides (Bullock and Murphy, 1976). Matte surfaces, rarely polished or covered with clay gouge, were found on joints in unlithified sediment thought to be caused by deformation (Chandler, 1973).

CONCLUSIONS

Fractures in unlithified sediment are caused by a variety of processes which can act alone or in combination, making genetic classification difficult. These processes include deformation from shear or compaction beneath glacier ice, unloading during deglaciation, erosion, or following overconsolidation, stress resulting from differential isostatic rebound, and shrinkage cracks from physical and chemical weathering processes. These processes can be superimposed on the same structures through time, and it is often unclear whether the final fracture form reflects every event that influenced fracture genesis, or just the most recent disturbances.

Description of the fracture style associated with the different processes of fracture formation can lead to a model of fracture formation given the fracture characteristics at a particular site; suggested methods of fracture measurement and characterization are the topics of the next paper (Chapter 2). Fracture style includes the orientation of fractures, along with their length, spacing, and surface characteristics. Caution is necessary, however, as different processes can form fractures with similar attributes. Information on the fracture characteristics should be combined with information on the lithology of the sediment, weathering profiles, associated deformation, and geologic history in order to learn as much as possible about the factors that influenced fracture genesis.

Table 1 Joint patterns in unlithified fine-grained sediment

Origin	Material	Orientation	Pattern and Characteristics
"Tensional" fracture	till or clay		uneven and hackly; low lustre; high asperity
(Sleeman, 1963)			irregular, affected by minor lithologic breaks
"Shear" fracture	till or clay		feather or plumate structures
(Price, 1966)			smooth, planar, and unaffected by lithologic changes
Depositional			
fluctuations in sedimentation	lacustrine and marine clay	near-horizontal in undeformed sediment	parallel to bedding closely spaced partings ("fissility")
inheritance of englacial foliation (Shaw, 1977)	meltout till	near-horizontal	

Table 1 Joint patterns in unlithified fine-grained sediment (continued)

Origin	Material	Orientation	Pattern and Characteristics
Deformation			
(Boulton, 1970)			perpendicular to ice flow
(Trainer, 1973)			parallel to ice flow
(Boulton, 1970)		Near-Horizontal	dipping up-ice or indeterminant dip slickensided, evidence of shear
Compaction (La Fluer, 1980)	Till or other overridden sediment	oblique to surface	dipping in range of $45+\phi'/2$ to $45-\phi'/2$
Unloading (through melting, erosion, or elevated pore pressures) (Boulton, 1970)	Till or clay	parallel to surface; radial around margin of the deposit	
Isostatic Rebound (Grisak and Cherry, 1975)	till or clay	Vertical	may be similar to bedrock jointing; propagation of bedrock joints or lineaments into overlying lacustrine sediment; consistent over wide area in similar material

Table 1 Joint patterns in unlithified fine-grained sediment (continued)

Origin	Material	Orientation	Pattern and Characteristics
inheritance of preexisting joints from parent material (Jennings, 1977)	residual soil	vertical and horizontal	coincidental with bedrock jointing; persistance of joints into underlying bedrock
synaeresis (White, 1961)	lacustrine and marine clay	random	parallel sided; sedimentation filling what were open fractures
Deformation			
Subglacial deformation (Pusch, 1973)	till or other overridden sediment		consistent preferred oriention matte joint surface, some polished or clay gouge-covered due to dilation
(Schwan and Ritzema, 1982)			associated with deformed sediment and compressive ice flow
(May and Thomson, 1978)			rectangular pattern
(Boulton, 1970)		Near-Vertical	Conjugate around ice flow

Table 1 Joint patterns in unlithified fine-grained sediment (continued)

Origin	Material	Orientation	Pattern and Characteristics
Lateral stress release (Fookes and Denness, 1969)	till or clay	strongly developed parallel to slope or cut; weak perpend. to cut	non-planar rather than planar
mass movement (McGown and Radwan, 1977)	flow till; mudflows	vertical with random orientation on surface of flow; radial around nose of flow	
Differential settlement (Selsing, 1981; Banham, 1975)	primarily till (deposition over buried ice)	vertical or oblique	parallel to normal faults and deformed bedding
Tectonic Stress (Sleeman, 1963)	till or clay	vertical?	consistant över wide area in similar lithology; related to regional folding and faulting; parallel sets of joints, smooth or matte

Table 1 Joint patterns in unlithified fine-grained sediment (continued)

Origin	Material	Orientation	Pattern and Characteristics
Earth Tides (Burford and Dixon, 1977)	till or clay		consistant over wide area; similar to underlying bedrock
Weathering			
Dessication (Boulton and Paul, 1976)	till or clay	vertical	polygonal
(Donovan and Foster, 1972)			polygonal and rectangular
White, 1966			spacing increases with depth
Bullock and Murphy, 1976, White, 1966)			stress cutans and/or slickensides from repeated shrink-swell
(Anderson, et al., 1982)			fracture length related to swelling potential
Birkland, 1974		low angle	

Table 1 Joint patterns in unlithified fine-grained sediment (continued)

Origin	Material	Orientation	Pattern and Characteristics
Freeze/Thaw (formation of segregation ice lenses) (Boulton and Paul, 1976)	till or clay	horizontal	closely spaced; restricted to depth of winter freezing or permafrost
freeze/thaw and wetting and drying (Bullock, 1978)	till or clay; upper B horizon of fine-textured soils	random	angular blocky structure
Chemical Alteration			
leaching of CaCO3 (Penny and Catt, 1967)	carbonate-rich till or lacustrine clays	random; vertical ?	polygonal ?
replacement of Na by Ca in smectites	marine clay	random; vertical ?	
dehydration due to nonpolar organic solvents (Green, et al, 1981; Green, et al, 1983)	landfill liners		

Chapter 2:

Measurement and Analysis of Jointing Characteristics in Fine-grained Glacial Soils

Peter J. Bosscher & Douglas E. Connell

ABSTRACT

Jointing can have significant effects on the permeability, shear strength, compressibility, and slope stability of fine-grained soils. Field measurement of jointing characteristics is an important step in the development of models that simulate fracture flow and directional strength. Preferred joint orientation, which can introduce anisotropy in permeability and strength, can be detected through independent and random measurement of joint orientation combined with graphical or statistical techniques. PATCH, a computer program based on the Poisson statistical test (Mahtab and others, 1972), is used to evaluate preferred orientation. It was tested for its ability to reject spurious concentrations of joint poles and calculate mean joint orientation and dispersion. From an analysis of till joint data, we conclude that the use of the statistical computer program alone provided spurious results and required supplementation with other methods to properly determine joint set orientations.

INTRODUCTION

Fine-grained till units are often regarded by geotechnical engineers as homogeneous and non-fractured examples of overconsolidated clay. Observations made by most of the authors cited in this paper demonstrates that fracturing can be common in fine-grained till such as that found in many areas surrounding the Great Lakes (Chapter 3). Jointing can also be found in expansive residual clay (Williams, 1975) as well as in most dessicated clay (e.g., Maher and O'Neill, 1983).

Numerical models are being used more frequently for simulation of geologic conditions at hazardous waste and low-level radioactive waste sites. Sophisticated models incorporating fracture behavior require an accurate description of joint systems. This paper reviews the methods of joint measurement and characterization that are necessary for input to numerical models of shear strength (McKinlay and others, 1975), and permeability (Long and others, 1982), and also evaluates techniques for the identification of joint sets and their orientation. From our evaluation of these techniques based on an analysis of till joint data and synthetic data sets, we conclude that the use of statistical methods alone can produce spurious results and require supplementation with other methods to properly determine joint set orientations.

IMPORTANCE OF JOINTING ON SOIL CHARACTERISTICS

Joints with preferred orientation can induce three dimensional anisotropy of permeability, shear strength, and compressibility. In addition, slope stability is affected by the orientation of planes of weakness within the shearing zone. Anisotropy of fractured clay is complicated both by fracturing of the clay and the localized variability of preconsolidation pressure produced during desiccation. The variable preconsolidation pressures result from spatially variable suction pressures in the pore spaces, with highest suction developing near the joints (Mahar and O'Neill, 1983), but additional research into this phenomena is needed. The spatially varying preconsolidation pressure contributes to spatially varying permeability, shear strength, and compressibility within the intact blocks between joints which further compounds the difficulty of modeling such systems.

1. Permeability

Joints in fine-grained soil have been cited as causing failures in earth dams (Khan, 1983) and contributing to solute transport through clay liners (Prudic, 1982). The formation of joints can lead to a higher permeability, causing a shear stress (owing to increased hydraulic gradients and fluid flow) that exceeds the shear resistance of the particles in the wall of the joint, and eventually this stress may lead to piping failures. Joints may open and close because of moisture and stress-level changes. However Anderson and others (1982) found that even after the cracks had closed, the permeability of soils in which cracks or jointing had occurred increased several orders of magnitude on a permanent basis.

McGown and Radwan (1975) indicate that at stresses greater than that exerted by the overburden, joints in fine-grained till tend to close, reducing the permeability, whereas at stress levels less than the overburden, joints remain open. Therefore, during stress release the mass permeability of jointed clay can be expected to significantly increase when exposed in trenches or cut slopes.

Kazi and Knill (1973) also support the concept of joints closing at higher confining stresses, tending to mask the differences between fractured and fracture-free soil. However, they indicate that at the stress levels typically found in till, the permeability of jointed soil is considerably higher than non-jointed soil. Interpretations of the influence of joints on hydraulic conductivity or permeability are consistent in the literature, but unfortunately the methods used in the determination of permeability may result in poor quantification of the coefficient of permeability. This problem may stem from the inherent variability of soil fabric and, hence, permeability typically present in fractured soils.

The consolidation test is the most common method used to determine the permeability of jointed soil. Calculating permeability parameters from consolidation data requires one to assume ideal behavior of soil deformation characteristics. For homogeneous, isotropic, uniform soils these assumptions are difficult to support; they become especially untenable for jointed soils. Because of this, direct measurement of the permeability of jointed soil is necessary to determine the coefficients of permeability in different directions and at different stress levels. Proven methods for sample collection, preparation and testing remain to be developed for jointed soils.

Other methods to predict hydraulic conductivity changes based on joint characteristics have been suggested but have not been widely applied. Long and her coworkers (1982) developed methods to determine if a porous-media equivalent exists for materials with networks of discontinuous fractures. They found that a fractured medium behaves more like a porous medium when fracture density is increased, when fracture openings are of a uniform size rather than distributed, and when fracture orientations are distributed rather than uniform. Andersson and others (1984) developed a method for modeling fractured media that takes into account the uncertainty of the fracture geometry as well as the uncertainty of measurements in the field. The method needs to be compared to actual field problems before its full potential can be assessed.

2. Shear Strength

The influence of jointing on anisotropic shear strength has received considerable study. As with all soil testing, the quality of sampling and the method of testing the soil control the significance and applicability of the test results. When collecting samples for strength tests, the size of the sample relative to the spacing between fractures is critical (Bishop and Little, 1967). In addition, proper analysis is essential to gain meaningful results. Lo (1970) suggests methods for analyzing data based on joint spacing, specimen size, and undrained strength.

McGown and others (1974) using large samples, found that differences in the strength of a jointed till in different directions were as high as 60%; that is, some strengths were only 40% of the maximum strength found. Lower strength values occurred where the maximum shear stresses were produced along the fissure planes. Higher strength values occurred where the maximum shear stresses were produced perpendicular to the joint plane. This suggests that for other than randomly jointed clay, simple relationships between undrained strength and sample size cannot be applied without consideration of sample and joint orientation and joint spacing. According to McKinlay and others (1975) the most reasonable method of strength testing is to test a vertically oriented specimen of representative size (where the representative size is approximately 20 times the minimum fissure spacing) and combine the results with computer-aided model predictions of directional strength anisotropy ratios based on measured joint characteristics. Radwan (1974) has designed computer models which permit an indication of the three dimensional drained and undrained strength anisotropy ratios of tills based on fissure orientations and spacings. An estimate of the directional strength properties could be determined using this technique. If it is predicted that a structure will be critically affected by low soil strengths in particular directions, then limited detailed testing could be carried out to test the computer-aided predictions.

Some of the testing in the past has made use of the direct shear box, placing the joint surface at the prescribed shear surface of the shear box. Measured strengths will change dramatically if the sample is not placed in the shear box so that the shear takes place only along the joint surface. The error that would result would be a greater strength along the joint than really exists. Therefore, the results of such testing underestimates the magnitude of the loss in strength along a joint.

3. Compressibility

Compressibility, or more specifically the consolidation behavior of jointed soils, has also been studied. The consolidation characteristics of a soil are a function of its permeability, as well as other parameters. Therefore, if the permeability of the soil were anisotropic, the consolidation behavior should be anisotropic. The major differences in the coefficient of consolidation seen in soils seem to be related to differences in the permeability of the soil because of stress history. As in most clay, the coefficient of consolidation, c_v, in jointed clay appears to be larger in the direction of bedding. It was determined by McGown and Radwan (1975) that the consolidation characteristics of jointed clay were not greatly influenced by the presence of the joints. Particularly at pressures higher than the present overburden pressure, the vertical consolidation characteristics of clay with mainly near vertical and near horizontal fissures, were very similar to non-jointed clay.

4. Slope Stability

The stability of slopes in jointed soils seems to be strongly influenced by the presence of joints. The stability of such slopes is related to joint orientation, spacing, length, and surface characteristics. Many slope failures are thought to occur because of low shear strength along slip surfaces formed by joints. For jointed clay, the most common failure is a structural type failure where blocks of soil fail as units. These failures are generally shallow initially, but can progress to greater depth. The controlling factor is the orientation of the cut or exposed slope as compared with the preferred joint orientation. In most of the till examined in this study (Chapter 3), two types of fractures were found, near vertical fractures and near horizontal fractures. The near horizontal fractures comprised one set, the vertical fractures comprised two or sometimes four sets.

Sometimes, inclined joints can cause one side of an excavation to remain intact even as the other side fails. Aerial photographs of areas of jointed clay have shown the slope failures to be oriented along the preferred joint orientations (McGown and Radwan, 1975). In the same article, McGown and Radwan indicate that failures of this nature occur rapidly after cutting and stress relief.

Rainfall or high humidity can accelerate slope failure because of increases in water content along the joints resulting in softening of the soil. Perched water in the fractures produces hydrostatic stress within the fracture and has been suggested as a cause of additional slope instability (Sterrett and Edil, 1982).

JOINT CHARACTERIZATION

Owing to the complexities of groundwater flow in fractures and the effects of fractures on the strength of soils, computer models of fracture behavior have come into increasing use as predictive tools. Characterization of joint orientation, length, intensity, and spatial variability is necessary for input into models of strength anisotropy (e.g., McKinlay and others, 1975); effective hydraulic aperture must also be estimated for modeling the permeability of fractured media (Long and others, 1982). Additional parameters may need to be measured for purposes such as slope stability analysis. This characterization can be entirely quantitative or can be based on descriptive classifications (Fookes and Denness, 1969). Commonly measured joint parameters, the error associated with their measurement, and their observed statistical distribution are discussed below.

"Error" is used here to mean the difference between a measurement and its true value. In general, the true value of any geologic parameter is never known, but can only be estimated from many measurements, all which are subject to error. Error resulting from inaccuracy of the measuring device is termed measurement error and is typically assumed to have a normal distribution around some mean value. Bias, or systematic error, can also occur during measurement, and alters the mean value of the measured parameter.

The accurate and unbiased estimation of joint parameters is dependent on careful and independent measurement of joints in the field. An ideal exposure allows the examination of joints in three orthogonal directions, decreasing the bias that can be introduced by the undersampling of joints oriented at a low angle to the sampling face (Terzaghi, 1965). Either line surveys, as in sewer trenches (Williams, 1975), or aerial surveys, as in an excavation, can be used. Excavation of a small cavity into the wall or floor of the exposure is desirable as it allows jointing to be measured in three dimensions as the cavity is excavated, and keeps the disturbance of the soil to a minimum (Fookes and Denness, 1969).

Measurement Parameters

Orientation

Joint orientation is readily determined using a compass and inclinometer. Either the strike and dip or the dip and direction of dip of the joint plane can be measured. Errors in the measurement of dip have been estimated to range from 2.9° for high angle joints in hard rock (Ronca and Chaivre, 1977) to 5.0° (West, 1979) for the dip of joints in consolidated sediments. Measurement error for planar joints in fine-grained soils is probably within this range. Error in the measurement of azimuth, either strike or dip direction, is influenced by the dip angle and increases with decreasing dip. Cruden and Charlesworth (1976) found that the standard deviation of repeated strike measurements (S) was related to the dip angle by the following equation:

$$S = D \csc (\delta) \tag{1}$$

where D is the standard deviation of repeated dip measurements and δ is the dip angle. This relationship assumes a normal distribution for strike and dip error, equivalent to the Arnold (Fisher) distribution on the sphere. Field data from the repeated measurement of joints in a dolomite quarry shows a somewhat similar relationship (Figure 1).

For near-vertical jointing, joint azimuth alone may be sufficient

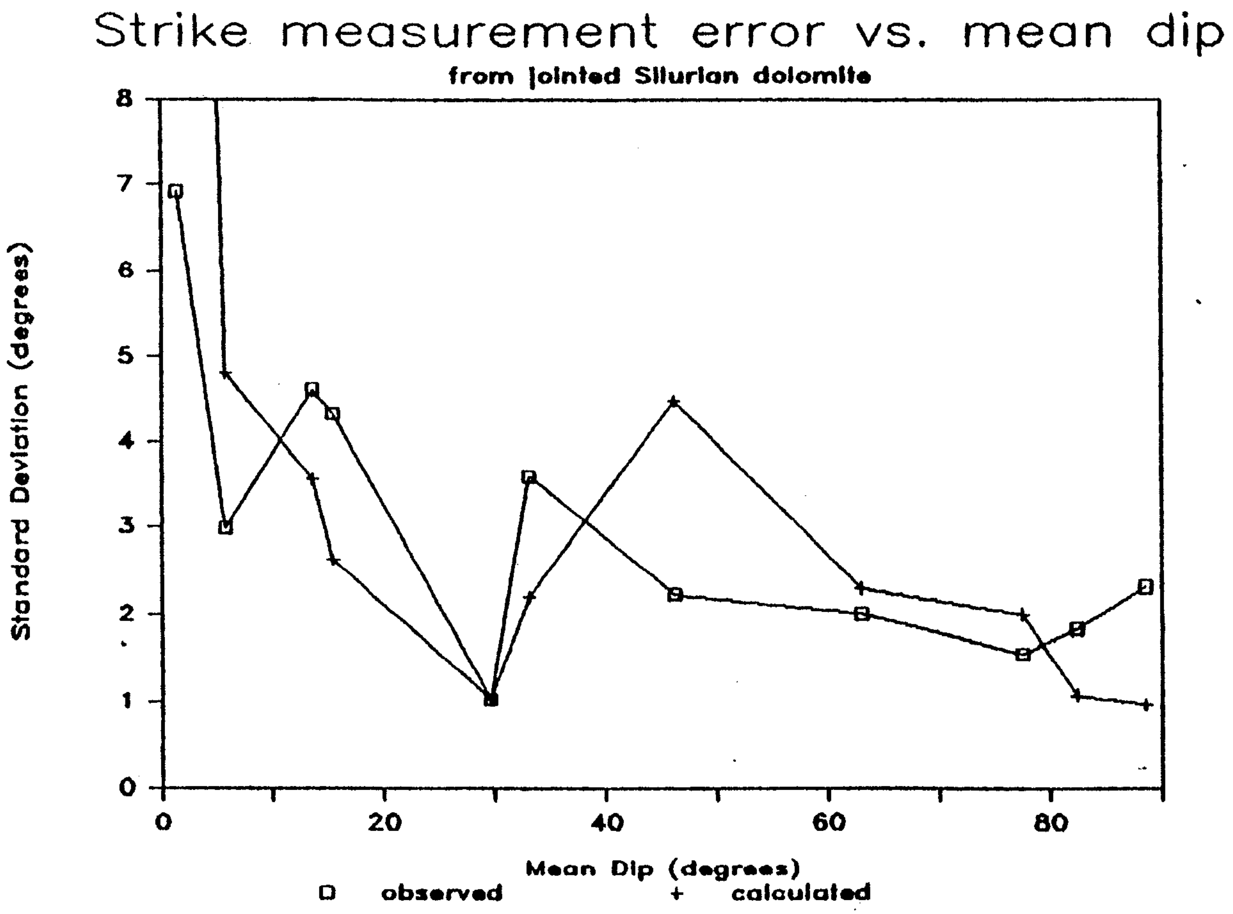


Figure 1: Standard deviation of measurement plotted against mean dip angle. Observed values from repeated measurement of joints at Door Quarry (Site #2) and Valders Quarry (Site #14), Wisconsin. Calculated values derived from the same data using Equation 1 (Cruden and Charlesworth, 1976).

to characterize joint orientation. One approach for two-dimensional joint characterization is the quadrat method (Abdel-Rahman and Hays, 1981). The presence or absence of joints with a given orientation (e.g., with a strike of 20 to 30 degrees) is recorded for many sampling localities, or quadrats. A rose diagram is then constructed from these data by plotting the number of sites containing joints in a given strike range. This differs from the usual method of plotting the number of joints in a given strike range. This method is little affected by exposure orientation because it does not depend on the number of joints of a given set at any one quadrat, and can rapidly be applied by using photographs of the quadrats (Abdel-Rahman and Hays, 1981).

Spacing

Joint spacing is measured as the perpendicular distance between adjacent joints or the overlapping ends of joints of a single set. Measurement should be done as close to the exposure face as possible to prevent bias of the measurement toward smaller spacings (Wheeler and Dixon, 1980; Wheeler and Holland, 1981). To avoid measuring the same joint spacing more than once, it is helpful to always measure in the same direction (Wheeler and Dixon, 1980). Spacing can also be estimated from location data if individual joints are mapped or tied to a coordinate system of measurement during excavation (McGown and others, 1974).

For unbiased estimation of joint spacing, the size of the exposure must be large. Scanline lengths of fifty times the mean spacing distance were found necessary for estimation of the distribution of joint spacings (Priest and Hudson, 1976).

Joint spacing can be described by an arithmetic mean and standard deviation, or by the trimean (Figure 2), which is not as affected by extreme values (Wheeler and Dixon, 1980). Error in the measurement of joint spacing when measurements are made from joint to joint are estimated to be low (Wheeler and Holland, 1981).

Joint spacing of joints within a single set usually have a positively skewed distribution (Figure 3), typically log-normal or exponential. Priest and Hudson (1976) state that joint spacings in rock, for all joints in an exposure (not only of a single set), commonly have a negative exponential distribution of the form

$$f(x) = \lambda \exp\{-\lambda x\} \tag{2}$$

where λ is the frequency of joints (joints/meter) and $f(x)$ is the percentage of joints with spacing x. For the negative exponential distribution, both the mean and standard deviation of $f(x)$ are equal to $1/\lambda$. Limited field data from fractured till ($n = 21$; Figure 3) indicate that joint spacings from a specific set are also not normally distributed.

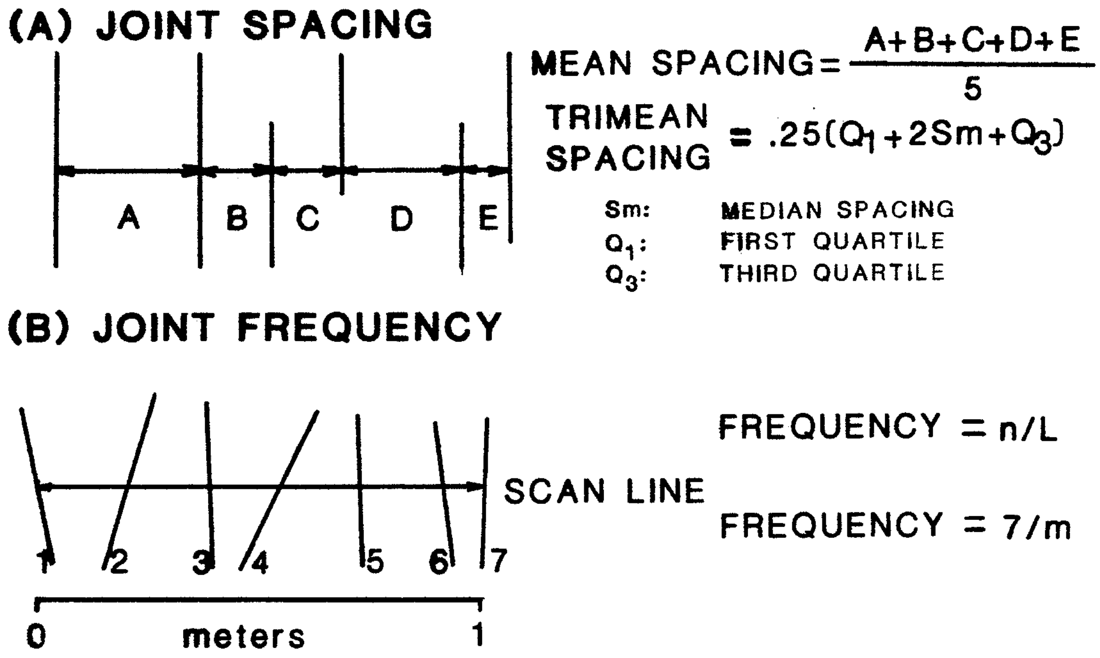
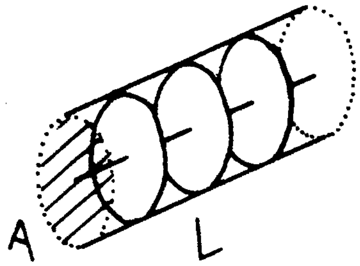


Figure 2a: Sketch of different measures of joint abundance. (A) Joint spacing, measured between adjacent joints of the same set. (B) Joint frequency, expressed as the number of joints per unit length.

(C) JOINT INTENSITY



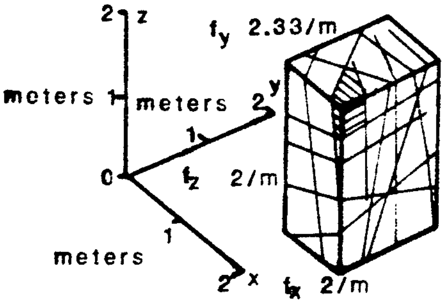
after Wheeler and Dixon,1980

$$\text{INTENSITY} = (nA)/(LA)\ \text{cm}^2/\text{cm}^3$$
$$= (n/L)\ 1/\text{cm}$$

$$\text{INTENSITY} = \sum_{j=1}^{m} \frac{1}{S_j} \qquad j=1,\ldots,m \text{ joint sets}$$

where A joint area
L scan line length
S mean joint spacing of a single set

(D) REPRESENTATIVE BLOCK SIZE



$$RBS = (1/fx)(1/fy)/(1/fz)$$
$$RBS = .11m^3$$

Figure 2b: Sketch of different measures of joint abundance. (C) Relationship between intensity and spacing. (D) Represenative block size, after McGown and others (1975).

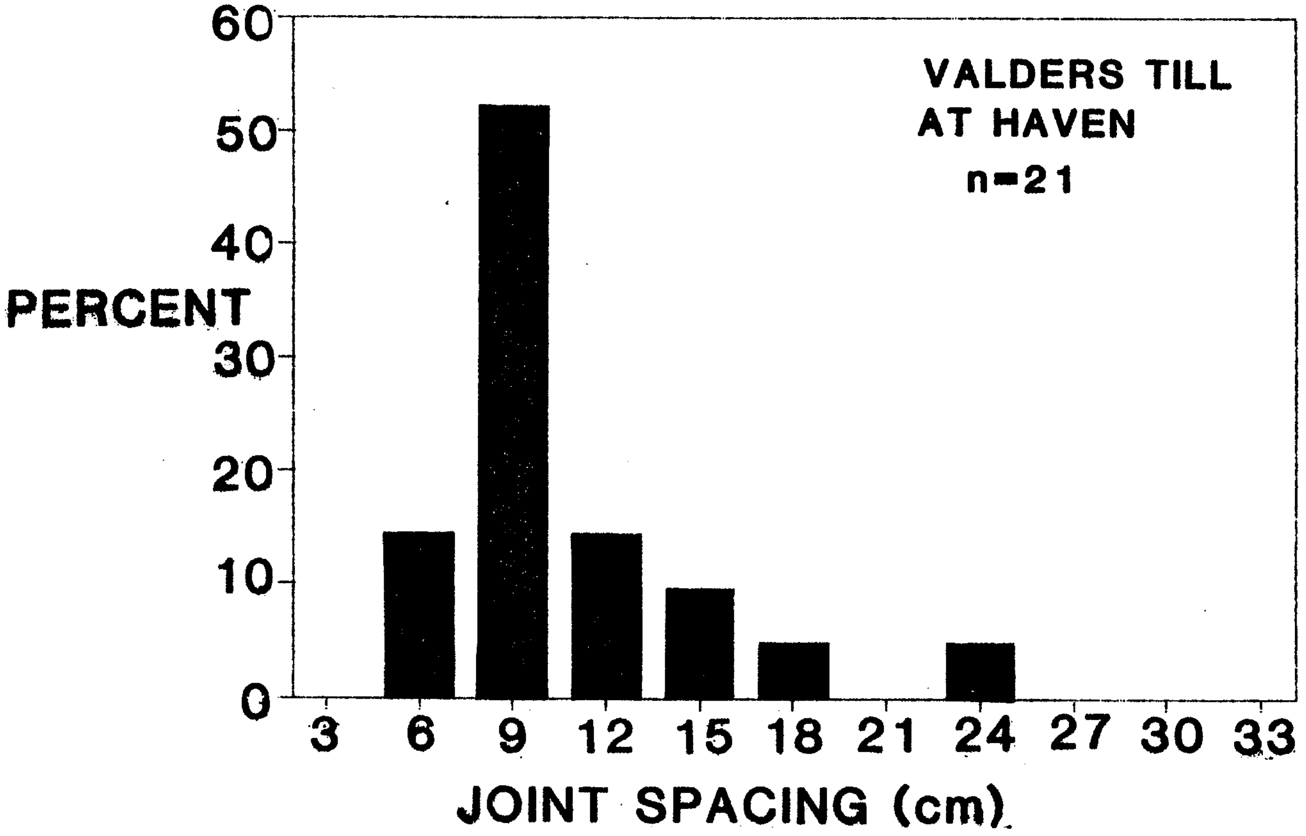


Figure 3: Joint spacing distance distribution for all joints (by set) in Valders till at the Haven site (Site #20-1).

Frequency and Intensity

Joint frequency, or the number of joints per unit length, is the inverse of the joint spacing if only a single joint set is present, or can be measured directly when more than one set occurs (Figure 2). Joint intensity, defined as the area of joints per unit volume (Fookes and Denness, 1969), can be calculated as the sum of the joint frequency of each joint set present (Wheeler and Dixon, 1980). Another measure of the degree of jointing is the representative block size (Figure 2), defined as the product of the mean joint spacings as measured in three orthogonal directions (McGown and others, 1974).

Measurement error of joint frequency can be as high as 100% because different operators do not always include the same features as joints (West, 1979). Because joint intensity depends on the number of joint sets as well as the frequency, no known statistical distribution exists for intensity (Wheeler and Dixon, 1980).

Length

Joint length is typically measured by noting the length of the intersection of the joint with the exposure surface. Error in measurement of joint length by this method is low (Wheeler and Holland, 1981). The distribution of joint length measurements is seldom normal, but usually has a skewed distribution (e.g., Figure 4). The exponential distribution is commonly assumed for joint length

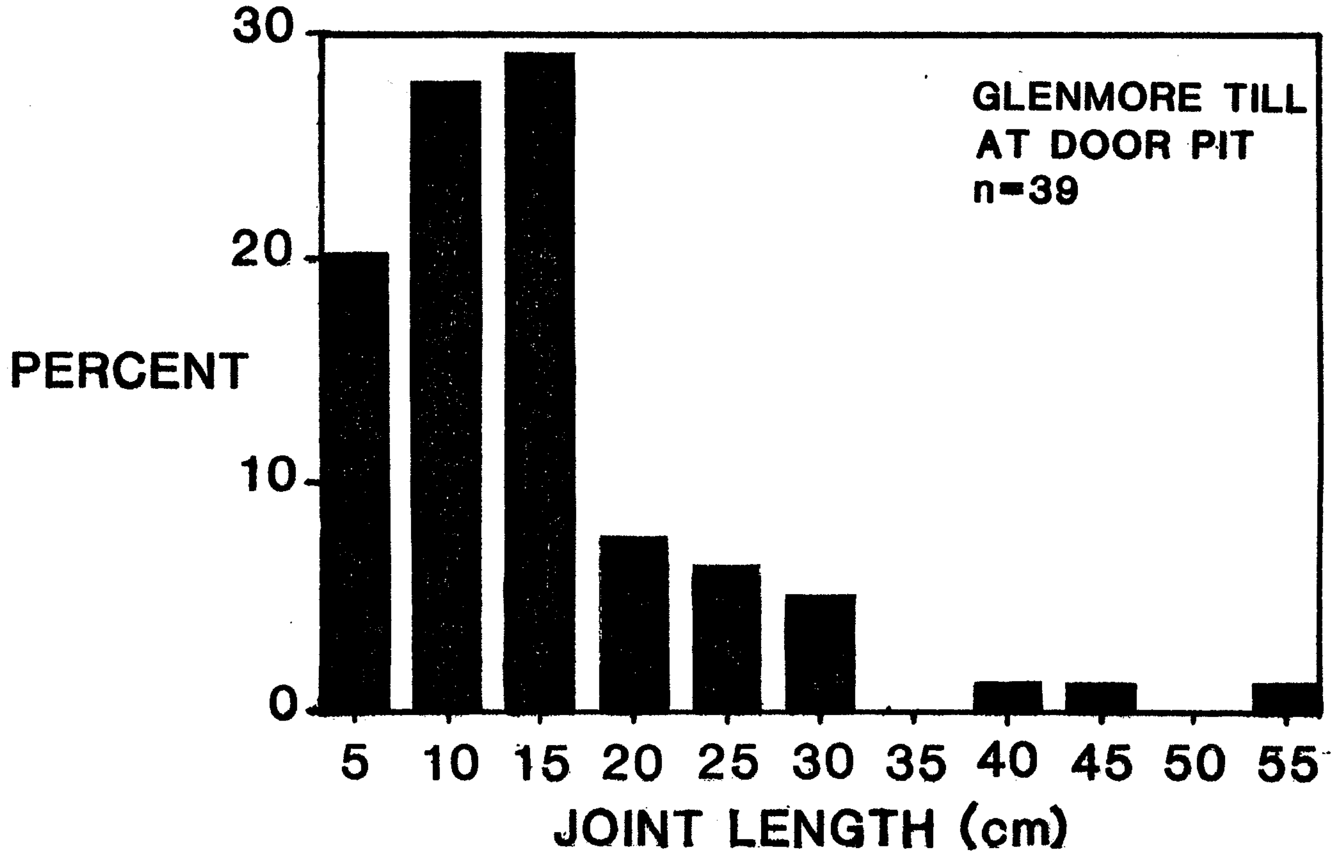


Figure 4: Apparent joint length distribution for Glenmore till at Door pit (Site #3).

measurements in rock, but the lognormal distribution may also be common (Baecher, 1980).

Measurement of joint length is problematic when only the apparent, rather than true joint length, can be measured, as when the joint is terminated by the exposure or extends into material beyond the exposure face (Laslett, 1982). Mean joint length and variance will be low compared to the actual values if only the apparent length is measured for longer joints, which have a greater likelihood for exceeding the dimensions of the exposure (Baecher and Lanney, 1978). If one or both ends of a joints intersection with the exposure are obscured, then the joint length measurement is said to be censored. By noting the number of joint ends that are visible, corrections can be made to the censored data which allow an unbiased estimation of mean joint length to calculated. Baecher (1980) presents a method for finding the mean of exponentially distributed joint lengths from censored (apparent) length data of the form

$$\hat{\mu} = (\sum X_i + \sum Y_j + \sum Z_k)/n \tag{3}$$

where $\hat{\mu}$ is the mean length, $\sum X_i$ the summation of joint length measurements for joints with both ends visible, $\sum Y_j$ the sum of joint length measurements for joints with one end visible, $\sum Z_k$ the sum of joint length measurements for joints with no visible ends, and n the number of joints with both ends visible. The variance of the joint length distribution is

$$V[\hat{\mu}] = \frac{\{\hat{\mu}\}^2}{n} \tag{4}$$

This method can only be applied to joints belonging to one set at a time, which may lead to large variances with small sample sizes.

Propagation of joints because of lateral unloading or other disturbance during excavation may bias the length measurements of older joints. Differentiation of older joints from joints stemming from unloading is possible through measurement of joint curvature, as joints in unconsolidated sediment resulting from unloading are usually less planar than joints resulting from other processes (Fookes and Denness, 1969). Joint curvature can be quantified by estimating the radius of curvature and the wavelength, as in the classification system of Fookes and Denness (1969). Older joints can also be distinguished by the presence of weathering products on the joint faces.

Spatial Distribution

The location of joints within an exposure can be established through mapping or photogrametric methods, or the position of every joint may be fixed by its intersection with a coordinate system of measurement (McGown et al, 1974). A recently developed technique, the azimuth versus traverse distance (AVTD) method (Wise and McCrory, 1982) can be used to characterize the changes of joint azimuth with distance. The AVTD method is based on the measurement of the azimuth of prominent joints at stations a few paces apart along a traverse. Joint azimuths are then plotted against the distance along the traverse (Figure 5).

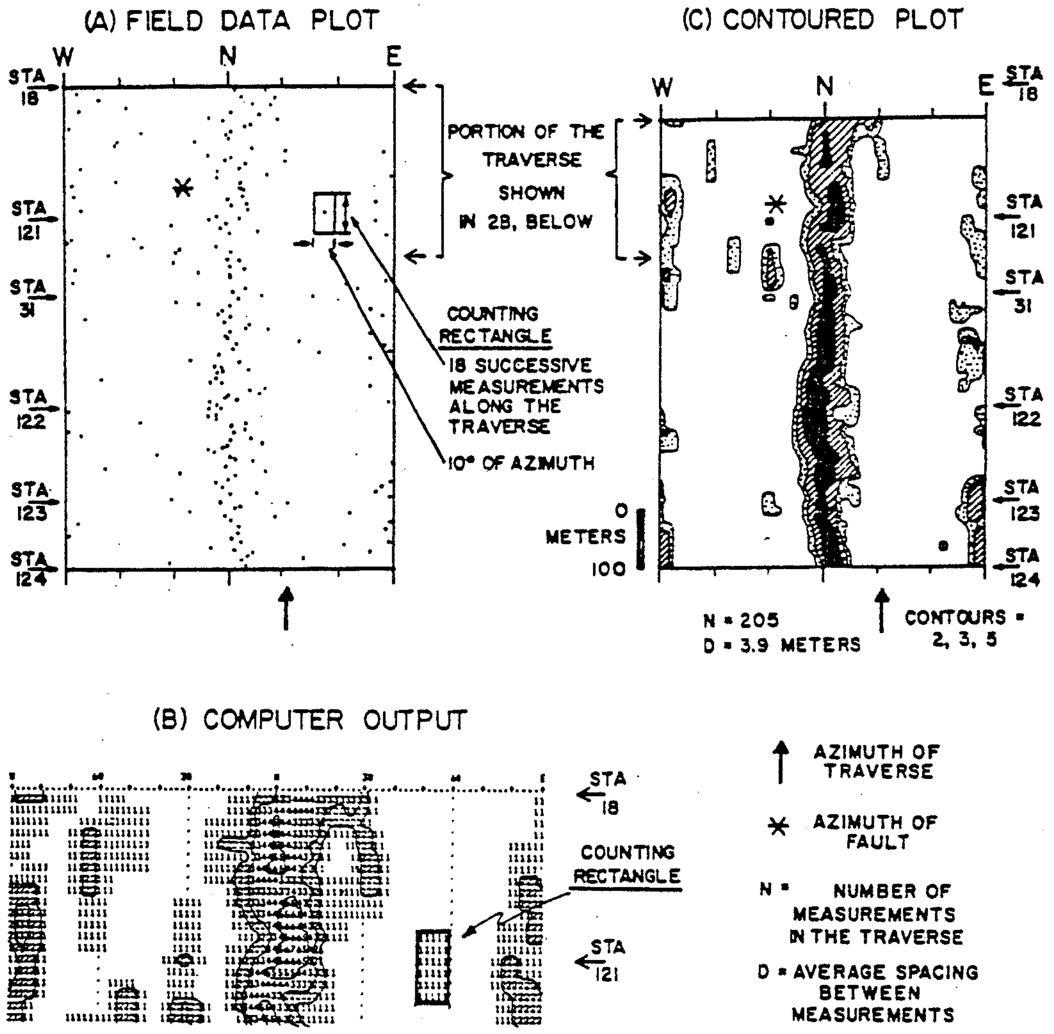


Figure 5: Example of azimuth versus traverse distance method, from Wise and McCrory (1982; Figure 2). Their caption reads "... Method of producing an azimuth versus traverse distance (AVTD) plot. (A) Plot of the data in field notebook style. Data are plotted sequentially from top to bottom as they are measured. Station locations at left are keyed to locations along traverse on map or air photo. (B) Computer output of contouring program for data in the upper part of A. A counting rectangle of the size shown in A is moved over the plot. Units are in concentration factors over and above that expected from a perfectly uniform distribution of the data. Effect of a single data point on the contours is shown by the numbers in the rectangle. (C) Finished contour plot with portion of B incorporated at top..."

Surface Characteristics

Asperity, or the roughness of the joint surface, is typically measured with a caliper or ruler (McGown and others, 1974). Smoother surfaces can be classified by comparison with material of known roughness such as grades of sandpaper (Fookes and Denness, 1969). Structures on the joint surfaces such as slickensides, feather fractures (Roberts, 1961), and other indicators of fracture genesis should be noted (Hodgson, 1961). As joint coatings and fillings can affect the strength of joints and give an indication of their relative age, identification of their thickness, composition, and abundance should be noted in the field. Typical joint coatings consist of weathering products such as iron and manganese hydroxides (Chandler, 1973); secondary carbonate, gypsum, and clay coatings also occur.

Detection of Preferred Orientation

Because directional variations of strength and permeability are influenced by preferred joint orientation, the detection of non-random joint patterns is an important step in investigations of jointed soil. Random and independent measurement of joints in an exposure allows the use of statistical methods to detect preferred orientation.

Most statistical methods for three dimensional orientation data are based on a plot known as a polar (or Pi) diagram. Polar diagrams are constructed by plotting the intersection of lines normal to the joint plane (termed poles) with the bottom of a hemisphere centered on the origin (Figure 6). The intersection of the poles and the hemisphere are then projected onto a two-dimensional surface known as a Schmidt net. The data can then be contoured to determine if preferred orientations exist. Contouring is typically done by computer, with programs based on an algorithm that counts the number of points that lie in an area equal to 1% of the area of the plot. The number of points is then converted into a percentage and contour lines drawn around areas of equal density (Figure 7). The data can also be converted into standard deviations away from a uniform data distribution, using a method developed by Kamb (1959) (Figure 8).

From an equal area plot, a three-dimensional version of the Poisson test, which tests directional data for non-uniform distribution, can be used to determine significant concentrations of points. The plot is broken down into small areas and the observed number of points in each area tested against the number that represents a statistically significant concentration of points. If many poles are concentrated in one portion of the plot (i.e. if there is a statistically significant departure from uniformity), the joints represented by the cluster of poles are assumed to have a preferred orientation. This method is the basis for the computer program PATCH developed at the Bureau of Mines (Mahtab and others, 1972) which is evaluated in this paper.

If more than one joint set exists at a site, the measurements of

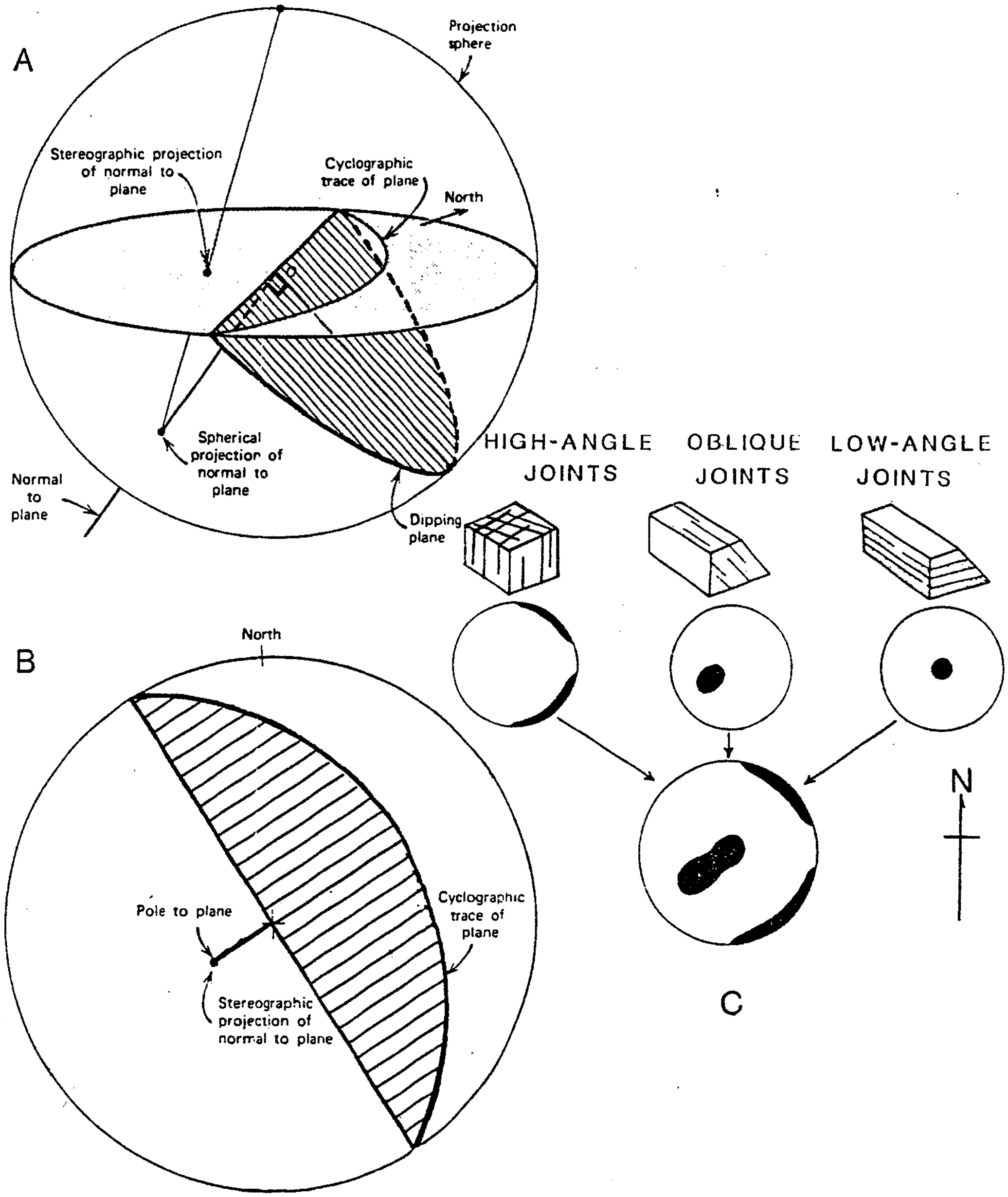


Figure 6: Method of constructing stereoplot. (A) Oblique view of dipping plane (ruled) showing position of projection sphere and the projection of the normal to the plane. (B) Projection plane (plan view) showing cyclographic trace of plane and the projection of the normal to the plane. After Hobbs, Means and Williams (1978). (C) Typical joint orientation and clusters of joint poles on a stereoplot. After Kazi and Knill, 1973.

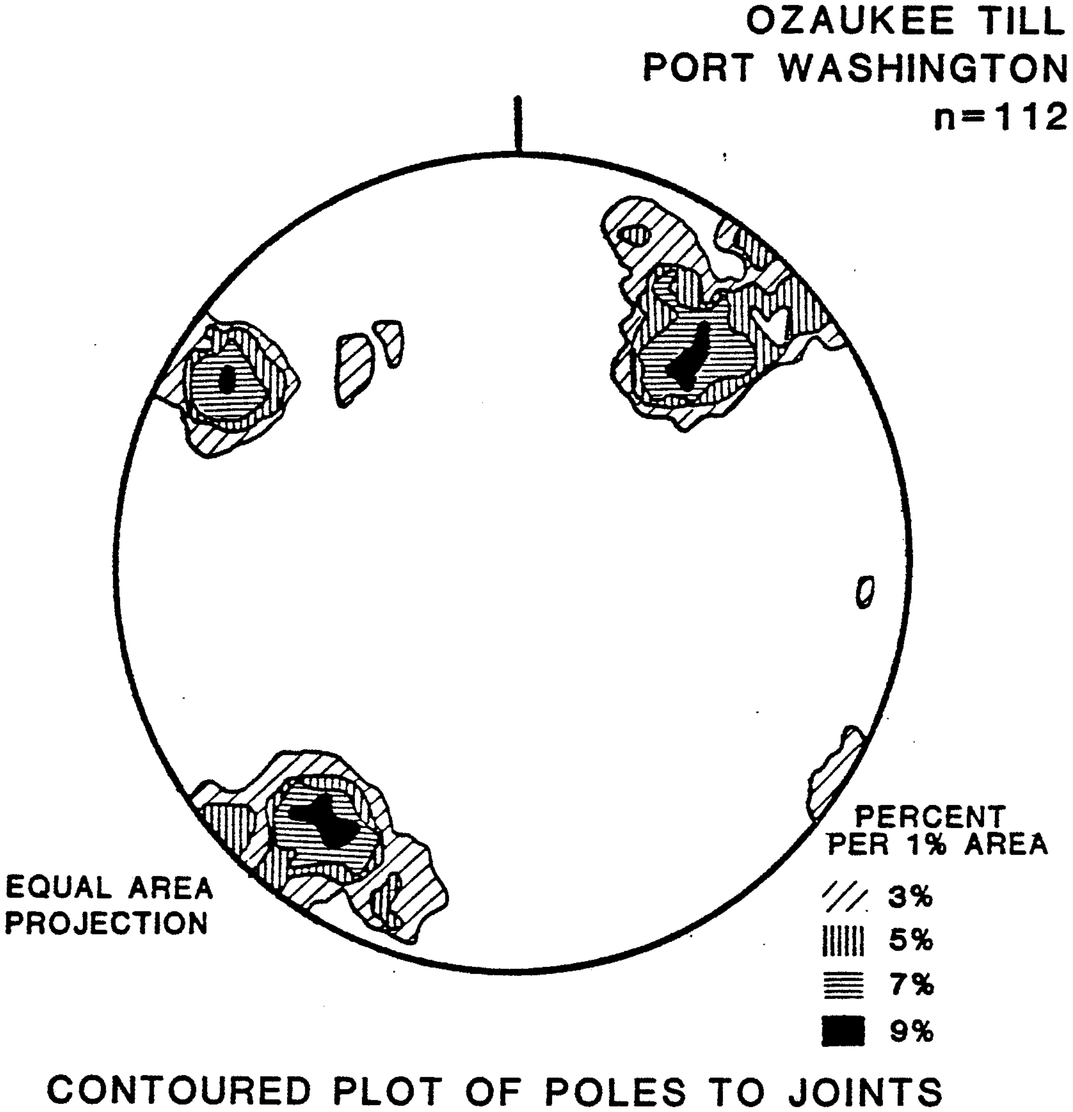


Figure 7: Contoured equal area stereoplot of poles to joints measured in the Ozaukee till at Port Washington (Site #21). Contoured from output of POINT (Warner, 1969).

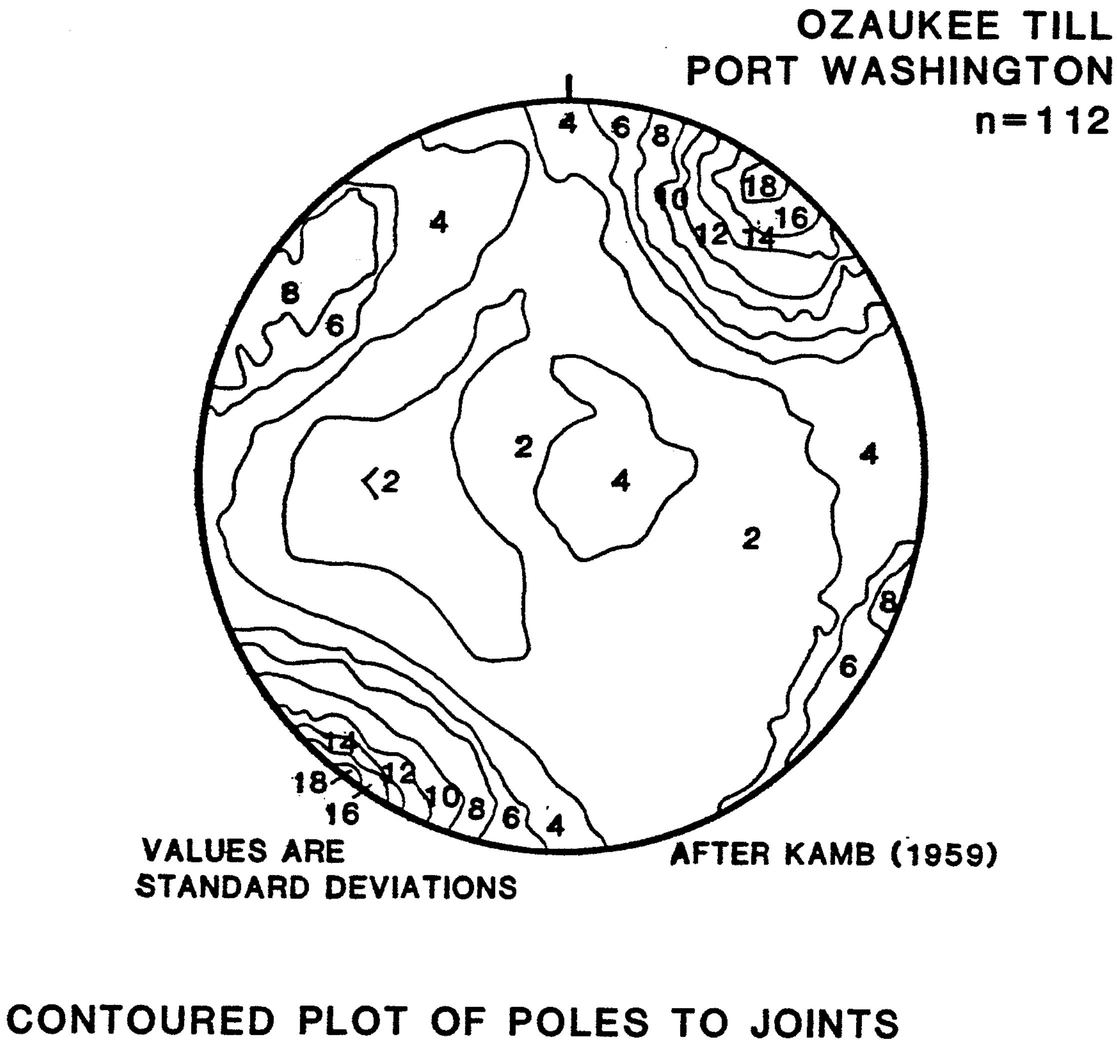


Figure 8: Contoured plot of value away from uniform distribution, in standard deviations. Plot of poles to joints in the Ozaukee till at Port Washington (Site #21). Output from FABRC (C. Corbato, Ohio State University) after the method of Kamb (1959).

joint orientation must be separated into sets before further analysis can take place (Bridges, 1975). Joints can be separated into sets using cluster analysis (Bailey, 1975), by sorting the orientation measurements via statistical tests (Ronca and Chaivre, 1977), or by combining adjacent areas that have significant numbers of points into clusters, as is done in PATCH (Mahtab and others, 1972).

Once the data are clustered, data from each joint set can be used to calculate a mean direction and dispersion (the three dimensional equivalent of variance) for the set. Different methods are used depending on the distribution that has been assumed for the joint measurements. For the radially symmetric Arnold distribution (also known as the Fischer or circular distribution), the mean orientation is calculated from the orientation of the vector sum of the projection of joint poles with unit length against three orthogonal axes (Mardia, 1972). The dispersion (k) of orientation around the mean direction for the Arnold distribution (analogous to standard deviation) is calculated from the magnitude of the resultant vector (Cheeney, 1983). For the more generally applicable Bingham (elliptical) distribution, the mean direction is estimated using the eigenvector of the greatest eigenvalue of the matrix of cosine values (Cheeney, 1983).

The distribution of joints around the mean direction is influenced both by measurement error and by natural variation (Figure 9). The goodness-of-fit of clustered data to theoretical statistical

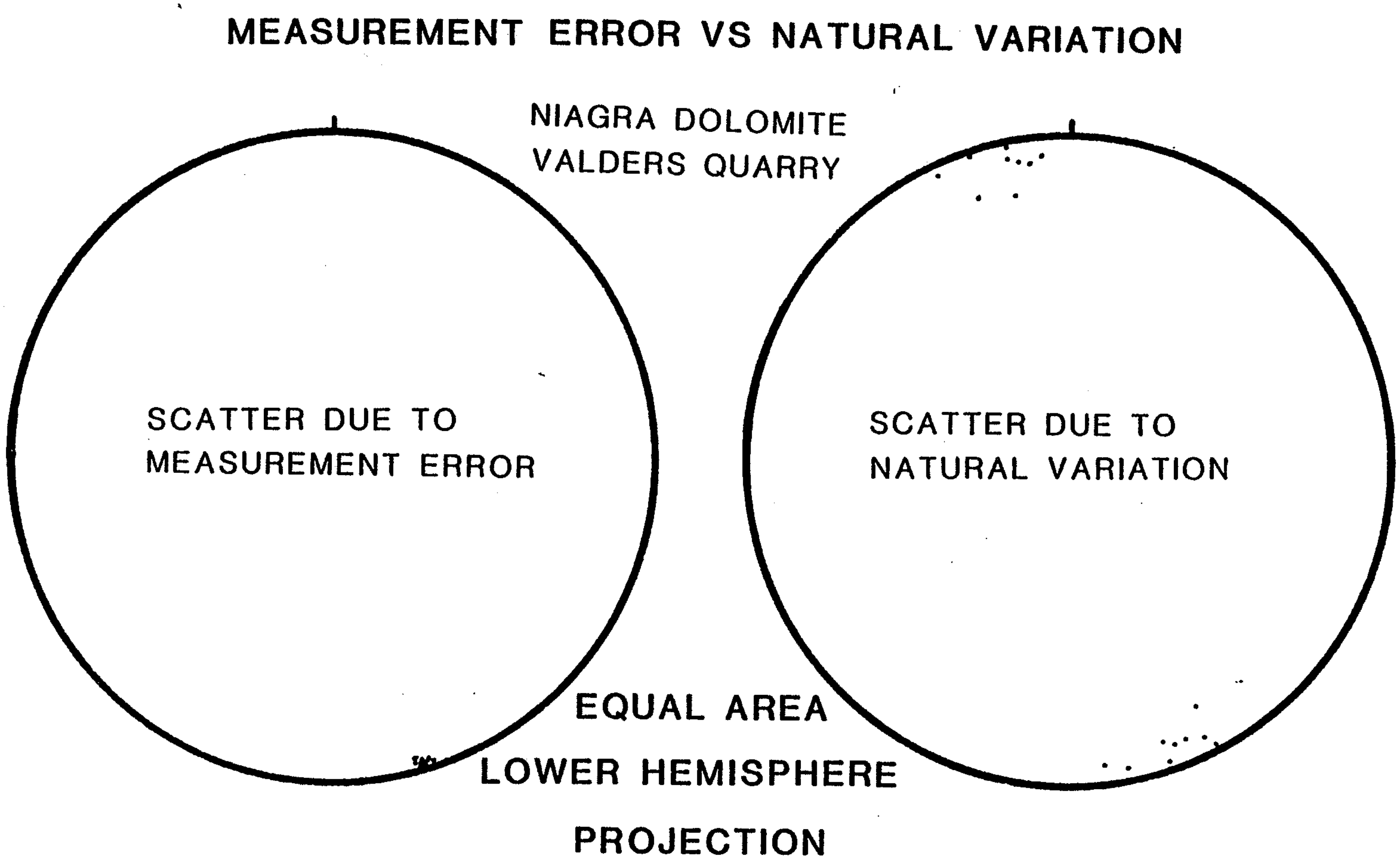


Figure 9: Comparison of scatter due to measurement error (from 10 measurements of one joint) and scatter from natural variation (from 18 measurements of joints in a single set in a 2 meter square area).

distributions can be tested by the chi-square method (Mahtab and others, 1972; Shanley and Mahtab, 1974).

Evaluation of Computer Programs using Field and Synthetic Data

As the methods outlined above entail tedious calculation, several groups of workers have developed programs to analyze orientation data (Mahtab and others, 1972; Shanley and Mahtab, 1975; Schuenemeyer and others, 1972). In the section below, the program PATCH (Mahtab and others, 1972) is evaluated following the methods of Stauffer (1966). The program was run on a Harris 800 computer and on an IBM Personal Computer.

Synthetic Data Results

Although it is advantageous to use computer programs to rapidly and objectively process orientation data, the investment in time and effort made to collect the data in the first place requires that the program provide significant quantification of the collected data. Five factors control the suitability of a specific algorithm to a particular field problem:

1) spurious concentrations of points in isotropic data sets;
2) the ability to resolve joint sets that have near-parallel orientations;
3) error in calculation of the dispersion around a mean

orientation;

4) error in the calculation of the mean orientation itself;

5) error in the test against three dimensional distributions.

Two independent factors control the importance of these potential problems: sample size and the confidence level chosen for statistical tests.

The influence of sample size and confidence level were analyzed by running the program PATCH with synthetic data sets. These data sets were comprised of random points distributed uniformly over a sphere. Several data set sizes were used: 10, 25, 50, 100, 200, 999 points with four different data sets for each size. Each data set was run at two different Poisson confidence levels, 99% and 95%. These confidence levels control the percentage of points required in 1% of total sphere area to be grouped together as a "cluster." Optimally, from uniform data, no clusters should be found as the data contains no natural clusters.

The results of the runs is shown in Figure 10 and Table 1. As can be seen, clusters were noted in 30 out of 48 runs, although 18 runs indicated no clusters. With a confidence level of 95%, the number of clusters per run was as high as 8 with the mean number of clusters increasing with increasing number of data points. A confidence level of 99% yielded a maximum number of clusters of 3, but the number of clusters was not a function of the number of data points. In addition, the number of clusters found for the 99% confidence level were usually about 8 times less than the number found

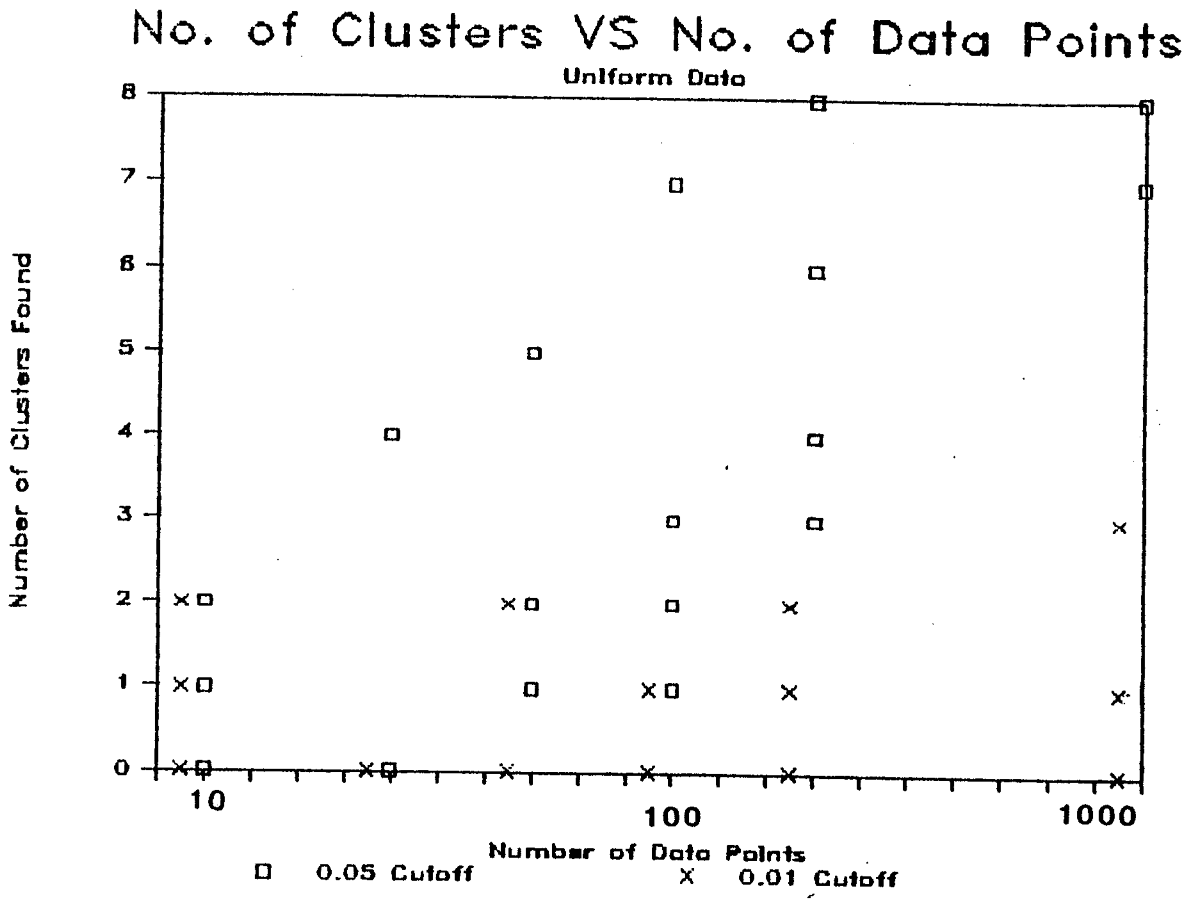


Figure 10: Number of spurious clusters found using PATCH from uniform, random synthetic data sets plotted against data set size at different Poisson confidence levels.

Table 1: NUMBER OF SPURIOUS CLUSTERS (SYNTHETIC UNIFORM DATA)

# of points	Confidence Level 95 %	99 %	
10	1.00	1.00	MEAN
25	1.00	0	
50	2.25	0.50	
100	3.25	0.25	
200	5.25	0.75	
999	7.25	1.25	
	95 %	99 %	
10	0.71	0.71	STANDARD DEVIATION
25	1.73	0	
50	1.64	0.87	
100	2.28	0.43	
200	1.92	0.83	
999	0.43	1.09	
	95 %	99 %	
10	0	0	MAXIMUM
25	0	0	
50	1	0	
100	1	0	
200	3	0	
999	7	0	
	95 %	99 %	
10	2	2	MINIMUM
25	4	0	
50	5	2	
100	7	1	
200	8	2	
999	8	3	

for the 95% confidence level. Our conclusion is that rejecting spurious clusters is best accomplished by using high Poisson confidence levels.

An analysis of the program's ability to calculate dispersion was done by analyzing synthetic data sets constructed using the Arnold distribution. Since PATCH compares clusters of poles to the Arnold distribution, results obtained from PATCH should match the parameters used to generate the data. Again, several sizes of data sets were run to determine the data set size required for accurate determination of dispersion. Five different dispersion values of each data set size were run. Figure 11 and Table 2 show the results. Figure 11 shows the error made by PATCH in computing the dispersion of the input data.

The figure seems to indicate that the error is a maximum at data sets of 25 points and is lower for data sets of 10 points. This is due to the manner in which the data points are clustered. To define significant concentrations of data points, a Poisson distribution is used to indicate the acceptable level of significance or "threshold" density. Clusters are then defined as collections of all points in adjacent patches, where each patch possesses a density that exceeds the threshold value. For data sets of 10 points and a Poisson confidence level of 99%, the required threshold density is 10% or 1 point, therefore each point meets the threshold density requirement. For data sets of 25 points and a Poisson confidence level of 99%, the required threshold density is 8% or 2 points and therefore the threshold density criterion is not as easily met. This produces a

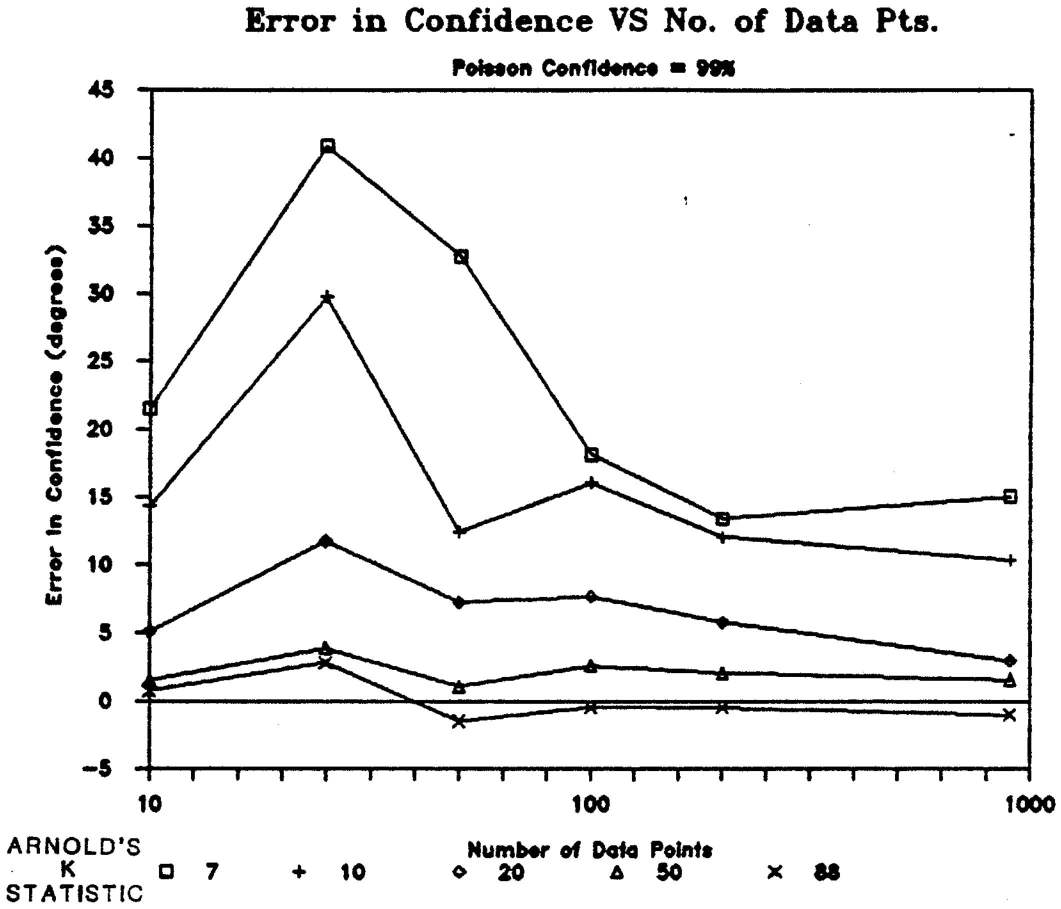


Figure 11: Plot of error in confidence limit versus the number of data points at different values of Arnold's dispersion (K). These results are from synthetic data sets run through PATCH (Mahtab and others, 1972). Data plotted here are shown in Table 2.

Table 2: Confidence Radius of Mean Cluster Orientation as a Function of Data Set Size at Different Values of Dispersion (K)

Calculated from Synthetic Data Based on the Arnold Distribution

# of points	Dispersion (Arnolds K)					
	7	10	20	50	88	
10	33.58	31.23	26.68	18.4	14.26	MEAN CONFIDENCE RADIUS
25	14.27	15.79	20.02	16.08	12.18	
50	22.38	33.05	24.55	18.86	16.48	
100	36.96	29.47	24.12	17.34	15.47	
200	41.71	33.48	26.02	17.92	15.49	
900	40.09	35.2	28.82	18.39	16.00	
10	8.5	5.85	4.48	2.84	2.47	STANDARD DEVIATION
25	7.26	8.11	5.24	2.21	2.12	
50	----	----	----	----	----	
100	----	----	----	----	----	
200	----	----	----	----	----	
900	----	----	----	----	----	
10	37	40	48	51	51	NUMBER OF DATA SETS
25	34	32	46	50	51	
50	2	1	1	1	1	
100	1	1	1	1	1	
200	1	1	1	1	1	
900	1	1	1	1	1	

seemingly larger error for data sets of 25 points compared to data sets of 10 points.

Additional observations of Figure 11 show that the error is lowest for values of k (Arnold's measure of dispersion) equal to 50 and 100. In general, the higher the k value, the denser the packing of the data points. The number of data points does not seem to change the error for the k equal to 50 or 100. In other words, PATCH calculates the dispersion of dense clusters much better than that of dispersed clusters for any sized data set. On the average, PATCH computed a confidence radius 20 degrees larger than the radius which was input. A possible explanation for this is the use of the maximum likelihood estimate method for determination of the confidence radius which overestimates the true confidence radius (underestimates the true dispersion) of the data. The larger data sets tended to produce smaller error and variation in the confidence radius as seen in the average and standard deviation values in Table 2.

The program also tests the distribution of clusters of poles by performing a Chi-square test against the Arnold distribution. Only 87% of the clusters from the synthetic (Arnold) data sets were found to match the Arnold distribution. Since all the input data sets were based on the Arnold distribution, one would expect PATCH to find 100% of the clusters to match the Arnold distribution. The results of the comparison which PATCH makes are not a function of the level of dispersion or scatter of the data, rather a function of the uniformness of the distribution of the data about the mean of the data.

The significance of the synthetic data with respect to the Arnold distribution was tested with the program PATCH and the test of the data against the Bingham distribution was done with the program FRACTAN, a computer program based on the Bingham distribution (Shanley and Mahtab, 1975). Unfortunately the last publicly available version of FRACTAN that we obtained from the Bureau of Mines incorporates every pole into a cluster (Shanley and Mahtab, 1975). Based on the data used in this analysis, FRACTAN identified more than 1.5 times the number of clusters than PATCH did for the same data. In addition, FRACTAN incorporated many more of the total number of poles in each cluster. The authors of FRACTAN have indicated that a new commercial version of FRACTAN is available which solves this problem.

In short, the applicability of the versions of programs which we used is limited and should not be relied upon as the only method used to determine the existence and orientation of joint sets.

Field Data Results

The methods of detection and characterization of preferred orientation described above were adopted for a study of fracture genesis in the fine-grained Late Wisconsin till units of northwestern and eastern Wisconsin (Chapter 3). Exposures of the till units were available in quarries, gravel pits, landfills, and on the actively eroding bluffs along Wisconsin's Great Lakes shorelines. At each site, the orientation of 100 to 250 joints was recorded, along with measurements of joint spacing and frequency. Maximum joint length was estimated from the longest joints exposed at each site. Disturbed samples were obtained for grainsize analysis and undisturbed oriented samples were taken for study of the till microfabric. The characteristics of the till units studied are shown in Table 3.

Joints in the clay till units studied commonly have preferred orientation, as shown by an example (Figure 7) contoured from output of the program POINT (Warner, 1969). The strength of the preferred orientation was estimated from the output of a computer program FABRC developed by C. Corbato of Ohio State University, which is based on the method of Kamb (1959). As shown by the contoured output from this program (Figure 8), the joint poles clustered around a mean value and represent a statistically significant concentration of points.

Mean joint orientation and the associated dispersion was required for models of joint behavior. The program PATCH was used to quantify the level of uncertainty in the estimation of mean joint orientation as it could handle individual joint sets separately. PATCH proved to be useful as it included subroutines not only to estimate mean orientation and dispersion, but also calculated the confidence limits around the mean for each statistically significant joint set.

Several problems were encountered when PATCH was used to analyze field data from jointed till. The first was a spurious concentration of points, which resulted from the large areas used for the Poisson test for non-uniformity. This error could have been corrected by

Table 3: Characteristics of jointed till units in Wisconsin

Unit	% Sand <2.0 mm	% Silt <.0625mm	% Clay <.002mm	% Expand ************	% Illite <.002mm	% Kaol/ Chlorite **********	% CO3 in coarse silt
Kewaunee Formation:							
Glenmore till	37.3(2.4) 9	33.8(1.7) 9	28.9(3.7) 9	25.6(11.8) 4	56.4(6.1) 4	18.0(6.4) 4	56.4(2.6) 7
Two Rivers till	26.7(7.5) 13	47.1(6.7) 13	26.2(7.3) 13	37.1(15.2) 7	55.4(16.0) 7	11.6(6.1) 7	43.8(10.3) 10
Middle Inlet till	23.0(6.6) 4	51.5(12.6) 4	25.5(6) 4	38.0(13.4) 3	45.7(8.6) 3	16.2(5.4) 3	26.8(6.6) 3
Valders till	21.7(9.2) 22	52.2(4.9) 22	26.1(10.7) 22	57.8(15.6) 11	34.5(12.5) 11	7.7(3.5) 11	42.9(3.5) 21
Haven till	14.8(10.4) 7	47.4(3.6) 7	37.8(10.4) 7	32.1(3.9) 3	54.5(3.1) 3	13.4(1.2) 3	37.8(3.0) 6
Ozaukee till	30.0(6.8) 6	46.7(5.2) 6	23.3(3.2) 6	48.9(-) 1	39.6(-) 1	11.6(-) 1	47.1(2.2) 6

Note: Values are listed as Mean (Standard Deviation)
# of samples

Table 3: Characteristics of jointed till units in Wisconsin (continued)

Unit	% Sand <2.0 mm	% Silt <.0625mm	% Clay <.002mm	% Expand ***********	% Illite <.002mm	% Kaol/ Chlorite **********	% CO3 in coarse silt
Oak Creek Formation:							
till 2C	17.1(2.6) 3	56.5(1.5) 3	24.6(4.4) 3	48.6(-) 1	11.8(-) 1	74.0(-) 1	14.2(-) 1
till 2B	10.0(4.6) 6	59.0(5.7) 6	31.3(5.6) 6	11.8(-) 2	74.0(-) 2	14.2(-) 2	48.6(-) 6
till 2A	23.6(12.6) 4	44.0(3.1) 4	32.4(10.6) 4	17.3(-) 2	61.4(-) 2	21.4(-) 2	48.6(1.2) 4
New Berlin till	14.3(-) 2	45.5(-) 2	40.2(-) 2	4.1(-) 1	76.1(-) 1	19.7(-) 1	41.8(-) 2
Miller Cr Formation:							
Douglas till	8.9(4.4) 14	21.2(10.2) 14	69.9(11.8) 14	75.0(-) 2	17.0(-) 2	8.0(-) 2	6.4(4.3) 3
Hanson Creek till	6.9(2.5) 5	22.9(6.0) 5	70.2(8.3) 5	52.2(13.9) 3	27.0(4.0) 3	20.8(12.8) 3	3.6(1.1) 5

Table 4: Geotechnical properties of jointed till units in Wisconsin

Unit	Liquid Limit (%)	Plasticity Index (%)	Water Content (%)	0' degrees)	c' (kN/m2)	OCR	Reference
Kewaunee Formation:							
Glenmore till							
Two Rivers till	26.3	12	15.6			4.8	Pulley, 1980
						31	Pulley, 1980
Middle Inlet till							
Valders till	22.8	10.4	13				Mickelson, et al., 1977
						9	Pulley, 1980
	28.4(5.0)	13.1(3.4)	17.4(4.1)	29.3(.6)	28.3(5.6)	4	Mickelson, et al., 1979
Haven till	30	13.5		29.5	30.21		Mickelson, et al., 1977
	30.3(6.0)	14.4(4.7)	16.5(2.5)	31.2(0.5)	23.8(5.6)	3.7	Mickelson, et al., 1979
Ozaukee till	30	14.4		30.5	0		Mickelson, et al., 1977
	30.6(2.6)	14.0(2.5)	17.6(1.2)	31.4(.8)	0	3.3	Mickelson, et al., 1979

Note: Values are listed as Mean (Standard Deviation)

Table 4: Geotechnical properties of jointed till units in Wisconsin (cont.)

Unit	Liquid Limit (%)	Plasticity Index (%)	Water Content (%)	O' degrees)	c' (kN/m2)	OCR	Reference
Oak Creek Formation:							
till 2C							
till 2B	20.3	6.4		31.4	0		Mickelson, et al., 1977
	25.1(2.8)	10.0(2.3)	15.2(1.4)	31.4	0		Mickelson, et al., 1979
till 2A	24.5(2.9)	9.9(1.9)	15.8(3.1)	31.1(.1)	0		Mickelson, et al., 1979
New Berlin till	Non-plastic		4.8	34.6	0		Mickelson, et al., 1979
Miller Cr Formation:							
Douglas till	69.5	45.1	47.3	22.2	2.55	3.1	Bagchi, 1982 Pulley, 1980
Hanson Creek till	63.7	32.4	34.8	30.5	0		Bagchi, 1982

Note: Values are listed as Mean (Standard Deviation)

decreasing the size of the discretization of the stereoplot, which would have involved extensive changes in the computer code and increased computer time. We chose instead to compare the clusters of joint poles from PATCH to the contoured field data as output by POINT. For our study on fractured till, only clusters of poles with 5% or more of the data per 1% area were considered significant and included in further analysis. Comparisons of plots of the same data set from POINT (Figure 7) and PATCH (Figure 12) illustrate how this correction was made. Cluster #3 in Figure 12 contained only 4 joints, 2 less than the 6 that would represent 5% of the total data. Thus, cluster #3 was assumed to be insignificant and removed from further analysis.

The second problem resulted from the inability of PATCH to resolve near-parallel joint sets. Again, comparison with the contoured data was used to confirm the output from PATCH. When close clusters of joint poles were incorrectly lumped together, the clusters were separated based on the contoured data and a mean orientation estimated from the contour plot. A decrease of four percentage points between adjacent clusters (e.g., from 7 to 3 percent per unit area) was assumed to indicate that adjacent clusters could be separated.

Increasing the confidence level from 99% to 99.5% was also used as a means to limit the number of clusters considered statistically significant. This technique improves the resolution and performs better in rejecting spurious concentrations of points from isotropic data sets, but sacrifices accuracy as it underestimates the dispersion of the joint set around a mean direction, and may distort the value

```
PORT WASHINGTON -- OZAUKEE TILL   DATA SET 22
THERE ARE  112 POINTS IN THE SAMPLE
EQUAL AREA LOWER HEMISPHERE PROJECTION
```

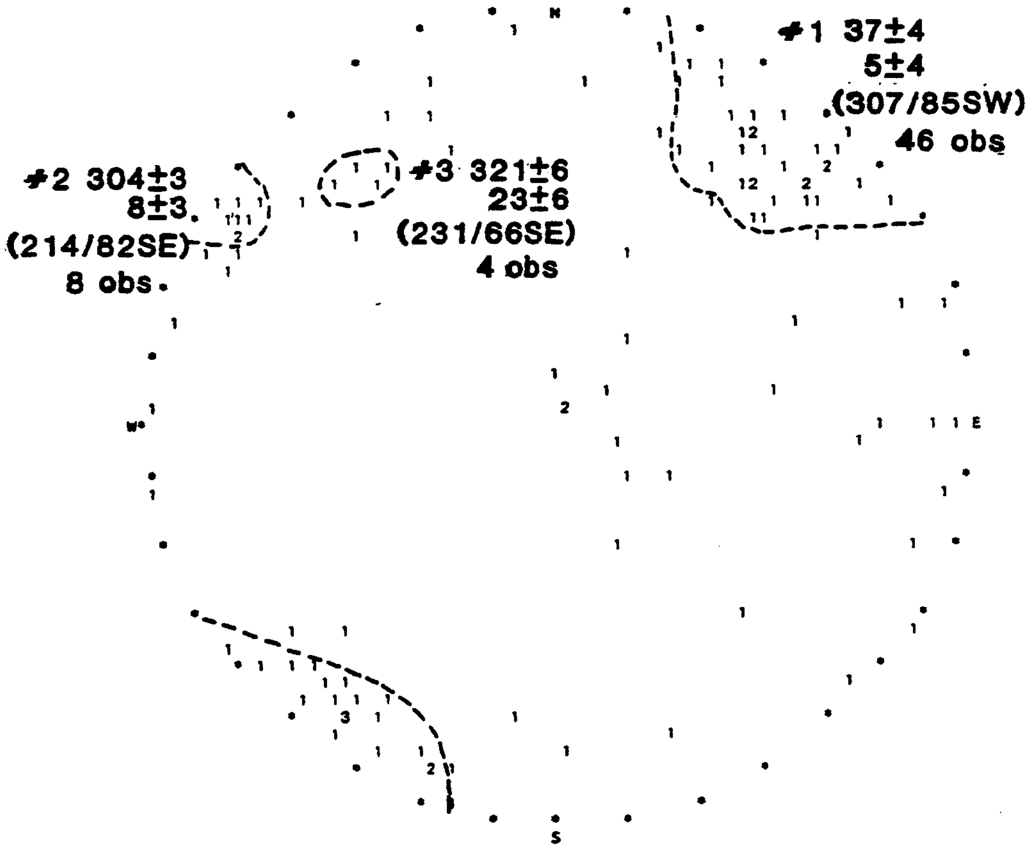


Figure 12: Output from PATCH (Mahtab and others, 1972) for poles to joints measured in Ozaukee till at Port Washington (Site #21). Plot is an equal area lower hemisphere projection. Small numerals within plot are number of points at each print position. Dashed line delineates clusters of points as grouped by PATCH. For clusters #1, #2, and #3, the trend and plunge of the pole to the mean orientation of the joint set are shown, along with the 95% confidence limit. Numbers in parenthesis are the strike and dip of the joint sets represented by the clusters of poles.

calculated for the mean joint set orientation if outlying points are not symmetrically excluded.

In data from fractured till at 28 sites, PATCH found 101 clusters of data points. Of these 101 clusters, only 73 clusters were determined to be correct. The others were excised for the reasons described above; low numbers of points per cluster or the inability of PATCH to resolve nearly parallel joint sets. POINT was used to identify an additional 32 clusters for a total of 105 joint sets.

Another advantage of PATCH was its subroutines to test the distribution of joints about the mean orientation of the set against the Arnold distribution. To test the distribution of clustered data, PATCH uses the Chi-square method, with a confidence level of 95%. Of the 73 joint sets identified using PATCH from the field data, 42 joint sets were determined to be significantly similar to the Arnold distribution. The remaining 31 joint sets identified by the program did not match the Arnold distribution. Therefore, about 58% of the joint sets identified by PATCH were significantly similar to the Arnold distribution. This percentage is almost identical to the 60% determined from the uniform, synthetic data. Although the similarities in these percentages may be coincidental, it is possible that they indicate that the distribution of joint poles about the mean joint set orientation is random and uniform, because of measurement error and natural soil variability. The 32 joint sets identified using the contouring program (POINT) were not tested against the Arnold distribution.

Data requirements

From the results of our evaluation of PATCH, some conclusions can be drawn about the size of a sample necessary to characterize joint orientation. It has been proposed that small sample sizes may not be adequate to allow the rejection of clusters in uniform orientation data, and that sample sizes of 300 to 400 measurements are required (Williams, 1975; Stauffer, 1966). As shown in Figure 10, when run at a confidence level of 99%, PATCH does as well at rejecting clusters of uniform data at sample sizes of 10 to 100 points as it does at sample sizes of 100 to 1000 points. For rejecting spurious clusters, large sample sizes do not appear necessary for use with PATCH.

There is, however, a trade-off between sample size and the accuracy of determination of mean orientation and dispersion. Figure 11 indicates that the error by PATCH in determining the confidence radius of a synthetic data set is smallest for large sample sizes and dense clustered data. Therefore the sample size should be tailored to the allowable error and desired accuracy of the intended study. For most studies, on the spot plotting of the data, even in its two dimensional form, can be sufficient to indicate when continued measurements become of marginal value (Fookes & Denness, 1969).

Decisions on sample size may also be influenced by the analysis requirements of joint characteristics other than orientation. Priest and Hudson (1976) found that 200 measurements of fracture spacing were required for analysis of the statistical distribution of spacing distances.

CONCLUSIONS

To apply the increasingly sophisticated models being developed to simulate flow through fracture systems and anisotropic stress behavior, careful and intelligent field work must provide basic data on the characteristics of joints in rock and soil. Techniques are available to measure and characterize joint length, spacing, and surface roughness, the important parameters for most fracture models. Detection of preferred orientation is of particular importance as it influences soil property anisotropy and can affect slope stability.

Methods of detecting preferred orientation must be combined to consistently test joint orientations against a random distribution. Even with the development of fracture analysis programs such as PATCH, Stauffer's (1966) conclusion that ". . . no single statistical test presently available is satisfactory by itself for determining the significance of weakly developed fabrics" remains true today.

The error in determination of mean joint orientation and dispersion is roughly governed by the sample size, and so the allowable error should dictate the sample size to most efficiently make the measurements in the field. Collecting large quantities of joint data has only marginal benefits for many applications. For some applications, the quantification of uncertainty available from statistical techniques is unnecessary and small sample sizes with simple contoured plots are enough. However, the use of statistical techniques alone can be misleading when applied to small data sets.

Chapter 3:

Distribution, Characteristics, and Genesis of Joints in Fine-grained Till and Lacustrine Sediment in Wisconsin.

INTRODUCTION

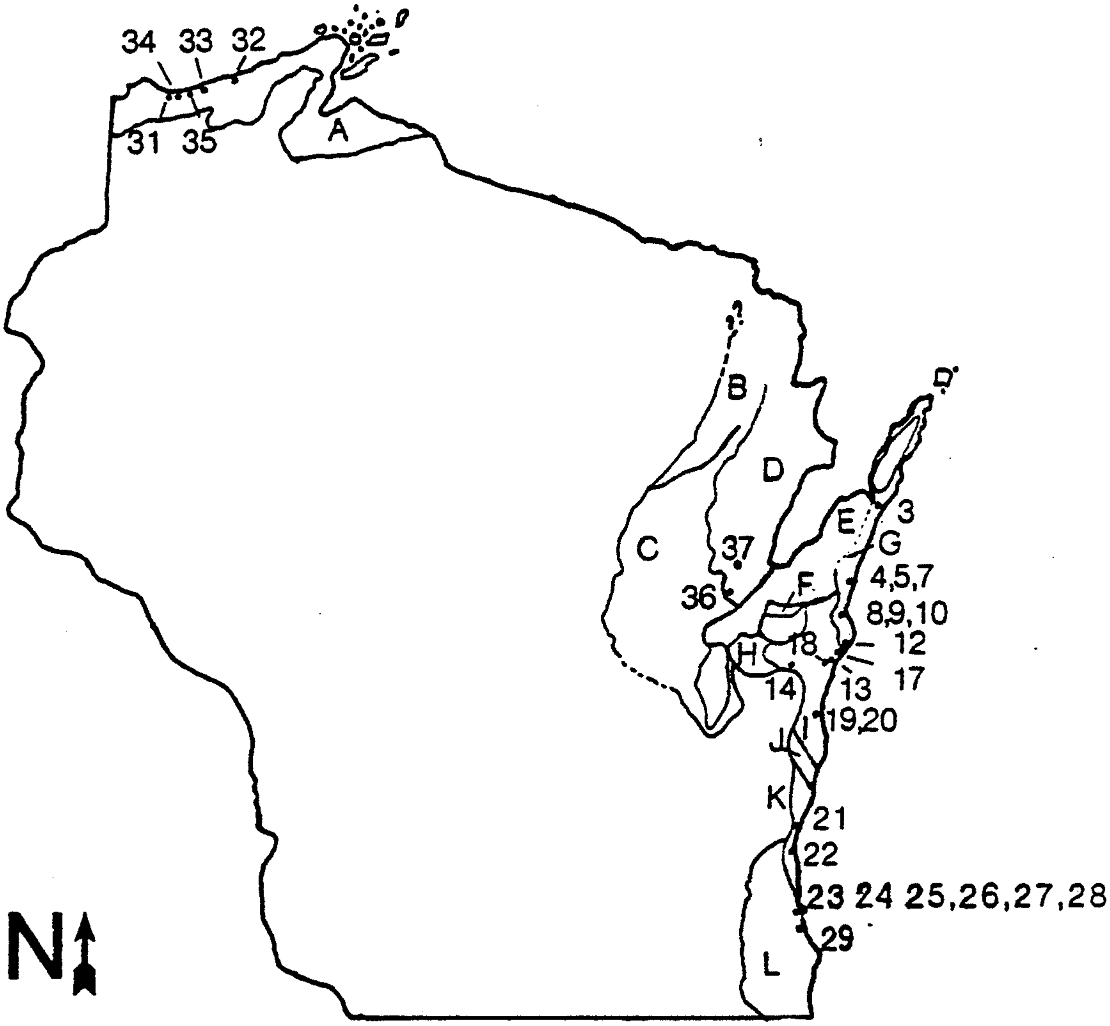
The importance of waste disposal and groundwater protection has prompted interest in the hydrogeologic behavior of unlithified low permeability materials such as fine-grained till. Discontinuities, particularly fractures, have been found to be an important factor in controlling the permeability of these materials, and can increase the complexity of groundwater flow systems (Gordon and Heubner, 1983; Grisak and Cherry, 1975; Grisak, Cherry, Vonhof, and Blumele, 1976; Hendry, 1982; Prudic, 1982; Sterrett and Edil, 1982).

Workers studying till stratigraphy and coastal erosion problems in Wisconsin have noted that many fine-grained till units are jointed in outcrop and in excavations (e.g. Sterrett and Edil, 1982). Their observations led to the present study, which was intended to be a reconnaisance of jointed till in Wisconsin. The goals of this study were to relate the common attributes of jointing in till to its physical properties from observations made at a large number of sites, to acquire data on the characteristics of jointing as a basis for further research, and to develop conceptual models for joint genesis from the results of the field investigation. This paper represents the first systematic examination of the characteristics and distribution of joints in till and the factors that may control joint genesis. Most previous studies on this topic have focused on the characterisitics of jointing at a single site.

METHODS

Study sites were chosen in representative locations for 10 fine-grained till units near and along Wisconsin's Great Lakes shoreline. Site locations and the distribution of clayey till units are presented in Figure 1 and in Table 1.

Sites were chosen in areas that had been studied by previous investigators, and were restricted to areas affording good exposures. The till units and lacustrine deposits included in this study have been previously investigated by Acomb (1978), Johnson (1980), McCartney and Mickelson (1982), Acomb, Mickelson and Evenson (1982), Need (1980), and Mickelson and others (1977). These stratigraphic units are summarized by Mickelson and others (1984). The physical characteristics of the units studied



A	Miller Creek Formation	G	Two Rivers Member
	Kewaunee Formation:	H	Chilton Member
B	Silver Cliff Member	I	Valders Member
C	Kirby Lake Member	J	Haven Member
D	Middle Inlet Member	K	Ozaukee Member
E	Glenmore Member	L	Oak Creek Formation
F	Branch River Member		(Group 2 Tills)

Figure 1: Site locations and distributions of fine-grained till units in Wisconsin. Numbers refer to sites listed in Table 1.

Table 1: Study site locations

Site #	Location	Unit (note 1)	Coordinates 1/4	1/4	1/4	Sec.	T	R	County	Depth (m)
1	Middleton Quarry	Dolo.	SW	NW	SW	10	7N	8E	Dane	5.0
2	Door Quarry	Dolo.	NE	NW	NW	1	27N	25E	Door	16.9
3	Door Sandpit	E	SW	SW	SE	17	27N	26E	Door	2.3
4	Sandy Bay	Lac.	SW	NW	NE	25	22N	24E	Kewaunee	10.0
5	Sandy Bay	J	NW	NW	SE	25	22N	24E	Kewaunee	2.5
6	Sandy Bay	Lac.	NW	NW	NE	25	22N	24E	Kewaunee	6.0
7	Sandy Bay	G	NW	NW	NE	25	22N	24E	Kewaunee	1.0
8	Highway BB	G	SE	SE	SE	35	22N	24E	Manitowoc	2.3
9	Highway BB	Lac.	SE	SE	SE	35	22N	24E	Manitowoc	4.5
10	Two Creeks	Lac.	NW	SE	NE	2	21N	24E	Manitowoc	3.1
11	Two Rivers Pit	Lac.	SE	SW	NW	31	20N	24E	Manitowoc	3.5
12	Two Rivers Pit	G	SE	SW	NW	31	20N	24E	Manitowoc	3.0
13	Manitowoc River	J	SW	SW	NE	24	19N	23E	Manitowoc	1.7
14	Valders Quarry	Dolo.	SW	SW	NE	32	18N	22E	Manitowoc	2.5
15	Valders Quarry	I	SW	SW	NE	32	18N	22E	Manitowoc	1.1
16	Manitowoc Suburb	I	SW	SW	NW	14	19N	23E	Manitowoc	7.0
17	Memorial Drive	J	NW	NW	NW	15	19N	24E	Manitowoc	4.5
18	Fricke Quarry	I		SW	NE	3	18N	23E	Manitowoc	1.2
19-1	Haven Type Sect.	I	NW	NW	NE	22	16N	23E	Sheboygan	1.9
19-2	Haven Type Sect.	I	NW	NW	NE	22	16N	23E	Sheboygan	0.8
20-1	Haven Type Sect.	I	NW	NW	NE	22	16N	23E	Sheboygan	1.3
20-2	Haven Type Sect.	J	NW	NW	NE	22	16N	23E	Sheboygan	3.0
20-3	Haven Type Sect.	I	NW	NW	NE	22	16N	23E	Sheboygan	1.3
21	Port Washington	K	NW	SE	NE	28	11N	22E	Ozaukee	3.4
22	Notre Dame	K	SW	SW	SE	8	9N	22E	Ozaukee	3.3
23	St. Francis	L(2A)	SW	SW	SW	24	6N	22E	Milwaukee	20.1
24	St. Francis	(note 2)	SW	SW	SW	24	6N	22E	Milwaukee	21.5
25	St. Francis	L(2A)	SW	SW	SW	24	6N	22E	Milwaukee	20.5
26	St. Francis	L(2B)	SW	SW	SW	24	6N	22E	Milwaukee	5.2
27	St. Francis	L(2C)	SW	SW	SW	24	6N	22E	Milwaukee	2.6
28	St. Francis	Lac.	NW	SW	SW	24	6N	22E	Milwaukee	2.5
29	Bender Park	L(2B)	SE	SE	SE	25	5N	22E	Milwaukee	3.1
31	Camp Amnicon	A	SE	SW	SW	27	49N	12W	Douglas	3.0
32	Corps Project	A	NE	NE	NW	4	49N	9W	Bayfield	1.0
33	Pearson Creek	A	NW	SW	NE	22	49N	11W	Douglas	1.5
34	Amnicon River	A	NE	NE	NE	34	49N	12W	Douglas	5.4
35	OSP-2	A	NW	SE	SW	18	49N	10W	Douglas	2.5
36-1	Outagamie Landfill	Lac.		SW	NE	17	21N	17E	Outagamie	1.4
36-2	Outagamie Landfill	D		SW	NE	17	21N	17E	Outagamie	2.6
37	Koenen Farm	D	NW	NW	NW	6	23N	19E	Outagamie	2.5

Note 1: Letters refer to lithostratigraphic units in Figure 1
other units are lacustrine sediment or dolomite.

Note 2: New Berlin Formation

are presented in Table 2.

Study sites contained exposures of one or more till units and associated fine-grained glaciolacustrine sediment, and were located both along actively eroding bluffs of the Lake Michigan and Lake Superior shorelines and in gravel pits, quarries, and other excavations. Joints were usually examined below the lower limit of translocated clay, and typically near or below the water table. At each site, measurements were made of joint orientation, length, and spacing, and samples collected for laboratory analysis. Oriented samples were obtained at most sites for preparation of thin sections to be used for microfabric analysis.

Joint characteristics were examined using a modification of the Cavity Technique (Fookes and Denness, 1969). An exposure of one to two square meters was stripped of slumped material and the orientation of all joints intersecting this surface measured. Several techniques were used to insure that the measurements were not biased by the exposure orientation. When possible, measurements from two near-orthogonal exposures were combined to give an aggregate joint orientation distribution. An effort was also made to excavate into the wall of the exposure in order to intercept joints oriented parallel to the exposure, but this was often

Table 2: Characteristics of jointed till units in Wisconsin

Unit	% Sand <2.0 mm	% Silt <.0625mm	% Clay <.002mm	% Expand ************	% Illite <.002mm	% Kaol/ Chlorite **********	% CO3 in coarse silt
Kewaunee Formation:							
Glenmore till	37.3(2.4) 9	33.8(1.7) 9	28.9(3.7) 9	25.6(11.8) 4	56.4(6.1) 4	18.0(6.4) 4	56.4(2.6) 7
Two Rivers till	26.7(7.5) 13	47.1(6.7) 13	26.2(7.3) 13	37.1(15.2) 7	55.4(16.0) 7	11.6(6.1) 7	43.8(10.3) 10
Middle Inlet till	23.0(6.6) 4	51.5(12.6) 4	25.5(6) 4	38.0(13.4) 3	45.7(8.6) 3	16.2(5.4) 3	26.8(6.6) 3
Valders till	21.7(9.2) 22	52.2(4.9) 22	26.1(10.7) 22	57.8(15.6) 11	34.5(12.5) 11	7.7(3.5) 11	42.9(3.5) 21
Haven till	14.8(10.4) 7	47.4(3.6) 7	37.8(10.4) 7	32.1(3.9) 3	54.5(3.1) 3	13.4(1.2) 3	37.8(3.0) 6
Ozaukee till	30.0(6.8) 6	46.7(5.2) 6	23.3(3.2) 6	48.9(-) 1	39.6(-) 1	11.6(-) 1	47.1(2.2) 6

Note: Values are listed as Mean (Standard Deviation)
\# of samples

Table 2: Characteristics of jointed till units in Wisconsin (continued)

Unit	% Sand <2.0 mm	% Silt <.0625mm	% Clay <.002mm	% Expand ************	% Illite <.002mm	% Kaol/ Chlorite **********	% CO3 in coarse silt
Oak Creek Formation:							
till 2C	17.1(2.6) 3	56.5(1.5) 3	24.6(4.4) 3	48.6(-) 1	11.8(-) 1	74.0(-) 1	14.2(-) 1
till 2B	10.0(4.6) 6	59.0(5.7) 6	31.3(5.6) 6	11.8(-) 2	74.0(-) 2	14.2(-) 2	48.6(-) 6
till 2A	23.6(12.6) 4	44.0(3.1) 4	32.4(10.6) 4	17.3(-) 2	61.4(-) 2	21.4(-) 2	48.6(1.2) 4
New Berlin till	14.3(-) 2	45.5(-) 2	40.2(-) 2	4.1(-) 1	76.1(-) 1	19.7(-) 1	41.8(-) 2
Miller Cr Formation:							
Douglas till	8.9(4.4) 14	21.2(10.2) 14	69.9(11.8) 14	75.0(-) 2	17.0(-) 2	8.0(-) 2	6.4(4.3) 3
Hanson Creek till	6.9(2.5) 5	22.9(6.0) 5	70.2(8.3) 5	52.2(13.9) 3	27.0(4.0) 3	20.8(12.8) 3	3.6(1.1) 5

hampered by the extreme toughness of the material.

Joint orientation was determined with a Brunton compass, either held directly on the joint surface or on a small aluminum shim inserted into the joint. Strike and dip and the direction of dip were recorded in the field using the azimuth (0-360 degrees) format. Joint data in its raw form is included in Appendix 3, and is also presented graphically in Appendices 4 and 5. Measurement error associated with this technique of joint measurement is approximately 2 to 4 degrees for dip and 1 to 5 degrees for strike, increasing at low dip values (Chapter 2).

Joint spacing was determined by measuring the perpendicular distance between joints of a given set (Wheeler and Holland, 1981). Measurements were made only between adjacent joints, or overlapping joint ends, in an effort to limit any bias towards larger spacings (Wheeler and Dixon, 1980).

Joint frequency, or the number of joints in a given length of scan line, was determined by counting the number of joints intersecting a 50-cm long scan line within the exposure. Four to eight scan lines in both the horizontal and vertical directions were used. The joint frequency may have been poorly characterized because the scan line length chosen may not have been of sufficient length to consistantly include joints of wide spacing. Priest and

Hudson (1976) state that a scan line length of fifty times the inverse of the mean frequency is required for accurate characterization of joint frequency: this would have entailed using scan lines of 5 to 10 meters in length, which was often larger than the exposure available.

It proved impractical to measure the length of every joint because of the extensive excavation required to trace joints their full length. For this reason, maximum joint length at a given exposure was estimated by measuring five to ten of the longest apparent lengths.

At each site sediment samples were taken at constant intervals vertically down the face of the exposure. Between 300 and 500 g of disturbed material per sample was collected for analysis of grain-size distribution and determination of clay mineralogy and carbonate content. Results of the lab analysis are contained in Appendix 1.

LABORATORY METHODS

Grain-size distribution was determined from the disturbed samples using sieve and hydrometer analysis. Techniques were those of the Quaternary Research Laboratory, University of Wisconsin -- Madison (Need, 1980). The coarse silt fraction (0.0625 mm to 0.0380 mm) was then digested in a Chittick apparatus for determination of calcite and dolomite content, as described in Need (1980). A semi-quantitative determination of the clay (less than .002 mm) mineralogy was made using x-ray diffraction of glycolated samples prepared by pipetting dispersed clay suspensions onto glass slides and then rapidly evaporating the dispersant by heating the slides in a small oven at 70 degrees C. Peak area atios were determined using the methods of Need (1980), with a baseline tied to non-refractive portions of the curve (Figure 2).

Microfabric analysis

Determination of ice-flow direction in the till units studied by traditional methods, such as macrofabric or striations, was difficult owing to the paucity of pebbles and the lack of exposed bedrock. For this reason, the orientation of sand grains and silt and clay bands in thin section (microfabric), was used to estimate ice flow direction. Microfabric was measured from thin sections cut from oriented blocks of till. Undisturbed samples were obtained in the field by careful excavation of a block so that it protruded from the the exposure wall. Orientation of the block was marked while still in place by inscribing its surfaces with arrows denoting north and top. The protruding block was then gently removed, wrapped in cellophane and aluminum foil, and returned to the lab.

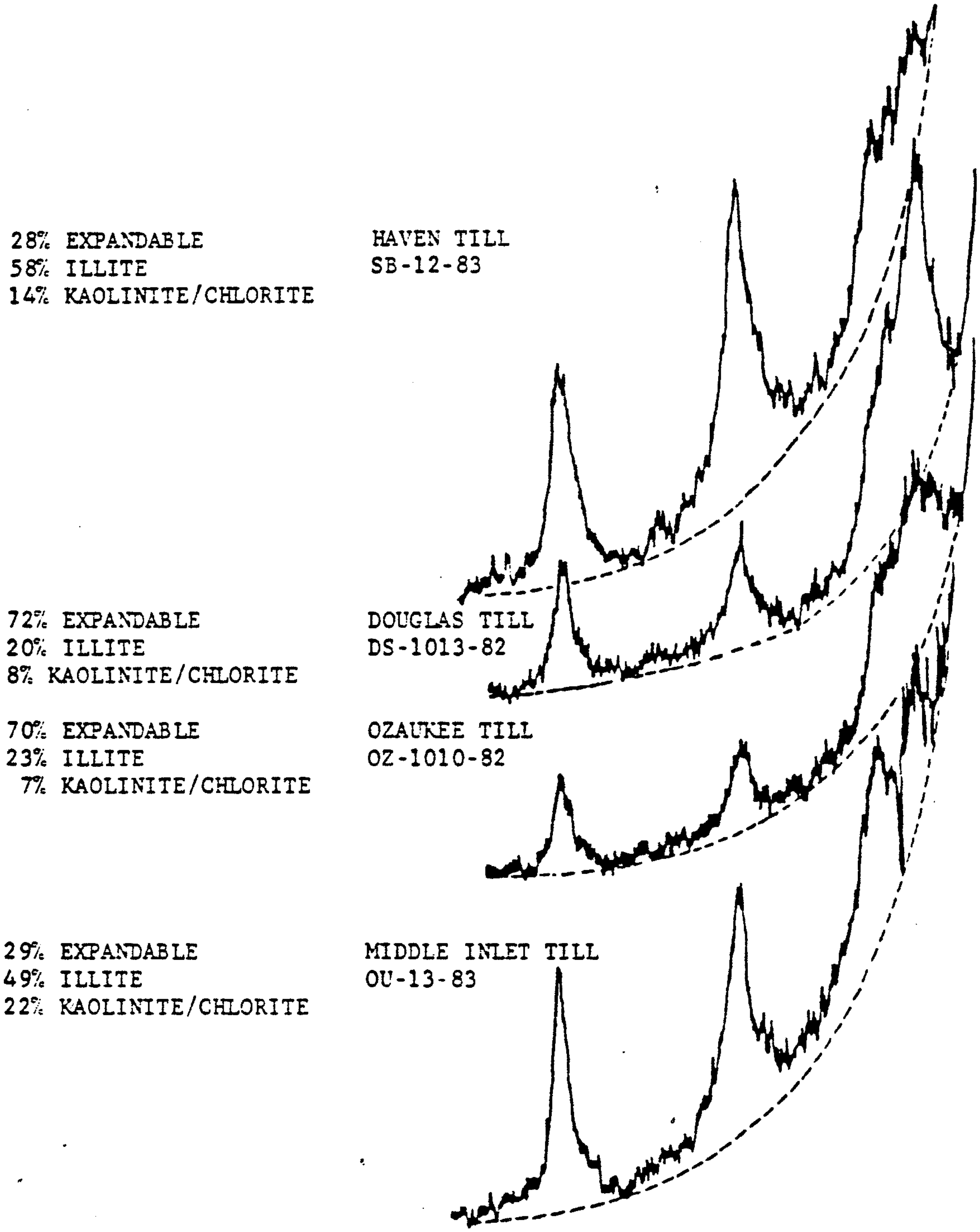


Figure 2: X-Ray diffractograms for semi-quantitative clay mineral determination illustrating baselines used for calculation of peak areas.

After the blocks had dried at 95 degrees C a belt grinder was used to trim the block to thin-section size while keeping disturbance of the sample to a minimum and maintaining a preferred orientation. The trimmed samples were then impregnated with low viscosity epoxy resin, mounted, and ground.

The resulting thin sections (Figure 3) were projected onto a large piece of paper and the long axis of elongate (length to width ratio of greater than 2 to 1) sand grains larger than 0.08 mm2 marked on the paper using the least projection elongation technique (Dapples and Rominger, 1945). The orientaton of the sand grains and other features were then measured with a protractor and plotted on a rose diagram, as shown in Figure 4.

Sand grain orientation was tested for statistically significant departure from a uniform orientation distribution. Significant peaks in the rose diagram were identified using a Poisson test for non-uniformity of directional data (Abdel-Rahman and Hays, 1981) at a 95% confidence level. The mean orientation of the significant peaks (or of the entire rose diagram if no significant peaks existed) was then calculated via a computer program based on the resultant vector of the direction cosines, as outlined in Cheeney (1983).

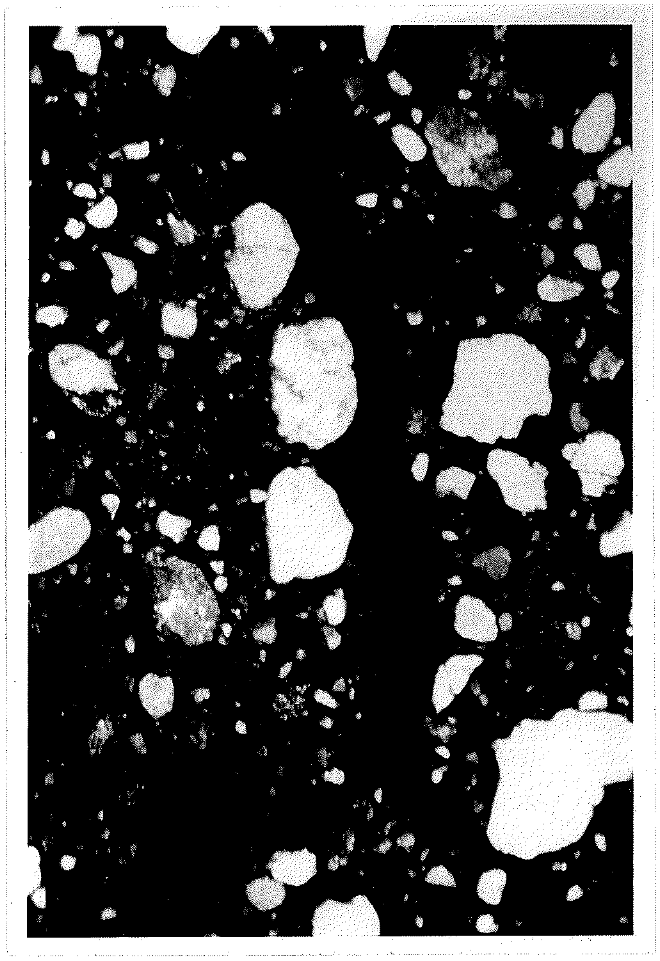
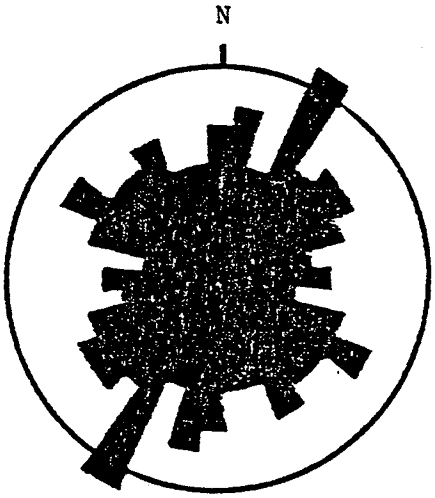


Figure 3: Photomicrograph of thin section cut from sample OS-1-83. Crossed Nichols, 20X magnification. North is up; horizontal section. Note joint passing through center of section. Joint is coated with opaque iron hydroxide.



MICROFABRIC ROSE DIAGRAM

GLENMORE TILL

DOOR PIT (SITE 3)

Figure 4: Microfabric of oriented sample of Glenmore Till shown in Figure 3. Rose diagram with 10 degree intervals. Sample OS-1-83.

The mean of the significant peak in a rose diagram of the orientation of elongate sand grains can be used as evidence of ice-flow direction, but it is not a unique indicator, as pointed out by Johnson (1983). Determination of ice-flow direction is complicated by the ambiguity that can be associated with the interpretation of fabric plots. Fabric plots can be difficult to interpret because of a lack of significant maxima or because the fabric may have maxima aligned both parallel and orthogonal to the direction of ice flow (Andrews and Smith, 1970; Johnson, 1983). Fabric may also be a non-unique estimator of ice flow direction due to large vertical variations (Andrews and Smith, 1970) or other natural scatter, as was observed by Harrison (1957); these may be the result of varying flow paths through time (Johnson, 1983).

Other means of ice flow determination, such as pebble fabric, were used in conjunction with the microfabric data when available. These usually tended to agree with the orientation calculated from the microfabrics, but were not always consistant. Along the Lake Michigan shoreline, microfabric data indicated a change of ice-flow direction as the ice came radially out of the basin over a given site, and later flowed dominantly down-basin as the ice thickened. Similar fluctuations in ice flow occured at

Glacier Bay, Alaska (Mickelson, 1971) and have been suggested for the Superior Lobe in northwest Wisconsin (Johnson, 1980).

Evidence of ice-flow direction was available from a limited number of pebble fabrics and striated boulders, and from striated bedrock and landform orientation. Sole marks at the contact between till and underlying sediment (Ehlers and Jurgen, 1979) were rarely available as ice flow indicators but were noted where appropriate. Folding or shearing of underlying material was also noted at some localities, but may not represent a unique direction of ice flow.

The stratigraphy and soil development at each site was recorded as is shown in Figure 5. Profile descriptions for all the sites are presented in Appendix 2. As part of the profile description, the thickness and composition of the underlying and overlying material was noted, as was the thickness of the till unit and the orientation of the exposure.

DATA ANALYSIS

Arithmetic means were calculated for the maximum joint length, joint frequency, the content of each size fraction, semi-quantitative abundance of clay minerals, and carbonate content. The trimean (Figure 2a in

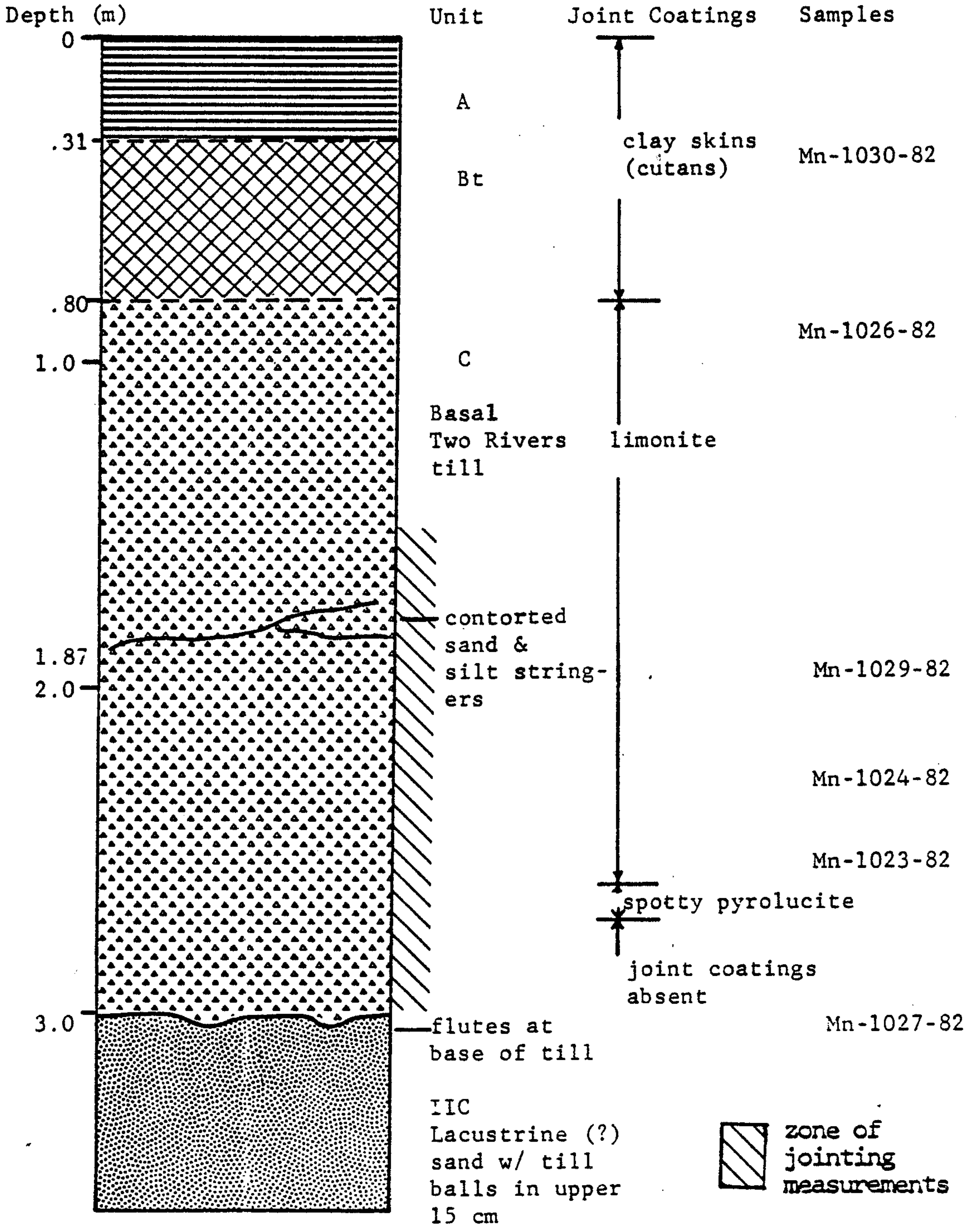


Figure 5: Measured Section at Two Rivers Pit (Site #12).

Chapter 2) was used to calculate a combined mean joint spacing for all joint sets at each site because it is less biased by outliers (Wheeler and Dixon, 1980).

Joint orientation data was initially plotted using two computer plotting programs. The first, POINT (Warner, 1969), was used to generate a PI plot and contour the poles to joints by percent per 1% of area (Appendix 5). The second, PATCH (Mahtab and others, 1972) was used to cluster poles to joints into statistically significant sets and then calculate mean joint set orientation, along with an estimate of the error associated with this determination. Examples of the output from these programs is shown in Figure 6. An evaluation of PATCH is presented in Chapter 2.

Comparison of the joint data contoured by POINT and the clusters of joint data produced by PATCH showed that neither method could be used alone to calculate the orientation of joint sets with the amount of data available. PATCH was unable to resolve joint sets that had near-parallel orientations, and generated spurious clusters of small numbers of points. Contoured POINT plots were used to evaluate clusters with small numbers of points; clusters that did not contain at least 5% of the total number of joints within a 1% area were considered spurious and removed from subsequent analysis. The

```
PORT WASHINGTON -- OZAUKEE TILL   DATA SET 22                           112  OBSERVATIONS
STEREOGRAPHIC PROJECTION OF AN AREA-TRUE COUNT ON THE HEMISPHERE
```

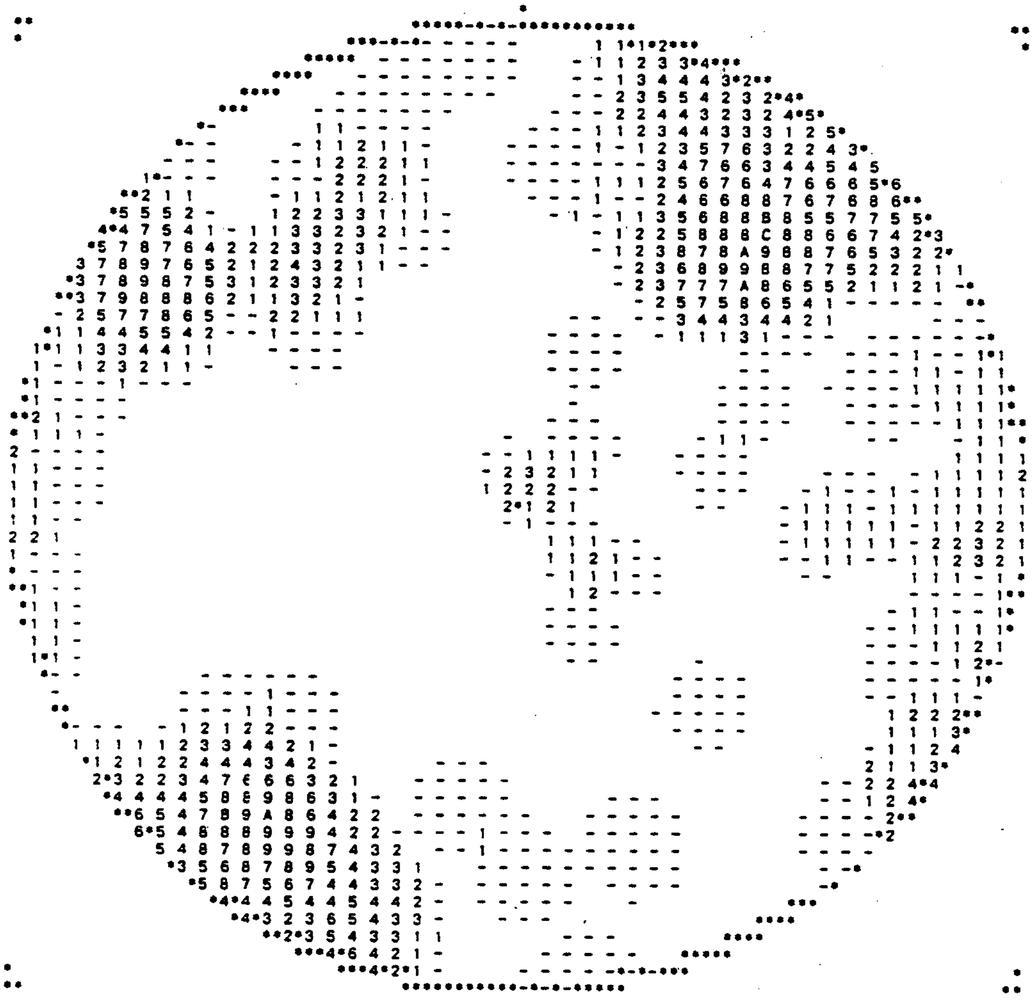


Figure 6a: Sample output from the program POINT (Warner, 1969). Equal area lower-hemisphere projection of poles to joints measured in Ozaukee till at Port Washington (Site #21). Numbers are percent of data per 1% of the area of the total plot.

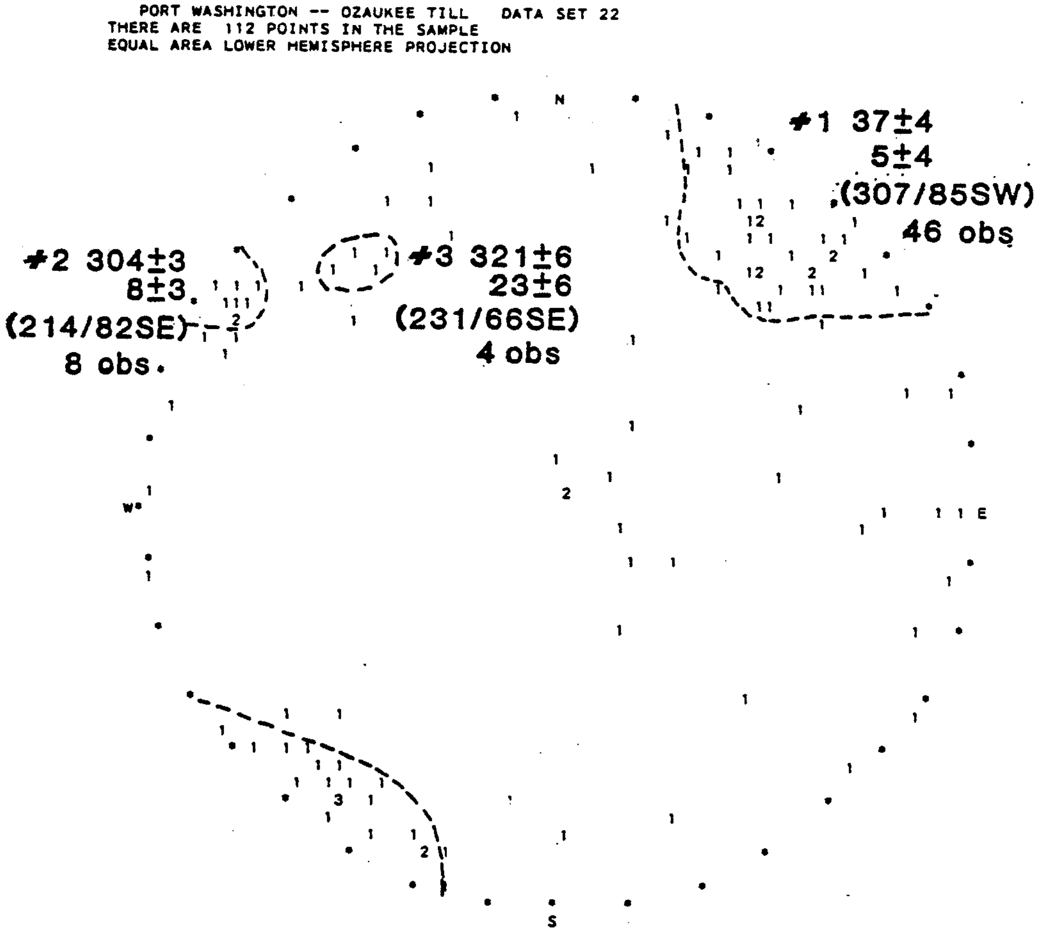


Figure 6b: Sample output from the program PATCH (Mahtab and others, 1972). Equal area lower-hemisphere projection of poles to joints measured in Ozaukee till at Port Washington (Site #21).

Small numerals within plot are number of points at each print position. Dashed line delineates clusters of points as grouped by PATCH. For clusters #1, #2, and #3, the trend and plunge of the poles to joint set are shown, along with the associated error. Numbers in parenthesis are the strike and dip of the joint sets represented by the clusters of poles.

clustering of more than one joint set into a single cluster was more problematic. One solution that met with limited success involved raising the Poisson confidence level used in PATCH's statistical routines from 99% to 99.5%, which decreased the number of significant points and thus kept PATCH from lumping adjacent areas into one cluster. For some joint sets this approach was not sufficient, and the contoured POINT data was used to estimate the orientation of individual joint sets, as explained in Chapter 2. Joint set orientation is summarized in Appendix 6.

Inferences about the direction of ice flow were based on the results of the microfabric plots and on other evidence of glacier movement. Pebble fabrics were analyzed with the aid of FABRC, a program developed by C. Corbato of Ohio State University, which utilizes the method of Kamb (1959).

RESULTS

JOINT CHARACTERISTICS

Orientation

Joint orientation was found to have two main modes. The most prominant joints at most exposures are arrayed in near-vertical sets, with dip angle greater than 70 degrees. Some exposures also contain undulating or planar near-horizontal joints, generally dipping less than 20 degrees.

As can be seen from histograms of joint orientation (Figure 7), the joints have a preferred orientation. Comparison of joint orientation to the direction of ice flow and to the exposure orientation (Figures 8 and 9) shows that both factors may have some control on joint orientation.

Relationship to Ice Flow

Jointing seems to have a consistant relationship with the inferred direction of ice flow, and in many sites the near-vertical joints were arrayed in sets that are conjugate around, perpendicular to, or parallel to the direction of ice flow as indicated by the microfabric data (Figures 10a, 10b, 10c, and 10d). Similar joint patterns have been reported by McGown and his coworkers (1974), for joints in fine-grained till in a drumlin near Glasgow, Scotland, and by Boulton (1970) from till beneath a modern glacier (Nordenskioldbreen) on Spitsbergen. Figure 11 presents an example of the joint sets present at Notre

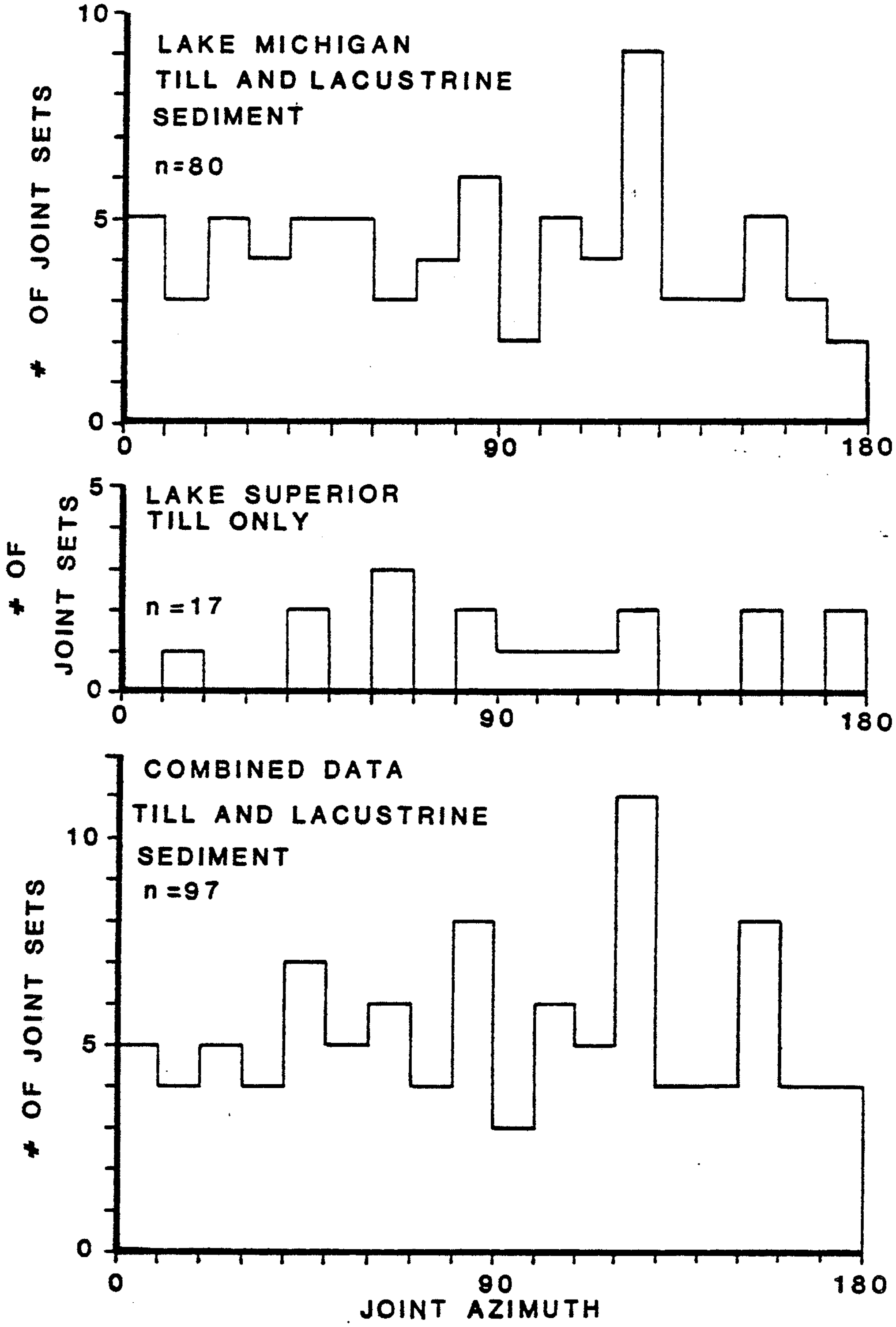


Figure 7: Histogram of joint-set orientation for near-vertical joints.

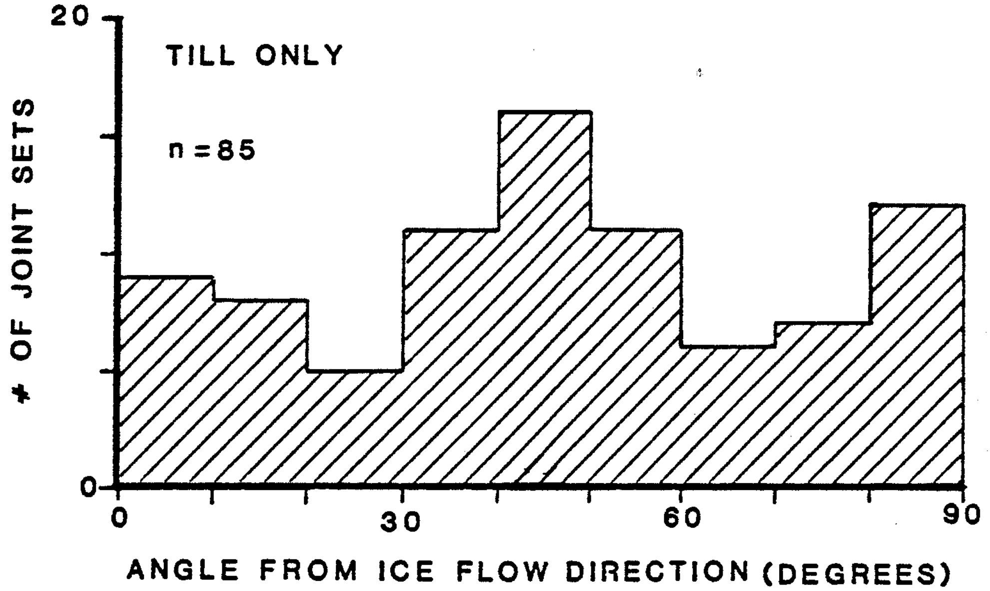


Figure 8: Histogram of Joint Sets versus Angle from Ice Flow Direction.

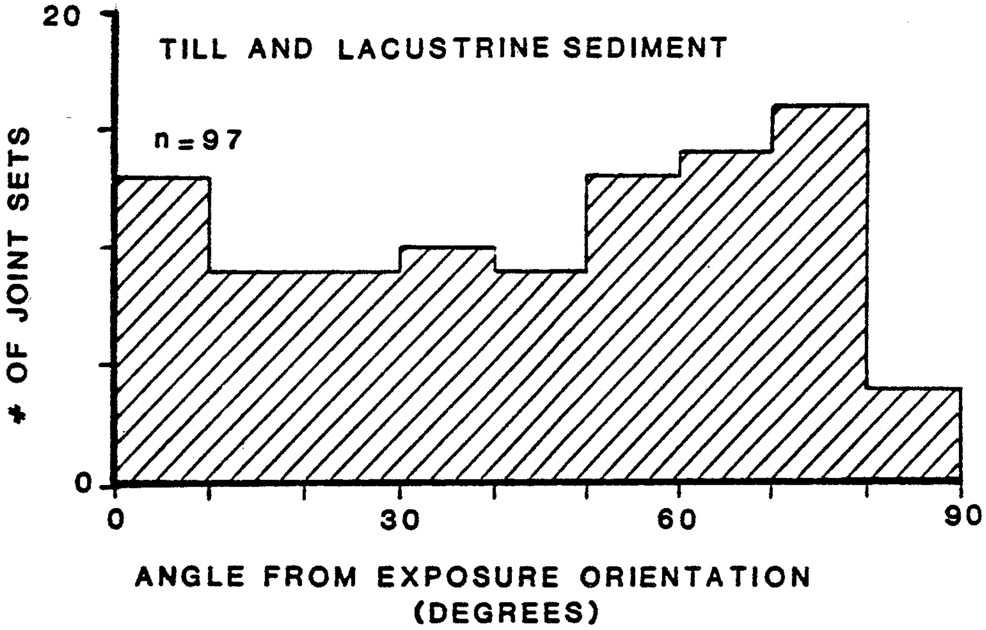


Figure 9: Histogram of joint sets versus angle from exposure orientation.

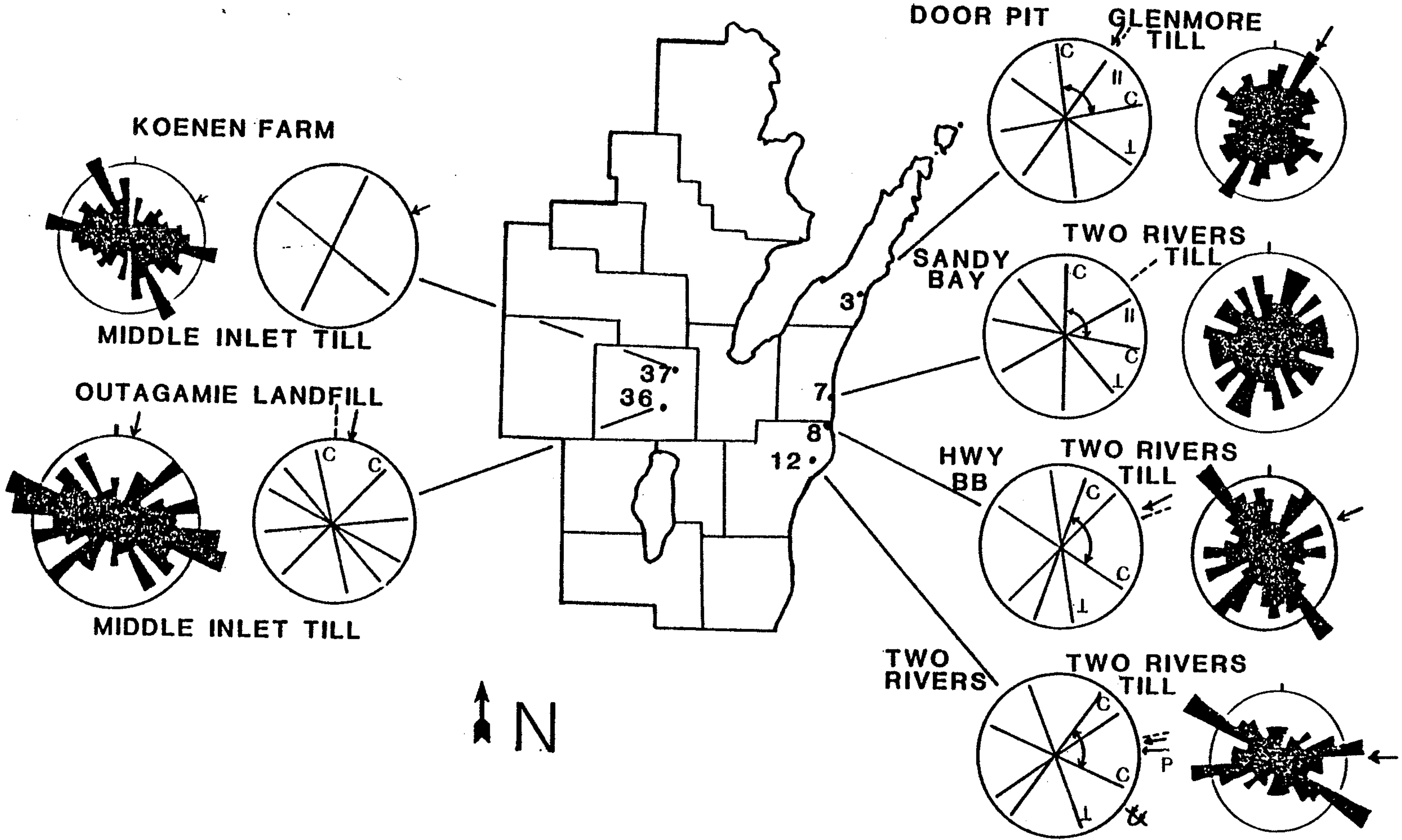


Figure 10a: Comparison of joint orientation with sand grain orientation (microfabric) for post Two Creeks till units. Arrow denotes microfabric trend, dashed line denotes conjugate bisector of conjugate joint sets, P denotes pebble fabric trend. Within joint plot, C denotes conjugate joints, 11 parallel joints, and ⊥ perpindicular joints when compared to ice flow.

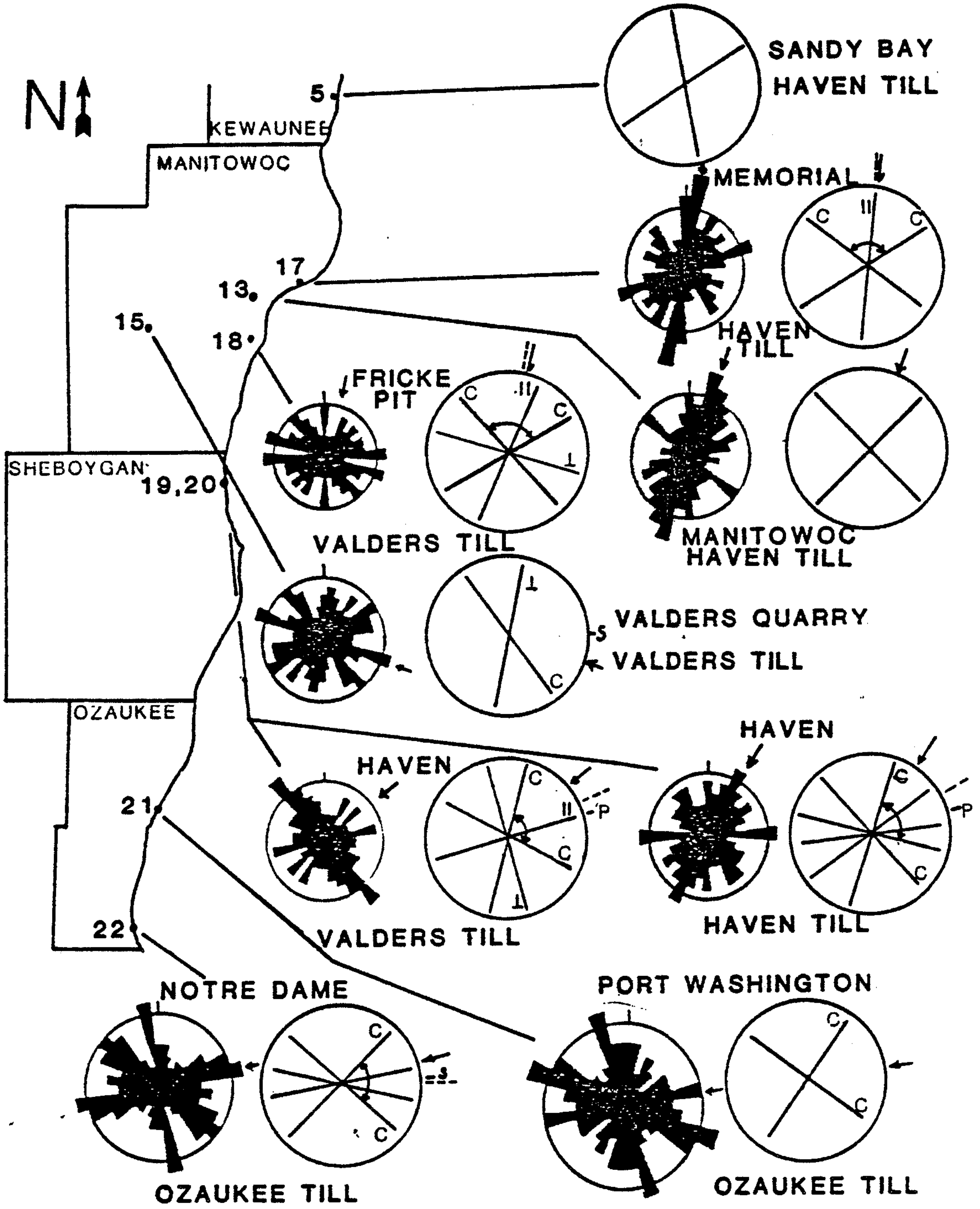


Figure 10b: Comparison of joint orientation and microfabric for Haven, Valders, and Ozaukee tills. S denotes striations.

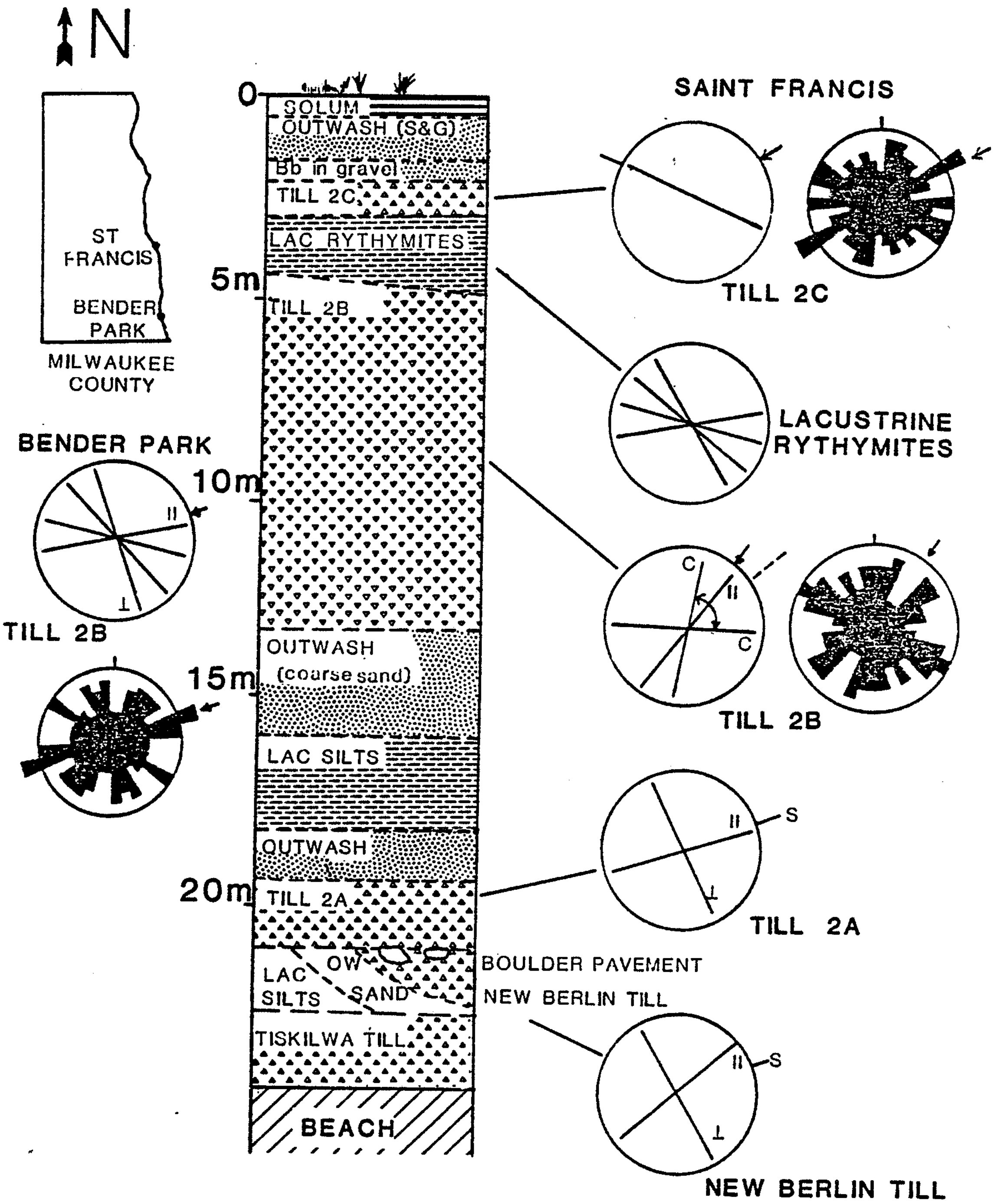


Figure 10c: Comparison of joint patterns and microfabric for Oak Creek and New Berlin tills.

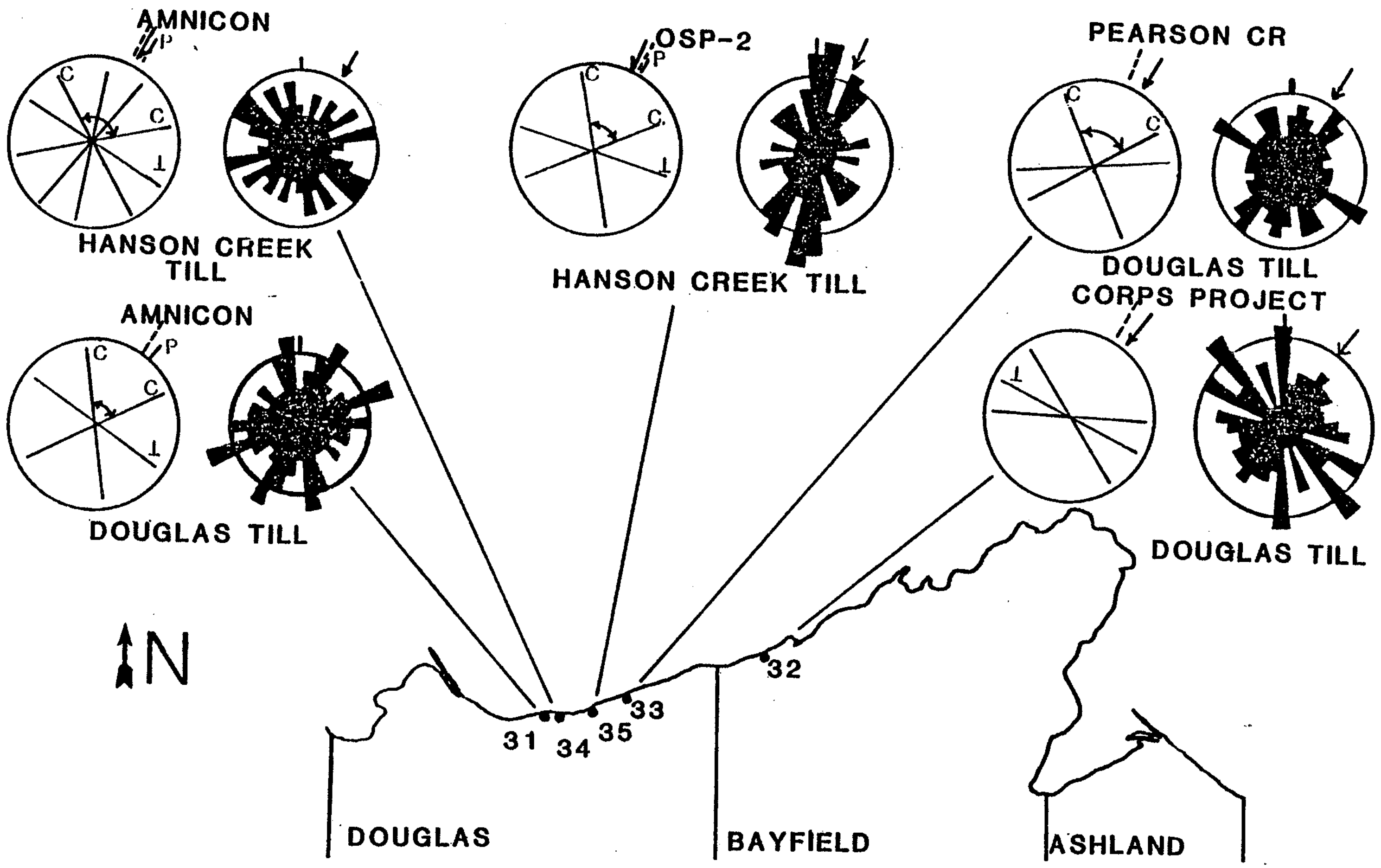


Figure 10d: Comparison of joint pattern and microfabric for Douglas and Hanson Creek tills.

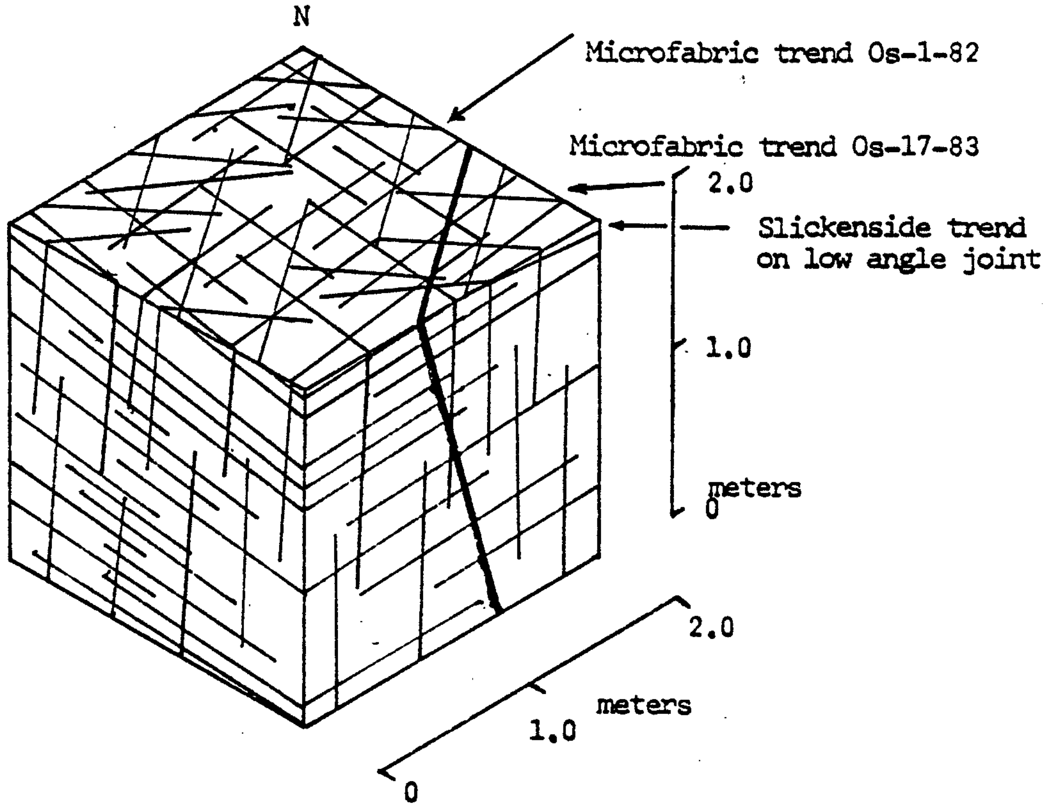


Figure 11: Schematic block diagram showing average spacing and orientation of joints in Ozaukee till at Notre Dame (Site #22). Heavy line is exposure orientation.

Dame (Site #22) along with the inferred direction of ice flow.

High-angle joints in many of the till units are concentrated at angles less than 10 degrees from ice flow (near parallel joints), at angles of 40 to 50 degrees (conjugate joints) and at angles of 80 to 90 degrees (near perpendicular joints). This relationship is evident in Figures 12, 13, and 14, which show the joint azimuth and the associated measurement error for joints in the different units studied.

Relationship to Exposure Orientation

A significant number of joints parallel the exposure or are concentrated at an angle of 70 to 80 degrees from it (Figure 9). Similar slope-parallel joints were found in till from the Norfolk coast of England (Kazi and Knill, 1973). Thus it seems that both subglacial deformation and lateral unloading may have influenced joint orientation. The large number of joints perpendicular to the exposure may be an artifact of sampling bias, as discussed by Terzaghi (1965). This sampling bias can occur when jointing striking nearly parallel to the exposure face is undersampled because few of the joints intersect the sampling face.

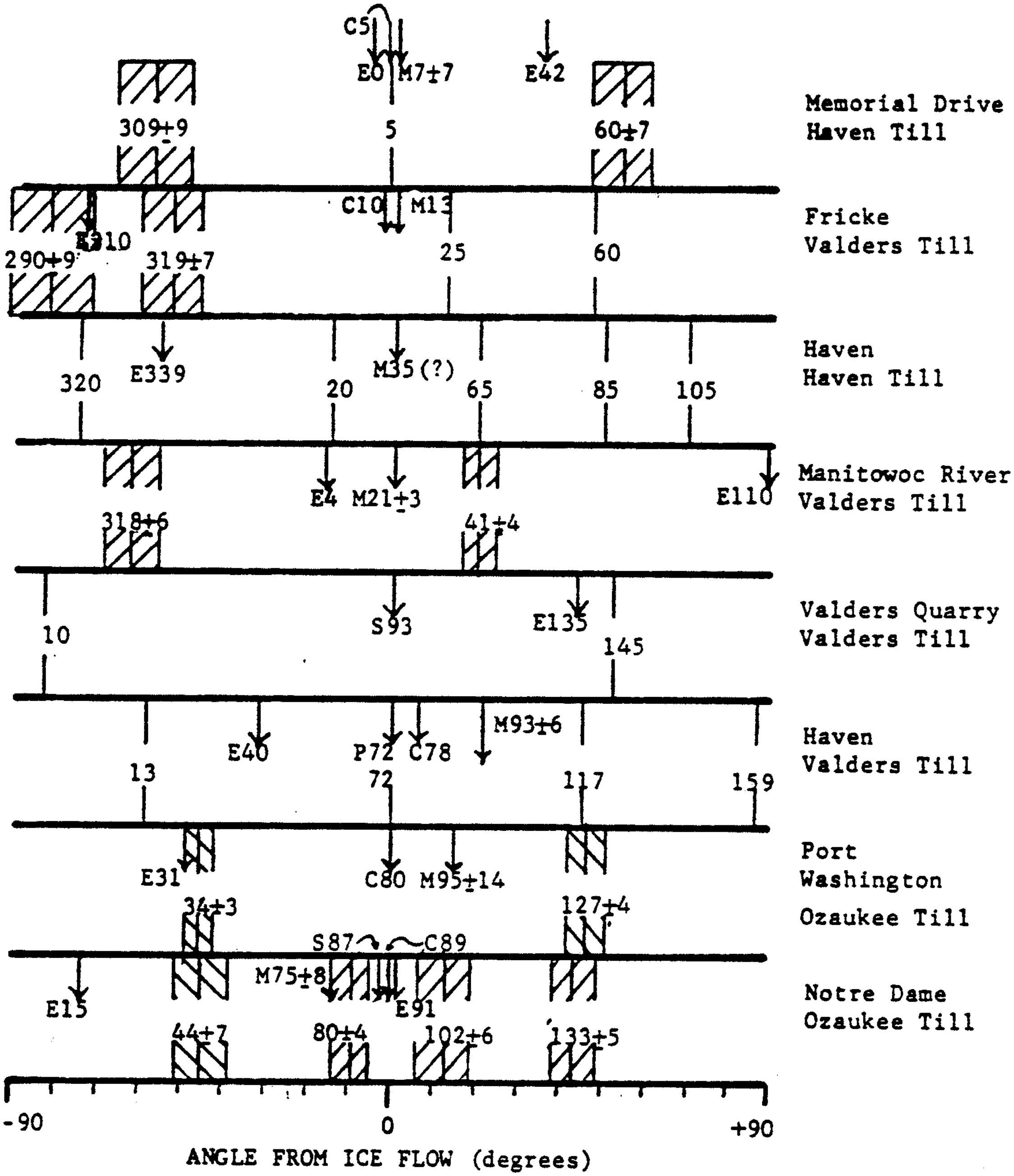


Figure 12: Summary plot of high-angle joint orientation, centered around the the inferred direction of ice flow. Hatched boxes represent the orientation and margin of error of joint sets as output by PATCH. Vertical lines represent the orientation of joints as estimated from POINT. Also shown are (C) bisector of conjugate joint sets, (E) exposure orientation, (M) microfabric trend, (P) pebble fabric trend, (S) striations, and (LA) low-angle joints (dip less than 20 degrees).

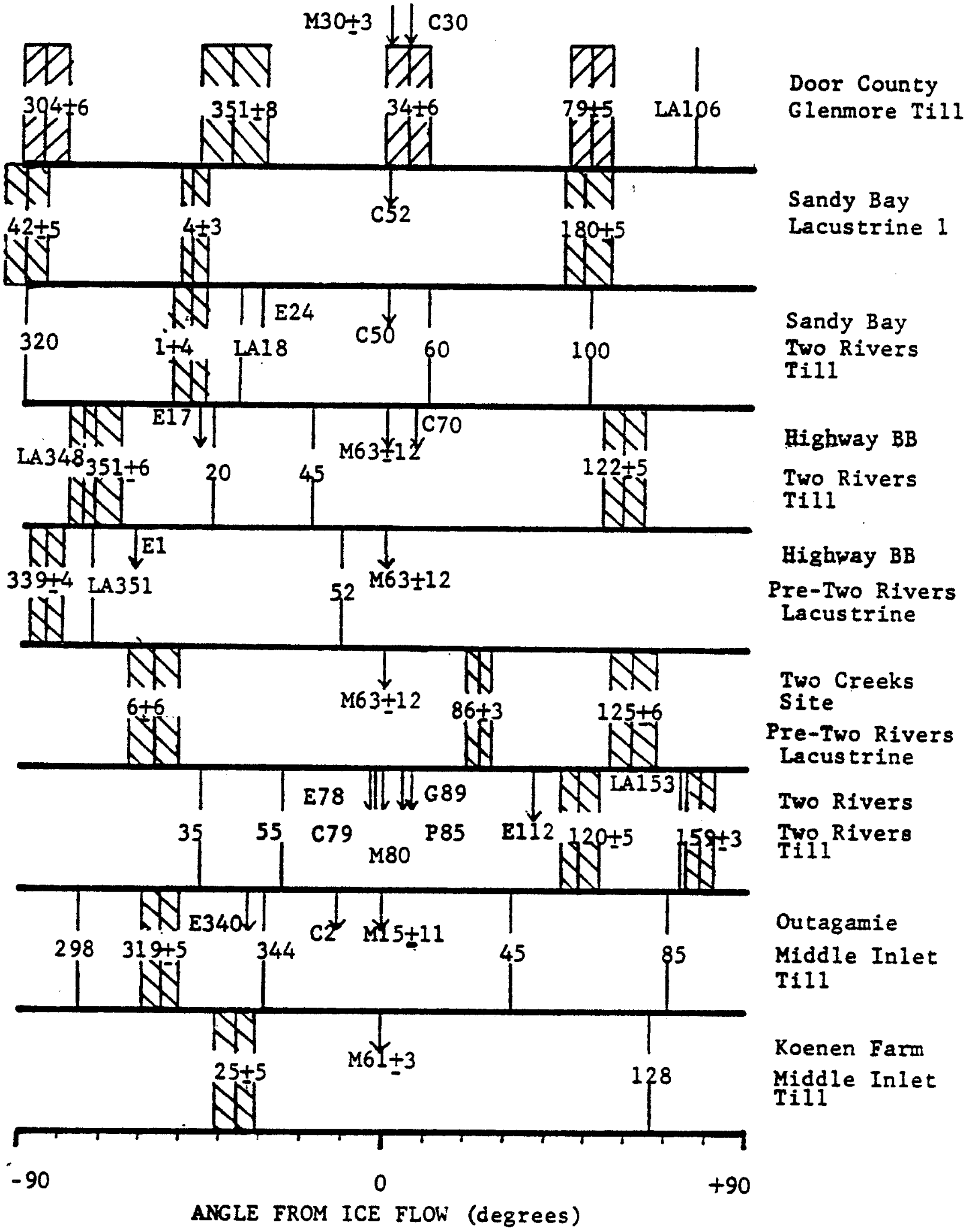


Figure 13: Summary plot of high-angle joint orientation, centered around the inferred direction of ice flow. Hatched boxes and vertical lines represent joint orientation. Other symbols are explained in Figure 12.

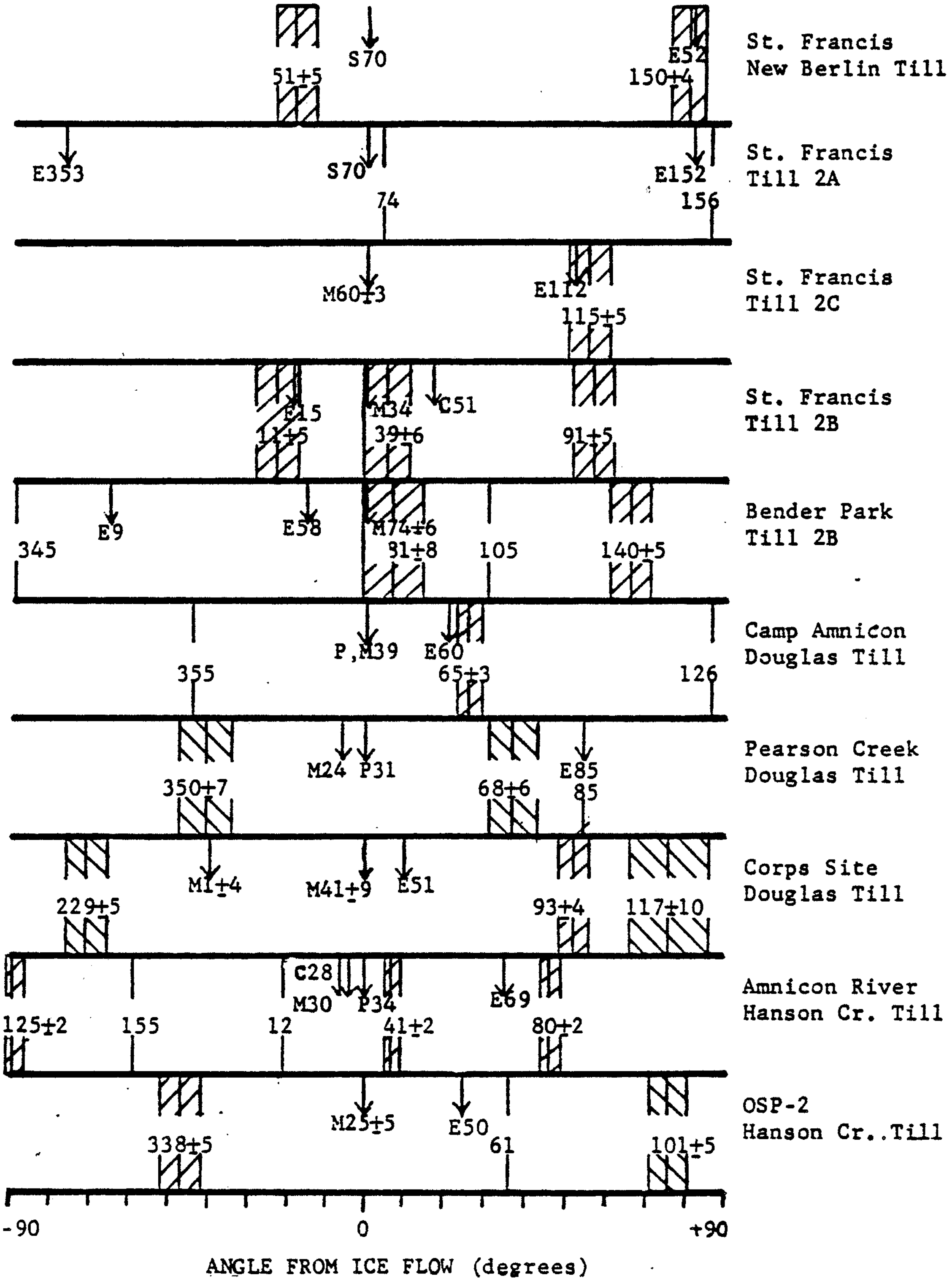


Figure 14: Summary plot of high-angle joint orientation, centered around the inferred direction of ice flow. Legend same as for Figure 12.

Spacing

Mean spacing of joints in a single set in till ranges from 1.6 to 29.0 cm. Near-vertical joints tend to have a wider spacing than the low angle joints when both are present. Wide variation in mean spacing was found from unit to unit, because spacing is affected by changes in lithology and till thickness.

Surface Characteristics

The near-vertical joints are typically planar, and showed no evidence of shearing or deformation. Surface roughness is seldom greater than 2 cm, and usually much less, but protruding pebbles sometimes span the joint plane. Structures commonly thought to indicate shear along joints, such as slickensides, plumose markings or feather structures (Figure 15) (Hodgson, 1961; Roberts, 1961), are absent from the high angle joints; slickensides were found only rarely on the low angle joints. The lack of failure structures on the high angle joints may be the result of limited displacment along the joint surface, repeated opening and closing of the joint during cycles of wetting and drying, or because of precipitation of weathering products on the joint surface.

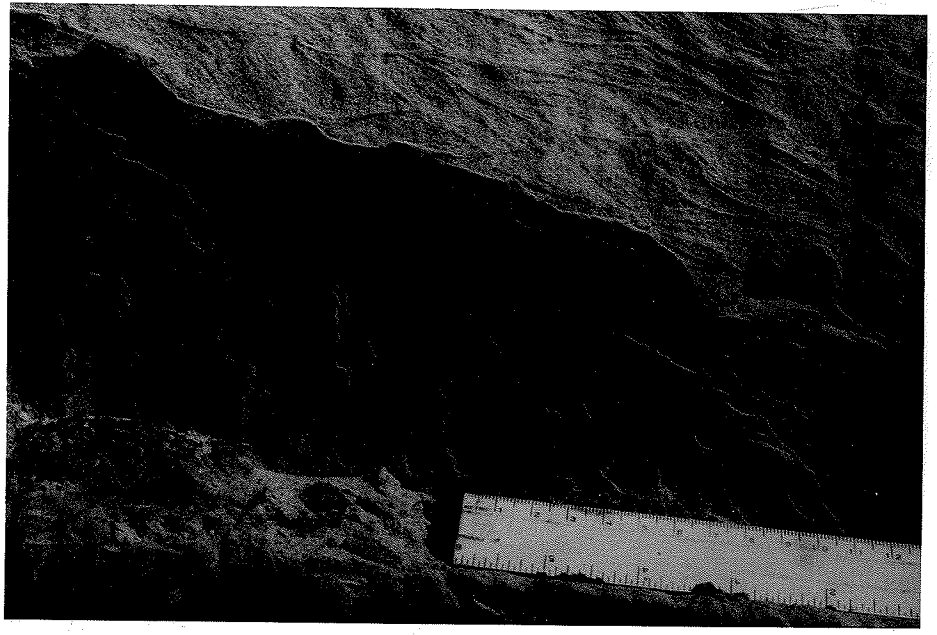


Figure 15: Plumerose structure from folded silt unit in deformed lacustrine sediment underlying Two Rivers till at the Two Rivers pit (Site #12). Joint is perpendicular to the fold axis.

The joint surfaces are often coated with weathering products, and this feature was used to distinquish joints from fresh discontinuites that may have formed during excavation of the exposure. Translocated clay (cutans) and secondary carbonates can be found at some sites to partially coat joint surfaces in exposures close to the surface (in the B horizon), while coatings or mottles of ferric hydroxides and dendritic or mottled manganese hydroxides are typically present at greater depth in a sequence similar to that discussed by Chandler (1973) (Figure 16). Reduction (gleying) was commonly observed along joints at some sites (Figure 17); this was interpreted as indicating that the exposure was below the water table for most of the year.

Joint Frequency

Joint frequency ranges from 5.0 to 28 per meter for near vertical joints, and from 3.0 to 42.4 per meter for low angle joints. Frequency is related to mean spacing (Figure 18), but is also a function of the number of joint sets. The influence of lithology on joint frequency is discussed below.

Figure 16: Iron and manganese hydroxides on joint face in Glenmore till exposed at Door pit (Site #3).

Figure 17: Gleyed joints conjugate around striated clast. Douglas till at Corps site (Site #32). Photograph by Ken Bradbury.

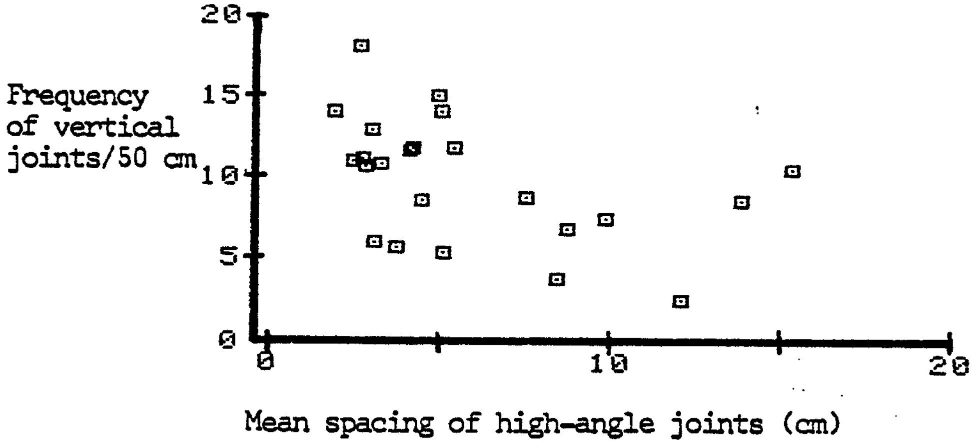


Figure 18: Frequency of vertical joints versus spacing of vertical joints

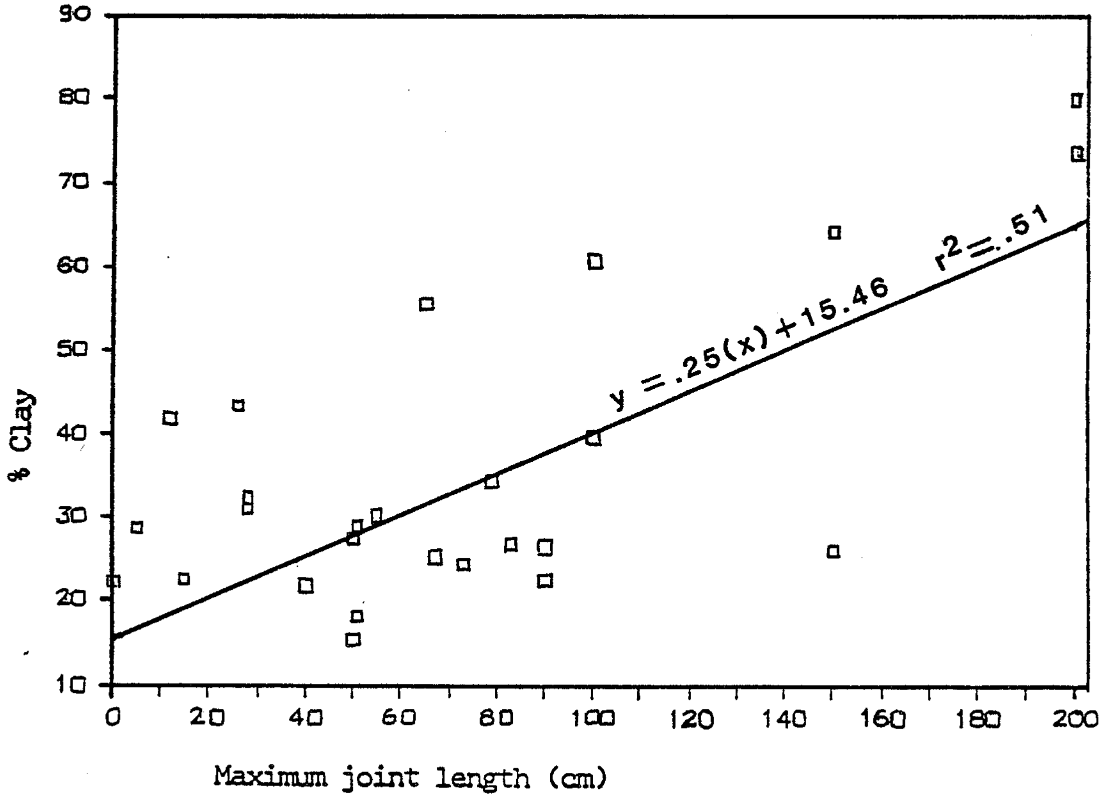


Figure 19: % Clay (less than 0.002um) versus maximum joint length

Joint Length

Excluding joints that have been significantly modified by slope failure, maximum joint length ranges from .28 to about 2.0 m. Maximum joint length shows a positive correlation with the clay content of the till units (Figure 19), and may reflect the greater shrinkage of clay-rich till during dessication.

The maximum length of the longest dimension of blocks or peds formed by intersecting joint systems ranges from 2.9 to 29.0 cm. The maximum ped length was observed to increase with depth at any one site, indicating the influence of dessication on the jointing characteristics.

Joint Pattern

Systematic mapping of joint pattern at two sites showed that joints frequently intersected with angles of 90 degrees. Polygonal joints due to dessication usually do not form acute or right angle intersections, but typically intersect at obtuse angles, commonly 120 degrees. The abundance of right angle joint intersections suggests that dessication alone is not responsible for joint genesis.

JOINT CHARACTERISTICS BY STRATIGRAPHIC UNIT

One goal of this study was to investigate what control the physical properties of the till units have on jointing characteristics. Some degree of jointing was found in all of the units investigated, but differences from site to site and from unit to unit mask any clear relationship between the till properties and the joints. While joint orientation stays nearly constant for nearby exposures of the same stratigraphic unit, joint spacing and length sometimes differ greatly between sites and from unit to unit. The jointing style associated with the tills studied is summarized below and in Figure 20.

Till of the New Berlin Formation

Only the exposure of the New Berlin till at St. Francis (Site #24) was examined as part of this study. This section may represent an anomalously fine-grained (silty-clay-loam) facies. The poorly developed jointing that is present is nearly parallel with that of the overlying Oak Creek till. Maximum joint length is the shortest of any unit, 28 cm. Joint spacing ranges from 1.9 to 6.8 cm.

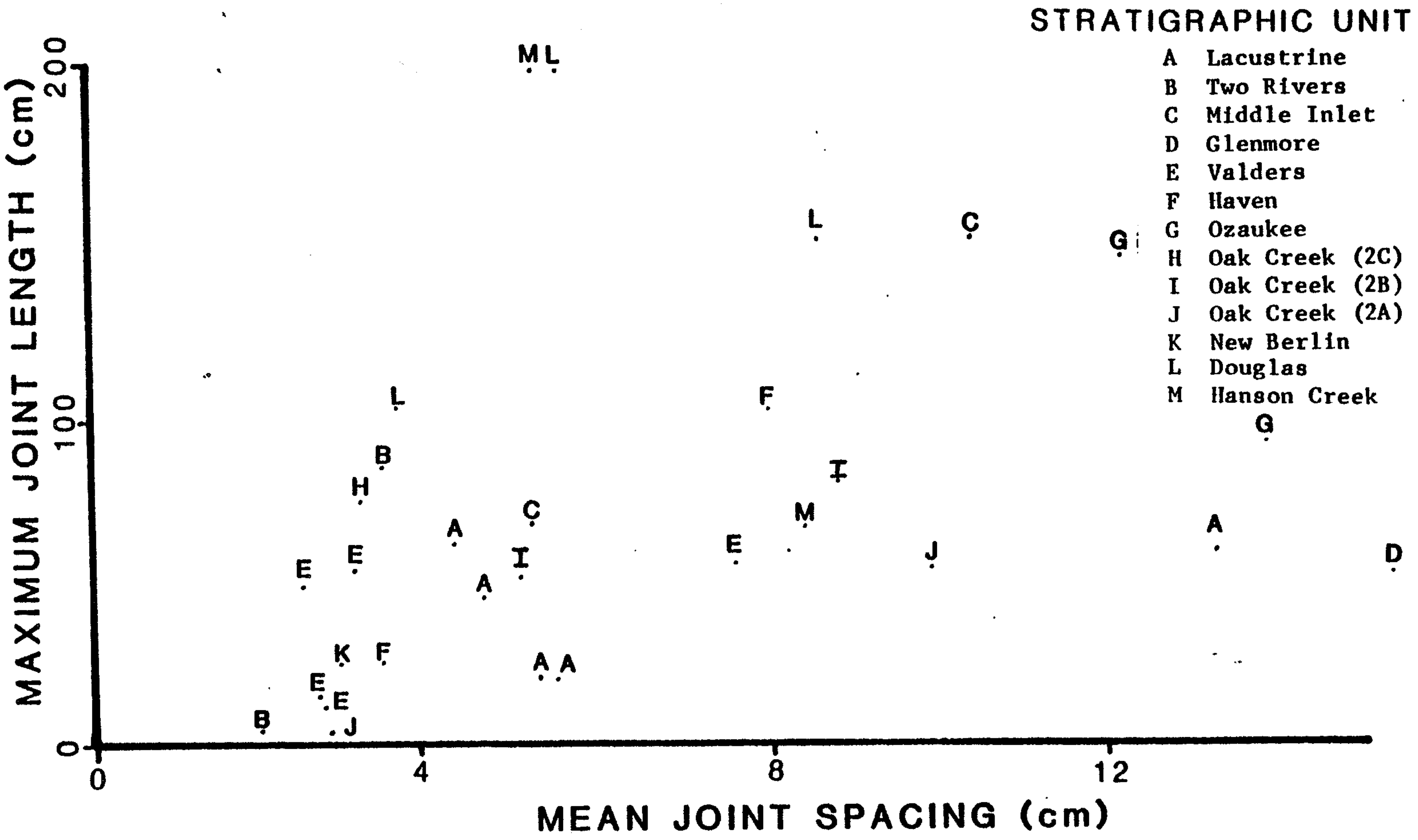


Figure 20 : Summary plot of joint length versus mean spacing for sites in the thirteen stratigraphic units investigated.

Tills of the Oak Creek Formation

Tills of the Oak Creek formation have been informally subdivided into three units -- 2a, 2b, and 2c -- by Mickelson and others (1977). The youngest of these units, till 2c, was found only at the St. Francis site (Site #27). Although it is somewhat similar in texture and mineralogy to the underlying till 2b, it is highly oxidized and contained only slope-parallel joints.

till 2a

Two exposures were available of the clay-loam till 2a, both at St. Francis (Sites #23 and 25). Observations from these two sites are combined in the data plots (Figures 10c and 14; Appendix 5). No microfabric samples were obtained from this unit, but ice-flow direction was estimated from striations on boulders in a boulder pavement between till 2a and the underlying New Berlin till. Mean joint spacing ranges from 3 to 1 cm, and maximum joint lengths are very short (less than 50 cm), perhaps because of a lack of previous dessication at the depth (approximately 20 m) where this till is found. The high-angle joints are parallel and perpendicular to the striations; conjugate high-angle joints were absent.

till 2b

Thick sections of this unit over outwash sand were studied at two bluff sites in the Milwaukee area (Sites #26 and 29). Till 2b is commonly silty-clay-loam and is oxidized at the surface; due to slumping, only oxidized exposures were available for study. Mean joint spacing ranges from 5 to 9 cm and maximum lengths are generally less than 100 cm. Three to four high-angle joint sets are present at the two sites, but low-angle joints are absent. Joint orientations measured in this study at Bender Park (Site #29) agree with those measured by Sterrett (1980), who excavated a test pit on level ground behind the bluff face.

Tills of the Kewaunee Formation

Ozaukee till

In contrast to the situation that may have existed during deposition of the younger Valders and Two Rivers tills, microfabric profiles indicate that ice coming out of the Lake Michigan basin during the deposition of the Ozaukee till shifted from a northerly to a more easterly direction as deposition proceeded. This is illustrated by the microfabric sample from Port Washington (Site #21) (Figure 10b), and by the striated, up-ice dipping, low-angle joints at Notre Dame (Figure 21). The jointing

Figure 21: Slickensides on low angle joint from Ozaukee till at Notre Dame (Site #22). Slickensides are roughly parallel to till microfabric and indicate ice flow from approximately due east.

seems to be related to the last (easterly) ice flow. The thick sequence of loam to clay-loam Ozaukee till over sand at Port Washington contains joints with the longest mean spacings measured during this study, ranging from 12 to 14 cm. The orientation of high-angle jointing is very similar at the two Ozaukee sites.

Haven till

Silty-clay to silty-clay-loam Haven till of unknown thickness is exposed at four sites (Sites #5, 13, 17 and 20-2), and has mean joint spacings of 4 to 8 cm and maximum joint lengths of up to about 1 m. Three to five high angle joint sets were found at the Haven sites, the largest number at the type section exposure (Site #20-2). The large number of joint sets at the type section may be a reflection of ice overide by both Haven and later Valders ice.

Valders till

The silt-loam to clay-loam Valders till is fairly thin in most exposures (Sites #15, 16, 18, 19 and 20), and was found over older till, bedrock, and outwash sand and gravel. Jointing in the Valders till was not well developed, particularly at Valders quarry (Site #15). Maximum joint lengths were usually less than 60 cm and mean spacings were in the range of 2 to 8 cm. Two to four high-angle joint sets were found in the Valders till; low-angle joints with preferred orientation were absent. Valders till has the highest expandable clay mineral content of tills of the Kewaunee Formation.

Two Rivers and Glenmore tills

At the four sites in the Glenmore and Two Rivers till of the Kewaunee Formation, thin till overlies fluvial or lacustrine sand (Sites #3, 7, 8, and 12). The till is commonly clay-loam to loam in texture, and has a clay mineralogy that is dominated by illite. At two sites, the underlying sand showed evidence of deformation such as shearing and folding (Figure 22).

Mean joint spacing in the Two Rivers till ranged from 4 to 6 cm and the maximum joint length is less than 1 m. Four high-angle joint sets were found at each of the sites; at three of the sites, prominant low-angle joints striking roughly perpendicular to ice flow are also present.

Middle Inlet till

Two exposures of the silt-loam Middle Inlet till were examined; both are in excavations in the Lake Winnebago (Glacial Lake Oshkosh) lowland. Gleyed joints at both exposures are very well developed, indicating that joints can remain open below the water table. Maximum joint

Figure 22: Sheared sand and overlying Glenmore till at Door pit (Site #3). Ice motion was from left to right.

lengths range from 75 to 150 cm, and mean spacings range from 6 to 12 cm. Two to five high-angle joint sets were found at the Middle Inlet sites; low-angle joints were prominant only at the Koenen farm site. The joints at the Outagamie County Landfill north of Appleton, Wisconsin, are thought to increase the permeability of the till by several orders of magnitude above the values measured from small laboratory samples (Gordon and Huebner, 1983). Post-glacial lacustrine sediment, which covers the Middle Inlet till at the Outagamie landfill, shows a similar joint pattern as the underlying till. This may reflect an upward propagation of joints (Figure 23).

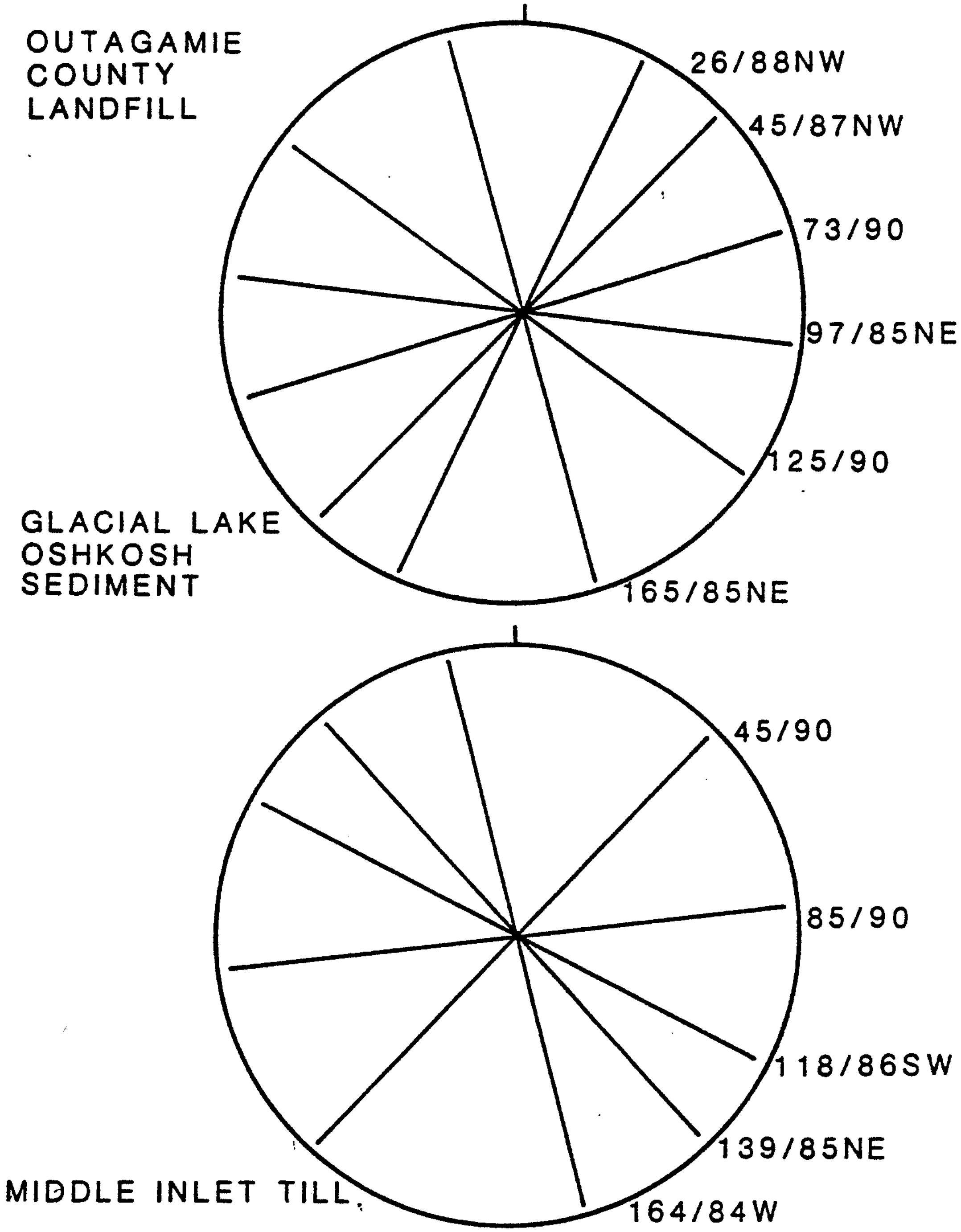


Figure 23 : Comparison of joint set orientation for Middle Inlet Till and overlying lacustrine sediment at Outagamie Landfill (Site #36).

Tills of the Miller Creek Formation (Douglas and Hanson Creek tills)

The thick, clayey, smectite-rich tills of the Lake Superior basin are very different from their Lake Michigan Lobe counterparts in eastern Wisconsin. Mean joint spacings do not differ greatly, typically ranging from 4 to 8 cm, but very long lengths (up to approximately 2 m) were found at some exposures. As ice-flow direction did not vary a great deal along the shoreline, the joint pattern is fairly consistent from site to site, and includes three to five high-angle joint sets. Three sites also have low-angle joints (Sites #33, 34 and 35). Johnson (1980) and Need (1980) found low-angle stringers of clay in the Hanson Creek till that they interpreted as resulting from shear during ice flow. There is a good agreement between the results of pebble fabrics at Sites #31, 34 and 35 and till microfabric. These tills were studied only at exposures along the bluffs of Lake Superior.

Glaciolacustrine Sediment

Glaciolacustrine sediment examined as part of this study shows a variety of jointing characteristics. Laminated silty clay exposed at the base of the bluff at Sandy Bay (Site #4) exhibits a joint pattern much like many of the till units, as do the rythymites exposed at St. Francis (Site #28) (Figure 10). The lacustrine unit exposed at the Highway BB and Two Creeks Buried Forest sites (Sites #9 and 10) has a quite different joint pattern, and has the only intermediate angle (40 to 60 degree) joints found in this study. These differences in joint style between the lacustrine units may reflect different histories of ice overridding or upward propagation of joints from underlying till, as may have occured in the stratified Glacial Lake Oshkosh sediment at Outagamie Landfill (Site #36-1) (Figure 23). Mean joint spacing for all the lacustrine units examined ranges from 4 to 13 cm, and maximum joint lengths are generally short, not exceeding 50 cm.

RELATIONSHIPS BETWEEN TILL PROPERTIES AND JOINT CHARACTERISTICS

In this study I have attempted to investigate empirical relationships between the physical properties of till and their jointing characteristics using results from laboratory measurements of texture, carbonate content, clay mineralogy, and from published and unpublished geotechnical data. Whether these correlations represent controls on jointing is unknown, but with the variety of tills examined, these generalizations may be applicable to other till units in the upper midwest.

TEXTURE

As was shown in Figure 19, maximum joint length is proportional to the clay content of the till, and it appears to be inversely proportional to silt and sand content (Figures 24 and 25). These relationships may reflect the importance of dessication in lengthening and exploiting prexisting joints, because dessication's effects are enhanced in materials with higher clay contents (e.g. Anderson and others, 1982).

Joint spacing is proportional to sand content for spacings of about 1 to 8 cm (Figure 26, line A). Joints with wider spacings (8 to 24 cm) are also proportional to sand content, but the relationship has a different slope

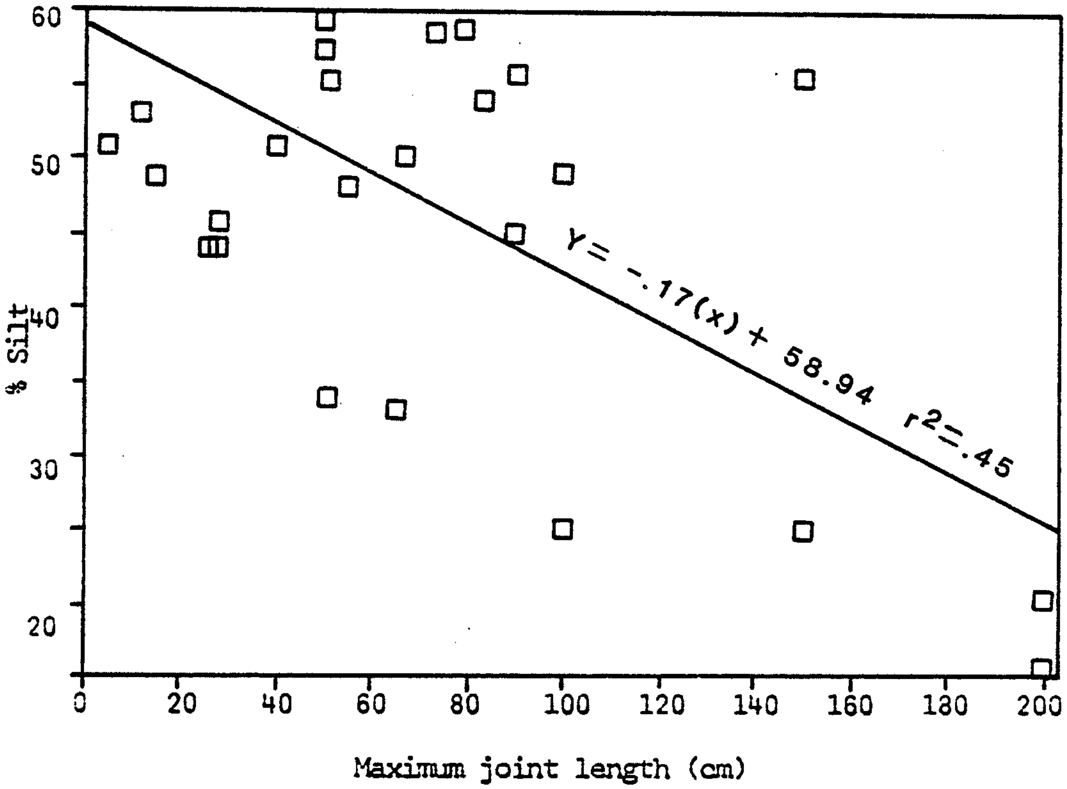


Figure 24: % silt versus maximum joint length. Silt content and maximum joint length show a negative correlation.

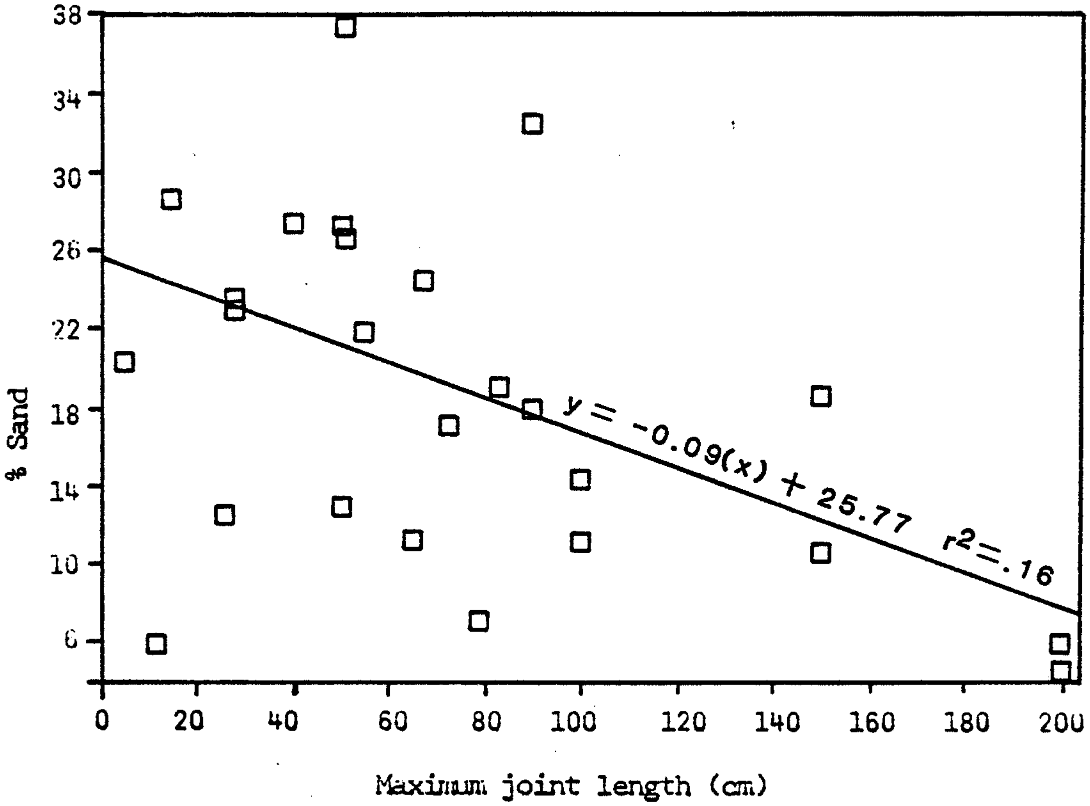


Figure 25: % Sand versus maximum joint length, showing weak inverse correlation between the variables

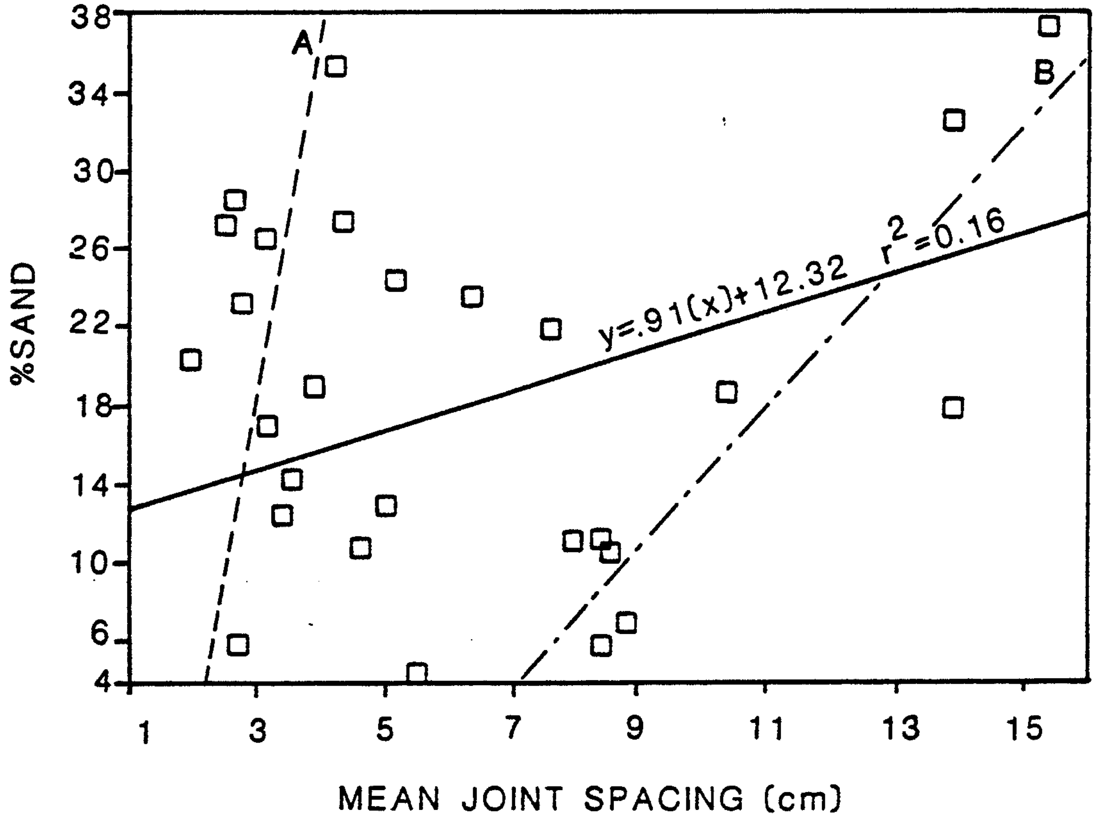


Figure 26: Percent sand versus mean spacing of high angle joints. Lines A and B are visual best fit lines for mean joint spacings less than 7 cm and greater than 7 cm, respectively. The solid line is the least-squares regression line for all data.

than the more narrowly spaced joints (Figure 26, line B). The least-squares regression line shown in Figure 26 masks this relationship.

CARBONATE CONTENT

The carbonate content in till of the same unit, when compared from site to site, exhibits little change in the horizons where joint measurements took place (for example, typical sections are shown in Figure 27). Till of the Miller Creek Formation (Superior Lobe) has much less carbonate than Lake Michigan Lobe till, and also has the longest joint lengths. The lack of carbonate, along with the large amount of expandable clay, creates greater shrink and swell upon wetting and drying (Rimmer and Greenland, 1976); this may lengthen joints in areas subject to moisture fluctuations.

CLAY MINERALOGY

The wide range in clay mineral composition, from the smectite-rich Miller Creek Formation to the illite-rich tills of the Oak Creek and New Berlin Formations, seems to have little effect on the jointing characteristics. As an example (Figure 28), while the Valders till (point A on Figure 8) has very high smectite contents, on the order of 58%, it has some of the shortest maximum joint lengths. Samples of the Hanson Creek till, with roughly similar

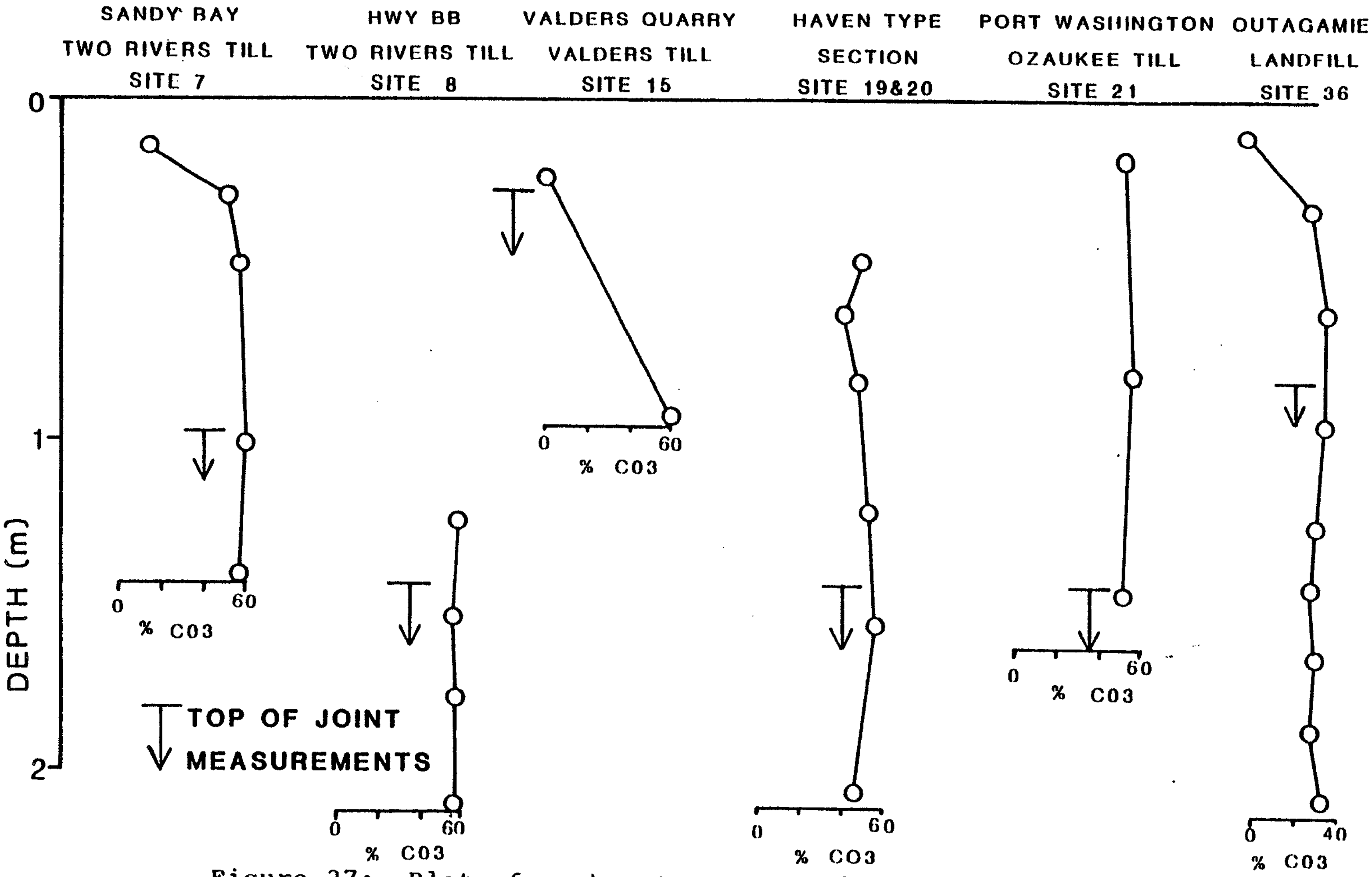


Figure 27: Plot of carbonate content in coarse silt fraction versus depth for six sites in till in eastern Wisconsin. Leaching appears to be restricted to the uppermost 30 to 40 cm of the till.

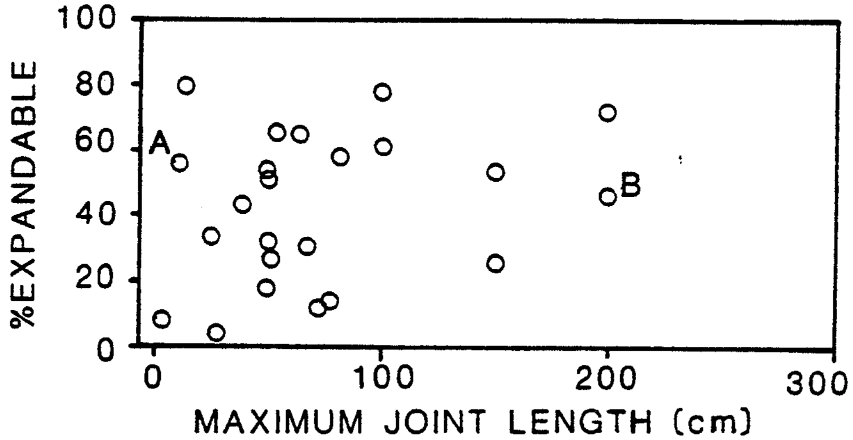


Figure 28a

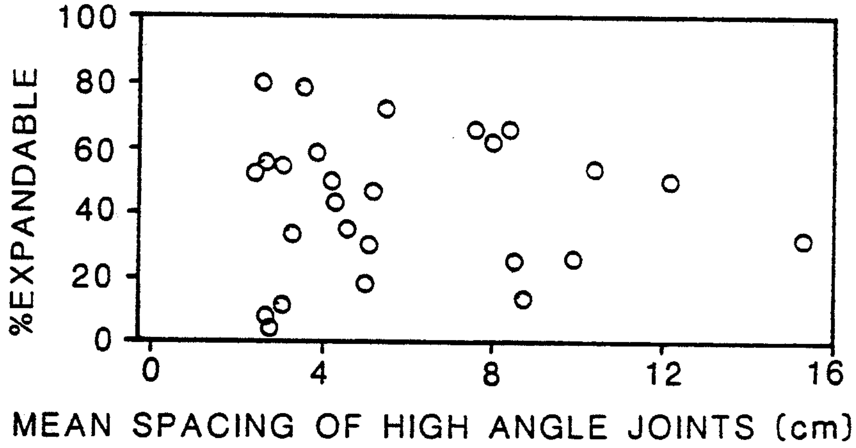


Figure 28b

Figure 28: (a) Percent expandable clay versus maximum joint length.
(b) Percent expandable clay versus mean spacing of high angle joints.

smectite content (point B on Figure 28), had much longer joints. Any effect of the clay mineralogy is probably masked by much larger differences in clay and carbonate content in the units studied.

GEOTECHNICAL PARAMETERS

Detection of any relationship between geotechnical parameters -- Atterberg limits, undrained strength (ϕ' and c'), and the overconsolidation ratio -- and the jointing characteristics is hampered by the limited data on the geotechnical properties of many of the tills. The data available show that the geotechnical properties do not vary greatly from unit to unit (Table 3) and so are not useful in distinguishing between the tills (Pulley, 1980). Caution must also be observed when trying to relate the geotechnical properties measured from small samples to the conditions that may have existed during deposition or deformation beneath the ice.

Overconsolidation Ratio

The limited data available indicates that many of the till units studied are moderately overconsolidated, with overconsolidation ratios of 3.0 to 4.5, although some individual values are as high as 31 (Pulley, 1980; Mickelson, and others, 1979). While this supports the relationship observed by many authors between jointing and

Table 3: Geotechnical properties of jointed till units in Wisconsin (cont.)

Unit	Liquid Limit (%)	Plasticity Index (%)	Water Content (%)	O' degrees)	c' (kN/m2)	OCR	Reference
Oak Creek Formation:							
till 2C							
till 2B	20.3	6.4		31.4	0		Mickelson, et al., 1977
	25.1(2.8)	10.0(2.3)	15.2(1.4)	31.4	0		Mickelson, et al., 1979
till 2A	24.5(2.9)	9.9(1.9)	15.8(3.1)	31.1(.1)	0		Mickelson, et al., 1979
New Berlin till	Non-plastic		4.8	34.6	0		Mickelson, et al., 1979
Miller Cr Formation:							
Douglas till	69.5	45.1	47.3	22.2	2.55	3.1	Bagchi, 1982 Pulley, 1980
Hanson Creek till	63.7	32.4	34.8	30.5	0		Bagchi, 1982

Note: Values are listed as Mean (Standard Deviation)

Table 3: Geotechnical properties of jointed till units in Wisconsin

Unit	Liquid Limit (%)	Plasticity Index (%)	Water Content (%)	0' degrees)	c' (kN/m2)	OCR	Reference
Kewaunee Formation:							
Glenmore till							
Two Rivers till	26.3	12	15.6			4.8	Pulley, 1980
						31	Pulley, 1980
Middle Inlet till							
Valders till	22.8	10.4	13				Mickelson, et al., 1977
						9	Pulley, 1980
	28.4(5.0)	13.1(3.4)	17.4(4.1)	29.3(.6)	28.3(5.6)	4	Mickelson, et al., 1979
Haven till	30	13.5		29.5	30.21		Mickelson, et al., 1977
	30.3(6.0)	14.4(4.7)	16.5(2.5)	31.2(0.5)	23.8(5.6)	3.7	Mickelson, et al., 1979
Ozaukee till	30	14.4		30.5	0		Mickelson, et al., 1977
	30.6(2.6)	14.0(2.5)	17.6(1.2)	31.4(.8)	0	3.3	Mickelson, et al., 1979

Note: Values are listed as Mean (Standard Deviation)

overconsolidation, it does not allow enough range to evaluate the effects of the overconsolidation ratio on jointing characteristics.

Strength Parameters

The angle of internal friction (ϕ') varies from 22.2 to 34.6 degrees for the units studied, but the majority of the units are in the range of 29 to 31 degrees. Cohesion is zero in all but three of the units, reflecting the low overconsolidation ratio (Mickelson and others, 1979). No relationship was detected between the angle of internal friction and jointing characteristics.

Atterberg Limits

There is a weak negative correlation between the plastic limit and the frequency of vertical joints (Figure 29). This may reflect the combined influence of clay content and clay mineralogy, because clay content and the plastic limit are positively correlated (Figure 30). From the relationship between joint length and clay content described above, it seems likely that sediment with high clay content and long joints will have fewer joints per volume and thus lower joint frequency.

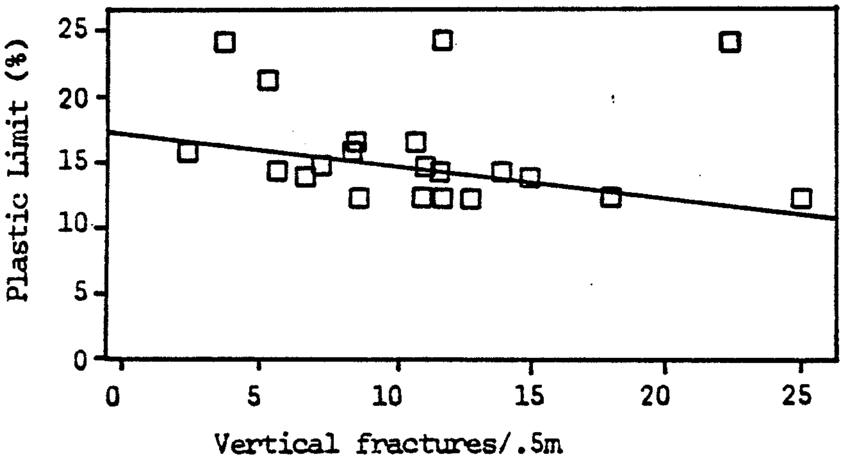


Figure 29: Plastic Limit versus frequency of vertical joints

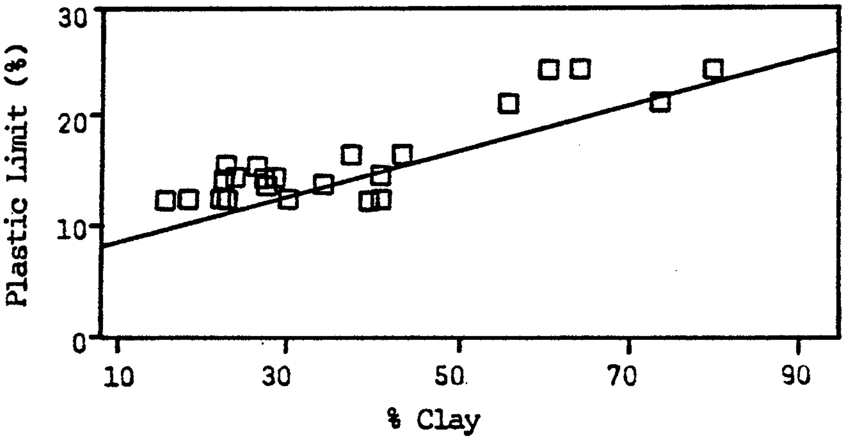


Figure 30: Plastic Limit versus clay content

THICKNESS

A weak correlation exists between joint spacing and the thickness of the till units (Figure 31). This relationship, while tenuous, may support a tectonic or deformational genesis, as suggested by Hills' (1972) statement that spacing is proportional to bed thickness for joints of tectonic origin. Joint length may also be indirectly controlled by the permeability of the material underlying the jointed sediment; those exposures underlain by thick sequences of fine grained-till or lacustrine sediment commonly had longer joints than the sites located over more permeable sand or gravel. This may reflect greater joint propagation in situations where excess pore pressure is not rapidly dissipated.

DEPTH

Joint length and ped size increases with depth in the upper-most 1 to 4 m of the tills, but then remains fairly constant. This probably represents the depth to which modern processes such as dessication, and in the upper horizons freeze/thaw, can currently operate.

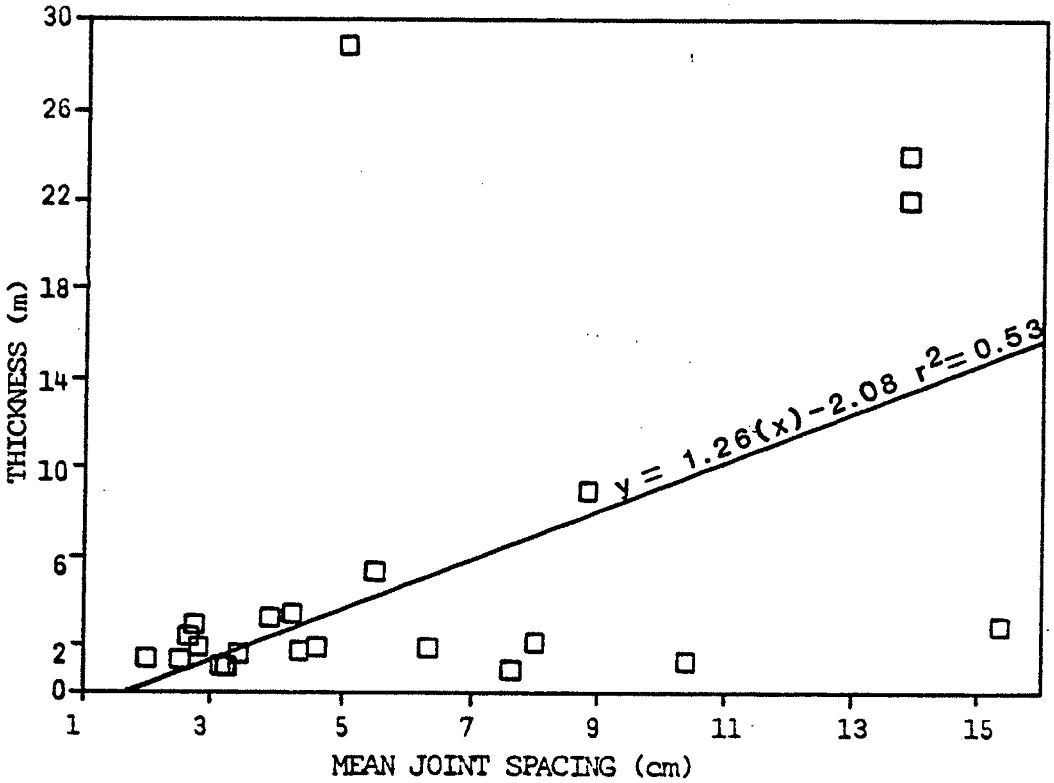


Figure 31: Minimum thickness of till units versus mean joint spacing.

There is some coincidence of bedrock jointing with the jointing in the overlying till (Figures 32 and 33). However, at Valders Quarry (Sites #14 and 15), the correlation between bedrock jointing and the jointing in the Valders till is not strong, despite the fact that jointed till is resting directly on the bedrock. Bedrock jointing is somewhat variable over the study area, which may foster spurious correlations. This was the case in Alberta, where Westgate (1976) found that jointing in till was consistent both with the direction of ice flow and with bedrock jointing.

In summary, texture seems to have the strongest relationship to jointing characteristics. Joint length is correlated with clay content, and spacing (weakly) with the proportion of sand. This may reflect the influence of dessication on the jointing characteristics. The till units have not undergone extensive leaching, as shown by the large carbonate content of their coarse silt fractions, so the influence of changes in carbonate content on jointing cannot be assessed. Geotechnical strength parameters do not vary widely between the till units and so their effect on jointing characteristics is uncertain. The weak positive realtionship between joint spacing and till thickness suggests a deformational origin for the jointing.

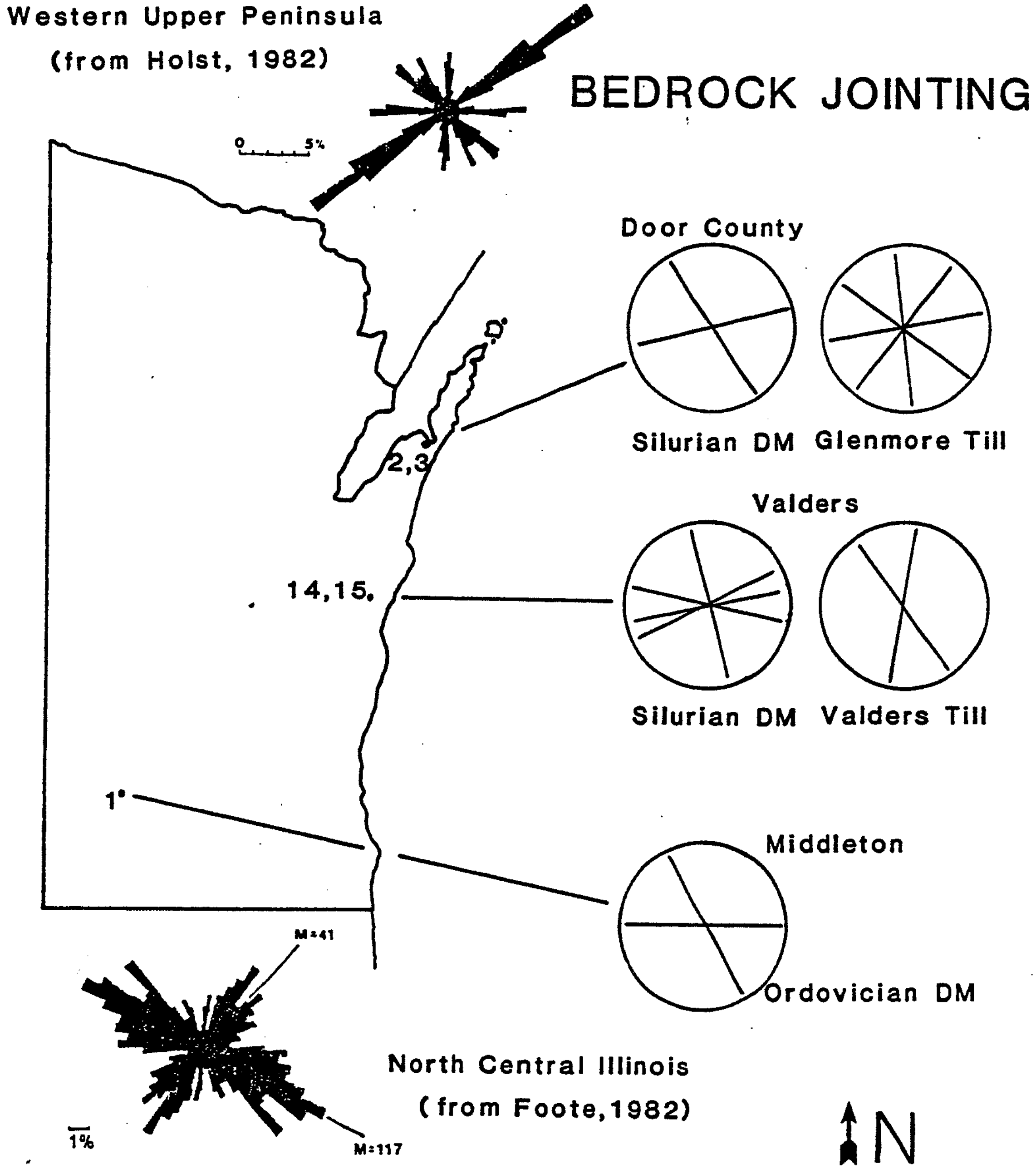


Figure 32: Comparison of bedrock and till jointing in study area and adjacent regions.

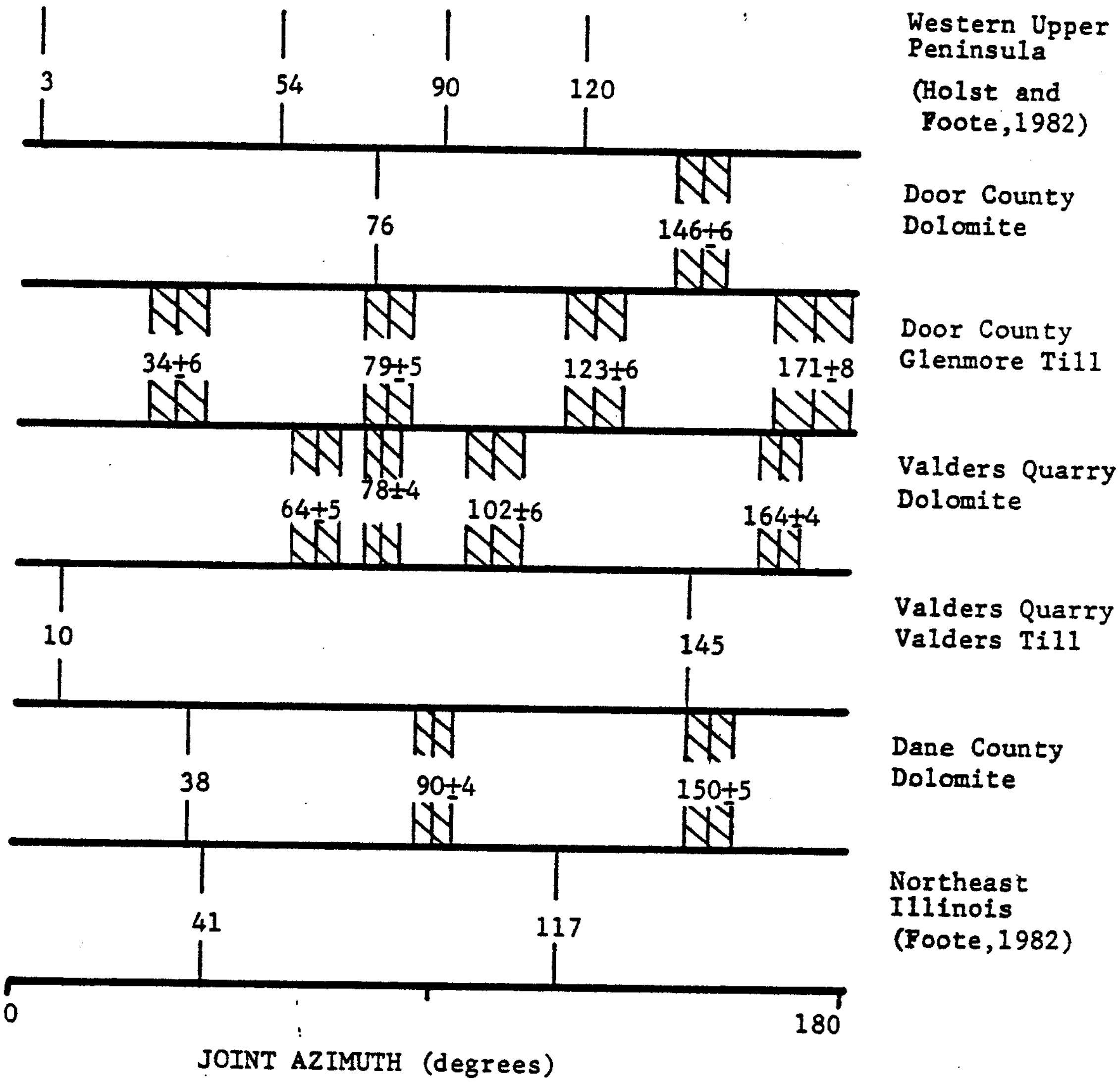


Figure 33: Summary plot of high-angle joint orientation in till and bedrock in eastern Wisconsin and adjacent Michigan and Illinois. Hatched rectangles are joint azimuth and associated error of estimation from PATCH. Single lines are joint azimuth as estimated from POINT.

JOINT GENESIS

The data on joint orientation and other characteristics form a basis on which to evaluate models for joint genesis. Much has been written on the ability of different processes to create joints in fine-grained till (summarized in Chapter 1). The joints examined in this study are undoubtedly polygenetic and may reflect a history of both deformation and weathering. Several models are proposed here for the genesis of the jointed tills in eastern and northwestern Wisconsin: deformation resulting from overriding by a moving glacier, regional deformation owing to isostatic rebound following deglaciation, and volume loss resulting from chemical change. While these models may explain the orientation of the joints, subsequent modification through weathering and unloading may control joint spacing, length, and surface characteristics.

SUBGLACIAL DEFORMATION

Formation of joints beneath moving ice has been proposed for till beneath a modern glacier (Boulton, 1970) and for joints in Pleistocene till (McGown and others, 1975). The thermal regime, flow regime, strain rate, ice thickness, pore pressure, and strength of the glacier bed are all factors that influence joint formation in the subglacial environment, and each bears discussion.

Thermal Regime

The temperature at the base of the glacier can influence joint formation in several ways. Basal temperature is not static, but may change both with respect to position and through time (Boulton, 1972). Permafrost conditions associated with polar glaciers may inhibit consolidation beneath a glacier, resulting in till that is normally consolidated to only moderately overconsolidated (for example Mickelson and others, 1979). This is in contrast to the moderate to large overconsolidation that can occur in till that bears the weight of the glacier while unfrozen (Harrison, 1957). The tills examined in this study are lightly to moderately overconsolidated, so it is unclear from this line of evidence what thermal regime existed during till deposition.

Thermal regime is also important in influencing the strength of fine-grained material at the glacier bed. Freezing generally increases the strength of fine-grained sediment (Tystovich, 1975), which otherwise has the lowest strength among glacial deposits (Muller, 1947, in Mathews and Mackay, 1960). Below the freezing point, the shear strength of frozen clay increases with decreasing temperature (Tystovich, 1975). Boulton (1972) states that the strength of debris/ice mixtures is greater than that of clean ice, and so failure taking place beneath a glacier frozen to its bed is restricted to clean ice or the base of the frozen sediment. However, debris with a high clay content, such as must have been the case with the fine-grained tills in this study, loses strength with increasing ice content and can approach that of clean ice (Muller, 1947, in Banham, 1975). Therefore it seems possible that deformation of fine-grained till can take place in either thermal regime, but the sediment must have been close to the melting point and contained a large amount of water or ice.

Thermal regime may also influence joint preservation. Fractures in frozen fine-grained till were observed to be filled with segregation ice lenses beneath the Nordenskioldbreen (Boulton, 1972). While Boulton interpreted the till as being refrozen after deformation, ice lenses may be an important factor in preserving fracture orientation during deglaciation, and in fact may be responsible for some of the fractures themselves.

Independent evidence for the thermal regime of ice of the Lake Michigan Lobe is sketchy, as the Late Woodfordian climate changed through time. The existence of drumlins and sub-glacial tunnel channels in the pre-Port Huron till

Strain Rate

Rapid loading, which may result from a surging glacier or from abrupt changes in pore pressure beneath the glacier, could have caused Mohr-Coloumb failure, yet not lasted long enough to heavily consolidate the sediment at the glacier bed. Overconsolidated clay, as it is a dilatant material, decreases in shear strength with more rapid loading (Hight and others, 1979). From a theoretical model of vertical consolidation, Poorooshasb and Yong (1983) calculated that during large loadings of a low permeabilty material, hydraulic fracturing could result if the loadings were rapid enough to cause elevation of the pore pressure that could not be dissipated through the pore spaces. The duration of loading is also important, as the shear strength of clay decreases with time after the initial load application (Tystovich, 1975). Thus, fracturing is favored in overconsolidated clay when the load is applied rapidly and the loading is of a long duration.

Confining Stress

Ice thickness and subglacial pore pressure control the effective confining stress at the glacier bed. High pore pressure, and thus low confining stress, inhibit consolidation and lower the strength of fine-grained sediment. At low confining pressure, brittle failure can occur (Hight and others, 1979), whereas high confining pressures may limit deformation to microfracturing or secondary creep. As overconsolidated clay behaves stiffly when deformed at confining pressures up to the overconsolidation pressure (Al- Shaikh-Ali, 1981), the decrease in confining pressure during ice retreat or down-wasting could enable deformation to occur as the ice margin retreated back over till that had been overconsolidated under thicker ice.

Unloading could also occur if high pore pressure built up. High pore pressure can be caused by overriding of impermeable units (Boulton, 1975), overriding of groundwater discharge areas, or freezing of water-bearing units at the glacier margin (Clayton and Mickelson, 1981). As mentioned above, it was observed that joints in thick till or in till underlain by relatively impermeable sediment typically were longer than those in till underlain by sand or gravel.

The importance of confining pressure on the deformation of the sediment at the glacier bed is illustrated by the work of Boulton and his coworkers (1974). From a model of the Breidermerkerjokull, Iceland, they found that deformation of subglacial till was limited to the area near the glacier margin, but could extend farther beneath the glacier if the glacier were to override impermeable units and and thus produce higher pore pressure beneath the ice. At higher confining pressures the till was thought to be unable to deform.

Flow Regime

Compressive flow, the slowing of ice movement along a flow path, occurs if ice is removed from the glacier by ablation, or if the flow lines are concave up as the glacier rises over an obstacle (Nye, 1952). Under compressive flow, the horizontal normal stress parallel to ice flow is greater (more compressive) than the vertical normal stress (Nye, 1952). Compressive flow appears to be associated with folding and deformation of frozen sediment at the glacier bed (Shaw, 1979); an example is the "brecciation" of fine-grained till described by Schwan and Ritzema (1982). Compressive flow is often associated with a transverse orientation of prolate pebbles in the ice (Paterson, 1981).

There may be a connection between some types of joints and the predominance of compressive flow. Nye (1952) states that during compressive flow, planes of maximum shear stress near the glacier sole are near-horizontal, dipping upglacier, or near-vertical and perpendicular to the direction of flow. Boulton (1970) described slickensided near-horizontal joints from beneath the Nordenskioldbreen, Svalbard, and found that the lenses formed by these joints were in turn cut by vertical joint sets perpendicular to ice-flow or conjugate around the principal horizontal stress. Similar vertical joints have been found to lie conjugate around the greatest horizontal stress in drumlinized Pleistocene till (McGown and others, 1975). The near-vertical conjugate joints, perpendicular joints, and the rarely striated near-horizontal joints discovered in this study are probably related to compressive ice flow.

Compressive flow is also associated with longitudinal crevasses at the ice surface, while extending flow is associated with transverse creavasses. As crevasses usually do not penetrate to the glacier base, their propagation into the till and subsequent preservation as joints is seems unlikely except near the glacier margin. No evidence such as crevasse fillings was found to suggest this type of origin for the joints examined in this study.

Failure Mode

Joints in the till units examined in this study have probably been modified by dessication and coated by weathering products. This modification may mask or alter evidence for shear, such as slickensides, or of tensional failure, such as feather or plumose structure. None of these features were found on vertical joints, and only a few near-horizontal joints were slickensided. Thus, it is not known whether failure occured because of shear or through tensile stress.

Transfer of tensile stress from glacier ice to the glacier bed is problematic if no cohesion exists between the glacier sole and the glacier bed. Cohesion is increased if the glacier is frozen in patches to its bed, and then sliding may produce tension on frozen ground (Derbyshire and Jones, 1980). This is illustrated by a figure from Boulton (1970) (Figure 34), which shows unfilled vertical tension cracks in ice at the base of a glacier moving over frozen till that is cut by vertical joints.

As no evidence of the failure mode of joint formation is available, it is also difficult to evaluate what mode failure may have taken. Failure may have been restricted to discrete zones, as in Mohr-Coloumb failure, or could have been pervasive and distributed throughout the till mass, as in ductile failure. Mohr-Coloumb failure is observed to occur at low confining pressures during triaxial tests of unfrozen, normally to slightly overconsolidated, reconstituted silty clay (Hight and others, 1979). Tertiary creep occurs at lower levels of

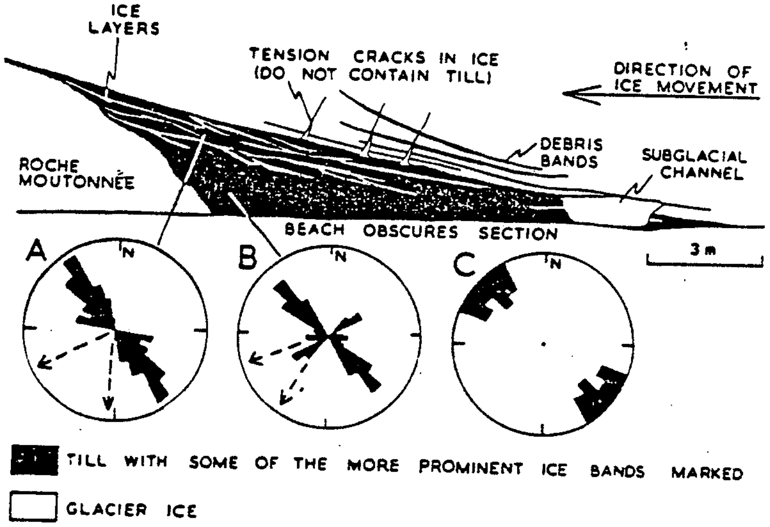


Figure 34: Conditions at base of Nordenskioldbreen, Svalbard, from Boulton (1970; Figure 5).

Boulton's caption reads "...Active glacier ice over-rides partly frozen till banked against the bank of a roche moutonnee. Note the ice layers lying along shear planes within the till. A and B are rose diagrams showing the orientations of long axes of rod- and blade-shaped till stones, projected on to the horizontal plane. 70 measurements at A and 68 at B. Dashed arrows show the variation in direction of slickensides on sub-horizontal shear planes at these two horizons, reflecting the direction of glacier over-riding. C shows the orientation of 24 high-angle joints measured throughout the till."

sustained stress but then ends in fracture as strain levels increase (Derbyshire and Jones, 1980).

Preservation

Preservation of joints caused by subglacial deformation through subsequent dessication and weathering seems to require a rearrangement of the fabric. Ductile failure of unfrozen material can form planes of weakness that control subsequent fracture orientation (Boulton and others, 1974), while Mohr-Coloumb failure can lead to dilation accompanied by a reduction of density in dense fine-grained materials. In fine-grained sediment, dilation is restricted to overconsolidated clays (Mead, 1925), and represents a breaking of the interparticle bonds that form during overconsolidation. La Fluer (1980) found a fabric of "conjugate, incipient shears shears expressed as clustered laths of illite" in vertical thin sections from the jointed Lavery Till from the West Valley site in New York. Orientation of elongate sand grains in horizontal thin sections from the Wisconsin till units sometimes show minor peaks parallel to joints that are conjugate around the direction of ice flow (e.g. at Outagamie, Memorial, Notre Dame, and Bender Park; Figure 10).

Models of subglacial deformation

Presented below are four models that represent combinations of the conditions of flow regime, thermal regime, effective stress, and overconsolidation that may have existed during subglacial deformation. Comparison of these models to the field evidence points out some discrepencies, which are discussed. It should be kept in mind that these models describe what might have happened during and shortly after deposition while the field data represents the superposition of processes through time, and may in fact represent only the actions of the most recent processes.

1. shear failure during compressive flow

Undrained compression of moderately overconsolidated till beneath a grounded, wet-based glacier undergoing compressive flow could result in shear fracture along near-horizontal, up-ice dipping shear planes, vertical joints perpendicular to ice flow, and vertical joints conjugate around the maximum horizontal stress. The deformation probably takes place not far behind the ice margin under conditions of low effective confining stress. Subsequent unloading during retreat could enhance the jointing as the sediment relaxes after loading. Joints formed in this situation would typically be associated with deformation of other materials beneath the glacier through faulting and folding, as well as with fluid escape structures that reflect high pore pressure. Preservation of joint orientation may have occured by dilation of the material during shear or through fabric rearrangement.

This model is supported by the evidence for compressive flow, which includes; concave uphill profiles for many of the sites (Figures 35a and 35b), transverse orientation of sand grain microfabrics, and folding and faulting of underlying lacustrine sediment. The sedimentology of the till units is consistent with a wet-based, grounded glacier; the tills of the Lake Michigan Lobe were termed "basal" by Acomb (1978), who also noted a facies that he described as "subaqueous flow tills" at the base of some of the units. Johnson (1983) stated that the till units of the Lake Superior Lobe were "subglacial". The overconsolidation ratio, which ranges from 3.0 to 4.5 for most of the tills, is consistent with this model. Evidence of shearing of till of the Miller Creek Formation was noted by Johnson (1983). The consistent orientation of ice-flow and vertical joints described earlier points to a relationship with a horizontal stress directed in the ice-flow direction.

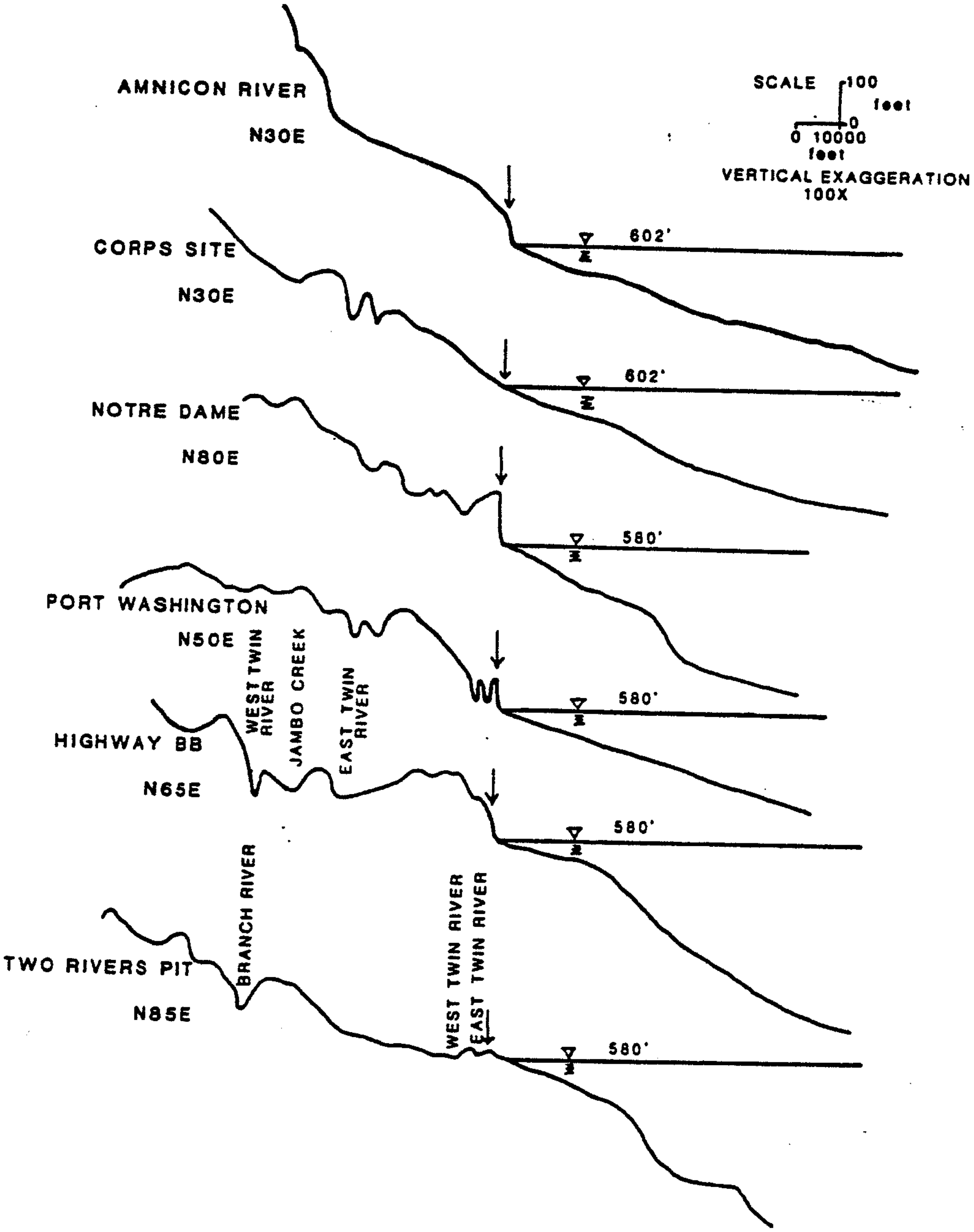


Figure 35a: Topographic profiles along ice flow direction as inferred from till microfabric. Bearing of ice flow is given below location name. Vertical arrow denotes site location. Reference elevation given in feet. Ice flow from right to left.

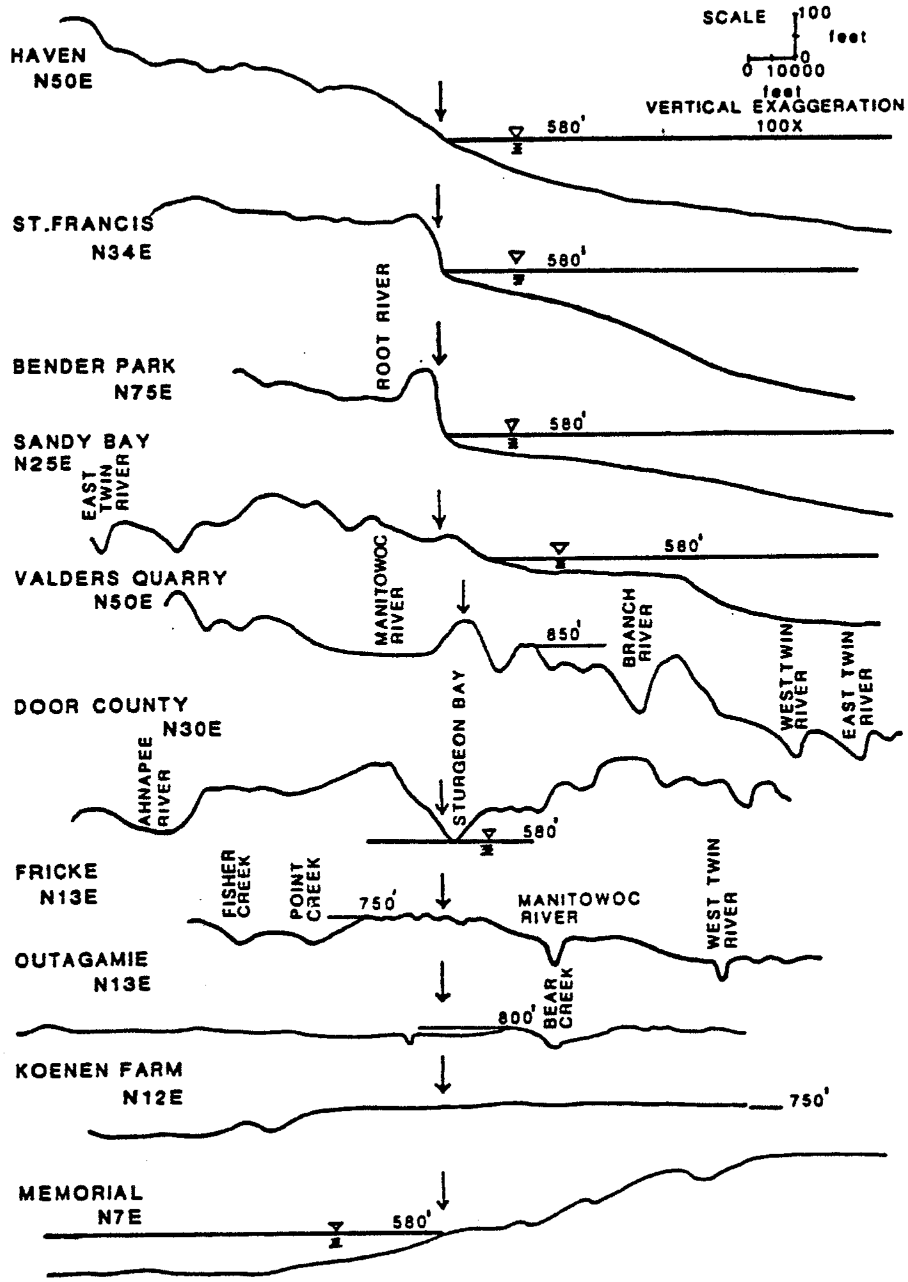


Figure 35b: Topographic profiles along ice flow direction. Ice flow from right to left.

The most obvious inconsistency between this model and the field evidence is the lack of any structures on the vertical joint surfaces indicative of shear during failure. Evidence of offset was also rare, and was found only in laminated lacustrine sediment overridden by ice. The lack of evidence for shear may be the result of modification during weathering. It is also possible that the measured joints may have inherited the orientation, but not joint surface chracteristics, of joints formed beneath the ice. The angle between the conjugate joints and the ice flow direction (presumed to represent the direction of greatest horizontal stress) is typically near 45 degrees, but the angle predicted by Mohr-Coloumb failure with an effective angle of internal friction (ϕ') of 30 degrees is 45 + $\phi'/2$, or 60 degrees. The angle between the conjugate joint and the ice flow may be less than predicted from the Mohr-Coloumb equation because the effective angle of internal friction represents consolidated undrained (CU) conditions. The actual deformation, however, may have been similar to unconsolidated undrained (UU) conditions.

While field evidence exists for compressive flow at many of the sites, some of these sites are located below the Glenwood level of Glacial Lake Chicago, or may have been beneath the lake surface of Glacial Lake Oshkosh.

Water depth in these lakes may have been sufficient to bouy the glacier margin, thus causing extending flow. No differences were detected between jointing formed above the Glenwood level from the jointing formed below lake level.

2. Failure due to subsequent release of locked stress

This model represents a slight modification of the conditions presented in Model 1. A grounded, temperate or polar glacier under compressive flow may have set up high lateral earth pressure in the glacier bed, as suggested by Lafluer (1980). This pressure may not have been sufficient to cause failure at the high confining stresses existing in back of the glacier margin. However, as the ice receeded, the confining pressure declined and the remnant earth pressure, being greater than the shear strength at lower confining pressure, could have caused the sediment to fail.

Lafluer's original model considered only vertical compaction. Under a vertical greatest compressive stress, fractures forming due to compaction or release of built up stress would have an dip of $45 + \phi'/2$ to $45 - \phi'/2$, or in the range of 30 to 60 degrees for most of the units examined. Joints with this orientation were rarely found. However, if the greatest compressive stress were horizontal, as could occur during compressive flow, the vertical joints could represent failure due to high remnant earth pressure.

3. Tensional failure while frozen

In this model, a grounded or semi-grounded glacier in extending flow, undergoing limited sliding, could create tension in frozen till beneath the glacier. Failure may have occured through tertiary creep at low strain rates because of the high strength of the frozen till, causing a rearrangment of the fabric. Elevated pore pressures would not be expected under these circumstances (Mathews and Mackay, 1960), therefore deformation would be limited to the ice margin where the confining stress is low. The frozen till/ice mixture would have to have to be composed largely of ice if it were to become as weak as the clean ice, which would limit any consolidation. Preservation of the tension cracks could occur if they filled with segregation ice, which could also reorient the fabric of the failure zones and perhaps persist while the till/ice mixture thawed after deglaciation.

The lack of shear structures on vertical joints support a tensional-type fracture genesis, but evidence for tensional failure, such as high asperity (Sleeman, 1963) is also lacking. Slickensided near-horizontal joints could be formed under these conditions, but their preservation through thawing is problematic unless ice lenses protected them from modification.

4. Hydraulic fracturing

This model requires a rather stringent set of conditions, but it is included to illustrate one of the many other possibilities that could not be considered. Creation of vertical joint planes by hydraulic fracturing could have taken place when the effective stress was less than the total stress as a result of elevated pore pressure. The model developed by Poorooshasb and Yong (1983) requires that the pore pressures drain only laterally, which could have occured in the active layer of a permafrost zone if overridden by ice, or if the groundwater discharge area was frozen or ice covered. Loading must have been applied in large steps so that the pore pressure did not have time to dissipate.

These models illustrate the complexity of subglacial deformation; it is difficult to find overwhelming evidence to support one over another. Near-vertical joints parallel to ice flow were seldom found together with low angle joints, which suggests that the two may form under different stress regimes. Low-angle joints were associated with concave up profiles (Figure 36) and

CONCAVE PROFILE	PERPENDICULAR MICROFABRIC	UPHILL PROFILE	PERPENDICULAR JOINTS	LOW ANGLE JOINTS	CONJUGATE JOINTS	LEVEL PROFILE	PARALLEL JOINTS	PARALLEL MICROFABRIC	DOWNHILL PROFILE	CONVEX PROFILE	SITE	STRATIGRAPHIC UNIT
X	X	X			X						21	OZAUKEE
X		X	X		X						31	DOUGLAS
X	X	X	X	X	X						33	DOUGLAS
X	X	X	X		X						32	DOUGLAS
X	X	X	X	X	X						34	HANSON CR
X	X		X		X	X					36-2	MIDDLE INLET
X	X	X	X	X	X		X				20-1	VALDERS
X		X	X	X	X		X				7	TWO RIVERS
X	X	X		X	X		X				22	OZAUKEE
X		X	X		X			X			20-2	HAVEN
X		X		X	X			X			35	HANSON CR
X			X	X	X	X	X	X			3	GLENMORE
	X	X	X		X		X			X	18	VALDERS
	X	X	X	X	X					X	8	TWO RIVERS
		X	X	X	X			X		X	12	TWO RIVERS
	X	X			X		X			X	26	2B
		X	X		X		X	X		X	29	2B
		X	X		X			X		X	15	VALDERS
		X			X			X		X	13	HAVEN
	X				X	X	X			X	37	MIDDLE INLET
X					X		X	X	X		17	HAVEN

Figure 36: Matrix of flow regime indicators and associated joint set orientation. Compressive flow regime is associated with features on left side of diagram, extending flow is associated with the features on the right.

perpendicular sand grain orienation, both of which suggest compressive flow. Flow-parallel vertical joints were associated with parallel sand grain orientation and convex profiles, both indicative of extending flow. Changes in flow regime through time may have occured because of fluctuating lake levels in Glacial Lake Chicago or differences in flow direction as ice moved radially out of and then along the Lake Michigan Basin.

It is also likely that the temperature regime was changing during the time period represented by the till units. The older units, such as the New Berlin and Oak Creek Formations, were likely deposited under frozen-bed conditions, while the younger Kewaunee and Miller Creek Formations may have been deposited under wet-based ice. Seasonal changes that may have occured in the thermal regime complicate the situation, making prediction of the thermal regime difficult. It is interesting to note that evidence of high pore pressure and subglacial deformation of coarse-grained sediment was only observed in association with the younger Valders, Two Rivers, and Glenmore tills. Joints in these units were not as long or as widely spaced as in some of the older units, but the influence of the lithology differences cannot be removed.

To conclude this discussion on joint genesis, other models for joint formation are considered. The models that follow are not supported as well by the field evidence, but the possibility that they occured in concert with subglacial deformation can not be eliminated.

STRESS FROM ISOSTATIC UPLIFT

During and shortly after deglaciation, crustal response to the weight of the ice sheet is at a maximum (for example Mörner, 1978). Tension created by differential isostatic uplift has been proposed as the cause of jointing in glaciolacustrine sediment and till in the northern Great Plains (Grisak and Cherry, 1976).

Isostatic uplift seems to be an unlikely explanation for the jointing observed in the fine-grained till in Wisconsin, because these till units are the youngest in the state and have experienced a relatively small amount of rebound. Quantification of the amount of rebound is not possible as there is no datum against which to measure absolute elevation changes, but warping of remnant shorelines in the Lake Michigan basin provides an indication of differential uplift. The shorelines also provide a time record of when the uplift took place. Evenson (1973, figure 3) presented isobases for the Glenwood level of Glacial Lake Chicago that indicate 6.1 m

(20 ft) of differential movement between Sheboygan and Two Rivers. The post-Two Rivers Algonquin stage shoreline is tilted approximately 5 m in the distance between Racine to Sturgeon Bay (Wolcott, 1972, figure 13). The younger Nipissing (approximately 3900 y.b.p.) and Algoma (approximately 3200 y.b.p.) shorelines have been considered to have undergone no differential uplift (Evenson, 1973). This assumption is challenged by the recent work of Larsen (in press, figure 4) which indicates that differential uplift from Racine to Sturgeon Bay have been as high as 5 m for the Nipissing II shoreline and 3m for the Algoma. The differential uplift experienced by the youngest of the Lake Michigan Lobe tills is nearly an order of magnitude less than that reported for the till and lacustrine sediment of the northern Interior Plains (Grisak and Cherry, 1976).

The response of unlithified surficial deposits to stress resulting from isostatic rebound is unknown. While these stresses are capable of causing siesmic activity, particularily along pre-existing bedrock faults (Stephansson and Carlsson, 1980), it is questionable if the stress causes deformation of the surfical materials. It seems more likely that adjustment takes place between large blocks of basement rock.

If the pattern of joints in surficial materials were controlled by regional deformation stemming from isostatic uplift, one would expect that the joint pattern would be consistent over a large area. Joint orientation from site to site is relatively consistent in a single stratigraphic unit, but is dissimilar from unit to unit. My conclusion is that isostatic uplift was probably a minor factor in the genesis of joints in Wisconsin till.

WEATHERING

Chemical change, such as loss of cations from clays or intense leaching of carbonates (Chpater 1), is also unlikely to have caused the formation of joints in Wisconsin till except in the uppermost horizons. The sediments were deposited in a fresh water environment (either subaqueously or subaerially) and their chemistry and pore water composition probably has not changed significantly since deposition.

Dessication is probably the largest factor influencing joint length, as shown by the relationship between maximum length and clay content. However, it is problematic to explain how dessication could be responsible for joint formation in exposures that are below the water table. Water table levels may not have been constant, especially along the Great Lakes, where isostatic depression lowered the northern outlet of Lake Michigan and Lake Superior during the Holocene. Soderman and Kim (1970) have proposed that lowered lake levels were responsible for dessication and formation of a jointed, overconsolidated crust on the clay till and lacustrine sediment of the Lake St. Clair Plain. It is also possible that inland ground water levels were lower during drier periods during the Holocene, which may have led to dessication at greater depths than occurs at present.

One factor which argues against dessication as being the sole cause of fracture genesis in the Wisconsin till units is the consistent preferred orientation of high-angle joints. Joints due to shrink and swell of clay soils are typically oblique to the ground surface (Birkland, 1974). It also seems likely that joints formed solely by dessication would have a random orientation. As the majority of the joints are neither oblique or random, dessication alone is probably not responsible for the formation of the joint sets.

PREDICTIBILIY

Fine-grained till in Wisconsin occupies a limited position in the regional landscape, restricted to the areas near the Great Lakes. The till units have a limited range of textural, mineralogical, and geotechnical properties. Because of these limitations, it is difficult to develop a predictive model for jointing in fine-grained till for other areas of the country based solely on the field data collected in Wisconsin. The data from this study needs to be combined with data from other research in order to be able to predict the presence of jointing on the basis of the physical properties of the tills. This is the goal of Chapter 4.

FURTHER RESEARCH

At present, shelby tube sampling provides the only method of detecting joints without the expense of digging test pits or other excavations. While this method has been successful in some instances (Johnson and others, 1983; Olimpio, 1982), application of geophysical techniques may improve the detection of joints without the unloading and possible propagation that would accompany excavation. Geophysical resistivity methods in particular may prove to be of great value in determining the intensity of jointing and in attempting to characterize the joint pattern.

The relationship between joint pattern and permeabilty is currently unknown, although progress is being made in predicting fracture porosity and permeabilty (Sledz and Hall, 1981). Models of two dimensional fracture permeability have been developed (Long and others, 1982), but the application of numerical models of fracture flow to field problems is hampered by a lack of adequate field data on the fracture characteristics and the difficulty of accurately verifying the models' predictions in the field. More research is needed to relate jointing in fine-grained till to permeabilty and to discover how the jointing influences anisotropy of permeabilty and strength.

CONCLUSIONS

Dominant genetic processes

The consistent orientation of joints with ice-flow direction, associated deformation of more competent underlying strata, and lack of intense weathering support a glaciotectonic origin for the joints in Wisconsin till. This evidence is augmented by the observation that joint spacing is proportional to the thickness of the till unit, a relationship common for joints thought to be formed by deformation in rock. Conditions of high pore pressure and/or low confining stress, as evidenced by fluid escape structures and ductile deformation of lacustrine silt at the Two Rivers site (Site #11), may have led to the deformation of frozen or unfrozen fine-grained sediment at high water or ice contents.

The joints may have formed as discrete fractures during brittle failure, or may have originally been zones of weakness formed under ductile deformation. The glaciotectonic joints or failure zones, by causing a rearrangement of the fabric, could then influence the orientation of subsequent failure caused by dessication or unloading. They may also have propagated upward into fine-grained sediment that had not been overridden by ice.

Evidence for joint formation because of isostatic uplift, earth tides, or regional tectonic activity is not sufficient to determine what influence, if any, these processes had on the joint characteristics. The area of the Michigan shoreline covered by this study has undergone at least 6 m of differential isostatic uplift, as indicated by tilting of the Glacial Lake Chicago shorelines, a relatively small amount of deformation over such a large area. The changes in joint orientation from site to site as ice-flow direction changes suggests that the joints are not caused by processes operating on a regional scale but are formed by local stress.

Dessication was probably an important factor in modifying joint length, and perhaps joint spacing. This is indicated by the positive relationship between joint length and clay content, and the fact that maximum ped size was observed to increase with depth. However, joints were found below the water table, and so may reflect processes other than dessication. While it is likely that ground water levels were lower at times during the Holocene than they are at present, the consistent preferred orientation of joint azimuth and the lack of obliquely dipping joints indicate that dessication alone did not cause the joints to form.

Jointing in the till units studied extends below the pedogenic horizons of leaching and translocated clay (the solum). Thus the joints probably preceeded pedogenic soil structure development, which typically masked joints in the uppermost horizons of the exposures. Joint surface weathering may have altered the joint characteristics, either through deposition of weathering products, leaching, or reduction (gleying). Freezing and thawing probably have had little effect on the joints at depth.

Chapter 4:

The Importance of Joints as Siting Criteria for Low-level Radioactive and Hazardous Waste Disposal Facilities in Fine-grained Glacial Sediment

INTRODUCTION

Fine-grained till and glaciolacustrine sediment have wide distribution in the glaciated regions of North America (Figure 1). Fractures have been observed to influence the hydrologic behavior of fine-grained unlithified sediment and have been found at operating waste disposal facilities and low-level radioactive waste disposal sites. Because fractures reduce the suitability of fine-grained glacial sediment for the disposal of hazardous or low-level radioactive wastes their identification is important during site selection and design. Fractured fine-grained sediment differs from unfractured sediment in that it typically has higher permeability, lower shear strength, and may have significant anistropy of strength and permeability. This anisotropy can limit the accuracy of characterization of fractured silt and clay, making modeling of flow systems or contaminant migration difficult.

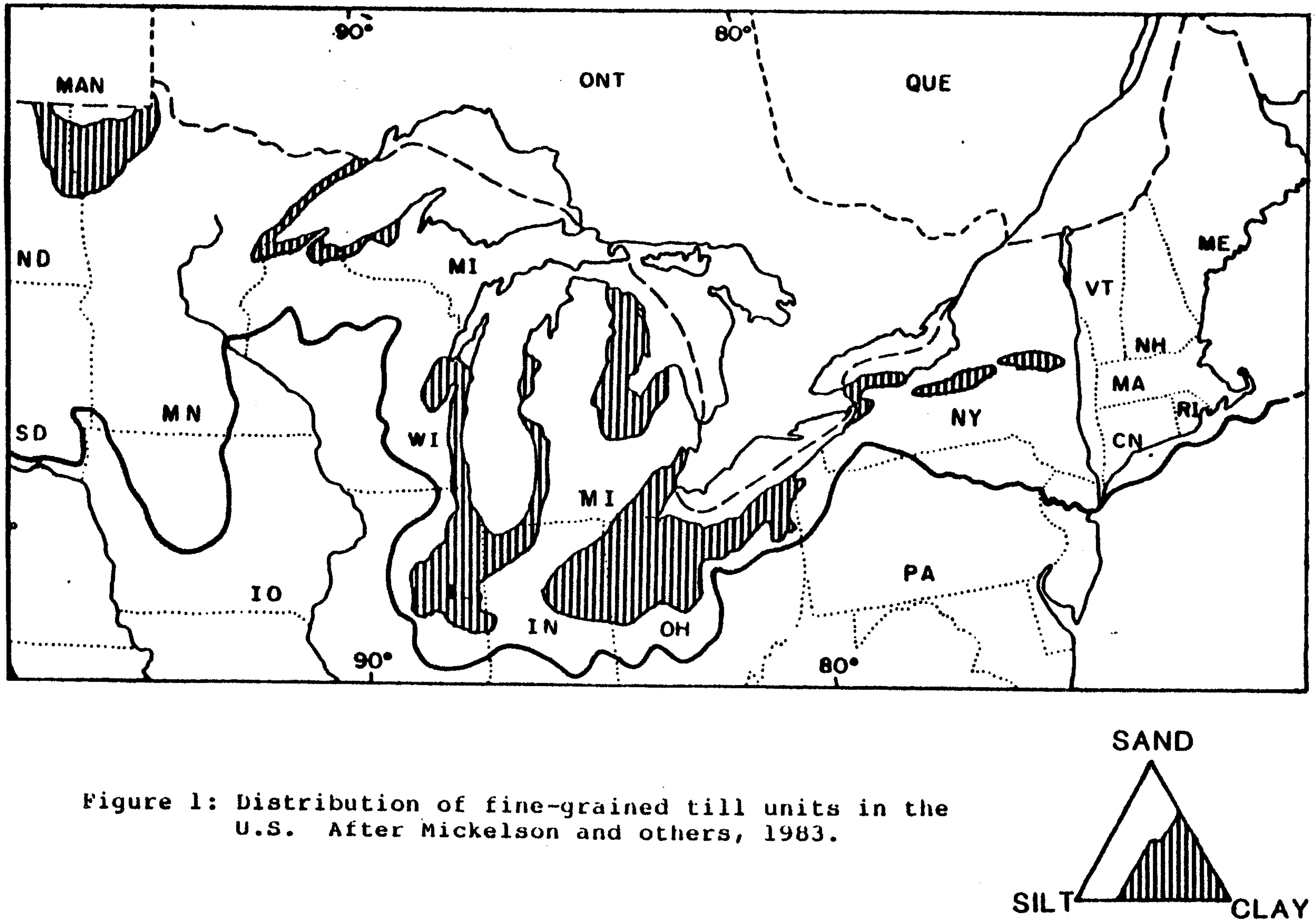


Figure 1: Distribution of fine-grained till units in the U.S. After Mickelson and others, 1983.

This paper considers the effects of fractures on the physical behavior of fine-grained glacial sediment, the implications of fracture-controlled behavior, and examines the factors that may control fracture genesis. The goal of this paper is to present a set of criteria for evaluating the potential for fracture development in fine-grained glacial sediment, based on a review of the literature and on the results of a study of fractured till in eastern and northwestern Wisconsin (Chapter 3).

INFLUENCE OF FRACTURES IN FINE-GRAINED SEDIMENT

HYDROLOGIC CHARACTERISTICS

The hydrologic behavior of clay till and other fine-grained sediment has not been studied as extensively as other water bearing materials because they are not capable of yielding significant amounts of water to wells. Instead, it is usually assumed that their low permeability makes them suitable for waste disposal, and that they act as protective caps over more permeable aquifers (Desaulnier and others, 1981). However, this permeability can be increased and "...ground water in aquifers overlain by glacial till in which waste disposal sites are located may become polluted if the till contains joints" (Williams and Farvolden, 1967).

Hydraulic Conductivity

Fractures in clay till have been observed to increase the hydraulic conductivity significantly above the conductivity of unfractured deposits or unfractured blocks of till. Hydraulic conductivity measurements from piezometer tests of fractured till were up to one thousand times greater than laboratory permeameter measurements in a study at a landfill in eastern Wisconsin (Gordon and Huebner, 1983). Sterrett and Edil (1982) found that slug tests performed in a piezometer installed in fractured till gave values of hydraulic conductivity three orders of magnitude higher than piezometers finished in unfractured till, but cautioned that their conclusions were tentative. Wells intersecting fractures in till respond more rapidly to recharge events than wells isolated from the fracture system (Williams and Farvolden, 1967; Sterrett and Edil, 1982).

Because fractures may open and close due to changes in moisture content or overburden pressure (McGown and Radwan, 1975), their ability to transmit water when closed is uncertain. Anderson and others (1982) found that dessication fractures that were visible only in X-radiographs created elevated permeability in a clay roadway embankment. This topic is treated in further detail in Chapter 2.

The most thorough study to date of the hydrogeologic behavior of fractured till is the work of Grisak and his coworkers (1975; 1976) at Whiteshell Nuclear Research Establishment (WNRE) in southeastern Manitoba. Fractures were found to be the dominant factor in controlling both hydraulic conductivity and specific storativity in a section of clay loam till and overlying lacustrine silt and clay. Bulk hydraulic conductivity of 1.8 X 10-7 cm/sec, as determined through the calibration of a finite element flow model of the site, was greater than that calculated from laboratory tests of undisturbed samples by one to two orders of magnitude. Analysis of a pumping test in an underlying sand artesian aquifer showed that some of the piezometers in the overlying till exhibited rapid response. This rapid response was thought to reflect a hydraulic connection between the piezometers and the fracture system (Grisak and Cherry, 1975). Specific storativity values for the affected piezometers were smaller than the intergranular storativity values measured by lab testing, and were typical of values observed in fractured rock.

From tracer tests at the WNRE site, Grisak (1975) calculated that fracture porosity controlled the dispersion of the tracer plume. The fracture porosity was approximately .0002, smaller by several orders of magnitude than the intergranular porosity of .25 to .4 estimated from laboratory tests.

Anisotropy of Permeability

When fractures cause increased permeability, they can also introduce anisotropy of permeabiltity, making characterization and modeling more complex. Strongly anisotropic systems may lack porous media equivalents, as suggested by the work of Long and her coworkers (1982). Long found that flow systems in fractured media behave less like porous media when fractures are of finite length, anisotropic orientation, low density, and nonuniform hydraulic aperature.

There is little information in the literature upon which to evaluate the effects of fractures on anisotropic permeability in clay till. McKinley and others (1975) noted that large diameter oriented samples of a fractured till showed anisotropy of permeability of up to thirty-six times when tested in large-diameter oedometers. On a larger scale, however, Prudic (1982) found that fractured till behaved isotropically in a groundwater flow model of the West Valley site in upstate New York.

Hydrogeochemical Effects

By decreasing the surface area exposed to that portion of the pore fluid flowing preferentially through the joints, jointing can restrict the ability of fractured sediment to react with contaminants in groundwater. An analogous effect occurs in natural systems in the form of oxidation along joints extending below the generally oxidized surface horizons (Williams and Farvolden, 1967). However, solute transport through fractured clay till differs from that of fractured impermeable media because there is some diffusion into the matrix from the fracture face, allowing the pore water and matrix materials to interact with a contaminant (Grisak and Pickens, 1980). Diffusion into the matrix is low when fracture aperatures are wide, fracture fluid velocities high, or the matrix pore volume low compared to the fracture volume (Grisak and Pickens, 1980).

GEOTECHNICAL CHARACTERISTICS

In addition to affecting the hydrogeological behavior of fine-grained glacial sediment, fractures can have significant effects on the material's geotechnical properties, and may introduce large error in predictions of strength and compressibility from laboratory tests.

In general, the in itu shear strength of fractured clay and clay till is lower than that of intact lab specimens (Terzaghi, 1936); drainage path lengths are shorter and thus consolidation rates faster than obtained from lab tests; and fractured clay exhibits very little tensile strength (Jennings, 1975). Fractured clay till has created significant problems during construction (e.g. Eisenstein and Thomson, 1978).

Strength Parameters

Kaczynski and Wyokinski (1970) found that samples of fractured lacustrine clay and clayey sand till had less strength than unfractured samples when tested in triaxial apparatus with fractures oriented oblique to the direction of deviatory stress. Their results indicated that unfractured clayey sand till was 1.35 times as strong as fractured clay till; for lacustrine clay the ratio was 1.5. McKinley and others (1975) calculated from triaxial testing that fractures decreased the strength of till by approximately eighty percent.

In addition to decreasing the bulk strength, fractures with preferred orientation may introduce anistropy of strength. McGown and Radwan (1975) report anisotropy of strength of up to fourty percent in triaxial tests of large diameter oriented samples.

Recognition in Subsurface

Fractures are not easily detected in the subsurface through common sampling techniques, but their presence can be indicated through careful collection of undisturbed samples (Olimpio, 1982). Their presence can also be indicated through indirect evidence such as anomalously high _in situ_ permeability test results, as at Outagamie Landfill in Wisconsin (Gordon and Huebner, 1983). Surface exposures or excavations are usually necessary to confirm the presence of fractures and are required for detailed study of their orientation and other characteristics (Chapter 2).

Representative Sampling

In order for meaningful results to be obtained from laboratory tests of fractured till and clay, undisturbed samples must be large enough to be representative of the material. Eisenstein and Thompson (1978) stress the importance of considering the _in situ_ behavior of large volumes of till rather than extrapolating from tests done on small samples; the measured strength of fractured till can be strongly dependent on the volume sampled (Pusch, 1973). McKinlay and others (1975) recommend that a block with dimensions twenty times the minimum fracture spacing be used to obtain representative strength measurements: if fracture spacing increased with depth a larger block would be needed to accurately characterize the material. Similarily, equidimensional blocks five times larger than the fracture spacing are required for accurate determination of permeability. McKinlay and others (1975) also stress the importance of testing oriented samples in order to evaluate any anistropy of strength or permeability that may exist.

Slope Stability

Fractures can have significant effects on slope stability through their influence on permeability and strength. Sterrett and Edil (1982) found that slope stability was affected by a perched water table developed in the fractured upper portion of a till, because "...water in the fractures will cause failure of the upper bluff as a sliding block when there is a critical depth of water". Their conclusions support those of Esu (1966), who found that the main cause of slope failure in a jointed overconsolidated clay was the hydrostatic pressure within the joints. Perched water table conditions associated with dessication fractures were also found to cause surficial slips in a clay embankment in southern England (Anderson and others, 1982).

The influence of fractures on the strength of a till was found by McGown and Radwan (1975) to decrease slope stability. The effect of the fractures was to cause slope failure in excavations when the cut was parallel to the fractures.

IMPLICATION OF FRACTURES IN FINE-TEXTURED SEDIMENT

In order to provide for isolation of wastes from groundwater flow systems, hazardous waste and low-level radioactive waste disposal sites in humid regions of the United States have typically been located in fine-grained unlithified sediment. Several of these sites been found to be underlain by fractured till and are discussed below.

FRACTURED TILL AND WASTE DISPOSAL

The potential problems of waste disposal associated with fractured till are suggested by examples of sites constructed without complete characterization of the materials at the site. Fractures in till at a zone-of-saturation landfill in Outagamie County, Wisconsin were first detected from large discrepencies between *in situ* and laboratory permeability tests (Gordon and Huebner, 1983). Subsequent excavation showed that large vertical fractures extended through lacustrine clay and clay till to a depth of 5.7 m, prompting the state regulatory agency to require excavation and recompaction of the floor and side walls of the affected cells.

Fracture flow, as a factor in the migration of organic chemicals from hazardous waste disposal sites, is one of the areas being evaluated by an interdisiplinary team from the Illinois Geological Survey at a hazardous waste site near Wilsonville, Illinois (Johnson and others, 1983). Preliminary indications are that a fractured zone in weathered basal Vandalia Till associated with a buried Sangamon soil has in situ permeabilities of approximately 2.3 X 10-6 cm/sec, one to two orders of magnitude larger than that of the unfractured till beneath it, and up to two orders of magnitude larger than remoulded samples tested in permeameters. Differences were also found between the results of slug tests in vertical and angled boreholes (Herzog and Morse, 1984). Average values of permeability were one order of magnitude higher in the angled boreholes (1.15 X 10-5) when the slug tests were analyzed by the method of Nyugen and Pinder (1984). Herzog and Morse tentatively attributed this to the intersection of more vertical fractures by the angled boreholes.

Jointed till has also been encountered at three low-level radioactive waste sites. The upper 4.5 m of till at the site of the low-level radioactive waste burial trenches at West Valley, New York is fractured and the fractures are thought to influence the permeability of the till (Prudic, 1982). Slug tests showed the hydraulic conductivity of the jointed till to range from that of the unjointed till to values two orders of magnitude higher. Elevated hydraulic conductivity was also suggested by a groundwater flow model of the site, the results of which indicated that the jointed oxidized till was ten times more permeable than the underlying unjointed till (Prudic, 1981). Analysis of fracture coatings has not shown any evidence for lateral migration of radionuclides along the fractures (Prudic and Randall, 1979), perhaps due to the strong vertical gradient at the site.

The low-level radioactive waste site at the Palos Forest Preserve in northwestern Illinois is located in fractured till. A tritium-contaminated plume has been released from the burial trenches and is thought to have contaminated a well finished in the bedrock 360 m away (Olimpio, 1982). Fractures have affected the hydrology of the site by increasing infiltration and causing lateral migration away from the trenches, but the main mass of the tritiated plume is moving downwards under the influence of a vertical gradient. Channelized flow through fractures and sand lenses is limited to the uppermost horizons of the till.

Fractures have been observed in the tills at the Sheffield site in northern Illinois but most of the investigations there focus on other aspects of the geology and hydrogeology (John N. Fischer, USGS, personal communication, 1984). Joints in the weathered shale at the Maxey Flats site in Kentucky are presenting problems for site closure (Zehner, 1983).

IDEALIZED SITING CRITERIA

Although there are no regulations specifically governing the siting of low-level radioactive waste disposal facilities in fractured unlithified sediment, the complexity of flow in some fracture systems increases the difficulty of characterizing disposal sites. Some feel that increasing effort should be made to site new facilities in "... relatively simple geohydrologic envrionments to facilitate the prediction of their containment properties" (Robertson, 1984). Papadopulos and Winograd (1974) state that "...dense fractured or soluble media, and poorly permeable porous media (aquitards) are not suitable for use as burial sites... ...because of media heterogeniety and difficulties of sampling, and consequently of predictive modeling...". The quest for a simple hydrogeologic setting is reflected in policy statements of the Nuclear Regulatory Commission (U.S. NRC, 1981) and has been included in recent federal legislation. The Licensing Requirements for Land Disposal of Radioactive Wastes (10 CFR 61) state that "The disposal site shall be capable of being characterized, modeled, analyzed, and monitored".

Current capabilities for modeling contaminant flow in fractured silt and clay is poor. University of Waterloo Professor John Cherry stated at a recent conference that prediction of contaminant movement through fractured fine-textured sediment is extremely difficult (Gass, 1984). As the success of modeling efforts depend on field calibration, the problems of sampling in fractured sediment can limit the level of characterization and calibration that can be acheived (e.g. Roberts and others, 1982).

Given the complexities introduced by fractures in unlithified sediment and the trend toward simple hydrogeologic settings for waste disposal siting, it is advantageous to be able to predict the presence of fracture systems, so that they can be anticipated when necessary or avoided if possible.

DISTRIBUTION OF JOINTED TILL AND LACUSTRINE CLAY

Development of predictive capability for fractures in unlithified sediment depends on previous observations of fractured material on which to base models of fracture genesis. The study described in Chapter 3 provides this data base and presents several theories for the genesis of the jointing observed. However, as the study in Chapter 3 was restricted to a limited geographic area and small number of till units, other occurances of jointed fine-grained till and lacustrine clay were reviewed in order to provide a larger data base. The results of this review are presented in Appendix 7.

Investigation of fractures in unlithified sediment first appeared in the literature in a paper by Karl Terzaghi (1936), who described the importance of fractures in controlling the strength of overconsolidated clays. Little was done after his initial work until Fookes (1965) described an investigation of fracture orientation in the Pliocene-Pleistocene Siwalik Clay of Pakistan. Esu (1966) investigated slope stabilty in overconsolidated lacustrine and marine Pliocene-Pleistocene clay in central and southern Italy, and observed that jointed clay acts as a discontinuous anisotropic medium similar in behavior to jointed rock. These initial investigations set the trend for the study of fractured, fine-grained, glacial and nonglacial sediments in Europe and North America. Many workers have since observed fractures in fine-grained till and glaciolacustrine sediment: their observations and interpretations are briefly summarized in Appendix 7. Much of the information available on jointed till was collected in the course of other studies; a relatively small number of workers have concentrated their investigations on fracture genesis or hydrogeologic behavior.

Fractured till deposits and glaciolacustrine sediment are found in several areas of the northern Great Plains. These tills are predominantly fine-grained owing to their derivation from Mesozoic sedimentary rocks. The glaciolacustine sediment is typically post glacial, resulting from depostion in large glacial lakes such as Glacial Lake Agassiz (Figure 1).

Jointed till units also occur in much of the region surrounding the Great Lakes. Tills in this area were derived from Paleozoic shale and from reworked lacustrine and marine sediment.

Several groups of workers have been active in investigating fractured till and clay in the United Kingdom. Modern research on jointed sediment began with studies of jointing in the Eocene marine London Clay of southeast England (Skempton and others, 1969; Fookes and Parrish, 1969).

PREDICTING JOINTS IN TILL

As has been stressed above, the effects of fractures on the hydrologic and geotechnical behavior of till makes it important that fractures are recognized before waste disposal facility construction begins. As the methods of fracture detection are expensive and inexact, the development of predictive models of fracture distribution is desireable. While it is a goal of this paper to present some components of a predictive model, I must state at the outset that my attempts have not been totally successful. The polygenetic nature of fracturing coupled with uncertainties about fracture genesis make simple interpretation difficult, let alone prediction. However, common characteristics of fractured fine-grained sediment do exist, and these merit discussion.

FACTORS AFFECTING FRACTURE DISTRIBUTION

Physical characteristics

Fracture distribution is related to the physical characteristics of the materials. Study of fractured till in Wisconsin (Chapter 3) shows that texture exerts the greatest control on fracture length and spacing; clay mineralogy and carbonate content have a much smaller effect.

Other factors not found in till units in Wisconsin, such as high organic content or sulphide oxidation, or the development of sensitivity (i.e. in clays of marine origin that are now exposed to fresh water) may have greater effects than that of texture in some situations.

lithology

The lithology of fractured fine-grained sediment appears to exert some control on fracture distribution. Fractures have been reported from a wide variety of fine-grained sediment ranging from marine clay to till to weathered shale (for references, see Appendix 7). Fookes and Denness (1969) found in studying fractures in a variety of sediment types that the influence of lithology was "...one of scale rather than of pattern" except in widely differing materials such as chalk versus clay. Joint size increased and the intensity of fracturing decreased with increasing "brittleness" (Fookes and Denness, 1969).

texture

For sediment of mixed lithology such as till, texture seems to play an important role in fracture development. Figure 2 summarizes the grain size distribution of fractured till and clay as found in the Wisconsin study (Chapter 3) and as reported by a number of other authors

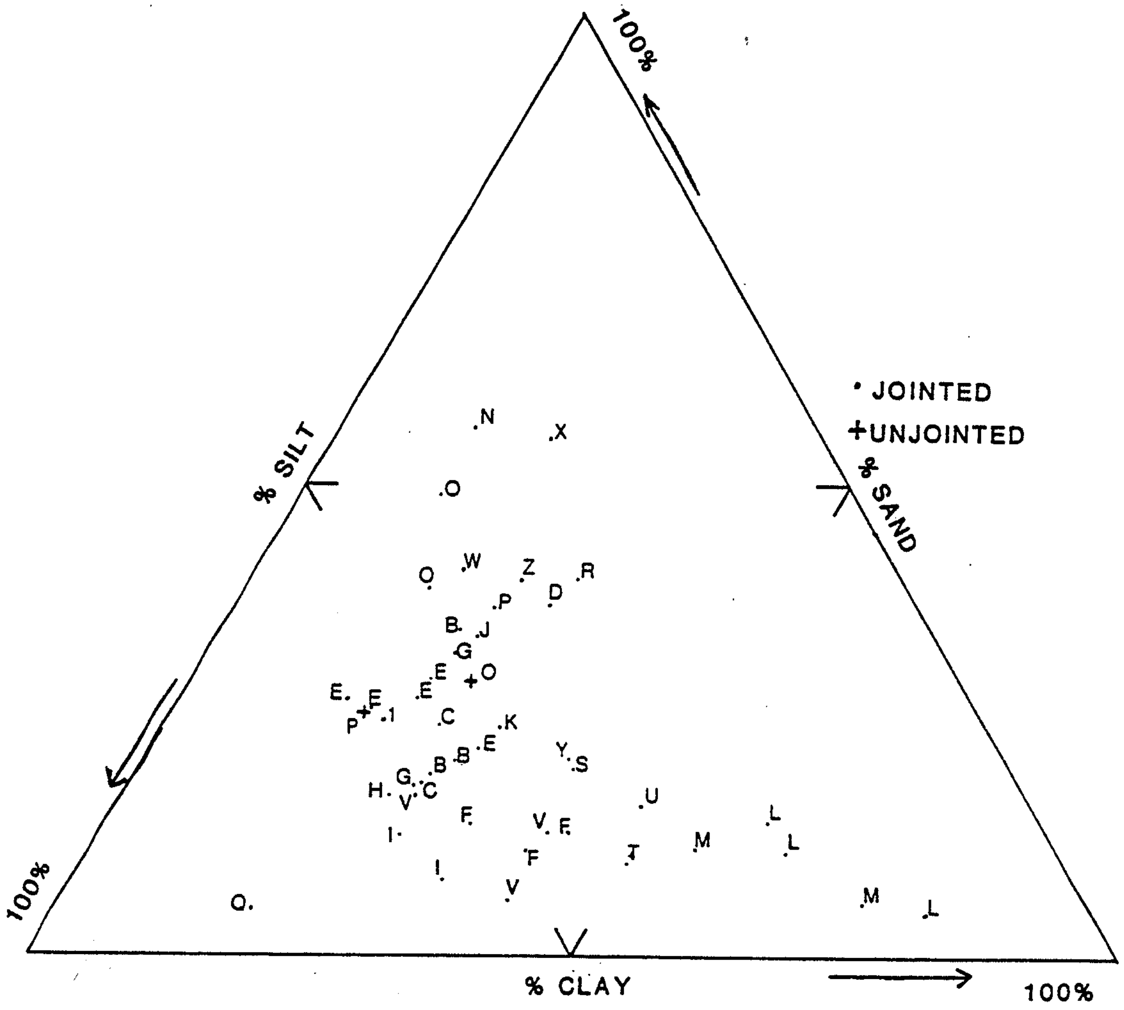


Figure 2: Sand-Silt-Clay diagram of fine-grained till units. Letters and numbers refer to references in Table 1.

Table 1: References of fractured till and clay

	Sediment type	Location	Reference
A	Lacustrine sediment	Wisconsin	This study (Ch. 3)
B	Two Rivers till	"	"
C	Middle Inlet till	"	"
D	Glenmore till	"	"
E	Valders till	"	"
F	Haven till	"	"
G	Ozaukee till	"	"
H	Oak Creek till (2A)	"	"
I	Oak Creek till (2B)	"	"
J	Oak Creek till (2C)	"	"
K	New Berlin till	"	"
L	Douglas till	"	"
M	Hanson Creek till	"	"
N	"Tazewell" till	North central Iowa	"
O	"Des Moines" till	North central Iowa	"
P	till	Maine	"
Q	lacustrine silt	Glacial Lake Drumheller, Alberta	Babcock, 1977
S	till	Welland Canal, south east Ontario	Owen, 1971
T	Black Shale till	Sarnia, Ontario	Soderman and Kim, 1970
U	Lavery till	West Valley, N.Y.	LaFluer, 1978
V	Wadsworth Till	Palos Forest Reserve, NE Illinois	Olimpio, 1982
W	Vandalia Till	Wilsonville, Illinois	Johnson, et al., 1983
X	Lodgement till	Hurlford, Scotland	McGown, et al., 1974
Y	weathered till	Northumberland, England	Eyles and Sladen, 1981
Z	unweathered till	"	"
1	till	south central sweden	Mickelson, et al., 1981
2	fractured till		Boulton and Paul, 1976
3	London Clay	Wraysbury, England	Skempton, et al., 1969
4	disturbed clay	south east England	Anderson, et al., 1982
5	glaciolacustrine clay	Norfolk, England	Kazi and Knill, 1973
6	Krakoweic Clay	Machow, Poland	Kaczynski and Wyokinski, 1970
7	Poznan Clay	Debrzyn, Poland	"
8	"Middle Poland Till"	Plock, Poland	"
9	Valdarno Clay	Southern Italy	Esu, 1966
10	Ancona Clay	"	"

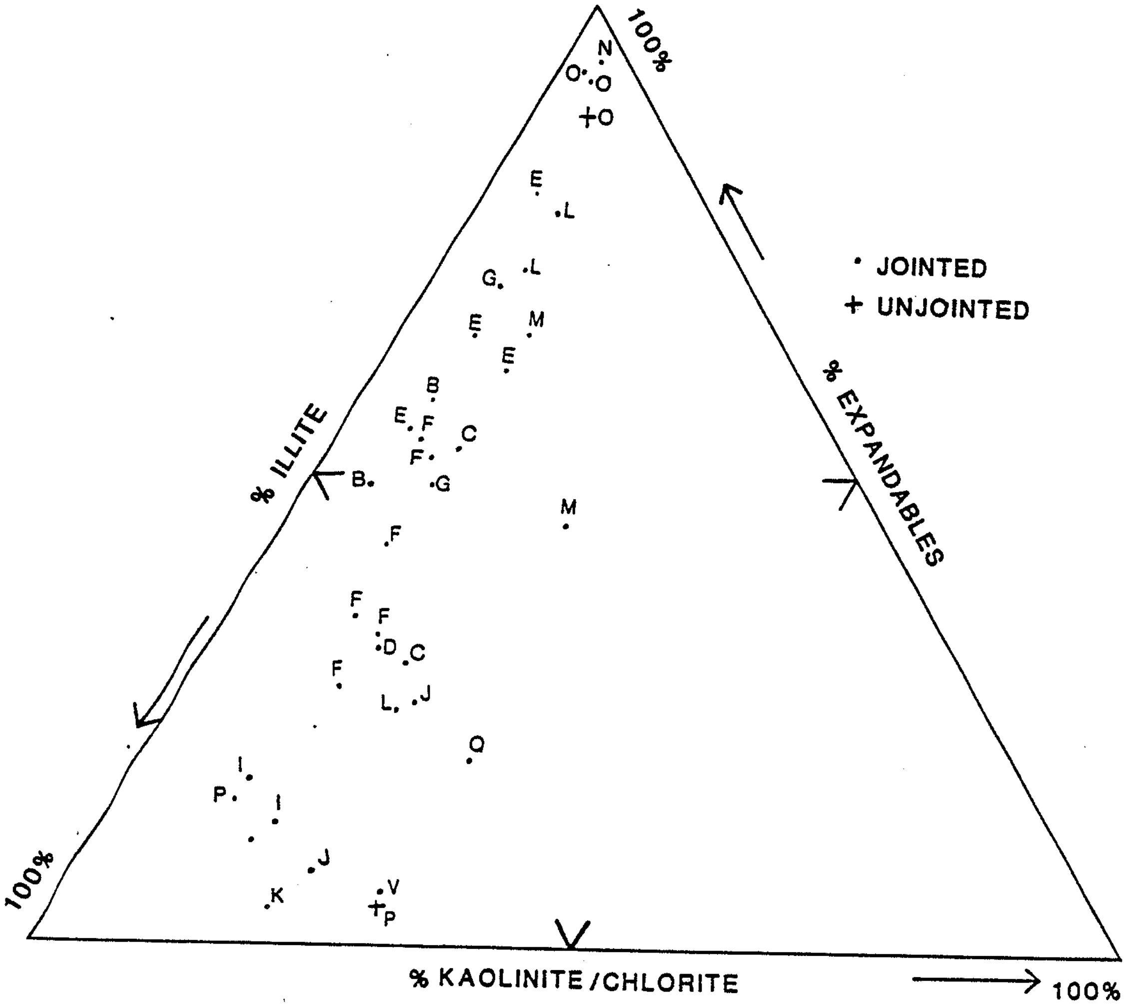


Figure 3: Ternary diagram of clay mineralogy for fine-grained till units. Labels on points refer to references in Table 1.

(Table 1). Fractures are restricted to sediment having less than about 50% sand and are concentrated in the USDA (USDA Soil Survey Staff, 1951) loam, silt loam, silty clay loam, silty clay, and clay textural classifications. It is interesting to note that relatively large amounts of silt are typical for jointed sediment; very little jointed sediment has been described with less than 35% silt except those with a very large amount of clay (greater than 50%).

clay mineralogy

Few authors have published quantitative analyses of the clay mineralogy of fractured sediment. Most of the till units shown on Figure 3 are from the Wisconsin study (Chapter 3) or from Illinois. Clay mineral abundance as used here refers to the results of semi-quantitative methods based on x-ray diffraction (for example Hallberg and others, 1978). The data available show that fractured sediment has variable amounts of illite and expandable clay, but none of the data show kaolinite + chlorite values above 30%. Illite and expandable clay exhibit shrinkage during dessication and this may be reflected in the clay mineralogy of fractured sediment. The influence of clay mineralogy on joint length could not be evaluated in the Wisconsin study because it was masked by large differences in clay content.

carbonate content

Fractures have been reported in sediment containing widely varying amounts of carbonate. Removal of carbonate during leaching has been thought to form shrinkage cracks in some materials (e.g. Figure 4 in Chapter 1), but fracturing was observed in many sites in Wisconsin to be prevalent below the leached zone (Chapter 3, Figure 27). Carbonate content may influence the shrink and swell of expandable clay minerals, and thus the response of fine-textured sediment to dessication (Chapter 1).

Geotechnical Characteristics

Following the work of Terzaghi (1936), fractures have been associated with overconsolidated clay. This relationship is not contradicted by the results of the Wisconsin study (Chapter 3), which found overconsolidation ratios ranging from 3.1 to 9 for the fractured till, but no data are available for the jointing characteristics of normally consolidated clay to test this relationship.

More information is available on the Atterberg limits of fractured sediment (Figure 4). The range of liquid limits and plasticity indices associated with fractured sediment usually plots above the Casagrande "A" line, and so these sediments will usually be classified as CH (high plasticty clay) or CL (low plasticity clay) by the Unified

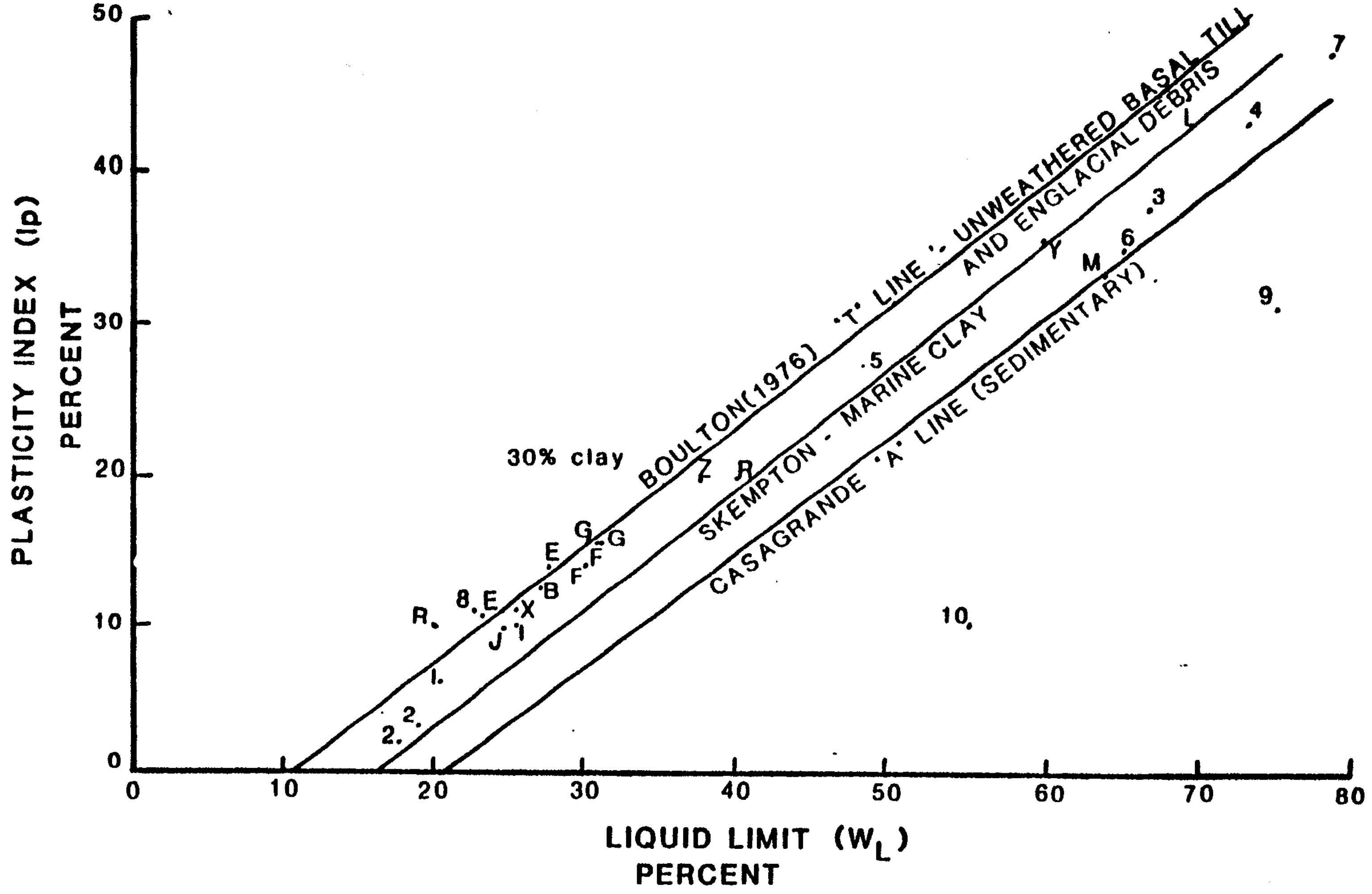


Figure 4: Plot of Plasticity Index versus Liquid Limit for fractured till and clay. Labels on points refer to references in Table 1.

Soil Classification System (Dunn, Anderson, and Kiefer, 1980). Most of the till units studied in Wisconsin have a low Liquid Limit (Chapter 3, Table 3), and come close to Boultons "T" line, which represents unweathered basal till and englacial debris (Boulton and Paul, 1976).

Limited data are available on the strength of fractured sediment; the Wisconsin study (Chapter 3) found till units with angles of internal friction (O) ranging from 22.2 to 31.4 degrees, which is within the range of 20 to 37 degrees reported in the literature. Cohesion in fractured fine-textured sediment ranges from 0 to 15 kPa (literature values) and was as high as 30 kPa for some of the Wisconsin till units (for references see Chapter 3).

RELATED FEATURES

Bedding, fabric, faulting, and other material anisotropy exerts an influence on fracture distribution. Characterization of these elements may make fracture prediction easier as they are more easily detected in smaller samples.

Fookes and Denness (1969) found bedding to be a dominant factor in controlling fracture orientation in tectonically formed joints in a variety of materials. Fracture propagation was perpendicular to the bedding planes even where folding had taken place.

It was suggested in Chapter 3 that fabric rearrangement taking place during subglacial deformation may control subsequent joint orientation. LaFluer (1980) found that incipient conjugate shears thought to result from subglacial compaction were prevalent in thin sections from jointed till. The identification of deformation-related fabrics in thin sections made from oriented samples may prove to be an important technique for fracture study. The samples necessary for thin sectioning are relatively small and can be obtained by shelby tube coring. Disturbance of the sample during coring can be evaluated from the thin sections, an advantage over the typical permeability testing of the cores. X-radiographs have also been used to detect jointing of shelby tube cores (Anderson and others, 1982).

Faulting of fine-grained sediment and other material can be recognized through a variety of techniques such as refraction seismic profiling and bore hole logging. As faulting can be related to joint formation and propagation, as in diapirism or collapse of buried ice blocks, faulting can be an important indicator of fractures.

HISTORY

Interpretation of the geologic history of fine-grained sediment is probably the most important factor in predicting fracture distribution. Knowledge of the genesis of a particular deposit is useful in ruling out certain modes of fracture formation. As an example, the focus of the Wisconsin study (Chapter 3) on glacial processes was in part due to an interpretation of the genesis of the sediment.

The geologic history of fractured sediment can be subdivided into three parts for purposes of discussion; deposition, deformation, and weathering.

Deposition

Fracture formation may take place in the subglacial environment during or shortly after deposition. In particular, the association of joints with subglacially deposited till makes identification of the position of till deposition an important component of fracture prediction. Syneresis cracks in marine clay are another example of fracture genesis related to depositional history.

Deformation

Identification of conditions leading to deformation of the sediment can be an indication of possible fracture genesis. For example, in the glacial environment, deformation is often related to compressive ice flow (Shaw, 1979). Indications of compressive flow and deformation include faulting and folding of associated sediment such as sand or gravel, or repetition of stratigraphy. Compressive flow is commonly restricted to areas of concave slopes directed toward the ice surface and to the ice marginal zone where ablation decreases ice velocity. Deformation in the subglacial environment is enhanced by elevated pore pressure existing during deformation, and evidence of elevated pore pressure (such as fluid escape structures) can be important indicators of the subglacial environment.

Other types of deformation, such as that associated with neotectonism or differential isostatic rebound, may also be related to fracture distribution. Evaluation of the magnitude of any deformation can assist in fracture prediction.

While not deformation per se, overconsolidation that may take place because of a drop in ground water levels or compaction by overlying sediment may be important if the material experiences later unloading. Unloading during deglaciation, erosion, or excavation can release the stress stored in the material and thus form fractures if the stress is greater than the strength of the material at lower confining pressures.

Weathering

Weathering of fine-grained sediment may exploit joints formed by other processes, and can form new joints through dessication or leaching. Identification of weathered horizons below the surface (paleosols) is especially important as they may have large impacts on behavior of the material, as illustrated by the results of the study at the Wilsonville site (Johnson and others, 1983). As weathering typically increases with age, older deposits typically will have better developed fracture systems.

dessication

Fractures and related structures in the solum may change their characteristics markedly due to dessication (Henin and Turc, 1949). The depth to which dessication can take place is an important factor, as jointing seems to be a greater problem at the surface and decrease in importance with depth. Figure 5 shows the maximum observed depth of jointing reported in the literature; it can be seen that fractures seem to be restricted to the

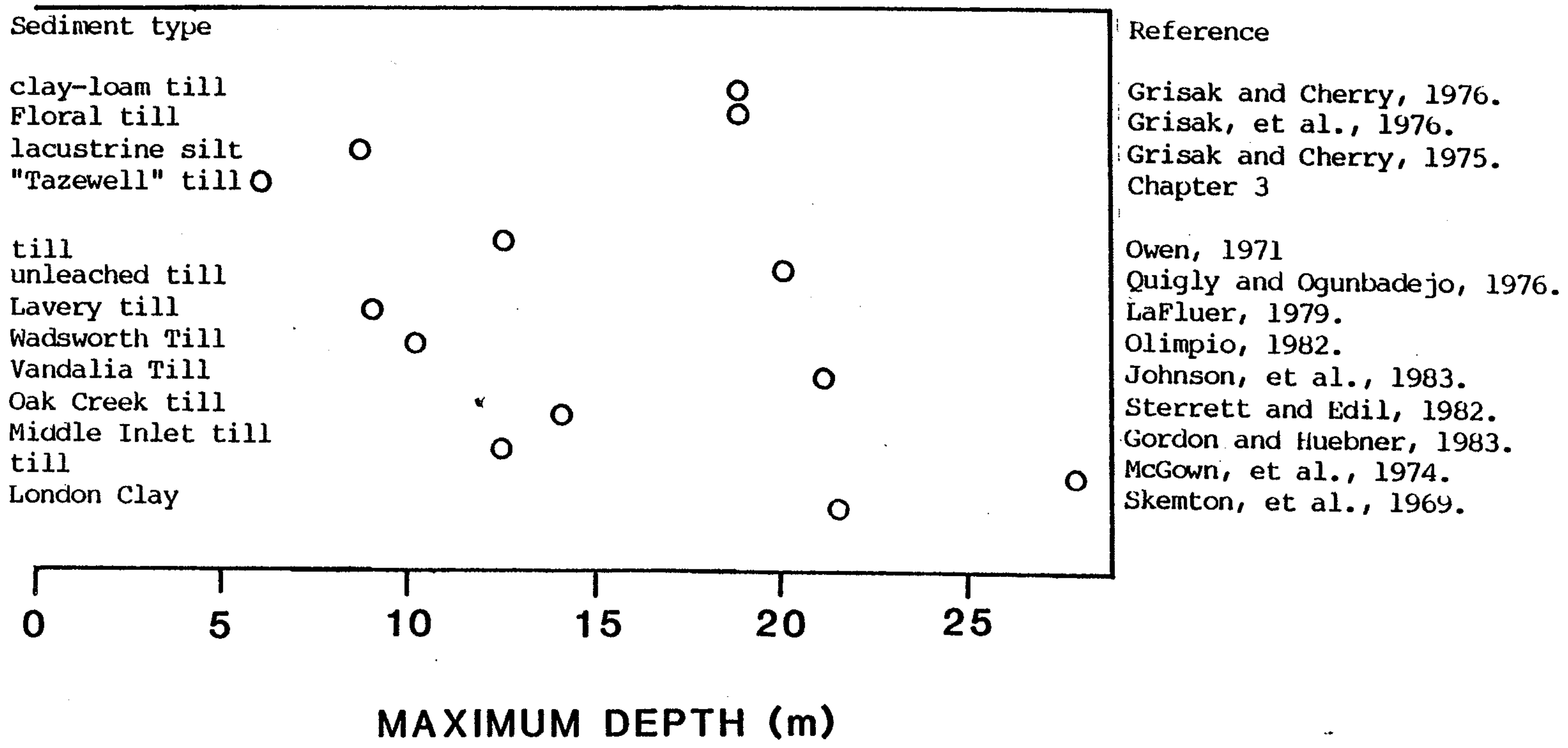


Figure 5: Maximum depth of fractures in fine-grained till and lacustrine sediment.

near-surface environment. In addition to dessication, overburden pressure may force fractures to close at depth. However, most of the jointing depth values listed in Figure 5 are based on excavation or borehole depth, and so the lower limit may reflect a lack of data for deeper deposits rather than a general occurance.

SUMMARY

Fractures can have significant impact on the hydrologic and geotechnical behavior of fine-grained till and clay. The complexity resulting from the presence of fractures makes accurate site characterization and simulation difficult and expensive. As current policy is to avoid complex hydrogeologic settings for siting of low-level radioactive waste disposal facilities, effort should be made to find ways to predict the existence of fractures before construction begins.

Fractures have been found in a variety of fine-grained sediment types, and are typically polygenetic. In till and lacustrine clay, they are associated with large amounts of silt and clay (greater than 50%), illite or smectite-dominated clay mineralogy, and are found primarily in the near-surface environment. The results of the Wisconsin study (Chapter 3) show that with increasing clay content and Plastic Limit, fracture length increases and the number of fractures (fracture intensity) decreases. Fracture orientation, as it can be controlled by ice-flow direction, slope failure, regional tectonism, or propagation of joints from underlying bedrock, is typically variable from site to site over large areas, but may be constant over smaller distances.

Fractures are often found in overconsolidated clay, and may be restricted to material that is classified as high or low plasticity clay by the Unified Soil Classification System. Insufficient data are available to evaluate the influence of the strength of the sediment on fracture form, but fractures have been observed in sediment with angles of internal friction ranging from 20 to 37 degrees, and cohesion values of 0 to 30 kPa.

Many authors have noted the association of fractures in till with subglacial deposition. Fracture formation may be related to deformation beneath thin ice at high pore pressure or low confining stress, perhaps in compressive flow regimes. The combination of thin ice and compressive flow was common in the marginal zones of continental glaciers, and so a knowledge of the regional geology and geologic history of an area can be important in predicting fracture distribution.

REFERENCES

Abdel-Rahman, M.A. and A.M. Hays, 1981. Sampling and statistical analysis of multi-modal orientation data. In O'Leary, D.W., (ed.), Proceedings of the 3rd International Conference on Basement Tectonics, 3, p. 73-86.

Acomb, L.J., 1978. Stratigraphic relations and extent of Wisconsin's Lake Michigan Lobe red tills. M.S. Thesis, University of Wisconsin -- Madison, 68 pp.

Acomb, L.J., Mickelson, D.M., and E.B. Evenson, 1982. Till stratigraphy and late glacial events in the Lake Michigan Lobe of eastern Wisconsin. GSA Bulletin, 93(4), p.289-296.

Adams, 1970. Effect of groundwater levels on stress history of the St. Clair till deposit: discussion. Canadian Geotechnical Journal, 7, p. 190- 193.

Al-Shaiku-Ali, M.M.H., Davis, G.H., and M.J. Lloyd, 1981. In situ measurement of Ko in a stiff fissured glacial till by hydraulic fracturing. Ground Engineering (Brentwood), 14(1), p. 19-24.

Anderson, M.G., Hubbard, M.G., and P.E. Kneale, 1982. The influence of shrinkage cracks on pore-water pressures within a clay embankment. Quarterly Journal of Engineering Geology, London, p. 9-14.

Andersson, J., Shapiro, A.M., and J. Bear, 1984. A stochastic model of a fractured rock conditioned by measured information. Water Resources Research, 20(1), p. 79-88.

Andrews, J.T., and D.I. Smith, 1970. Statistical analysis of till fabric: methodology, local and regional variability. Quarterly Journal of the Geological Society of London, 125, p. 503-542.

Babcock, E.A., 1977. A comparison of joints in bedrock and fractures in overlying Pleistocene lacustrine deposits, Central Alberta. Canadian Geotechnical Journal, 14, p. 357-366.

Baecher, G.B., 1980. Progressively censored sampling of rock joint traces. Mathematical Geology, 12, p. 33-40.

Baecher, G.B., 1980. Progressively censored sampling of rock joint traces. Mathematical Geology, 12, p. 33-40.

Baecher, G.B. and N. Lanney, 1978. Trace length biases in joint surveys. Proceedings, Symposium on Rock Mechanics, 19(1), p. 56-65.

Bagchi, A., 1982. Safe bluff inclination for zoning purposes along Lake Superior shoreline of Wisconsin. Independent study report, Department of Civil and Environmental Engineering, University of Wisconsin -- Madison, 107 pp.

Bailey, A.I., 1975. A method of analyzing polymodal distributions in orientation data. Mathematical Geology, 7(4), p. 285-293.

Banham, P.H., 1975. Glaciotectonic structures; a general discussion with particular reference ot the contorted drift of Norfolk. Geology Journal, Special Issue 6, p. 69-94.

Barber, E.S., 1958. Effect of water movement on soil. Highway Research Board Special Report 40, p. 212-225.

Bassarab, D., 1975. Geological discontinuities in rocks and soils. in Jointed Geological Materials (specialty session E), 6th Regional Conference for Africa on Soil Mechanics and Engineering, Durban, South Africa, September 1975, p. 215.

Bates, R.L., and J.A. Jackson, 1980. Glossary of Geology. Second Edition. Falls Church, VA: American Geological Institute, 749 pp.

Baver, L.D., Gardner, W.H. and W.R. Gardner, 1972. Soil Physics. 4th edition. New York: John Wiley and Sons.

Beard, C.N., 1959. Quantitative study of columnar jointing. Geological Society of America Bulletin, 70(3), p. 379-381.

Birkland, P.W., 1974. Pedology, Weathering and Geomorphological Research. New York: Oxford University Press, 265 pp.

Bisal, F., and K.F. Nielsen, 1967. Effect of frost action on the size of soil aggregates. Soil Science, 104(4), p. 268-273.

Bishop, A.W., and A.L. Little, 1967. The influence of the size and orientation of the sample on the apparent strength of the London Clay at Malden, Essex. Proceedings, Geotechnical Conference, Oslo, 1, p. 89-96.

Bjornbom, S., 1979. Clayey basal till in central and northern Sweden. Sveriges Geologiska Undersokning, NR 753, 62 pp.

Blackmore, A.V., 1973. Aggregation of clay by the products of iron. III. Hydrolysis. Australian Journal of Soil Research, 11, p. 75-82.

Bohn, H.L., McNeal, B.L., and G.A. O'Conner, 1978. Soil Chemistry. New York: John Wiley and Sons, 329 pp.

Boulton, G.S., 1975. Processes and patterns of subglacial sedimentation: A theoretical approach. in Wright, H.E., and F. Moseley, (eds.), Ice Ages Past and Present, p. 7-42.

Boulton, G.S., 1970. The depostion of subglacial and meltout tills at the margins of certain Svalbard glaciers. Journal Glaciology, 9(56), p. 231-245.

Boulton, G.S., 1972. The role of thermal regime in glacial sedimentation. in Price, R.J., and D.E. Sugden, (eds.), Polar Geomorphology. Institute British Geography Special Publication No. 4, p. 1-19.

Boulton, G.S., Dent, D.L., and E.M. Morris, 1974. Subglacial shearing and crushing, and the role of water pressures in tills from southeast Iceland. Geographiska Annaler, 56(A), p. 135-145.

Boulton, G.S., and M.A. Paul, 1976. The influence of genetic processes on some geotechnical properties of glacial tills. Quarterly Journal of Engineering Geology, 9(3), p. 159-194.

Brewer, R., 1964. Fabric and mineral analysis of soils. New York: John Wiley and Sons. 470 pp.

Bridges, M.C., 1975. Presentation of fracture data for rock mechanics. Proceedings, Australian -- New Zealand Conference on Geomechanics, 2, p. 144- 148.

Brykczynski, M., 1980. Valley-side glaciotectonics in the Warsaw Basin and the Plock Basin. Warsaw: Museum of the Earth, Polish Academy of Sciences.

Bullock, P., and C.P. Murphy, 1976. The microscopic examination of the structure of sub-surface horizons of soils. Outlook on Agriculture, 8(6), p. 348-354.

Burford, A.E., and J.M. Dixon, 1977. Systematic fracturing in clay and its relation to bedrock joints and drainage segments. GSA Abstracts with Programs, 9(7), p. 916.

Burford, A.E., and J.M. Dixon, 1978. Systematic fracturing in young clay of the Cuyahoga River Valley, Ohio, and its relation to bedrock jointing and drainage segments. Geotechnique, 28(2), p. 201-206.

Carlsson, A., and T. Olsson, 1982. High rock stresses as a consequence of glaciation. Nature, 298, p.739-742.

Cawsey, D.C., 1977. The measurement of fracture patterns in the Chalk of southern England. Engineering Geology, 11, p. 201-215.

Chamberlain, E.J., 1981. Overconsolidation effects of ground freezing. Engineering Geology, 18, p. 97-110.

Chandler, R.J., 1973. A study of structural discontinuities in stiff clays using a polarizing microscope. Proceedings, International Symposium on Soil Structure, p. 78.

Cheeney, R.F., 1983. Statistical Methods for Geologists. Allen and Unwin, London, 169 pp.

Cruden, D.M., and H.A.K. Charlesworth, 1976. Errors in strike and dip measurements. Geological Society of America Bulletin, 87, p. 977-980.

Czeratzki, W., and H. Frese, 1958a. Importance of water in formation of soil structure. Highway Research Board, Special Report 40, p. 200.

Czeratzki, W., and H. Frese, 1958b. Water and its conduction in soils. Highway Research Board, Special Report 40, p. 200.

Dapples, E.C., and J.F. Rominger, 1945. Orientation analysis of fine-grained clastic sediments. Journal of Geology, 53, pp. 246-261.

Derbyshire, E., McGowan, A., and A. Radwan, 1976. "Total" fabric of some till landforms. Earth Surface Processes, 1(1), p. 17-26.

Derbyshire, E., and P.F. Jones, 1980. Systematic fissuring of a matrix-dominated till at Church Wilne, Derbyshire, England. Geology Magazine, 117(3), p. 243-254.

Desaulnier, D.E., Cherry, J.A., and P. Fritz, 1981. Origin, age, and movement of pore watr in argillaceous Quaternary deposits at four sites in southwestern Ontario. Journal of Hydrology, 50, p. 231-257.

Donovan, R.N., and R.J. Foster, 1972. Subaqueous shrinkage cracks from the Caithness Flagstones. Journal Sedimentary Petrology, 42, p. 195-206.

Dreimanis, A., 1981. Till: terminology, definitions of till and related terms, genetic classification. INQUA Commision on genesis and lithology of Quaternary deposits, Till Work Group, Circular No. 23.

Ehlers, J., and H. Stephan, 1983. Till fabric and ice movement. in Ehlers, J., (ed.), Glacial Deposits in North-West Europe. Rotterdam: A.A. Balkema, p. 267.

Eisenstein, Z., and S. Thomson, 1978. Geotechnical performance of a tunnel in till. Canadian Geotechnical Journal, 15(3), p. 332-345.

Esu, F., 1966. Short-term stability of slopes in unweathered jointed clays. Geotechnique, 16(4), p. 321-328.

Evenson, E.B., 1973. Late Pleistocene shorelines and stratigraphic relations in the Lake Michigan basin. Geological Society of America Bulletin, 84, p. 2281-2298.

Eyles, N., and J.A. Sladen, 1981. Stratigraphy and geotechnical properties of weathered lodgement till in Northumberland, England. Quarterly Journal of Engineering Geology, 14, p. 129-141.

Flint, R.F., 1961. Two tills in southern Connecticut. Geological Society of America Bulletin, 72(11), p. 1687-1691.

Fookes, P.G., 1965. Orientation of fissures in stiff over-consolidated clay of the Siwalik System. Geotechnigue, 15, p. 195-206.

Fookes, P.G. 1975. Terminology of discontinuities. in Symposium on the Engineering Behavior of Glacial Materials. Birmingham, U.K.

Fookes, P.G., and B. Denness, 1969. Observational studies on fissure patterns in Cretaceous sediments of south east England. Geotechnique, 19, p. 453-477.

Fookes, P.G., and D.G. Parrish, 1969. Small scale discontinuities in London Clay. Quarterly Journal of Engineering Geology, 1, 217-240.

Frei, E., 1950. Genesis of various types of soil structure. Fourth International Congress Soil Science, 1, p. 46-48.

Foote, G.R., 1982. Fracture Analysis in Northeastern Illinois and Northern Indiana. Masters Thesis, University of Illinois at Champaign-Urbana.

Gass, T.E., 1984. Ground water in the news: Risk assessment frailties dominate water conference talk. Ground Water, 22(3), p. 386.

Gordon, M.E., and P.M. Huebner, 1983. An evaluation of the performance of zone of saturation landfills in Wisconsin. Proceedings, Sixth Annual Madison Waste Conference.

Green, W.J., Lee, G.F., and R.A. Jones, 1981. Clay - soils permeability and hazardous waste storage. Journal Water Pollution Control Federation, 53(8), p. 1347-1354.

Green, W.J., Lee., G.F., Jones, R.A., and T. Pallt, 1983. Interaction of clay soils with water and organic solvents: Implications for the disposal of hazardous wastes. Environmental Science and Technology, 17(5), p. 278-282.

Grisak, G.E., 1975. The fracture porosity of glacial till. Canadian Journal Earth Science, 12(3), p. 513-515.

Grisak, G.E., and J.A. Cherry, 1975. Hydrologic characteristics and response of a fractured till and clay confining a shallow aquifer. Canadian Geotechnical Journal, 12, p. 23-43.

Grisak, G.E., Pickens, J.E., and J.A. Cherry, 1980. Solute transport through fractured media, 2. Column study of fractured till. Water Resources Research, 16(4), p. 731-739.

Grisak, G.E., Cherry, J.A., Vonhof, J.A., and J.P. Blumele, 1976. Hydrogeologic and hydrochemical properties of fractured till in the Interior Plains Region. in Legget, R.F., (ed.), Glacial Till. Special Publication 12 of the Royal Society of Canada, p. 292-303.

Gustafson, Y., 1950. Genesis of soil structure. Fourth International Congress of Soil Science, 4, p. 29-31.

Hallberg, G., Lucas, J.R., and C.M. Goodmen, 1978. Semi-quantitative analysis fo clay mineralogy. in Standard Methods for Evaluation of Quaternary Materials in Iowa. Iowa City, Iowa: Iowa Geological Survey, Technical Information Series #8.

Hanna, T.H., 1966. Engineering properties of the glacial lake near Sarnia, Ontario. Ontario Hydroelectric Power Commission Research Quarterly, p. 1-12.

Hanson, W.W., 1982. Mechanical and chemical modification of till used in liner construction. paper presented at Minnesota Soil Mechanics Conference, January, 1982.

Harrison, P.W., 1957. A Clay-till fabric: its character and origin. Journal of Geology, 65, p. 275.

Hendry, J.M., 1982. Hydraulic conductivity of a glacial till in Alberta. Ground Water, 20(2), p. 162.

Herzog, B.L., and W.J. Morse, 1984. A comparison of laboratory and field determined values of hydraulic conductivity at a waste disposal site. Proceedings, Seventh Annual Madison Waste Conference, Madison, Wisconsin, September 11-12, 1984 (in press).

Hight, D.W., El-Ghamrawg, M.K., and Gens, A., 1979. Some results from a laboratory study of a sandy clay and implications regarding its in-situ behavior. Second International Conference on Behavior of Off-shore Structures, Paper 13, p. 133-150.

Hobbs, B.E., W.D. Means, and P.E. Williams, 1976. An Outline of Structural Geology. New York: John Wiley and Sons, 571 pp.

Hodgson, R.A., 1961. Classification of structures of joint surfaces. American Journal of Science, 259, p. 493.

Holst, T.B., 1982. Regional jointing in the northern Michigan Basin. Geology, 10(5), p. 273-277.

Holst, T.B., and G.R. Foote, 1982. Joint orientation in Devonian rocks in the northern portion of the Lower Peninsula of Michigan. GSA Bulletin, 92(2), p. 85.

Jennings, J.E.B., (convenor), 1975. Jointed geological materials. Specialty Session E, 6th Regional Conference for Africa on Soil Mechanics and Foundation Engineering, Durban, South Africa,September 1975, p. 215-229.

Johnson, M.D., 1980. Origin of Lake Superior red clay and glacial history of Wisconsin's Lake Superior shoreline west of Bayfield Peninsula. M.S. Thesis, University of Wisconsin -- Madison, 104 pp.

Johnson, M.D., 1983. The origin and microfabric of Lake Superior red clay. Journal of Sedimentary Petrology, 53(3), p. 859-873.

Johnson, T.M., Griffin, R.A., Cartwright, K., Fullmer, L.R., Herzog, B.L., Morse, W.J., DuMontelle, P.B., Killey, M.M., Stohr, C.J., and R.E. Hughes, 1983. Hydrogeologic Investigations of Failure Mechanisms and Migration of Organic Chemicals at Wilsonville, Illinois. Proceedings of the Third National Symposium on Aquifer Restoration and Ground-Water Monitoring, p. 413-420.

Jordan, R., and D.E. Andrews, 1978. Possible joint control of glacial meltwater channels, an indicator of immediate crustal response to ice unloading in north central New York state. GSA Abstracts with Programs, 10(2).

Kaczynski, R., and L. Wyokinski, 1970. The influence of the weakness surfaces on the strength of cohesive soils. Proceedings 2nd Seminar Soil Mechanics and Foundation Engineering, Lodz, 1, p. 219-228.

Kahn, I., 1983. Failure of an earth dam: a case study. Journal of Geotechnical Engineering, 109(2), p. 244-259.

Kamb, W.B., 1959. Ice petrographic observations from Blue Glacier, Washington, in relation to theory and experiment. Journal of Geophysical Research, 64(11), p. 1891.

Kazi, A., and S.L. Knill, 1973. Fissuring in glacial lake clays and tills on the Norfolk Coast, United Kingdom. Engineering Geology, 7, p. 35-48.

Krishna Murti, G.S.R., and S.J. Richards. Some effects of sesquioxides on soil structure. Indian Journal of Agronomy, 19(2), p. 141-147.

Kruger, J., 1979. Structures and textures in till indicating subglacial deposition. Boreas, 3(3), p. 323-340.

Kruger, J., and I. Marcussen, 1976. Lodgement till and flow till: a discussion. Boreas, 5, p. 61-64.

Kupsch, W.O., 1955. Drumlins with jointed boulders near Dollard, Saskatchewan. Geological Society of America Bulletin, 66, p. 327.

LaFluer, R.G., 1979. Glacial Geology and Stratigraphy of Western New York Nuclear Service Center and Vacinity, Cattaragus and Erie Counties, New York. United States Geological Survey Open File Report 79-989.

LaFluer, R.G., 1980. Petrographic analysis of the West Valley burial ground till. Rensselaer Polytechnic Institute, Try, N.Y., 12 pp..

Lagerlund, E., 1977. Till studies and neotectonism in northwest Skane, south Sweden. Boreas, 6, p. 159.

Larsen, C.E., (in press). Lake level, uplift, and outlet incision; a new perspective on the Nipissing- and Algoma-stage Great Lakes.

Laslett, G.M., 1982. Censoring and edge effects in areal and line transect sampling of rock joint traces. Mathematical Geology, 14(2), p. 125-140.

Lo, K.Y., 1970. The operational strength of fissured clays. Geotechnique, 20, p 57-74.

Long, J.C.S., Remer, J., Wilson, C., and P.A. Witherspoon, 1982. Porous media equivalent for a network of discontinuous fractures. Water Resources Research, 18(3), p. 645-658.

MacClintock, P., and A. Dreimanis, 1964. Reorientation of till fabric by overriding glacier in the St. Lawrence Valley. American Journal of Science, 262, p. 133-142.

Maher, L.J., and M.W. O'Neill, 1983. Geotechnical characterization of dessicated clay. ASCE Journal of Geotechnical Engineering, 109(1), p. 56-71.

Mahtab, M.A., Bolstad, D.D., Alldredge, J.R., and R.J. Shanley, 1972. Analysis of fracture orientation for input to structrual models of discontinuous rock. U.S. Bureau of Mines, Report of Investigations 7669, 76 pp.

Mahtab, M.A., and T.M. Yegulalp, 1982. A rejection criteria for definition of clusters in orientation data. in Goodman, R.E., and F.E. Hueze, (eds.), Proceedings 23rd symposium on Rock Mechanics, University of California, Berkeley, August 25-27, 1982, p. 116.

Mardia, K.V., 1972. Statistics of Directional Data. London: Academic Press.

Mathews, W.H., and J.R. Mackey, 1960. Deformation of soils by glacier ice and the influence of pore pressures and permafrost. Transactions of the Royal Society of Canada, Section 4, N54.

May, R.W., and S. Thomson, 1978. The geology and geotechnical properties of till and related deposits in the Edmonton, Alberta, area. Canadian Geotechnical Journal, 15(3), p. 362-370.

McCartney, M.C. and D.M. Mickelson, 1982. Late Woodfordian and Greatlakean history of the Green Bay Lobe, Wisconsin. GSA Bulletin, 93(4), p. 297-302.

McGown, A., Anderson, W.F., and A. Radwan, 1975. Geotechnical properties of the tills in west central Scotland. in The Midlands Soil Mechanics and Foundation Engineering Society, p. 89-99.

McGown, A., and E. Derbyshire, 1977. Genetic influences on the properties of tills. Quarterly Journal of Engineering Geology, 10, p. 389-410.

McGown, A. and A.M. Radwan, 1975. The presence and influence of fissures in the boulder clays of west central Scotland. Canadian Geotechnical Journal, 12(1), p. 84-97.

McGown, A., Salvidar Sali, A., and A.M. Radwan, 1974. Fissure patterns and slope failures in till at Hartford, Ayrshire. Quarterly Journal of Engineering Geology, 7(1), p. 1-26.

McKinlay, D.G., McGown, A., Radwan, A., and D. Hossan, 1975. Representative sampling and testing in fissured lodgement tills. in Symposium on the Engineering Behavior of Glacial Materials. Birmingham: Midland Soil Mechanics and Foundation Engineering Society, p. 143-155.

Mead, W.J., 1925. The geologic role of dilatancy. Journal of Geology, 33, p. 685.

Menzies, J., 1979. The mechanics of drumlin formation with particular reference to the change in pore-water content of the till. Journal Glaciology, 22(87), p. 373-384.

Mickelson, D.M., 1971. Glacial Geology of the Burroughs Glacier area, southeastern Alaska. Ohio State University Institute of Polar Studies Report 40, 147 pp.

Mickelson, D.M., Acomb, L.J., and T.B. Edil, 1979. The origin of preconsolidated and normally consolidated tills in Eastern Wisconsin, USA. in Schluchter, C., (ed.), Moraines and Varves: Origin, Genesis, and Classification. Rotterdam: Balkenia, p. 179-188.

Mickelson, D.M., and L. Clayton, 1981. Subglacial conditions and processes during Middle Woodfordian time in Wisconsin. GSA North Central section Abstracts with Programs, p. 30.

Mickelson, D.M, Clayton, L., Baker, R.W., Mode, W.N., and A.F. Schneider, 1984. Pleistocene Stratigraphic Units of Wisconsin. Wisconsin Geological and Natural History Survey Miscellaneous Paper 84-1.

Mickelson, D.M., Clayton, L., Fullerton, D.S., and H.W. Borns, Jr., 1983. The Late Wisconsin glacial record of the Laurentide ice sheet in the United States. in Porter, S.C., (ed.), Late-Quaternary Environments of the United States, vol. 1. Minneapolis: University of Minnesota Press, p. 3-37.

Mickelson, D.M., Johnson, M.D., Waldemarson, D., and H.G. Johansson, 1981. Jointing in till: An example from south-central Sweden. INQUA Commission on the genesis and lithology of Quaternary Deposits, Symposium on the genesis and lithology of morainic deposits in an alpine environment, Jackson, Wyoming, August 1981.

Mickelson, D.M, and others, 1977. Shore Erosion Study: Technical Report. Madison: Wisconsin Coastal Management Program.

Mitchell, J.K., 1976. Fundementals of Soil Behavior. New York: John Wiley and Sons.

Mörner, N., 1978. Faulting, fracturing, and seismicity as functions of glacio-isostacy in Fennoscandia. Geology, 6, p. 41.

Muller, S.W., 1947. Permafrost. Ann Arbor, Michigan: J.W. Edwards, Inc.

Muller, E.H., 1977. Dewatering during lodgement of till. Tenth INQUA Congress, Birmingham, Abstracts, p. 318.

Need, E.A., 1980. Till stratigraphy and glacial history of Wisconsin's Lake Superior shoreline: Wisconsin Point to Bark River. M.S. Thesis, University of Wisconsin -- Madison, 140 pp.

Nelson, J.D., 1973. Influence of clay fabric on bonds and dilation. International Symposium on Soil Structure, Gothenberg, p. 153.

Nye, J.F., 1952. The mechanics of glacier flow. Journal Glaciology, 2, p. 82-93.

Nyugen, V., and G.F. Pinder, 1984. Direct calculation of aguifer parameters in slug test analysis. in Rosenheim, J., and Bennet, G.D., (eds.), Ground Water Hydraulics. American Geophysical Union Water Resources Monograph 9, p. 222-239.

Olimpio, J.C., 1982. Low-level radioactive-waste burial at the Palos Forest Preserve, Illinois; Part II, Geology and hydrology of the glacial drift as related to the migration of tritium. USGS Open-file Report 82-78, 57 pp.

Owen, E.B., 1969. Stratigraphy and engineering description of the soils exposed on a section of the Welland Canal By-pass Project (Contract 743), Ontario, Canada. Geological Survey of Canada paper 69-31.

----, 1972. Geology and engineering description of the soils of the Welland -- Port Colborne area (Contract 743), Ontario, Canada. Canadian Geological Survey paper 72-49, 7 pp.

Popodopulos, S.S., and I.J. Winograd, 1974. Storage of low-level radioactive wastes in the ground -- Hydrogeologic and hydrochemical factors with an appendix on the Maxey Flats, Kentucky, radioactive waste storage site: Current knowledge and data needs of a quantitative hydrogeologic evaluation. USGS Open File Report 74-344, 49 pp.

Passl, F.B., and J.P. Schafer, 1968. Two-till problem in Naugatuck-Torrington Area, Western Connecticut. Connecticut State Geological and Natural History Survey Guidebook #2, p. B1-25.

Paterson, W.B., 1981. The Physics of Glaciers. Oxford: Pergamon Press.

Penney, L.F., and S.A. Catt, 1967. Stone orientation and other structural features in tills in East Yorkshire. Geology Magazine, 104, p. 344-360.

Poorooshasb, H.B., and R.N. Yong, 1983. Hydraulic fracturing from induced pore pressures in clay soils. Canadian Geotechnical Journal, 20, p. 546-555.

Post, J.L., 1981. Expansive soils: volume change and expansion pressure of smectites. California Geology, 34(9), p. 197-203.

Price, N.J., 1966. Fault and Joint Development in Brittle and Semi-Brittle Rock. Oxford: Pergamon Press.

Priest, S.D., and J.A. Hudson, 1976. Discontinuity spacings in rock. International Journal Rock Mechanics, Mining Science and Geomech Abstracts, 13, p. 135-148.

Prudic, D.E., 1981. Computer simulation of ground-water flow at a commercial radioactive-waste landfill near West Valley, Cattaragus County, New York. in Little, C.A., and L.E. Stratton (eds.), Modeling and Low-Level Waste Managment -- an Interagency Workshop. Oak Ridge National Laboratory, Oak Ridge, Tennessee. ORO-821, p 215-248.

Prudic, D.E., 1982. Hydraulic conductivity of a fine-grained till, Cattaraugus County, New York. Ground Water, 20(2), p. 194.

Pruidic, D.E., and A.O. Randall, 1979. Ground-water hydrology and subsurface migration of radioisotoopes at a low-level, solid radioactive-waste disposal site, West Valley, New York. in Carter, M.W., Kahn, B., and A.A. Moghissi, (eds.), Management of Low-Level Radioactive Wastes, volume 2. New York: Pergamon Press, p. 853-882.

Pulley, S., 1980. Determination of preconsolidation pressures of Wisconsin glacial tills by various methods. unpublished M.S. Thesis, Department of Civil and Environmental Engineering, University of Wisconsin -- Madison, 140 pp.

Pusch, R., 1973. Structural variation in boulder clay. Proceedings International Symposium on Soil Structure, Gothenburg, p. 113-121.

Quigly, R.M., 1975. Weathering and changes in strength of glacial till. in Yatsu, E., Ward, A.J., and F. Adams, (eds.), Mass Wasting. Norwich: Geo Abstracts Ltd., p. 117-131.

Quigly, R.M., and T.A. Ogunbadejo, 1976. Till geology, mineralogy, and geotechnical behavior, Sarnia, Ontario. in Legget, R.F., (ed.), Glacial Till. Special publication 12 Royal Society of Canada, Ottawa, p. 336-345.

Ramsden, J., and J.A. Westgate, 1971. Evidence for reorientation of a till fabric in the Edmonton Area, Alberta. in Goldthwait, R.P., (ed.), Till : A Symposium. Ohio State University Press, Columbus: p. 335-344.

Rehm, B.W., Groenewold, G.H., and W.M. Peterson, 1982. Mechanisms, distribution, and frequency of groundwater recharge in an upland area of Western North Dakota. North Dakota Geological Survey Report of Investigations 75.

Rimmer, D.L., and D.J. Greenland, 1976. Effects of CaCO3 on the swelling behavior of a soil clay. Journal of Soil Science, 27(2), p. 129-139.

Roberts, J.C., 1961. Feather fracture and the mechanics of jointing. American Journal of Science, 259, p. 481-492.

Roberts, J.R., Cherry, J.A., and F.W. Schwartz, 1982. A case study of a chemical spill: Polychlorinated Biphenyls (PCBs) 1. History, distribution, and surface translocation. Water Resources Research, 18(3), p. 525-534.

Robertson, J.B., 1984. Geologic problems at low-level radioactive waste-disposal sites. in Groundwater Contamination. Washington, D.C: National Academy Press, p. 104-108.

Ronca, L.B., and K.R. Chaivre, 1977. Errors in measurements of oblique and vertical joints. Modern Geology, 6, p. 61-65.

Russell, E.W., 1934. The interaction of clay with water and organic liquids as measured ny specific volume changes and its relation to the phenomena of crumb formation in soils. Phil. Transactions Royal Society London, 233A, p. 361-389.

Schuenemeyer, J.H., Kock, G.S., and R.F. Link, 1972. Computer program to analyze directional data, based on the methods of Fisher and Watson. Journal Mathematical Geology, 4(3), p. 177-202.

Schwan, J., and W. Ritzema, 1982. Till variation in a Weichselian glacial section along the coast of Southwest Funen, Denmark. Geol. Mijnbouw, 61, p. 163.

Selsing, L., 1981. Stress analysis on conjugate normal faults in unconsolidated Weichselian glacial deposits from Brorfelde, Denmark. Boreas, 10(3), p. 275.

Shanley, R.J., and M.A. Mahtab, 1974. A computer program for clustering data points on the sphere. U.S. Bureau of Mines Information Circular 8624, 58 pp.

----, 1975. FRACTAN: a computer code for analysis of clusters defined on the unit hemisphere. U.S. Bureau of Mines Information Circular 8671, 41 pp.

Shaw, J., 1977a. Till body morphology and structure related to glacier flow. Boreas, 6, p. 189-201.

----, 1977b. Tills deposited in arid polar environments. Canadian Journal of Earth Science, 14, p. 1239.

----, 1979. Genesis of the Sveg Tills and Rogen Moraines of central Sweden: a model of basal melt out. Boreas, 8, p. 409.

----, 1982. Melt-out till in the Edmonton Area, Alberta, Canada. Canadian Journal of Earth Sciences, 19, p. 1548-1569.

Skempton, A.W., and R.D. Northey, 1952. The sensitivity of clays. Geotechnique, 3(1), p. 30-53.

Skempton, A.W., Shuster, R.L., and D.J. Petley, 1969. Joints and fissures in the London Clay at Wraysbury and Edgeware. Geotechnique, 19, p. 205-217.

Sledz, J.J., and D.D. Huff, 1981. Computer model for determing fracture porosity and permeability in the Conasauga Group, Oak RIdge National Laboratory, Tennessee. Oak Ridge, Tennessee: Oak Ridge National Laboratory, Environmental Science Division Publication No. 1677. (ORNL/TM-7695)

Sleeman, J.R., 1963. Cracks, peds, and their surfaces in some soils of the Riverine Plain, N.S.W. Australian Journal of Soil Research, 1, p. 91-102.

Soderman, L.G., and Y.D. Kim, 1970. Effect of groundwater levels on stress history of the St. Clair clay till deposit. Canadian Geotechnical Journal, 7, p. 173-187.

Stauffer, M.R., 1966. An empirical-statistical study of three dimensional fabric diagrams as used in structural analysis. Canadian Journal of Earth Sciences, 3, p. 473-498.

Stanford, S.D., and D.M. Mickelson (in press). Till fabric and deformational structures in drumlins near Waukehsa, Wisconsin.

Stephansson, O., and H. Carlsson, 1980. Siesmo-tectonics in Fennoscandia. in Mörner, Nils-Axel (ed.), Earth Rheology, Isostacy, and Eustacy. Wiley and Sons, N.Y: p. 327-337.

Sterrett, R.J., 1980. Factors and mechanics of bluff erosion on Wisconsin's Great Lakes shorelines. Ph.D. Dissertation, University of Wisconsin -- Madison,, 372 pp.

Sterrett, R.J., and T.B. Edil, 1982. Groundwater flow systems and stability of a slope. Groundwater, 20(1), p. 5.

Terzaghi, K., 1936. Stability of slopes in natural clay. Proceedings 1st International Conference of Soil Mechanics, Harvard University, 1, p. 161-165.

Terzaghi, K., and Peck, R.B., 1967. Soil mechanics in engineering practice. Second Edition. New York: John Wiley and Sons.

Terzaghi, R.D., 1965. Sources of error in joint surveys. Geotechnique, 15, p. 287-304.

Thompson, W.B., and G.W. Smith, 1983. Pleistocene Stratigraphy of the Augusta and Waldboro Areas, Maine. Maine Geological Survey Bulletin, 27, 37 pp.

Trainer, F.W., 1973. Formation of joints in bedrock by moving glacial ice. USGS Journal of Research 1(2), p. 229-236.

Tystovich, N.A., 1975. The mechanics of frozen ground. (edited by G.K. Swinzow). Scripta Book Company, Washington: 426 pp.

United States Code of Federal Regulations (10 CFR 61) Licensing Requirements for Land Disposal of Radioactive Waste.

United States Department of Agriculture, Soil Survey Staff, (1951). Soil Survay Manual. Washington, D.C: U.S. GPO, 503 pp.

U.S. Nuclear Regulatory Commission, 1981. Licensing Requirement for land disposal of radioactive waste and associated amendment: Draft of 10 CFR 61 parts 2, 19, 20, 21, 30, 40, 50, 51, 70, 73, and 150.

Walcott, R.I., 1972. Late Quaternary vertical movements in Eastern North America: quantitative evidence of glaciostatic rebound. Review of Geophysics and Space Physics, 10(4), p. 849-884.

Warner, J., 1969. FORTRAN IV program for construction of PI diagrams with the UNIVAC 1108 Computer. Computer Contribution 33, State Geological Survey, The University of Kansas, Lawrence.

West, G. 1979. A prelimary study on the reproducibility of joint measurements in rock. Transport and Road Research Laboratory (U.K.), Supplementary Report

West, G. 1979. A prelimary study on the reproducibility of joint measurements in rock. Transport and Road Research Laboratory (U.K.), Supplementary Report 448.

Westgate, J.A., 1976. Hydrogeological and hydrochemical properties of fractured till in the interior plains region: discussion. in Legget, R.F. (ed.), Glacial TIll. Royal Society of Canada Special Publication 12, p. 334.

Wheeler, R.L., and J.M. Dixon, 1980. Intensity of systematic joints: Methods and application. Geology, 8, p. 230-233.

Wheeler, R.L., and S.M. Holland, 1981. Style elements of systematic joints: An analytical procedure with a field example. in O'Leary, D.W., (ed.), Proceedings of the 3rd International Conference in Basement Tectonics, 3, p. 393-404.

White, E.M., 1961. Volume changes in some clay soils. Soil Science, 94(3), p. 168-172.

---------------, 1966. Subsoil structure genesis: theoretical considerations. Soil Science, 101(2), p. 135-141.

Williams, A.A.B., 1975. Measurement of joint parameters in soils. in Jointed Geological Materials (Specialty Session E), 6th Regional Conference for Africa on Soil Mechanics and Foundation Engineering, Durban, South Africa, September 1975., p. 220.

Williams, R.E., and R.N. Farvolden, 1967. The influence of joints on the movement of groundwater through glacial till. Journal of Hydrology, 5, p. 163-170.

Wise, D.U., and T.A. McCrory, 1982. A new method of fracture analysis: Azimuth versus traverse distance plots. Geological Society of America Bulletin, 93(9), p. 889-897.

Zehner, H.H., 1983. Hydrogeologic investigation of the Maxey Flats radioactive waste burial site, Fleming County, Kentucky. USGS Open- FIle Report 83-133.