CONTEMPORARY FRONTAL MORAINE FORMATION IN THE YOHO VALLEY, BRITISH COLUMBIA

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by

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B.A. (Hons.), University of Wales, 1978

A Thesis Submitted in Partial Fulfillment of the Requirements for the Degree of Master of Science.

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1980

Abstract

The northern terminus of Emerald Glacier ($51^{\circ}31'$N, $116^{\circ}32'$W) in the Yoho Valley, British Columbia was bordered by a small, actively forming frontal moraine during summer 1979. Stratigraphic and morphological contrasts existed round the ice front, which primarily resulted from a contrast in the distribution of supraglacial debris.

Sedimentological and geotechnical techniques were utilised to determine the origin of stratigraphic units within the moraine ridge.

Moraine A, at the margin of heavily debris covered ice, exhibited a complex stratigraphy. At most sites a lens of subglacially derived till was evident, between units of supraglacially derived material. It is proposed that the moraine forming process involved the initial development of an ice-front talus apron, which was subsequently pushed and overridden. A plastic subglacial till was squeezed from beneath the supra-morainal ice margin, and overlain by a sorted supraglacial unit during glacier retreat. The moraine was actively advancing during the field season due to the maintenance of glacier-moraine contact resulting from the retardation of ice-melt afforded by the supraglacial debris cover.

Moraine B is located at the margin of debris-free ice. The stratigraphy is less complex, although an upper unit representing a younger depositional phase was observed. The process of formation involved the melt-out of subglacial deposits during the summer months (July to August), which were bulldozed into a ridge during winter advance. Successive accretions of till onto the proximal moraine side, perhaps on an annual basis, are suggested.

Deterioration of climate, resulting in positive mass balances is the main cause of glacier advance. Positive balances have been recorded from nearby glaciers between 1973 and 1976. This suggests that small glaciers are sensitive indicators of periods of climatic deterioration.

The thesis concludes that more than one moraine forming process may be observed around a single ice margin, one of which may be 'annual' in nature; and that complex moraines may be formed by depositional processes operating at the margin of a temperate glacier. If complexities exist in presently forming moraines, then such a possibility must be considered when examining deposits of past glaciations.

Acknowledgements

The research on Emerald Glacier, British Columbia was made possible by an N.S.E.R.C. operating grant awarded to Dr. R.J. Rogerson of the Geography Department, Memorial University of Newfoundland. Thanks are extended to Parks Canada, and the friendly and co-operative personnel of Yoho National Park for permission to collect samples.

I am considerally indebted to Richard Kodybka of the Geography Department at Memorial University who assisted me with the fieldwork. He was a constant source of inspiration and companionship, and was always willing to discuss any facet of the work, despite the hardships involved in the field. My thanks are also extended to Alison Stenning who helped with the surveying.

I wish to take this opportunity to thank the Geology Department at Memorial University who allowed me the use of their facilities, and especially to Mr. Jaan Vahtra who completed the X-ray diffraction analysis, identified the clay and non-clay mineral peaks and was also always readily available to discuss results and problems. I wish to express my thanks for all his help.

Advice on the cartography was provided by Charles Conway of the Memorial University Cartography Laboratory and I am certainly thankful for that. I would also like to thank Dr. G.J. Young of the Snow and Ice Division of Environment Canada for meteorological data from Peyto Glacier, and Dr. W.B. Whalley of the Geography Department, University of Belfast for permission to include elements of his doctoral dissertation in this thesis.

To my patient supervisor, Dr. Robert J. Rogerson I can only offer my heartfelt sympathy and admiration. Throughout he has provided inspiration and guidance. In the initial planning of the expedition his experience was invaluable and during the field work was of enormous help. During the laboratory and analysis period his experiences were again of tremendous value and through the writing stage he has been a constant source of advice and discussion.

I would like to offer my thanks for the interest and encouragement expressed by many of the faculty and graduates of the department - especially Gillian Mellars who was always a great help in all aspects of the thesis.

Last, but by no means least I must thank my patient typist Sue Sharpe, who although faced with the difficult task of reading my writing managed to do a superb job of things.

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Introduction

1.1 Rationale for Research

Within the last decade an increasing number of glaciers in Western Canada have been advancing or are at least stationary following some 30 years of rapid recession (Gardner 1972). This condition may reflect climatic deterioration or a better mass equilibrium, with reduced glacier size and higher ablation zone elevations.

Where conditions are suitable at the termini of such glaciers, contemporary moraines may be developing, and the opportunity to study active terminal moraine formation therefore exists. In such locations, it should be possible to observe precise formative processes and the resultant sediments, structures and morphologies, rather than assume process from the characteristics of long abandoned moraines. This latter approach has typified much previous work (e.g. Clayton and Moran 1974).

Contemporary research in glacial geomorphic processes has been primarily led by Boulton (e.g. Boulton 1967, 1968, 1970a, 1970b), although it has largely centred on sub-polar and polar glaciers, and highlighted complex depositional processes. Temperate glacier depositional environments were considered to be 'simpler', resulting in simple moraine types.

At the northern terminus of Emerald Glacier in the Yoho Valley, British Columbia however, a situation exists which negates a simple depositional model. Most of the terminal ice surface is dominated by a supraglacial till cover, at the margin of which a small, yet complex moraine was observed to be forming in 1978 (Rogerson, personal communication). The south-eastern section of the glacier has no supraglacial till element, yet also possesses an active terminal moraine; one contrasting in structure from the other moraine.

A situation exists where contrasting morphological and sedimentary environments both produce a moraine, but with some substantial differences in component sediments, morphologies and perhaps formative processes.

It is the objective of this thesis to quantify, describe and analyse these components. In achieving this end the study will contribute to knowledge on the depositional processes of temperate glaciers.

1.2 Definition of Terms

Considerable confusion and ambiguity of terminology exists within glacial geomorphological literature. A discussion of terminology to be incorporated within this thesis is therefore a necessary component.

The term 'moraine' is often assigned both morphological and sedimentological meaning (e.g. by Hewitt 1967, Minnel 1977, Ahmed 1979, Nakawo 1979), with a consequent loss of precision (Price 1973, Worsley 1974, Boulton and Eyles 1979). To avoid such confusion 'moraine' will possess a purely morphological significance in this thesis.

The composing material or sediment of moraines is till; "an aggregate whose components are brought together by the direct agency of glacier ice, which though it may suffer deformation by flow, does not undergo subsequent disaggregation and re-deposition" (Boulton 1972).

Prepositives to the term 'till' abound in the literature (Boulton 1976, Boulton and Eyles 1979), most of which lack any genetic significance. The schema proposed by Boulton (1976) will therefore be utilised here i.e. a subglacial component comprising lodgement, melt-out, flow and lee-side till elements; and those deposits aggregated from supraglacially derived debris for which the term 'supraglacial till' will be used. Detailed characteristics of each type are described by Boulton (1976).

1.3 Classification of Moraines

"Classification and terminology are primarily tools in communication, not an end in themselves" (Aario 1977). A classification system often reflects the state of scientific knowledge and as such must be continually refined and modified. Nevertheless, classification systems have become entrenched in the literature, inevitably leading to a breakdown in the 'communication' system (Stromberg 1965).

The final goal of classification should be a schema which is 'explanatory', based upon genesis. Often though, terminology is descriptive rather than genetic, and indeed one descriptive term may characterise a polygenetic landform. Alternatively, one moraine type may be described by more than one term.

Embleton and King (1975) for instance, sub-divide moraines on the basis of their state of activity into active, ice-cored and inactive moraines. As such, a single moraine type (e.g. a push moraine) could be included under all three headings, although position and mode of formation may remain essentially the same.

In reaction to the confusion that classification systems instil, several new schemas have recently been developed, most notably by Aario (1977). He attempts to attach genetic adjectives to his terms e.g. 'hummocky disintegration moraine' and 'hummocky squeezed-up moraine'. Genetically similar forms are regarded as 'associations' e.g. Rogen moraine, flutes and drumlins, whereas forms of no genetic similarity within an area are termed 'complexes'.

Alternatively, Borgstrøm (1979) proposes a classification based on positional influences, and considers a four fold subdivision into frontal and sub-glacial subaqueous, and frontal and sub-glacial supraaqueous features. The classification is essentially a modification of Prest's (1968) system, which considers position relative to the ice front, independent of further genetic implications.

The Prest classification of unitary nomenclature is the least ambiguous, if not the most satisfactory system to use. It leaves process open to investigation and refinement without subsequent reclassification, and has thus become widely accepted (e.g. Sugden and John 1976) as well as providing a core for Boulton's sediment/landform associations. A slightly modified version of Prest's system will be utilised in this thesis (Fig 1-1).

Sedimentary and geomorphic feature variation in terrestial glacial environments can be explained within one of three principal sediment/landform associations (Boulton 1976, Boulton and Paul 1976). The supraglacial, and subglacial/proglacial sediment/landform associations are distinguished by their mode of deposition manifest in distinct sediment differences

Linear features		Non-linear features
Parallel to ice flow (controlled deposition)	Transverse to ice flow (controlled deposition)	Lacking consistent orientation (controlled or uncontrolled deposition
Subglacial forms with streamlining: a) Fluted and drumlinized ground-moraine b) Drumlins and drumlinoid ridges c) Crag and tail ridges	Subglacial forms: a) Rogen or ribbed moraine b) De Geer or wash board moraine c) Kalixpinnmo hills d) Subglacial thrust moraines e) Sublacustrine moraines	Subglacial forms: a) Low-relief ground moraine b) Hummocky ground moraine
Ice-pressed forms: Longitudinal squeezed ridges	Ice-pressed forms: minor transverse squeezed ridges and corrugated moraine	Ice-pressed forms: random or rectilinear squeezed ridges
Ice marginal forms: lateral and medial moraines, some interlobate and kame moraines	Ice front forms: a) End moraines b) Push moraines c) ice thrust/shear d) Some kame and delta moraines	Ice surface forms: a) Disintegration moraine (controlled) b) Disintegration moraine (uncontrolled)

Figure 1-1 Classification of moraines modified after Prest (1968)

Field is located within the Kicking Horse Valley, a major east-west trending valley which 'funnels' predominant westerly airflows, and therefore cyclonic depressions affecting the region. Although the Little Yoho valley is also orientated east-west, it is sufficiently removed both altitudinally and laterally from the main valley to be less influenced by major storm events. Similarly, the valley at Field rises to the east, whereas the Little Yoho valley declines to the east. Westerly airstreams therefore subside down the Little Yoho valley becoming increasingly stable. Periods of less intense rainfall, but of longer duration tend to characterise the area.

Winter is dominated by Continental Arctic and Maritime Polar air masses and consequently climatic conditions are harsh. The lowest mean monthly temperatures at Field are recorded in January (-11°C), although this masks daily variations which may be extreme, in the -30 to -40°C range. The variation of temperature with altitude is less marked during the winter months; approximately 3.5°C/1000 metres. Temperatures of -48 to -54°C at the 1500 to 3000 metre levels are possible, but quite rare (Janz and Storr 1977), although wind chill may considerably increase heat loss. The influence of the wind chill factor increases with elevation. No winter climatic data are available for Peyto Glacier.

Inversions are common due to the ponding of cold Arctic air in major valleys, displacing milder air to higher elevations.

1.4 Frontal Moraines: A discussion of their genesis.

Ice front depositional forms have been described by Prest (1968). However, the process of deposition, although implicit within some of his terminology is not a prime consideration. This apparent shortcoming requires resolution in order to undertake any discussion on the genesis of frontal moraines. As such, Prest's classification system will be augmented by that of Price (1973) who recognises 3 processes as being dominant in the formation of moraines: dump, push and squeeze mechanisms.

a) Simple 'Dump' Moraines.

These 'dump' moraines (Okko 1955) are proglacial constructs of supraglacially transported (although not necessarily supraglacially derived) material. They are essentially ice front talus slopes (Hewitt 1967, Whalley 1974, and Eyles 1979).

Supraglacially derived debris is characteristic of valley glaciers or ice-caps with nunataks. Debris falling onto the ice surface in the accumulation area from extraglacial sources (e.g. valley sides and headwalls), is transported passively in an englacial layer, termed a 'bed-parallel debris septum' (Boulton and Eyles 1979). Material collected below the firn line is transported supraglacially.

Upon melt-out in the ablation zone these englacial septa concentrate as supraglacial morainic till (Boulton 1976, Eyles 1979). The thickness of this deposit increases towards the terminus, as a result of ablation and increasing longitudinal compressive strain (Eyles and Rogerson 1978a).

Deposition at the ice margin may be as a small dump moraine possessing no internal structure (Facies A, Boulton and Eyles

1979), or as a larger feature with a well defined coarse clast fabric and crude internal bedding (Facies B, Boulton and Eyles 1979). Within sub-polar moraines melt-out and flow till elements may also be present (Boulton 1967).

b) Complex 'Dump' Moraines.

Characteristic of cold-based glaciers, complex 'dump' moraines have been described under a wide variety of terms: ice-cored moraines, shear moraines, Thule-Baffin moraines, ablation cones and ice-moraine ridges. The term Thule-Baffin is preferred here, since it offers no genetic implications, simply being in recognition of the type sites.

The genesis of Thule-Baffin moraines fundamentally differs from simple 'dump' moraines in the mechanisms bringing material to the glacier surface; processes which are the centre of considerable debate. Three separate mechanisms have been proposed: the influence of shear planes; basal freezing and compressive flow; and the incorporation of a frontal apron, or possibly a combination of factors.

The shear plane hypothesis, supported by Goldthwait (1951), Bishop (1957), Swinzow (1962) results from the shearing of active over stagnant ice. Bed material is most actively eroded at the stagnant/active ice inter-face and is subsequently transported to the surface. A series of upwarping shear planes may develop up-glacier of the original shear. Opponents to this mechanism however, cite sorted and stratified units found within debris laden bands, plus the non-sliding character of the formative ice as being evidence against such a concept (Weertman 1961; Hooke 1970, 1973; Drozdowski 1977).

Instead, a 'freezing-on' of debris in near terminal ice is proposed. Ice melt, through geothermal and frictional heat at the glacier sole, is forced toward the ice margin under existing pressure gradients, where it refreezes. Basal debris, which previously contained unfrozen water is now incorporated into the ice by the expansion of interstitial water as it freezes. This is similar to the 'large scale block inclusion' process of Moran (1971). Subsequent transportation of debris to the ice surface, rather than the result of shearing, is facilitated by ablation and compressive strain within the terminal zone. This tendency for flow lines to return to the surface near the terminus has been described by Boulton (1970b), Hooke (1973), Drozdowski (1977) etc. Shear structures may be associated with upwarping debris bands (Souchez 1967, 1971).

A third method of debris entrainment is proposed by Shaw (1977a, 1977b), who considers overriding of a frontal debris apron by advancing glacier ice as an effective mechanism. Sorting processes within the apron itself explain sorted units within debris bands. Again, englacial flow patterns return material to a supraglacial position.

Once the composing material arrives at the ice surface, the development of Thule-Baffin type moraines is little disputed (Price 1973, Embleton and King 1975, Sugden and John 1976). It is a process outlined by Goldthwait (1951) (Fig. 1-2).

Formation of the trough (see Fig. 1-2) is a critical factor, with its subsequent development resulting in a complete separation of a debris covered ice-core from the glacier. Initiating processes are due to differential ablation between areas of thick debris cover on near terminal ice and the discontinuous cover ('black ice' (Goldthwait 1951)) on the proximal glacier side. The influence of a debris cover on ablation rates is to enhance melting where the sediment cover is thin (Ward 1952, Boulton 1967), whereas thick debris layers retard the melt rate (Østrem 1964, Carrara 1975, Eyles and Rogerson 1978b). Embleton and King (1975) report ablation rates of 32.5 cm/week for 'dirty' ice (a debris thickness of less than 0.5 cm.), compared to 20 cm/week for clean ice. Rates of less than 5 cm/week were recorded under a greater than 20 cm. debris cover. Such a retardation of melt explains the steep ice front of debris covered termini .

Once initiated, the trough is deepened by a variety of processes: the warming of air due to protection from katabatic winds (Goldthwait 1951); and perhaps of more importance, by supraglacial stream action (Kozarski and Szupryczynski 1973).

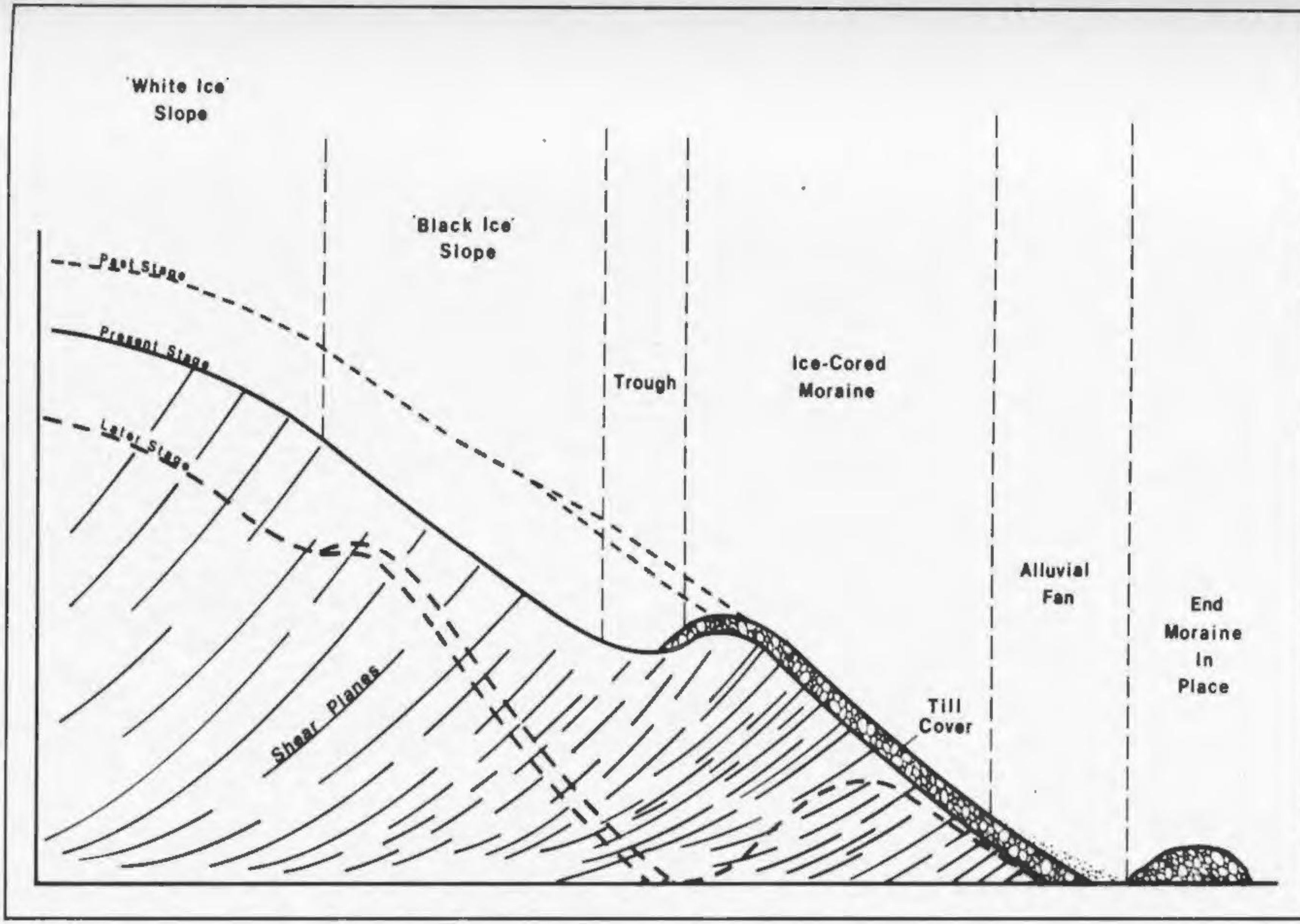

Figure 1-2 Development of Thule-Baffin moraines (after Goldthwait 1951)

Upon detachment from the main ice body, destruction of the ice-core begins, through the development of thermokarst (Østrem 1959, Clayton 1964, Johnson 1971, Driscoll 1974, Healy 1975).

The morphological expression of Thule-Baffin type moraines has been discussed by Østrem (1964), and Østrem and Arnold (1970). In their formative stage, the moraines are characterised as 'misfit' features when compared to the size of the forming glacier. This results from the magnitude of the ice-core rather than the overlying debris thickness, which is often small (Bishop 1957, Johnson 1971, Souchez 1971, Carrara 1975). A range of 0.01 to 1.10 metres is common.

Upon eventual melt-out of the ice-core, the resultant moraine may possess little relief, although it may be of considerable lateral extent. Composing sediments are subject to considerable post-depositional modification and will exhibit little internal structure or fabric.

c) Push Moraines

Push moraines have been defined by Chamberlain (1894): "A glacier pushes matter forward mechanically, ridging it at its edges, forming what may be termed a push moraine".

Although push moraines have been considered small, insignificant features of the glacier terminus, and unlikely to survive a prolonged period of pushing (Flint 1971, Price 1973, Whalley 1974a) large contrasts of scale are evident.

Kälin (1971) describes a push moraine on Axel Heiberg Island as 2.1 kilometres long; 700 metres wide and at a mean height of 45 metres above the outwash plain. Similarly, Rutten (1960) reports moraines 50-200 metres high and 10-100 kilometres long on the North-West European Plain. The composing sediments, which are derived from previously deposited till or glacial outwash deposits, were described in each case as being, or having been within zones of permafrost. The mode of formation was thus inferred as being due to the bulldozing of permanently frozen detritus. Indeed, Haeberli (1979) stated that "there is no doubt that push moraines are morphological expressions of the deformation of permafrost".

Horizontal and vertical stresses imposed by advancing ice cause the structurally weakened permafrosted soil to deform and buckle into large blocks, which are subsequently thrust upwards, often in a stacked sequence. This is similar to processes described by Moran (1971).

The thermal characteristic of the forming glacier is regarded as being polar or sub-polar, to facilitate the ploughing process (Embleton and King 1975). Tectonic structures are often evident within push moraines.

In contrast to the large-scale features described above, small moraines often less than 5 metres high, have been observed in many areas e.g. Norway (Worsley 1974, Matthews et al. 1979), Yukon (Bayrock 1967, Johnson 1972), Argentina (Rabassa et al. 1979), Antarctica (Birnie 1977) and the Himalayas (Hewitt 1967).

In each case, interstitial water was unfrozen. Material therefore deforms more readily, precluding the dislocation of large blocks of sediment that so characterise the larger push moraine features.

The pushing mechanism may not be a continuous process, and Hewitt (1967) stresses the importance of the seasonal effect in moraine construction. Summer is dominated by a recession of the glacier snout, when glacier melt is greater by volume than the supply of ice through glacier movement. Surface ice-melt ceases with the first significant fall of snow, while downglacier movement continues long after. A readvance of the glacier terminus results in the bulldozing of summer deposited material into a moraine ridge. The precise formative period has been suggested as being late in the accumulation season (Worsley 1974, Birnie 1977), but would depend on when, if at all, the glacier terminus became frozen to its bed.

d) Squeeze Moraines

Similar in morphology and genesis to washboard moraines (Elson 1969, Neilsen 1970), frontal squeeze moraines have been described by Gravenor and Kupsch (1959), Thorarinsson (1967), Price (1969, 1970) and Birnie (1977). Formation is the result of the migration of water-soaked till developed within the high pressure subglacial environment, towards areas of low pressure i.e. the ice margin.

Andrews (1963) comments that temperate glacier seepage pressures are toward the margin, whereas under arctic glacier conditions they are directed away from the ice front. This fact may explain the variability in position of squeeze developed moraines i.e. between washboard moraines formed in near terminal crevasses, and the small frontal moraine features that are evidenced from temperate environments.

Confining pressures are towards the proximal side of the moraine, whereas the distal side is free of overburdening ice pressures. As such, a distinct asymmetric profile characterises squeeze moraines. Similarly, confining pressures result in a strong fabric, normal to the moraine crest on the proximal face with an upglacier imbrication pattern, compared to the weakly developed fabric of the distal side (Price 1970).

Squeeze moraines are small in size. Price (1970) describes Fjallsjokull moraines as being up to 5 metres in height, but generally smaller values have been recorded. The pattern of deposition follows crenulations in the ice margin, although similar patterns have also been described for dump moraines (Boulton and Eyles 1979) and push moraines (Matthews et.al. 1979).

1.5 'Annual' Moraines

It has been suggested that some moraines are 'annual' features of the ice front (e.g. Hoppe 1959, Thorarinsson 1967, Price 1970, Kucera 1972, Mickelson and Berkison 1974, Worsley 1974).

Push moraines and squeeze moraines have been described as annual resulting from winter readvances of the terminus, often superimposed on a general recession. The moraines form part of a recessional sequence and direct correlations between the number of moraines and the length of the recession period have been found. Kucera (1972) on Athabasca Glacier for instance, evidences a series of annual photographs. However, absolute proof of this type is lacking for other areas (Stromberg 1965 , Embleton and King 1975), and often correlations are regarded as inconclusive evidence for an 'annual' term to be applied. Varve analysis undertaken by Stromberg (1965) in Sweden shows that a suggested 'annual' moraine sequence over a considerable time span was an invalid characterisation of the situation. 'Annual' moraines may however be identifiable over the short term.

The relative size and number of recessional moraines has been considered a function of the geomorphic setting, rather than glacier behaviour (Luckman and Osborn 1979). Gently sloping proglacial areas reveal a series of recessional units, whereas short, steep or heavily debris covered glaciers exhibit only one, larger moraine. Accretions to these moraines would be the 'annual' event.

1.6 The Emerald Glacier Moraine: Introduction

The small, actively forming moraine at the terminus of Emerald Glacier has developed within the Rocky Mountain alpine glacier system. The temperate thermal regime negates the Thule-Baffin moraine type formation as a primary mechanism.

However, the remaining moraine forming processes are all potentially applicable i.e. a push, squeeze or dump genesis (Fig. 1-3), or possibly a combination of processes.

Confirmation or rejection of these models is dependant upon glacier dynamics and sediment characteristics, as well as the structure and morphology of the moraine itself. Small 'dump' moraines characterise inactive glaciers (Boulton and Eyles 1979), while push moraines suggest frontal movement, if only on a seasonal basis. Dump moraines are composed of supraglacial debris whereas squeeze moraines comprise sub-glacial material. Push moraines may be formed of either debris suite, as well as proglacial sediments within an appropriate setting.

It is the applicability of each of these modes of genesis that will be the focus of subsequent chapters.

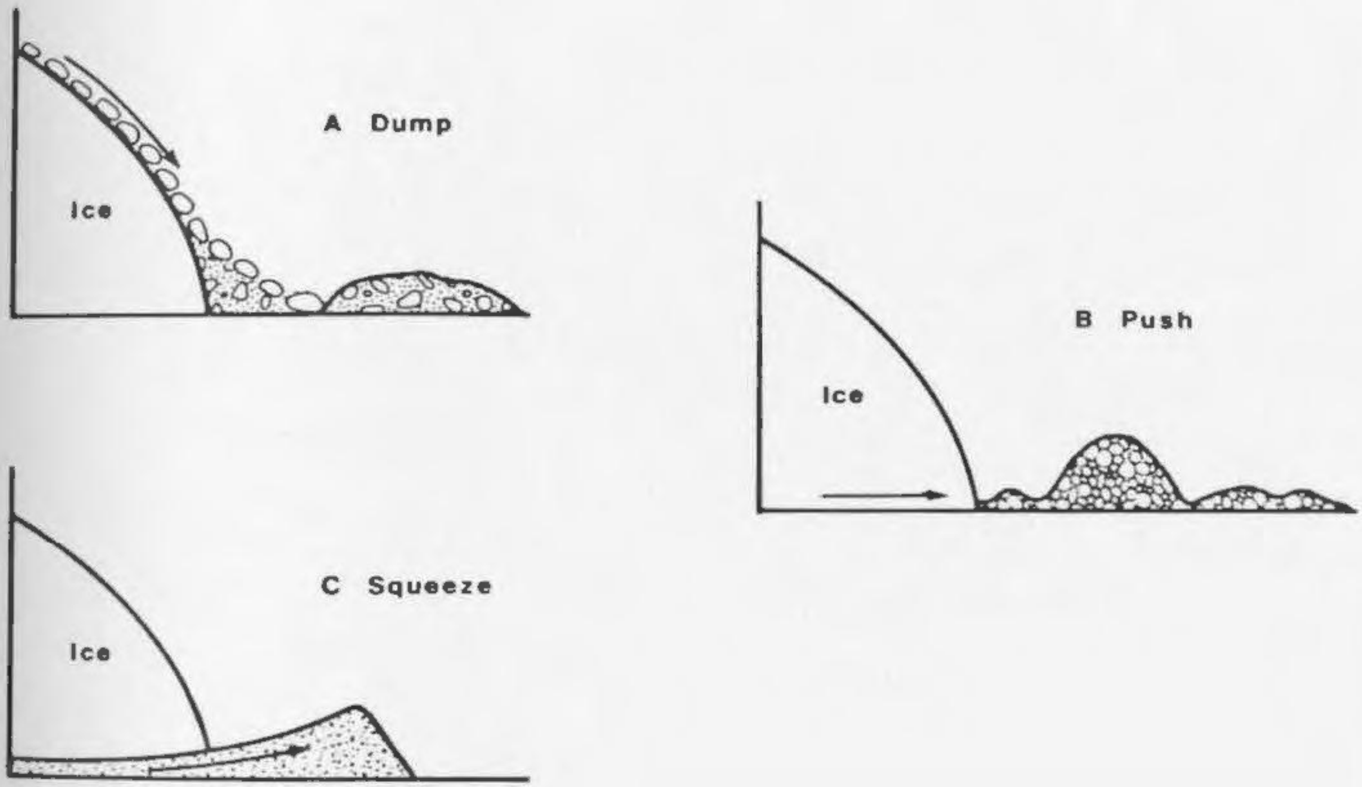

Figure 1-3

A summary of possible moraine forming processes

Chapter 2

Emerald Glacier: The Physical Setting

2.1 Location

Emerald Glacier is situated on the north-eastern slope of the President Range in Yoho National Park, British Columbia ($51^{o}31'$N, $116^{o}32'$W) (Fig. 2-1). In general it may be described as a mountain-shelf type glacier, with a high breadth-length ratio (about 6.3: 1).

The small (less than 1 km^2) northern section of Emerald Glacier is the area towards which research was directed. It is a shallow cirque with the terminus at an altitude of 2286 metres (7495 feet) a.s.l. this northern section is seperated from its south-easterly extension by a narrow ice-fall (Fig. 2-2).

2.2 Geology

Emerald Glacier is perched on a topographic shelf, which is an expression of the Wapta Mountain Thrust (Cook 1975). The shelf is located on the eastern downthrow side of the Cathedral Crags Anticline, and is bounded on its south-western side by a steep arête rising to President Peak (3140 m.a.s.l.),

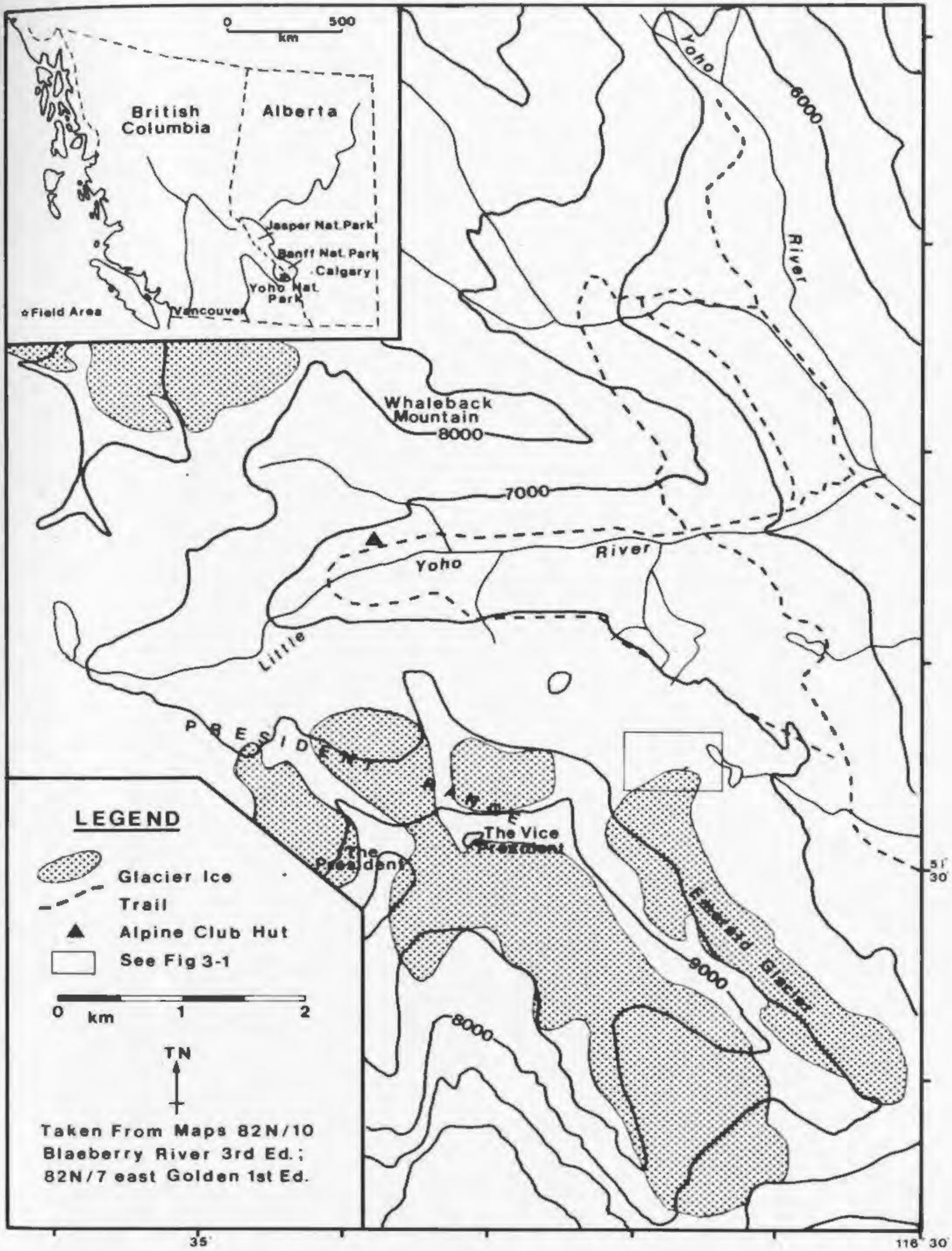

Figure 2-1 Location of the study area

and to the north-east by slopes dipping towards the Little Yoho and Yoho valleys.

Stratigraphically, four major formations of Cambrian age outcrop in the Emerald Glacier area (Fig. 2-3). The oldest, the Pika formation is an argillaceous limestone, composed of thin bedded to flaggy limestone with argillaceous partings. Contact with overlying sediments is sharp, with the more resistant dark grey Pika contrasting with the varicoloured Arctomys formation. This is a laminated, platy shale, siltstone and dolomite with ripple marks, mudcracks and scattered salt casts (Cook 1975). The Arctomys becomes progressively interbedded with the massive dolomites of the Waterfowl formation. The contact is largely arbitrary and of a type which has been described as part of the Sullivan-type Grand Cycle (Aitken 1978).

This Cambrian depositional cycle represents increments to a stable-shelf, low-latitude depositional platform that was bounded by a wide, practically unbreached carbonate-shoal complex which had high tidal energy (Aitken 1978, p. 515 and p. 519).

The Waterfowl formation is composed of bedded limestone and dolomite with some silty laminae and shale interbeds near the base, and characterised by alternating yellow and brown weathering beds.

Figure 2-2

Ice-fall demarking the south-easterly boundary of North Emerald Glacier. The geological stratigraphy can be noted along the backwall of the glacier.

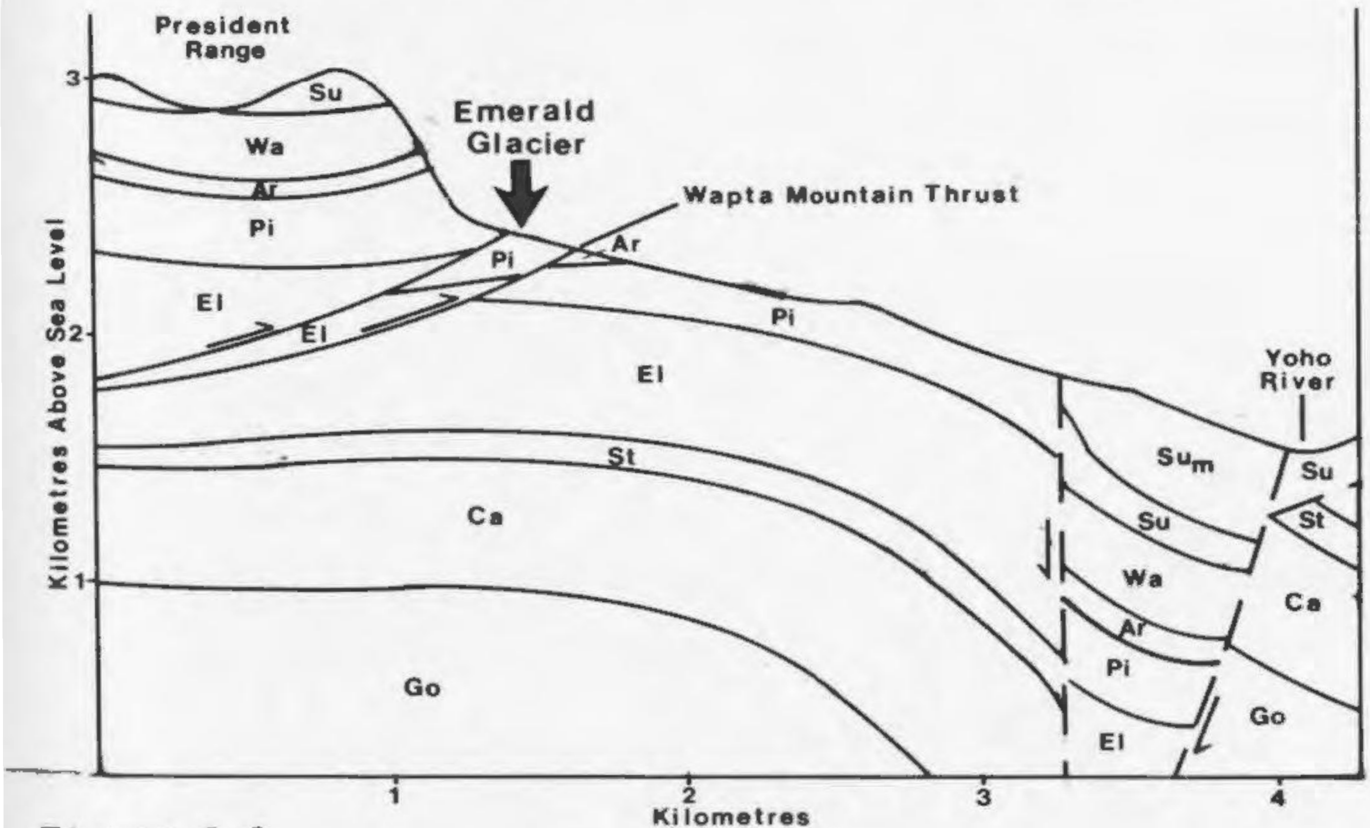

Figure 2-3

Cross section through the President Range showing the geological formations (Su = Lower Sullivan, Wa = Waterfowl, Ar = Arctomys, Pi = Pika, El = Eldon, St = Stephen, Ca = Cathedral, Go = Gog, Su_m = Middle Sullivan).

The Lower Sullivan formation is Upper Cambrian in age, and is a shale unit with calcareous interbeddings. It has an abrupt contact with the underlying Waterfowl formation.

Emerald Glacier itself rests on bedrock of the Pika and Arctomys formations. These define the parent material for the subglacial component of the glacier's sedimentary system. The Pika, Arctomys, Waterfowl and Lower Sullivan formations are exposed subaerially in the backwall of the shallow cirque, and constitute the supraglacial debris component. The broad differentiation of a limestone dominant subglacial unit and a shale/limestone supraglacial deposit can thus be made.

3.3 Climatic Regime

Emerald Glacier is dominated by a mid-latitude mountain climatic regime, one exhibiting both great temporal and spatial variations (Janz and Storr 1977).

The summer climate is characterised by three air masses: Maritime Arctic and Maritime Polar, which are introduced to the area by the predominating westerly airstream; and Continental Tropical air originating from South-Western North America, and penetrating British Columbia along southerly and south-westerly upper winds.

Average monthly temperatures during the summer (June to August) range from 12°C (June) to 15°C (July) at Field (1240 m.a.s.l.); the nearest meteorological station to Emerald Glacier.

However, vertical lapse rates of 7-8°C/1000 metres,considerably reduce temperatures at Emerald Glacier itself. Average monthly temperatures would therefore range between about 5 and 7°C for the summer months. Local micro-climatological variations however, of which aspect and slope are the most significant in mountainous areas will produce daily figures widely departing from the mean values. Similarly, in the immediate near-glacier environment, the influence of katabatic winds will be to reduce temperatures by 2-3°C. Nevertheless, elevation is the principal control on average temperatures (Janz and Storr 1977).

At Peyto Glacier (2200 m.a.s.l.), a meteorological station is run by the Snow and Ice Division of the Water Resources Branch. The similar terminal elevation to Emerald Glacier means that average monthly temperature and precipitation data may correspond to the Emerald Glacier situation. Mean monthly summer temperatures of between 4.6°C (June) and 7.3°C (July) have been recorded, averaged over a 13 year period (1966-1978) (Young, in press).

Mean monthly summer precipitation reaches a peak in June (2.7 mm at Peyto Glacier; 2.73 mm at Field), whilst in July/August values fall to 2.3 mm and 1.5 mm at Peyto Glacier and Field respectively. Precipitation increases with elevation as a result of the orographic effect, as air is forced upwards towards the continental divide.

Field is located within the Kicking Horse Valley, a major east-west trending valley which 'funnels' predominant westerly airflows, and therefore cyclonic depressions affecting the region. Although the Little Yoho valley is also orientated east-west, it is sufficiently removed both altitudinally and laterally from the main valley to be less influenced by major storm events. Similarly, the valley at Field rises to the east, whereas the Little Yoho valley declines to the east. Westerly airstreams therefore subside down the Little Yoho valley becoming increasingly stable. Periods of less intense rainfall, but of longer duration tend to characterise the area.

Winter is dominated by Continental Arctic and Maritime Polar air masses and consequently climatic conditions are harsh. The lowest mean monthly temperatures at Field are recorded in January (-11°C), although this masks daily variations which may be extreme, in the -30 to -40°C range. The variation of temperature with altitude is less marked during the winter months; approximately 3.5°C/1000 metres. Temperatures of -48 to -54°C at the 1500 to 3000 metre levels are possible, but quite rare (Janz and Storr 1977), although wind chill may considerably increase heat loss. The influence of the wind chill factor increases with elevation. No winter climatic data are available for Peyto Glacier.

Inversions are common due to the ponding of cold Arctic air in major valleys, displacing milder air to higher elevations.

Solid precipitation is directly related to variations in altitude. Field receives a mean total snowfall of 0.24 metres (water equivalent), with peaks in December/January of 0.063 metres (water equivalent). At Peyto Glacier perhaps the best indicator of winter precipitation can be obtained from the net specific winter glacier accumulation. Average figures for increasing altitude range from 0.63 metres (water equivalent) at the 2100 to 2200 m.a.s.l. level to 2.44 metres (water equivalent) between 3100 to 3200 m.a.s.l. (Young, in press).

The Emerald Glacier firn basin reaches an altitude of less than 2800 m.a.s.l., and thus a mean net accumulation of about 1.64 metres (water equivalent) might be anticipated, according to corresponding Peyto Glacier data. Averaging for the 2100 to 2800 m.a.s.l. range, a value of 1.2 metres (water equivalent) may be expected for Emerald Glacier as a whole, about five times that of Field.

In direct response to both winter and summer climatic conditions, a glacier's mass-balance is a reflection of its 'health'. Mass balance data is available for Peyto Glacier over a period of 13 years from 1966 to 1978 (Young and Stanley 1976, Young in press).

Results suggest that the position of the mean equilibrium line at Peyto Glacier may be fixed at approximately 2630 m.a.s.l. On this basis positive mass balances have been recorded in 1966, 1967, 1968, 1973, 1974 and 1976. On each occasion net mass balances were small, ranging from 0.01 metres (water equivalent) (1967) to 0.64 metres (water equivalent) (1976). In contrast, large negative balances have been recorded e.g. -1.70 metres (water equivalent) in 1970 and -1.05 metres in 1977.

The altitude of the Emerald Glacier firn basin backwall is approximately 2750 m.a.s.l., and should therefore exhibit strongly negative mass balances. However, the existence of the glacier at the base of cliffs rising up to approximately 3109 m.a.s.l. will have a direct affect on accretions of snow to the glacier surface, through the development of leeside snowdrifts and avalancing. Mass balance figures may therefore be comparable to the Peyto Glacier situation.

Clearly however, climate/glacier relationships are very complex and as such the transposing of Peyto results directly to Emerald Glacier may be a dubious practice (Young, personal communication). Although conclusions are of a guarded nature, the least to be expected is that the trends exhibited at Peyto Glacier may be evident at Emerald, although absolute values may differ.

2.4 Vegetation

Vegetation is sparse in the near terminal area, being developed in till exposed since the neoglacial maximum 150 - 270 years B.P. (see section 2-5). Bray (1964) conducted a vegetation study on President Glacier, approximately

5 kilometres to the west of Emerald, within the area defined by the last ice advance. Picea engelmannii Parry, Abies lasiocarpa Hook, and several varieties of mosses and flowering plants of which Dryas drummondii Richards, Dryas hookeriana Juz and Salix arctica Pall dominated, were identified.

Due to the proximity of President Glacier and the similarity of aspect and climate, it is assumed that the succession of colonising species will be similiar at Emerald Glacier.

The area immediately outside the Neoglacial limit is one of alpine meadow, which is gradually colonising the outer moraines. The meadows lie above the tree line, which is at 2280 m.a.s.l. on the south facing slopes of the Little Yoho valley and 2160 m.a.s.l. on north facing slopes. Below it is an interior subalpine forest type (Rowe 1977). Western White Spruce, Engelmann Spruce and their intermediate forms coexist with minor stands of larch and Alpine Fir, which increase in abundance at higher altitudes and is dominant at treeline.

2.5 Recent Glacial History

Final retreat of late Wisconsinan ice from the main ranges of the Rocky Mountains has been assigned a date of 9300 ± 170 years B.P. (Westgate and Dreimanis 1967). The age was derived from a charcoal deposit overlying till at Saskatchewan Crossing located between Yoho National Park and the Columbia Icefield. However, this may be regarded as a minimum date, since Harrison (1976) provides evidence for the Kananaskis Valley being ice free since 12000 years B.P..

In the Holocene, 2 major readvances have been evidenced: the Crowfoot and Cavell Advances (Luckman and Osborn 1979). Although others are suggested e.g. a Chateau Lake-Louise advance (Harris and Howell 1977), and 4 advances in the Cataract Brook Valley in Yoho National Park (Fox 1974), evidence is refutable or inconclusive (Luckman, Osborn and King 1978, Luckman and Osborn 1979).

The placing of the major advances within a temporal framework has been facilitated by tephro- and dendro-chronology. The older Crowfoot Advance has been assigned an age of greater than 6600 years B.P., based on the Mazama Ash fall which overlies Crowfoot deposits.

Morphological evidence has come from till deposits underlying the later Cavell Advance deposits, rock glaciers and moraines where they lie outside the Cavell Advance, and vegetation and/or lichen contrasts between the older and younger deposits.

The Cavell Advance is characterised by well-developed, fresh moraines close to present day ice margins, such as those at Emerald Glacier (Fig. 2-4). Dendrochronology has provided temporal limits e.g. Bray (1964) on President Glacier dated the outermost moraine at 1714 A.D., and the innermost at 1832 A.D. These data correlated well with those from other areas (Heusser 1956, Bray and Struick 1963, Gardner 1978) which suggest several periods of major advance, or advance and readvance: late 17th to early 18th Century; early to mid

Figure 2-4 Emerald Glacier as viewed from Whaleback Mountain. The extent of the Cavell neoglacial advance can be seen in the form of large frontal moraines. The extent of debris covered ice on the current glacier surface is also visible.

19th Century; and Late 19th to early 20th Century. These periods of activity may be expressed either as a series of closely spaced relatively small terminal moraines (e.g. President Glacier) or one, single massive moraine (e.g. Emerald Glacier), depending on the geomorphic setting (Luckman and Osborn 1979).

The last century has been characterised by general glacier recession, although fluctuations have occurred (Gardner 1972). The pre-1930 period was marked by gradually increasing recession rates (between 5-15 metres/year), which peaked between 1930 and 1950 at 30-50 metres/year. Since the 1950's a general decrease in recession rates is reported (Gardner 1972). Meier and Post (1962) recorded that 52% of glaciers they studied were retreating, 43% active (i.e. uncertain), 5% stagnant and 0% advancing.

Since then, recent estimates have suggested that up to 50% of Rocky Mountain glaciers are now stationary or advancing. Indeed, 5 out of the last 7 years have recorded increasing or stationary values for mean net specific annual balance at Peyto Glacier (Young in press). Similarly, the contemporary nature (see p. 99) of the Emerald Glacier moraine suggests a current stillstand or readvance of glacier ice.

Fluctuations in glacier activity have been explained in terms of mean annual temperature and precipitation variations (Heusser 1956).

Chapter 3

Investigative Procedures and Methodology

Field Research

Field oriented studies were conducted over a 33 day period from July 24 to August 25 1979, although some time prior to this was spent in reconnaissance. A programme of intensive, as well as extensive data collection was initiated, with individual components focusing on particular aspects of the Emerald Glacier terminus and environs.

Contrasts between debris covered ice to the west of the proglacial lake and the clean ice surface to the south were of special interest (See Fig. 3-1).

3.1 Morphological Description

The terminal region was the subject of a plane-table survey with a self-reducing alidade on August 15. Although the intention was to survey ice slopes, the ice margin and moraine morphology, poor weather conditions during this stage of the field season allowed only the position of the moraine to be established (Fig 3-1). The morphology of the proximal

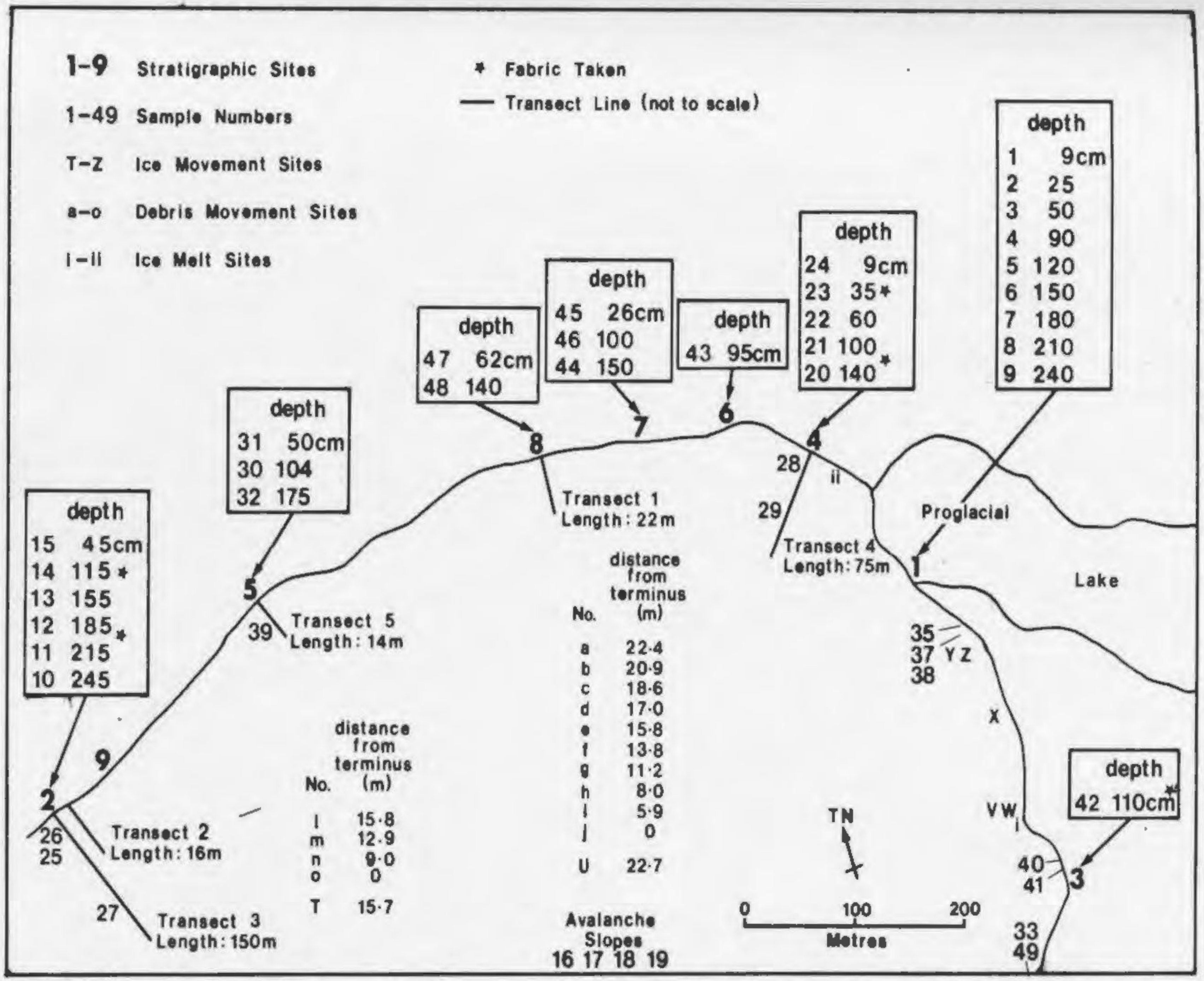

Figure 3-1 Map showing the position of the glacier margin and sample, stratigraphic, ice and debris movement, and ice melt sites.

ice surface was described from field sketches and ground photographs.

Morphological mapping of the moraines themselves was completed using a Brunton compass and graduated tape. Since the debris free area was one of constant terminal fluctuation, mapping was not attempted until late in the field season. Contrasts of moraine form led to the proposition of the names Moraine A, for that section west of the proglacial lake, and Moraine B to the south. Moraine A is the most extensive.

Stratigraphic sectioning was undertaken at 9 points in the moraine; 7 in Moraine A and 2 in Moraine B. Vertical sections were cut normal to the moraine axis, and extended laterally into the moraine to negate any disturbances as a result of slumping. Stratigraphic units were noted and described according to thickness, colour (Munsell soil colour chart), pH (portable pH kit) and the general textural appearan including the size and angularity of component clasts. Any obvious fabric orientation or imbrication was also recorded.

3.2 Fabric

Five detailed fabric diagrams were constructed, from within Moraine A and Moraine B stratigraphic units. Imbrication and orientation of clasts were measured by a Brunton compass. Avoidance of any stratigraphic unit margins was maintained to minimise disturbance. Fabric was taken within a small area (approximately 15 cm square) to reduce in-site variability (Young 1969, Andrews 1971). For each three-dimensional fabric, 50 clasts with an 'a' axis longer than 1.5 cm, and with an a:b axis ratio of greater than 3:2 were selected.

Results were plotted on a Wulff net and contoured on a stereonet, following the procedures outlined by Ragan (1968). The statistical significance of each fabric was tested using the Poisson test against randomness (Briggs 1977).

3.3 Ice Movement and Ablation

In order to establish the rate and pattern of ice movement and longitudinal strain, seven stakes were positioned in the near terminal area; two proximal to Moraine A and five proximal to Moraine B (constraints on the sampling procedure are discussed in Appendix D). Measurement was by a 30 metre tape relative to a fixed marker (e.g. a boulder) in the proglacial area. Rapid ablation meant that stakes had to be constantly resurveyed, but several days' measurements were lost as a result of melt-out. Data was collected on a daily basis whenever possible, from July 29 to August 25 inclusive. Direction of movement was consistently normal to the ice slope.

Accompanying the ice movement studies, contrasts in the rate of surface ice-melt between debris-rich and debris-free ice were also measured. Two sites were selected, one in each area (Fig 3-1). Measurement was against a fixed stake.

3.4 Supraglacial Sediment Investigations

Measurements of supraglacial sediment movement were made along two transects, one proximal to stratigraphic site 8 (transect I), and the other proximal to site 2 (transect II) (Fig. 3-1). Both transects were normal to the ice slope, and involved the marking of clasts of variable shape and size at increasing distance (up to 25 metres) upglacier of the moraine. Their movement was monitored relative to a fixed marker outside the glacier margin. In total, eleven clasts were selected along transect I and four along transect II, included in which were clasts situated on the moraine crest. An ice movement stake was also surveyed along each transect.

The depth and stratigraphy of supraglacial sediments were measured along three transects; proximal to site 2 (transect III), site 4 (transect IV) and site 5 (transect V). Each traverse was normal to the ice slope, and continued up to where the supraglacial debris cover became discontinuous. This occurred at approximately 150 metres along transect III and 75 metres along transect IV. Transect V was terminated at a distance of 14 metres upglacier of the moraine, due to the unstable condition of the supraglacial debris mantle.

Vertical sections were cut at 2 metre intervals within the first 20 metres proximal to the moraine, and subsequently at intervals varying from 5 to 20 metres. Within each sectio the depth of sediment was measured and any stratigraphic units noted and described on the basis of texture. Surface slope was measured using a Brunton compass, as were imbrications of surface clasts.

Increasing downglacier deviation of clast imbrication away from the angle defined by the surface slope was noted within the 10 metres immediately proximal to the moraine. Clast imbrication was therefore measured at 1 metre intervals from the moraine crest to a position about 10 metres up the ice slope. Twenty clasts were selected along a horizontal plane delineated by a tape and their imbrications measured by a Brunton compass.

3.5 Sediment Sampling

Each stratigraphic unit within the moraine was carefully examined and sampled. The procedure involved the clearing of a face in order to limit the influence of slumping, the identification and description of component units and the removal of approximately 0.5 kilograms of till matrix. Samples were stored within double sealed plastic bags and carefully labelled. In several cases, a series of samples were extracted vertically through a single stratigraphic unit Any variation in texture through the profile could therefore be determined.

Similarly, a series of sediment samples were taken from various glacier sediment/transport environments. In situ supraglacial samples were not only extracted along transects III to V, but also from avalanche slopes located along the backwall of the firn basin (Fig. 3-2). Samples from this active sedimentary environment could be compared with the near-terminal debris.

In situ subglacial samples were collected from locations proximal to Moraine B, where the ice had decoupled from the moraine during the melt season.

Laboratory Analyses

Techniques centred on a variety of tests undertaken on the forty-nine collected samples. In the laboratory, samples were split leaving approximately 300-400 grams for grain-size analysis.

3.6 Grain-size analysis

Analysis involved dry sieving of the coarser than 4 Ø fraction and pipette analysis of the finer than 4 Ø fraction, following standard techniques outlined by Griffiths (1967) and Bowles (1978) (Appendix A).

The total grain-size distribution was plotted on a Hewlet Packard desk top calculator-plotter using a programme devised by Slatt and Press (1976). Plots of a grain-size distributior histogram and cumulative curve were made, as well as the computing and printing of statistical parameters outlined

Figure 3-2 Avalanche slopes located along the backwall of the firn basin, July 1979.

by Folk and Ward (1957) (Mean, standard deviation, skewness, kurtosis and normalised kurtosis). Percentages of gravel (-8 Ø to - 1 Ø), Sand (-1 Ø to 4 Ø), silt (4 Ø to 8 Ø) and clay (finer than 8 Ø) were also presented.

Although a potential range of values from -8 Ø to finer than 9 Ø could be measured, only the finer than -2.5 Ø range was considered. Several reasons for this exist : firstly, the character of glacial sediments is best exemplified by the texture of the composing matrix, and as such the very coarse fraction is largely superfluous; secondly, when sampling a matrix in which large clasts are included, biasing by including either too much or insufficient of the coarse fraction can occur. A large coarse fraction has the effect of masking any subtle variations of texture in the fine size ranges; and thirdly, the practical consideration that inclusion of a large coarse fraction would necessitate a large sample to be collected and subsequently carried from the field area. As such, the size range examined was -2.5 Ø to finer than 9 Ø.

Standard statistical parameters of the grain-size distribution curve were analysed according to the 'method of moments' (Friedman 1961). Friedman's four moments are mean (first moment), standard deviation (second moment), skewness (third moment) and kurtosis (fourth moment). A graphical combination of these moments was observed to differentiate between the depositional environments of sand (beaches and dunes), although the technique has been applied to the investigation of glacial deposits e.g. Schlucter (1977), Vorren (1977) and Eyles and Rogerson (1978). It is in an attempt to differentiate between glacier sub-environments that this method will be used.

3.7 Atterberg Limits

Investigations into these geotechnical parameters were prompted by Boulton and Paul (1976), who characterised glacier sub-environments by their Atterberg limits. This follows from the consideration that different glacial environments e.g. supraglacial versus subglacial, are characterised by different proportions of clay-sized material and will therefore exhibit contrasting Atterberg limit values.

Atterberg proposed five states of soil consistency (Terzaghi and Peck 1967), of which three are most used: the liquid limit, which may be described as the 'water content above which the soil behaves as a viscous liquid' (Bowles 1979); the plastic limit, defined as 'the water content below which the soil no longer behaves as a plastic material but as a solid' (Bowles 1979); and the plasticity index, 'the range of water content within which a soil possesses plasticity' (Terzaghi and Peck 1967), which is defined by the difference between the liquid and plastic limits.

Atterberg limits were calculated using the standard techniques outlined by Bowles (1978) (Appendix B).

3.8 X-Ray Diffraction

Analysis of the finer than 4 Ø fraction of till and artificially crushed parent rock samples using X-ray diffraction was completed for two prime reasons. Firstly, the type of clay mineral affects the plasticity of a soil (Seed et al. 1964, Wu 1976, Bowles 1979), and therefore the identification of clay minerals would be a useful asset; and secondly, a study by Rieck et.al. (1979) differentiated till from two seperate ice lobes on the basis of the intensity of the 7Å: 10Å peak ratio. It was thus considered plausible that some differentiation of glacier sub-environments could be made on this basis. The potential for releasing clay-minerals, as well as clay-sized material, would presumably differ between a subglacial and supraglacial environment.

The analysis of clay minerals was facilitated by a Phillips diffractometer and followed the procedures outlined by Grim (1962), Carroll (1970) and Brown (1972) (Appendix C). Slides were heat treated to test for the presence of kaolinite, and glycolated to detect the presence of montmorillonite.

The identification of clay minerals followed the criteria outlined by Carroll (1970).

The laboratory analyses presented above do not represent the sum total of possible tests. Often it was considered that additional procedures would not add to the characterisations already inferred, but simply be 'noise'. Nevertheless, certain potentially useful techniques were not utilised and thus an explanation is warranted.

Clast shape analysis, although undertaken to a minor degree in the field was not expanded in the laboratory to include further samples. It was considered that shape, although a useful descriptive aid, would carry few genetic implications for the short-transported (approximately 1 kilometre) material characteristic of Emerald Glacier. Dreimanis and Vagners (1972) note the important effect of transportation on clast shape, as does Boulton (1978) but other aspects are equally important. Lithology is especially significant in the Emerald Glacier situation, where a large proportion of the source rock is shale, which disaggregates into platy clasts. They are prone to a sliding rather than a rolling process of movement in both subglacial and supraglacial transport systems. Given these conditions clast shape was considered not to be a significantly variable factor between glacier sub-environments to be included in this investigation.

Scanning Electron Microscopy (SEM) of quartz grains has been used in attempts to characterise glacial sediments (e.g. Whalley and Krinsley 1974, Eyles 1978, Whalley 1978).

However, although certain characteristics can be attached to a 'glacial' grain and a differentiation between supraglacial, high-level and basal, low-level transport systems may be possible (Eyles 1978), the technique was considered as perhaps a tool for corroborating data derived by other means. Similarly, the short transport distance of grains within the Emerald Glacier system may not be sufficient to reveal a distinct 'texture' on grains. Furthermore, quartz is present, but scarce in the lithologies studied. The inclusion of SEM analysis was therefore considered unwarranted.

The calculation of the shear strength of Emerald Glacier deposits was a proposed measure. Indeed, a differentiation between supraglacial and subglacial deposits on the basis of preconsolidation pressures may be expected. However, the unavailability of field shear test instrumentation, and the impracticality of conducting shear tests on disturbed samples, meant that this investigation could not be undertaken

Chapter 4

The Characterisation of the Ice Margin

4.1 Morphology

The margin of Emerald Glacier can effectively be subdivided into two distinct units. To the west of the proglacial lake, the ice surface is covered by a continuous mantle of supraglacial debris. To the south, a discontinuous cover or no cover at all was noted. Similarly, contrasts in moraine morphology were evident, the proglacial lake being situated approximately at the junction between the two areas. The lake accounted for a 50 metre moraine-free section in 1979, where calving of the ice margin occurred although this had decreased to less than 5 metres in 1980 (Rogerson, personal communication).

In 1979, Moraine A[1] (Fig. 4-1) ranged from 1.7 to 2.7 metres in height measured from the distal base to the crest. Immediate proglacial slopes were slight between sites 4 and 8 (9°), increasing to 23° further to the south-west. At its maximum, Moraine A was 67.8 metres (site 2) above the level of the proglacial lake. Glacier ice consistently lay immediately

[1] See page 35 and Figure 3-1 for location of Moraine A and Moraine B.

Figure 4-1 Moraine A and heavily debris covered proximal ice. This margin had advanced by several metres in summer 1980.

proximal to Moraine A throughout the field season. Moraine A exhibited pronounced asymmetry with distal slopes of about 50° compared with proximal slopes of only 17°-30°. Ice proximal to Moraine A was debris covered, with surface slopes of 25° - 30°.

In 1979, Moraine B (Fig. 4-2) was 1.8 to 2.4 metres high and increasingly separate from the proximal, debris-free glacier margin, throughout the field season. The moraine possessed abnormally steep slopes (maximum 69°), although no asymmetry was encountered. Oversteepened slopes of this type are often noted as characteristic of moraines (e.g. Whalley 1975). Total relief above the proglacial lake was less than for Moraine A, reaching only 17 metres. The southern end of Moraine B is marked by an ice cliff resting on a topographic shelf; an expression of the Wapta Mountain Thrust.

4.2 Stratigraphy

Sections cut in Moraine A revealed distinct stratification (Fig. 4-3). Four major units were identified on the basis of colour and textural variation.

A thick (19-90 cm), dark grey (5Y/4/1) to light olive grey (5Y/6/2) coarse to fine gravel deposit, lacking any visually obvious fabric overlay a thin (9-19 cm) olive grey (5Y/4/2) sandy-silty lens. This in turn rested on a 25-82 cm light brownish grey (2.5Y/6/2) to pale olive (5Y/6/3) silt-clay

Figure 4-2 Moraine B at the margin of debris free proximal ice. The ice cliff marks the southern end of the moraine.

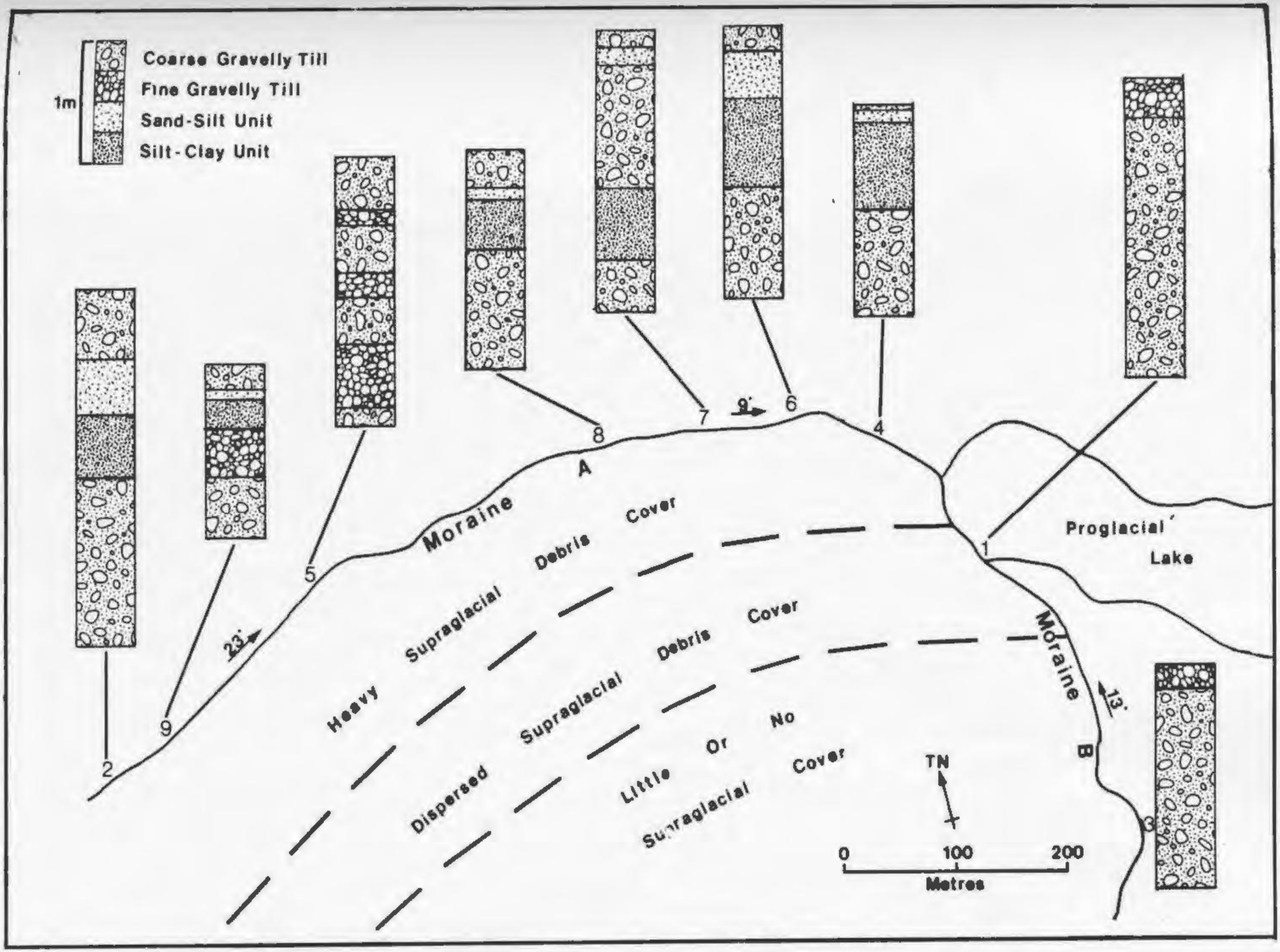

Figure 4-3 Map showing the moraine stratigraphy and proximal ice debris cover.

deposit. A distinction on the basis of inherent fabric and particle shape could also be made, with preferred imbrication and roundness being exhibited by the silt-clay layer, whereas the upper two deposits possessed no such characteristics.

The lowermost horizon was a 40-136 cm dark grey (5Y/4/1) to light olive grey (5Y/6/2) coarse to fine gravel. It graded downwards into large clasts (many greater than 0.5 metres in diameter), which created large cavities at the base of the moraine's distal side. These were often filled with snow-bank ice and snow.

Moraine B exhibited an essentially simple stratigraphy. A colour change from olive (5Y/5/3) to pale yellow (5Y/7/3) was however evident in the upper 17 cm of site 3 and the upper 33 cm of site 1. Component clasts within Moraine B showed no significant preferred roundness but a strong fabric, on the basis of visual interpretation. No snow was found at the base of the stratigraphic units, although evidence of snow accumulation within the moraine itself existed, in the form of small scale thermokarst features.

Stratigraphic sectioning was also undertaken within supraglacial debris proximal to Moraine A. The possibility of stratigraphic divisions within supraglacial debris has been emphasised by Eyles (1979), who outlined a washing process whereby fines are concentrated immediately above the ice surface as a 'skeletal soil'. The existence of such an

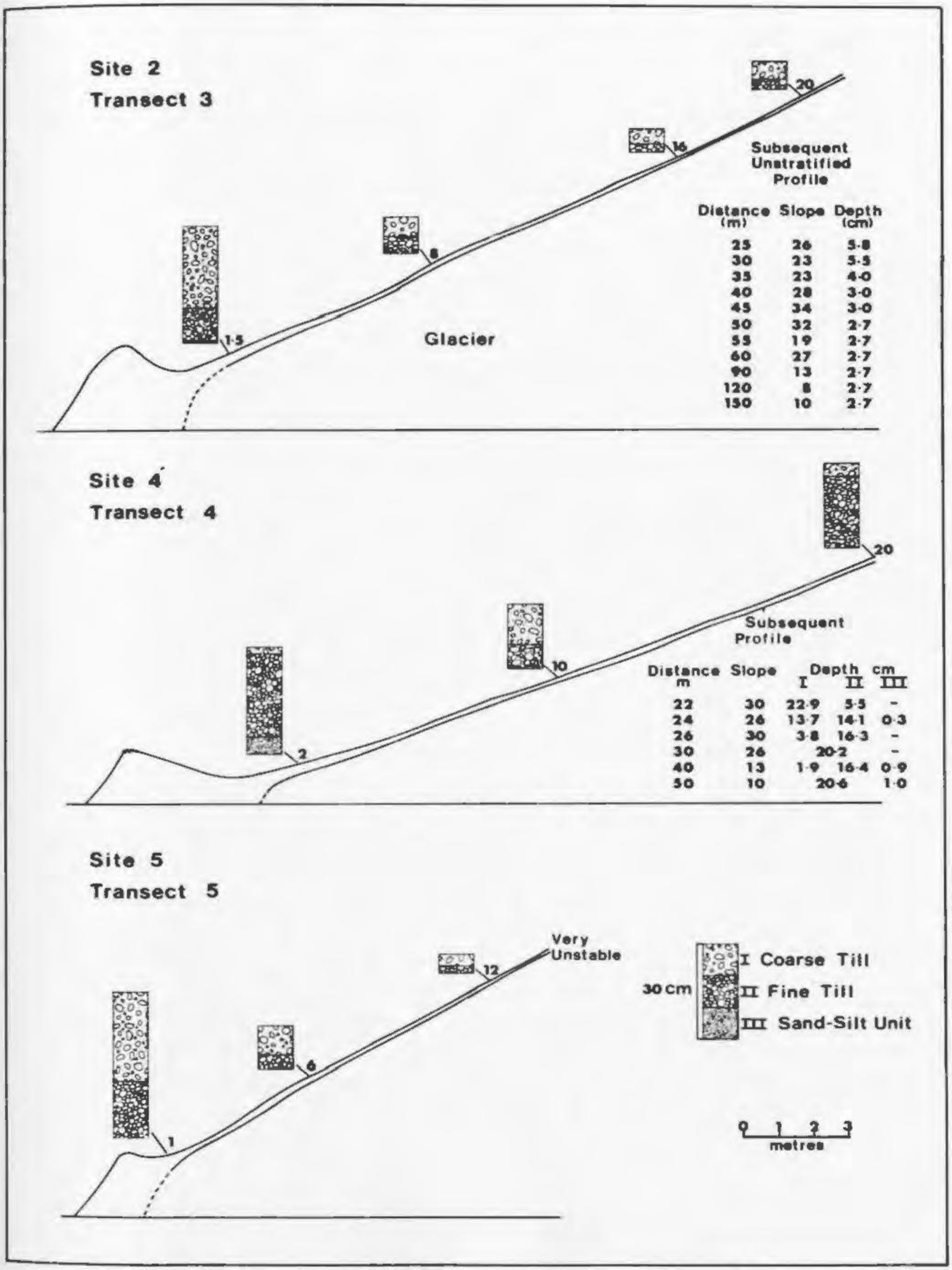

Figure 4-4 The stratigraphy of supraglacial debris at Emerald Glacier

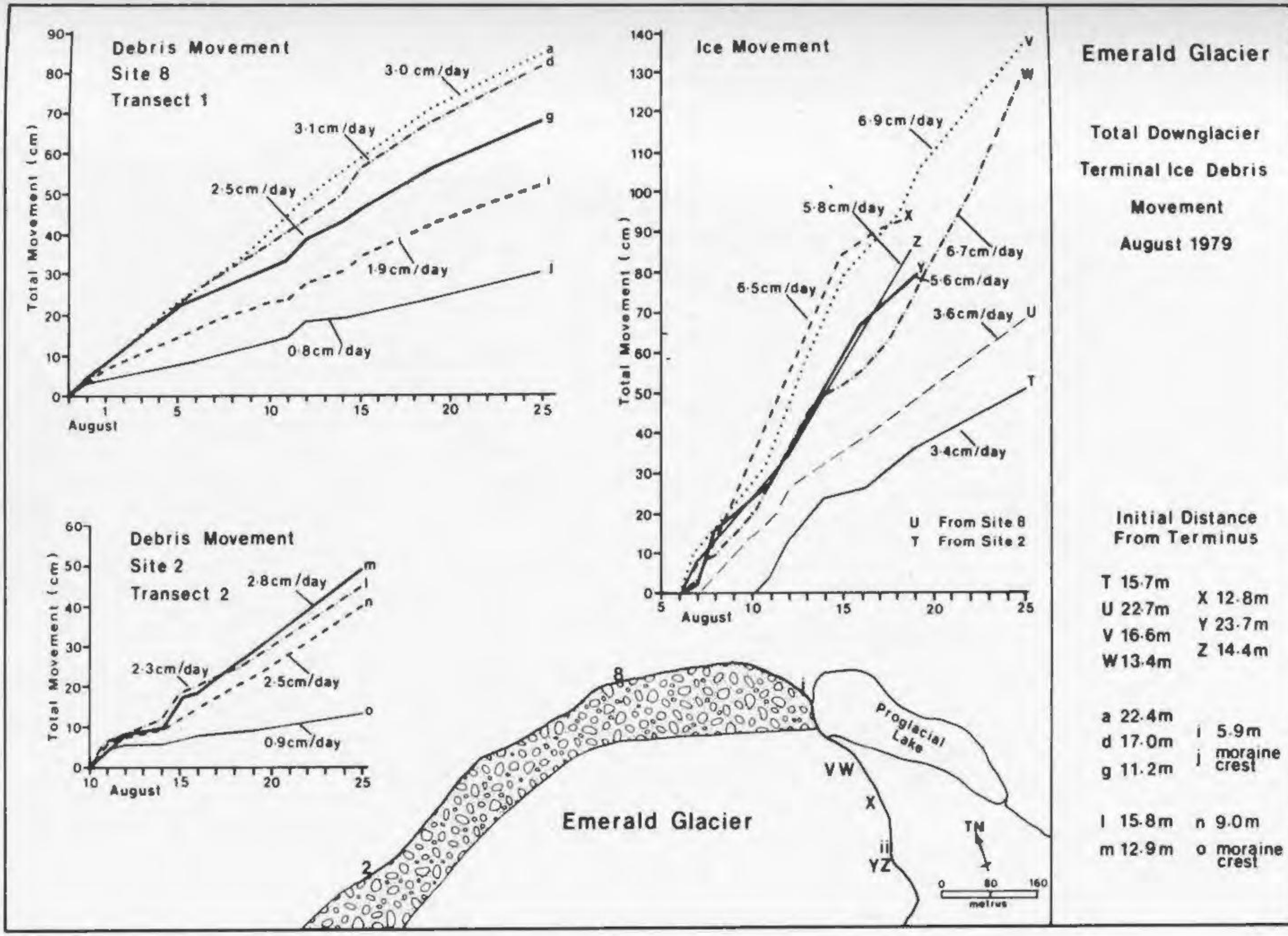

Figure 4-5 Terminal ice and supraglacial sediment dynamics, July to August 1979.

Values of 3.58 cm/day and 3.44 cm/day were recorded for stakes proximal to Moraine A, whilst increased rates were measured on debris-free ice (6.90, 6.65, 6.50, 5.64 and 5.77 cm/day for stakes V to Z respectively).

Debris movement results proximal to sites 2 and 8 indicate decreasing movement towards the margin. Velocities decreased from 3.0 cm/day at 22 metres upglacier of the proximal side of the moraine, to 0.8 cm/day on the moraine crest itself. The similarity of debris and ice movement results suggests that surface clasts do not move independent of glacier ice. As such, debris movement data may be considered as minimum ice movement rates, since ice velocity cannot exceed that of overlying debris. Increasing compressive longitudinal strain towards the terminus is thus confirmed.

4.4 Grain-Size Distributions

Comparison of the textural characteristics of in situ deposits with those composing the moraines should provide insights into possible formative processes, or at least genetical environments.

a) In situ supraglacial debris

In general, samples may be characterised as poorly to very poorly sorted, very positively skewed and leptokurtic. Mean size is coarser than 0 Ø and the silt/clay content is less than 11% (Fig 4-6 and Table A-2).

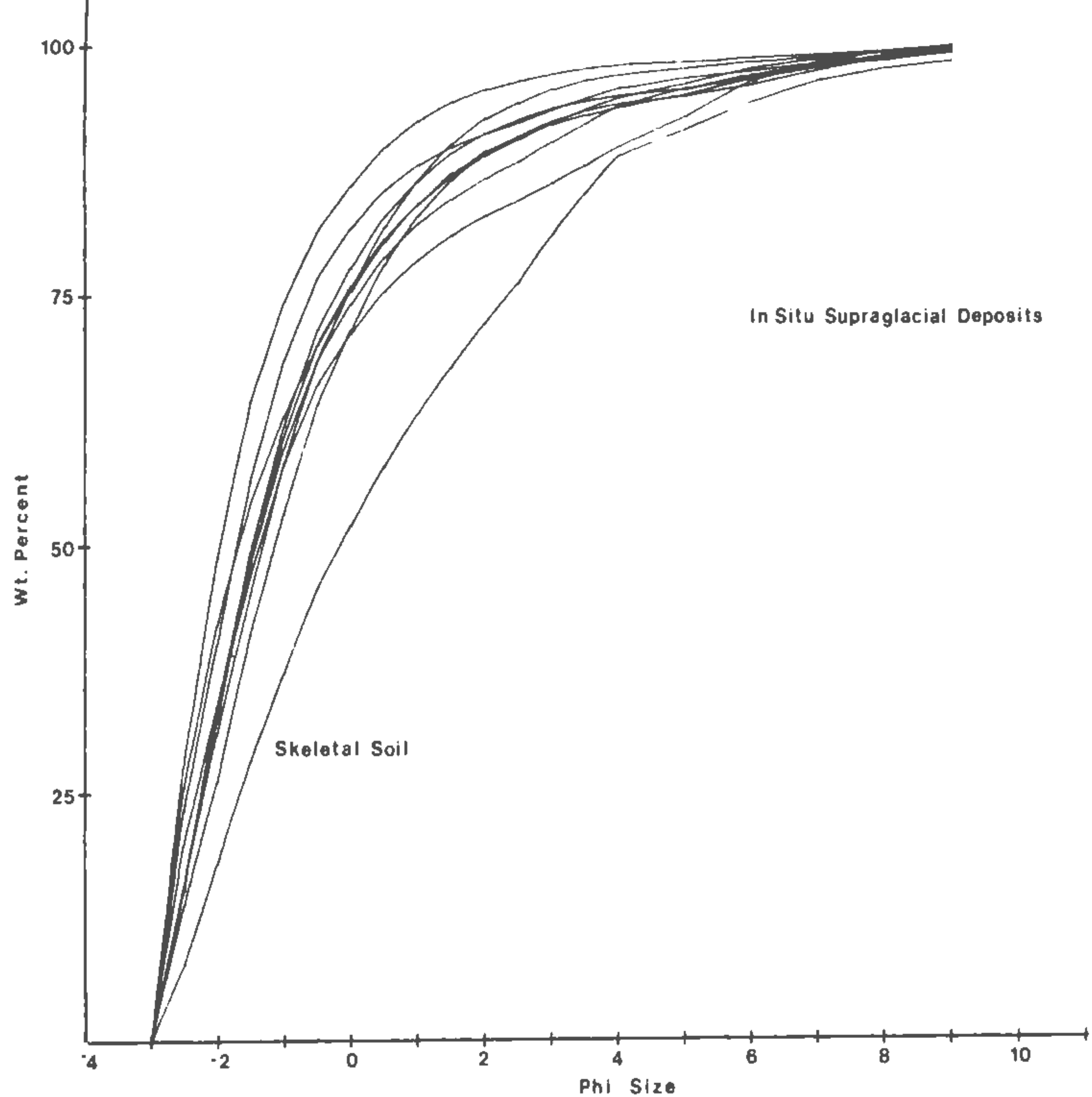

Figure 4-6 Grain-size distribution of in situ supraglacial deposits from Emerald Glacier

Eyles (1979) compounds descriptions from various authors and defines a supraglacial morainic till as possessing "...... a uniform particle-size distribution with the mode generally at about -7 Ø, a coarse mean size, poor sorting, fine skewness, a silt-clay content generally below 15%, predominance of angular oxidised clasts that lack preferred orientation, glacial shaping and striations". These characteristics have been described by Flint (1971), Knighton (1973), Boulton (1978), Gardner (1978), Boulton and Eyles (1979), Nakawo (1979) etc. Considering the constraints placed upon the sampling procedure (Appendix A), supraglacial debris found at Emerald Glacier encompasses many aspects of the definition.

The grain-size distribution of supraglacial morainic till is derived from its mode of origin and subsequent transportation. Within temperate glacier systems, supraglacial material is generally derived from valley sides or headwalls. Transport, either supraglacially or as englacial debris septa, entails minimal inter-particle activity and thus little comminution (Eyles and Rogerson 1978b, Boulton and Eyles 1979).

Nakawo (1979) however, working on the G2 Glacier, Nepal notes decreasing particle-size downglacier. This he explains by the addition of fines contained within glacier ice to the overlying supraglacial mantle upon melt-out. Essentially, this is the same process outlined by Andersen and Sollid (1971) and Boulton (1977), who evidence the outcropping of debris-rich thrust planes near the terminus following marginal freezing of the glacier to its bed in winter.

Nevertheless, the grain-size distribution of supraglacially derived and en- or supra-glacially transported material remains that of the parent rockfall. Eyles and Rogerson (1978b) and Slatt (1971) consider this to be independant of lithology, although it is disputed by Mills (1977a).

Boulton (1978) has suggested a crushing to abrasion (C/A) ratio to differentiate between material which has undergone a glacial tractive phase and that which has not. Crushing derivatives are coarser than 1 Ø, while abraded material dominates finer than 1 Ø. Thus, the higher the ratio, the greater the crushing and the more likely is the material to have been supraglacially transported. Values are in the range zero to infinity. Results from Søre Buchananisen and Breidamerkurjokull (Iceland) and Glacier d'Argentière (French Alps) however, suggest values of 0.5 to 4.2 characterise tractive phase material (peaks at 3.0, 0.9 and 1.4 respectively), whilst 2.0 to 10 is representative of high level transport (peaks at 5.3, 3.1 and 5.7 respectively) (Boulton 1978 p. 793-794). In situ supraglacial samples from Emerald Glacier ranged from 3.6 to 12.0.

An anomaly in the supraglacial grain-size distribution curves is sample 25, which has a mean size of 0.39 Ø and an 11.3% silt/clay content. This sample corresponds to the 'skeletal soil' described earlier.

b) In Situ Subglacial Debris

These deposits may be characterised as very positively skewed, very poorly to extremely poorly sorted, and platykurtic to mesokurtic. They have silt/clay contents ranging from 18.4 to 44.9% and an average mean size of 1.5 Ø (Fig. 4-7 and Table A-2).

Characteristics of subglacial tills are a result of their patterns of transport and deposition. Within temperate glaciers sub-glacial transport is by a thin (10-20 cm) layer immediately above the glacier sole and deposition is normally by a lodgement process (Boulton 1976). This occurs "when the frictional drag between clasts in traction and the bed over which they move is such that the force imposed on them by the moving ice is insufficient to maintain them in motion" (Boulton 1975).

In contrast to the unimodal distribution of supraglacial debris, in situ subglacial deposits from Emerald Glacier all illustrate poorly developed bimodal or multimodal tendencies. A major peak occurs between 6 and 7 Ø, and minor ones are occasionally evident at about -2 Ø and 3 Ø.

Bimodality in tills has been discussed by Dreimanis and Vagners (1971, 1972) and Elson (1961) as being a function of the nature and distance of transport, and the structure of the parent rock. They note for instance, that limestone

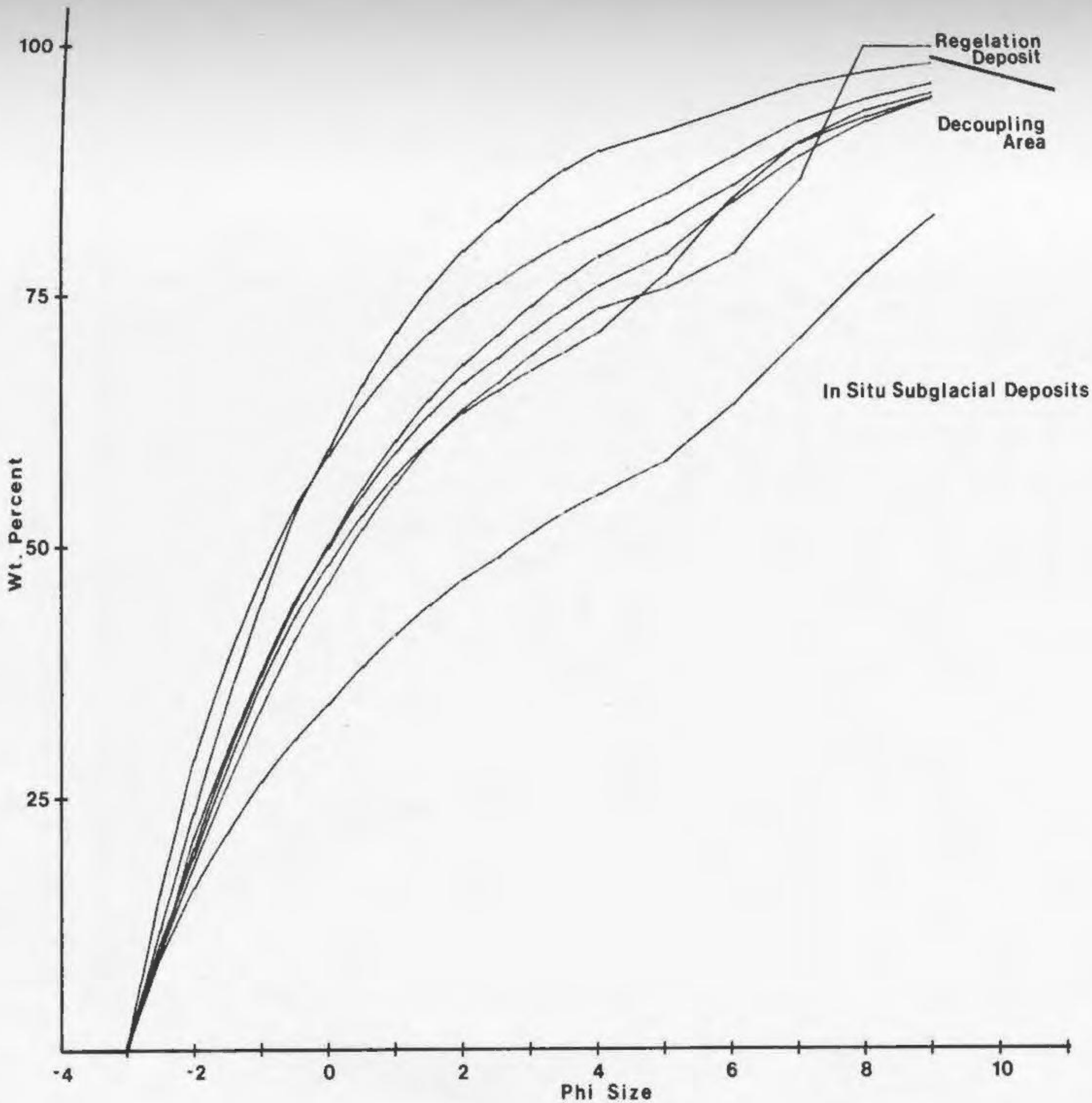

Figure 4-7 Grain-size distributions of in situ subglacial deposits from Emerald Glacier.

and dolomite derived tills have a silty matrix; predominantly calcite (7-10 Ø) and dolomite (4-6 Ø) minerals (Dreimanis 1969). Shale derived tills have a silt-clay matrix. Slatt (1971) however, de-emphasises the importance of lithology with respect to tills from small valley glaciers, and asserts the transportational environment provides the major control. Eyles and Rogerson (1978b) in general support this finding with respect to the coarser than 4 Ø fraction.

Concentration of minerals around the 'terminal grades' is an ongoing process from incorporation into the glacier system (Dreimanis and Vagners 1971). Distance of transport and thus traction, is an important factor. Nonetheless, the breakdown of soft sedimentary rocks under conditions of basal sliding may be rapid, as suggested by Rogerson and Kodybka (1980). Thus, even the Emerald Glacier tractive system, which is only approximately 1 kilometre in length may produce initial terminal grade concentration and thus bimodality.

An anomaly in the in situ subglacial deposits is the regelation layer sample. Although it has very similar grain-size distribution characteristics to other subglacial deposits, the regelation sample only contains 0.25% clay-sized material, at least 10 times less than any other in situ subglacial sample.

Regelation ice in temperate glaciers is a seasonal phenomenon, occurring during the accumulation season. It thus contrasts with polar and sub-polar ice masses in which regelation ice exists as a spatial zone. The mechanism of regelation has been described as a freezing-on process (Boulton 1975, Sugden and John 1976), effective only for small particles. Grain-size analyses have, to the author's knowledge, not been presented. As a constituent of cold based ice, regelation material is presumably associated with melt-out till, bu as a distinct unit of temperate glaciers this element has received little attention.

Regelation material is derived from underlying subglacial debris, and thus the question must be posed as to why the regelation layer clay-size proportions are far less than those of other subglacial debris samples. In response, Rogerson (personal communication) suggests the transfer of the winter freezing plane into underlying debris may have the effect of 'repelling' fine particles, eventually expelling them. However, until this proposal has been confirmed it remains tentative.

c) <u>Moraine A</u>

Grain-size distributions correspond to the four distinct stratigraphic units noted in section 4-2 (Fig. 4-8 and Table A-3).

The lower coarse gravel unit is a poorly sorted, very positively skewed, very leptokurtic deposit. It has a mean size of about -1.4 Ø, a silt/clay content of 6.9% and a C/A ratio of 6.4. This will be termed facies a.

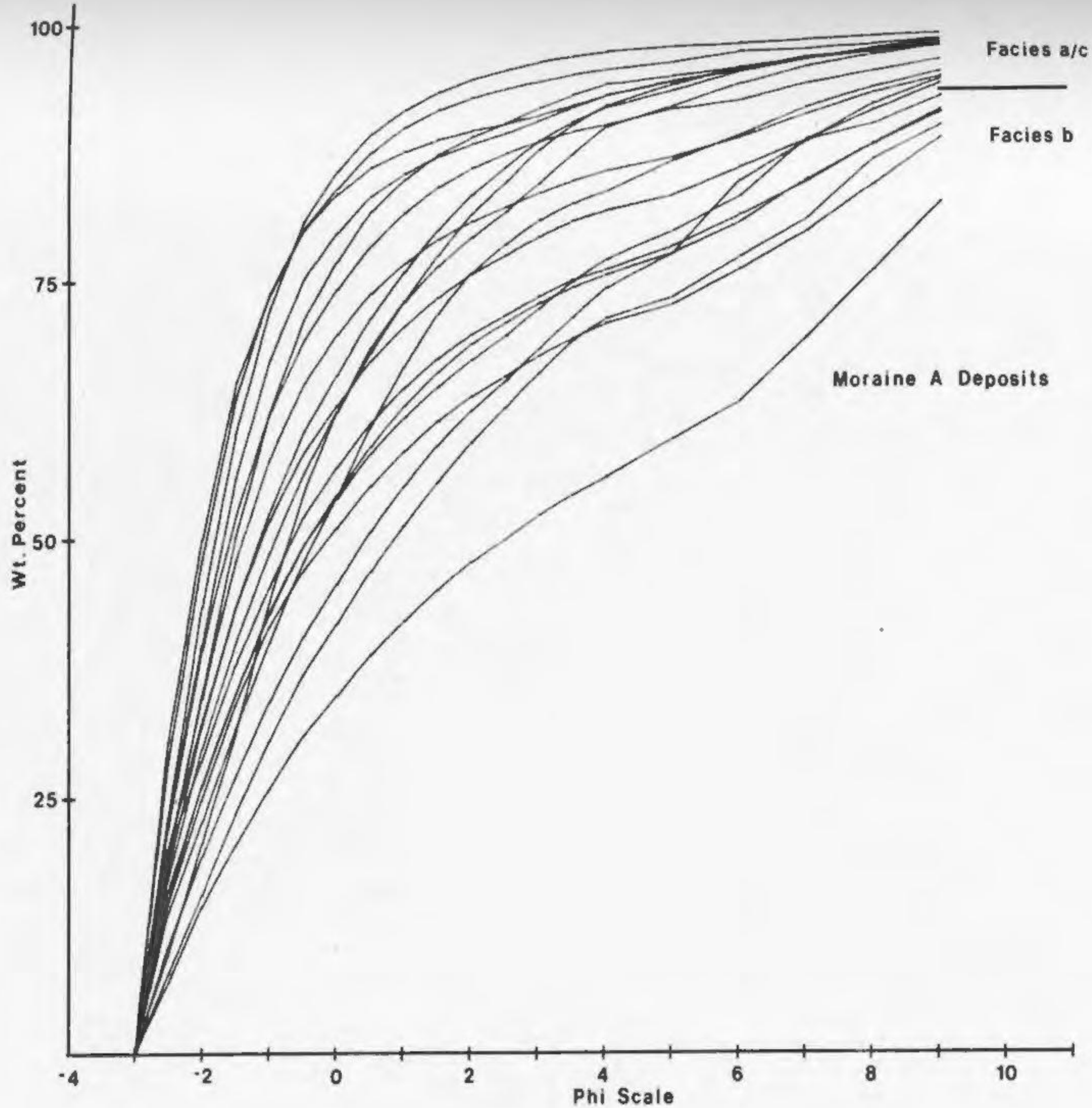

Figure 4-8 Grain-size distributions of Moraine A deposits

In marked contrast, the slit-clay lens is extremely poorly sorted, very positively skewed and platykurtic. It has a mean size of 2.1 Ø, a silt/clay content of 28.6% and a C/A ratio of 1.05. Since it exhibits a character significantly different from other units in the section, it is designated as a separate facies: facies b.

The overlying coarser deposit can be sub-divided into two; a lower sandy-silty lens (facies c) with a mean size of -0.26 Ø, a silt/clay content of 9.2% and an average C/A ratio of 2.91; and an upper coarse gravel deposit with characteristics similar to facies a. This upper unit will therefore be termed facies a_1.

Detailed analysis of sites 2 and 4 was undertaken. Both were stratigraphically very similar, although contrasts in corresponding facies units were evident. Facies a was more finely skewed and more poorly sorted at site 2 compared to the corresponding unit at site 4. A higher mean size (average 1.05 Ø greater silt/clay content (average 28.8%) and a lower C/A ratio (average 1.97) were recorded from site 2. Similarly, facies b is finer textured within site 2, although facies c shows little between site variability. Nevertheless, the within site variability maintains the facies distinctions outlined above. In no case were progressive variations down through a single facies unit noted.

d) Moraine B

Sedimentologically two distinct units are evident (Fig. 4-9 and Table A-3). Both are very poorly sorted and very positively skewed, but the upper (0-70 cm) unit is mesokurtic to leptokurtic, with a mean size of 0.91 Ø, a silt/clay content of 27.8% and a C/A ratio of 1.41. The lower unit, although not distinct in the field is platykurtic, with a coarse mean size (-0.3 Ø), lower silt/clay content (10%) and a higher C/A ratio (3.01). The lack of a complete separation however, necessitates the sub-division of a single facies. Thus, facies d and d_1 will denote the upper and lower units respectively.

e) Synthesis

Distinct and separate distributions are evident for in situ supraglacial and in situ subglacial deposits, based on the method of moments (Friedman 1961).

Kurtosis (third moment) and sorting (second moment) reveal increasing kurtosis with decreasing sorting (Fig. 4-10). The generally platykurtic and very poorly sorted (2-4 Ø) subglacial sediments are discrete from the essentially leptokurtic and poorly sorted (1-2 Ø) supraglacial sediments. This difference is related to the enrichment of fines due to subglacial comminution, compared to the lack of fines and predominance of sand and gravel-sized material in the supraglacial environment, where comminution is minimal. Kurtosis and mean (first moment) show a similar differentiation (Fig. 4-1

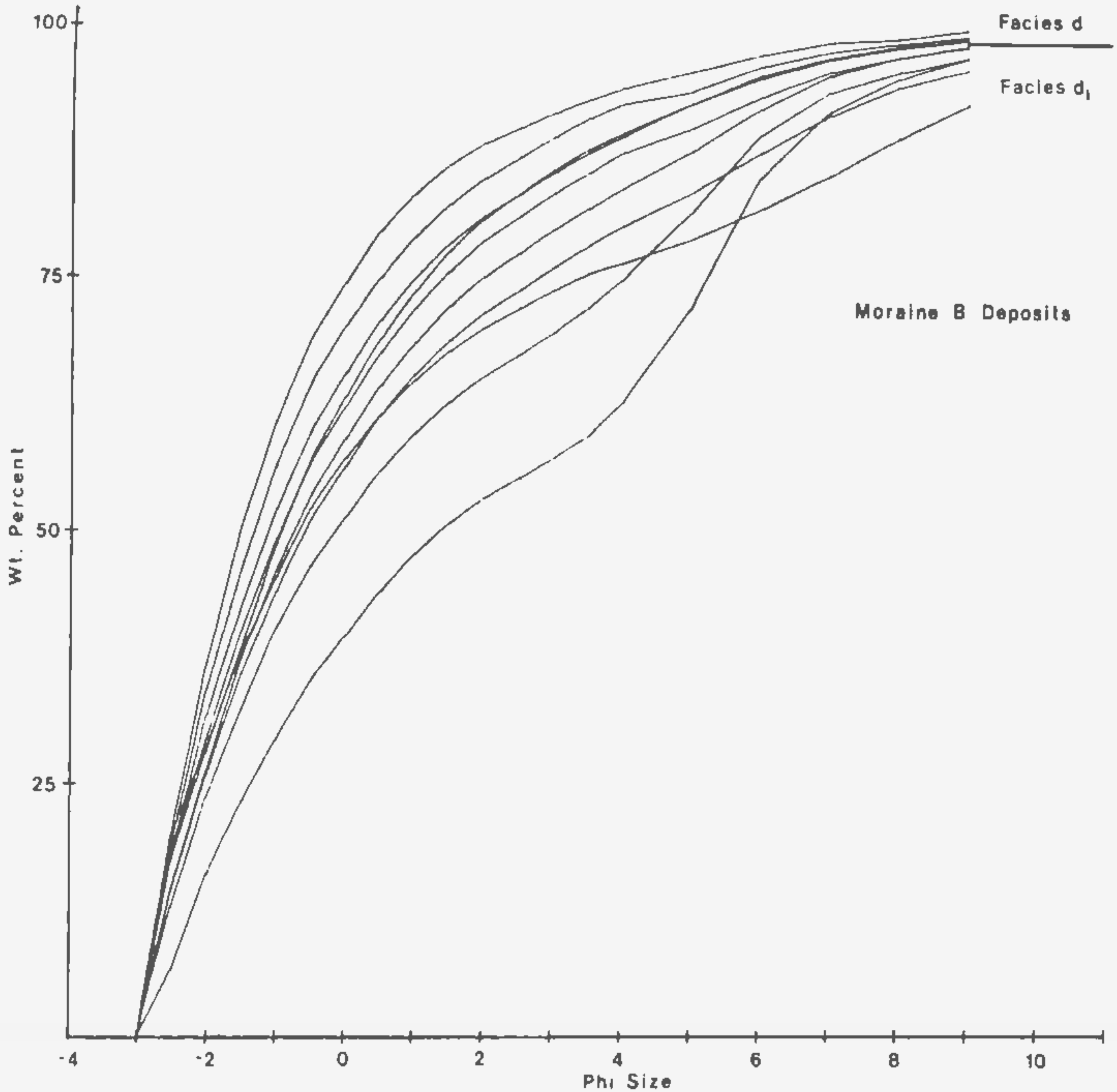

Figure 4-9 Grain-Size distributions for Moraine B deposits.

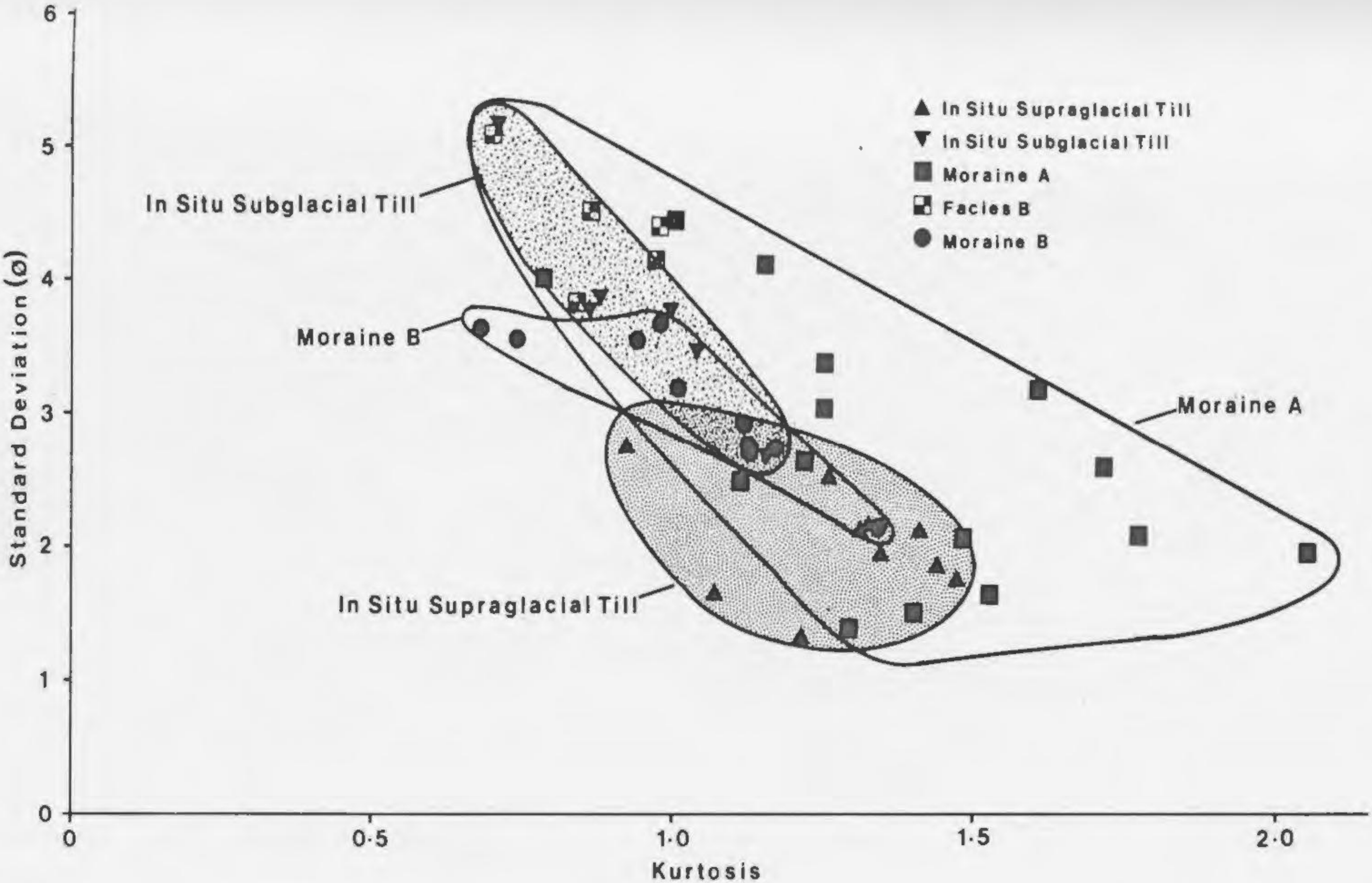

Figure 4-10 Plot of Friedman's third moment (kurtosis) against the second moment (sorting) for Emerald Glacier deposits.

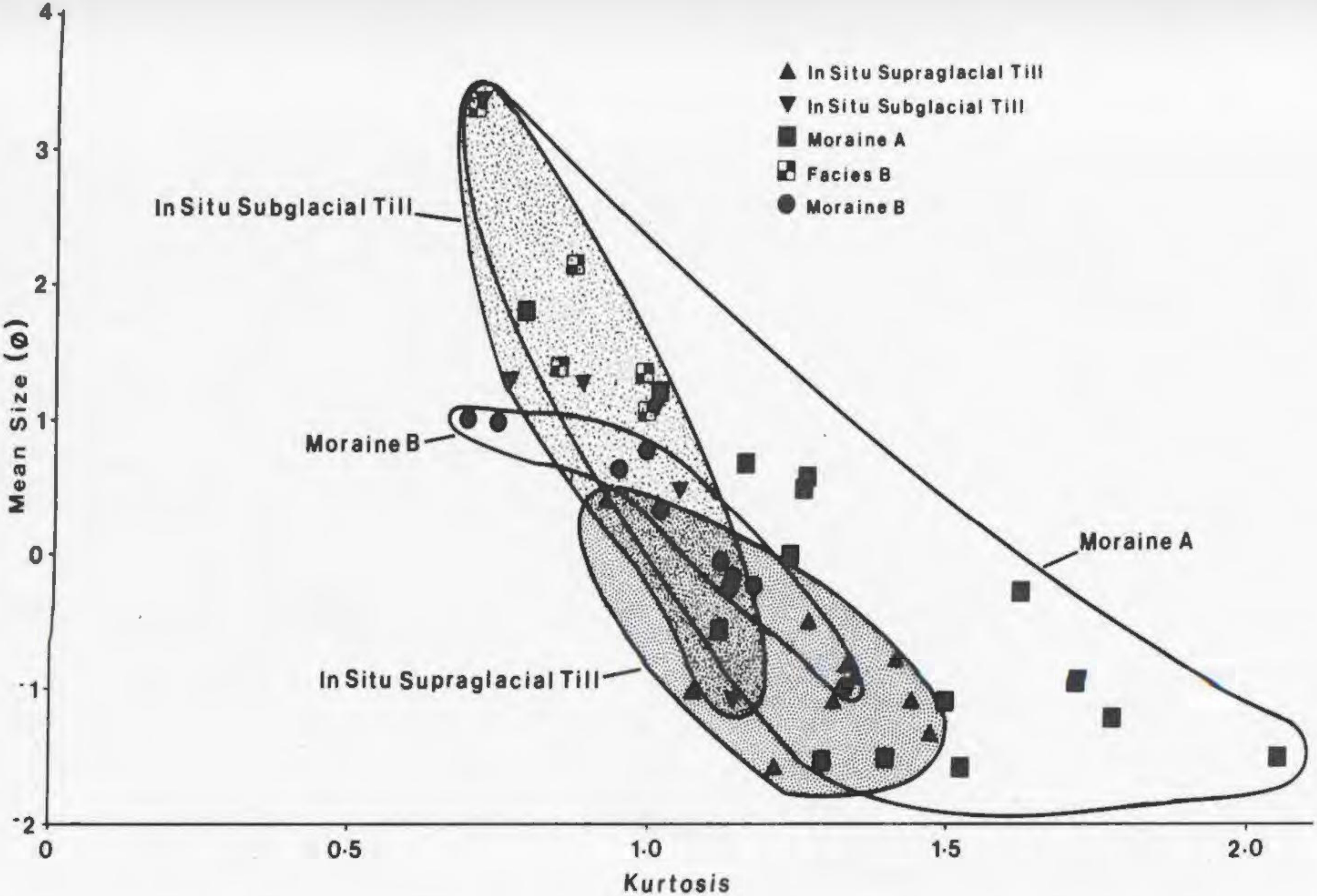

Figure 4-11 Plot of Friedman's third moment (kurtosis) against the first moment (mean) for Emerald Glacier deposits.

However, the clearest relationship was found between sorting and mean size (Fig. 4-12). In situ supraglacial deposits show relatively better sorting and coarser mean sizes, compared to the poorer sorted and finer mean sizes of subglacial samples. Again, differentiation is related to the mode of transportation within the glacier system.

Skewness (fourth moment) was insensitive to environment.

Moraine A samples were widely scattered, encompassing both in situ subglacial and supraglacial elements. This may provide an insight into a complexity of genesis.

Alternatively, Moraine B samples exhibited a smaller range of values, generally falling between the discrete in situ subglacial and supraglacial deposit distributions.

A similar differentiation of environments can be noted from the crushing-to-abrasion ratio results (Fig. 4-13). In situ subglacial deposits have ratios less than 2.6, whilst supraglacial elements are generally greater than 3.6, although the 'skeletal soil' appears as an outlier. Moraine A samples reveal a wide range of values from 0.6 to 11.0, extremes representing facies b and facies a/a_1 deposits respectively. Moraine B deposits have a much smaller range of values (0.9 to 4.8), although skewed towards the finer, subglacial elements.

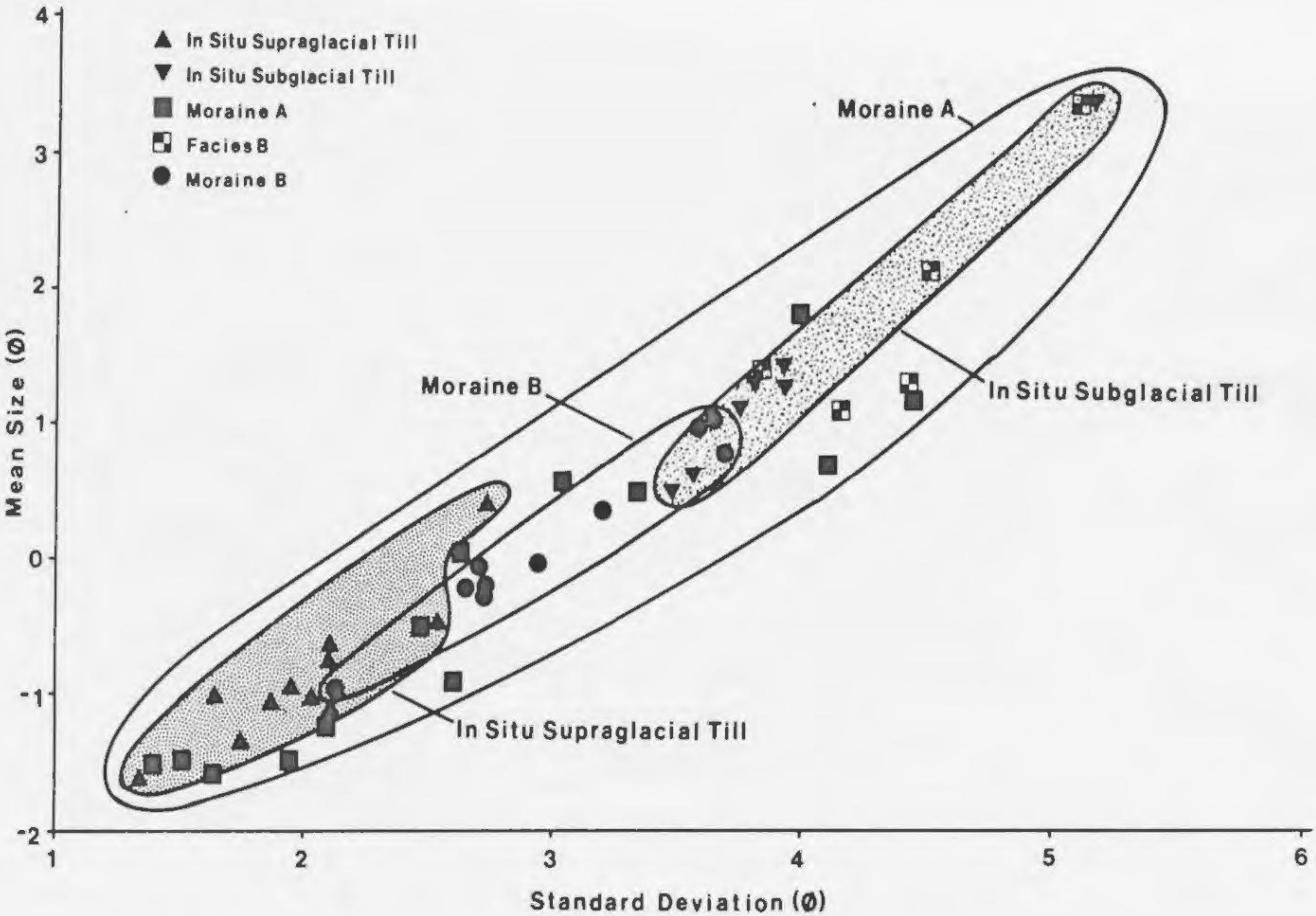

Figure 4-12 Plot of Friedman's second moment (standard deviation) against the first moment (mean) for Emerald Glacier deposits.

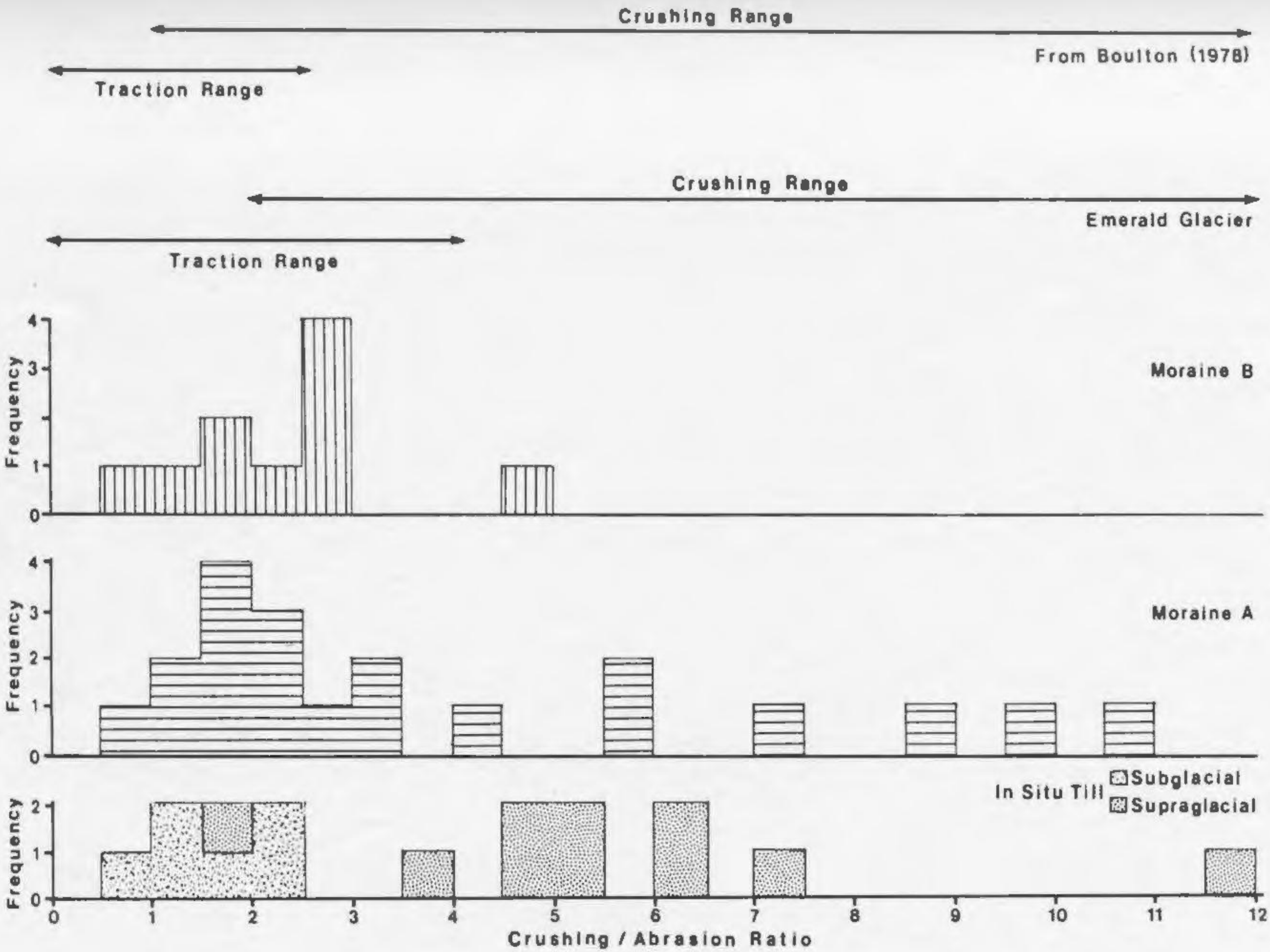

Figure 4-13 The crushing: abrasion ratio of Emerald Glacier deposits

4.5 X-Ray Diffraction

Clay minerals were identified as chlorite and illite/mica (e.g. Fig. 4-14 to 4-17).

Peaks at 14 Å, 7 Å, 4.7 Å and 3.5 Å represent chlorite minerals, of which 7 Å and 3.5 Å were major peaks. Carroll (1970) recognises these strong second and fourth order reflections as corresponding to iron-rich chlorite minerals; an expected situation in view of the parent rock (Pettijohn 1975).

Peaks at 10 Å, 5 Å, 4.4 Å and 3.7 Å correspond to illite/mi This composite term is preferred because of the difficulty of differentiating the two. Illite is often used as a field term for mica (Carroll 1970), and possesses a very similar diffraction pattern to mica, which is mainly muscovite.

Detailed examination of the characteristics of chloritic and micaceous clay minerals is presented by Grim (1962) and Brown (1972) chapters 6 and 5 respectively.

Glycolation and heat treatment showed that no kaolinite or montmorillonite was present in any of the samples.

Non-clay minerals were also identified. Quartz (4.2 Å and a major 3.3 Å peak) and dolomite (2.9 Å) were present in large quantities, dominating the samples. On average, combined quartz and dolomite peaks were between 2.4 and 4.6

times larger than clay mineral peaks. Smaller percentages of feldspar (3.2 Å) and calcite (3 Å) were also detected.

Analysis of crushed bedrock samples revealed that most clay minerals were derived from the local shale (Fig. 4-14). Very minor peaks for chlorite and illite/mica were present in limestone (Fig. 4-15) and dolomite/mudstone samples. Quartz dominated the non-clay element from shale, whilst dolomite dominated calcareous bedrock types. The findings are in keeping with the mineralogical analysis of bedrocks presented by Pettijohn (1975). Kodybka (personal communication) also concurs with this characterisation, based on thin sections from bedrocks adjacent to Yoho Glacier.

Emerald Glacier is within a cold, mountainous area with minimal organic matter. Under such conditions, Millot (1979) notes that hydrolysis would be generally insignificant and that fragmentation is the dominant weathering process, except in an active glacio-fluvial environment, where chemical leachin may be important (Reynolds 1971). The clay mineral content wil therefore be predominantly composed of disintegrated and slightly decomposed micas and chlorites. This is substantiated by Carroll (1970), who notes that mica commonly found in soils and unconsolidated sediments is mostly illite, and that chlorit is detrital in sedimentary environments, especially related to glacial activity.

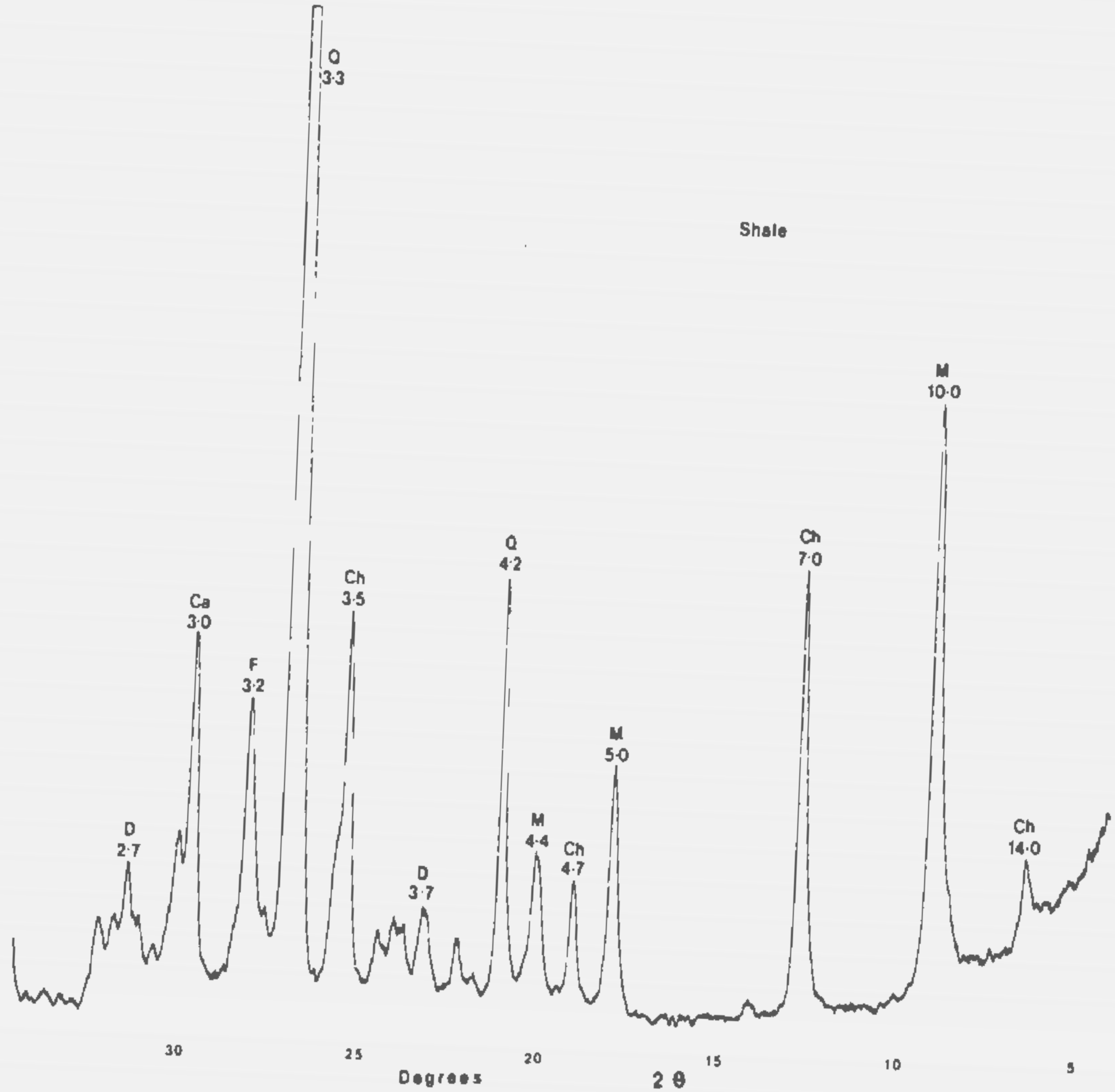

Figure 4-14 X-ray diffraction pattern of shale from Emerald Glacier. (Ch = Chlorite, M = Mica, Q = Quartzite, F = Feldspar, D = Dolomite, Ca = Calcite).

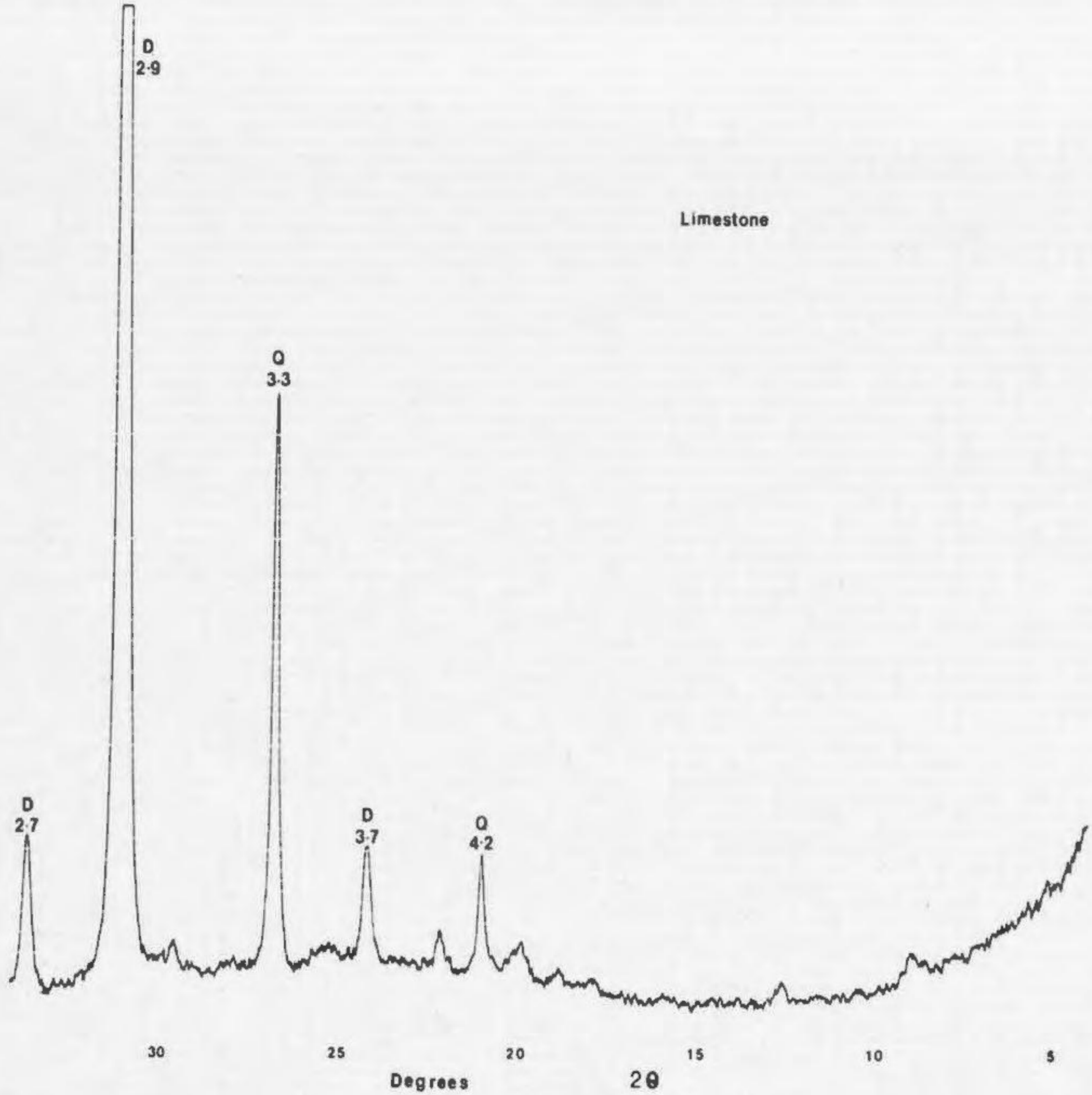

Figure 4-15 X-ray diffraction pattern of limestone from Emerald Glacier. (Ch = Chlorite, M = Mica, Q = Quartzite, F = Feldspar, Ca = Calcite, D = Dolomite).

Rieck et al. (1979) successfully differentiated between tills of the Saginaw and Huron-Erie glacial lobes in Michigan on the basis of 7 Å/10 Å diffraction ratios. It was thus hoped that such a differentiation could be made between glacier sub-environments, based on the varying production of clay minerals in the subglacial compared to supraglacial environment. However, results showed that no differentiation on the basis of clay mineral ratio values was forthcoming. This indicates the long time span required for clay mineral alteration.

Nevertheless, a differentiation between in situ supraglacial and subglacial deposits was made, on the basis of the proportions of major non-clay minerals (i.e. major dolomite and quartz peaks) and clay minerals (i.e. 10 Å illite/mica and 7 Å chlorite peaks). Ratio values averaged 0.42 for supraglacial samples and 0.22 for subglacial deposits. Furthermore, Moraine A samples showed a wide range of values (0.14 to 0.91), whereas Moraine B had a very narrow scatter (0.11 to 0.18) (Table C-1).

Based on the results of the bedrock analysis however, this may merely reflect different proportions of bedrock types within each environment: shale dominating the supraglacial environment (Fig. 4-16), while lacking subglacially (Fig. 4-17). Nevertheless, the derived ratio pattern accords with those evidenced by other means and may therefore be significant.

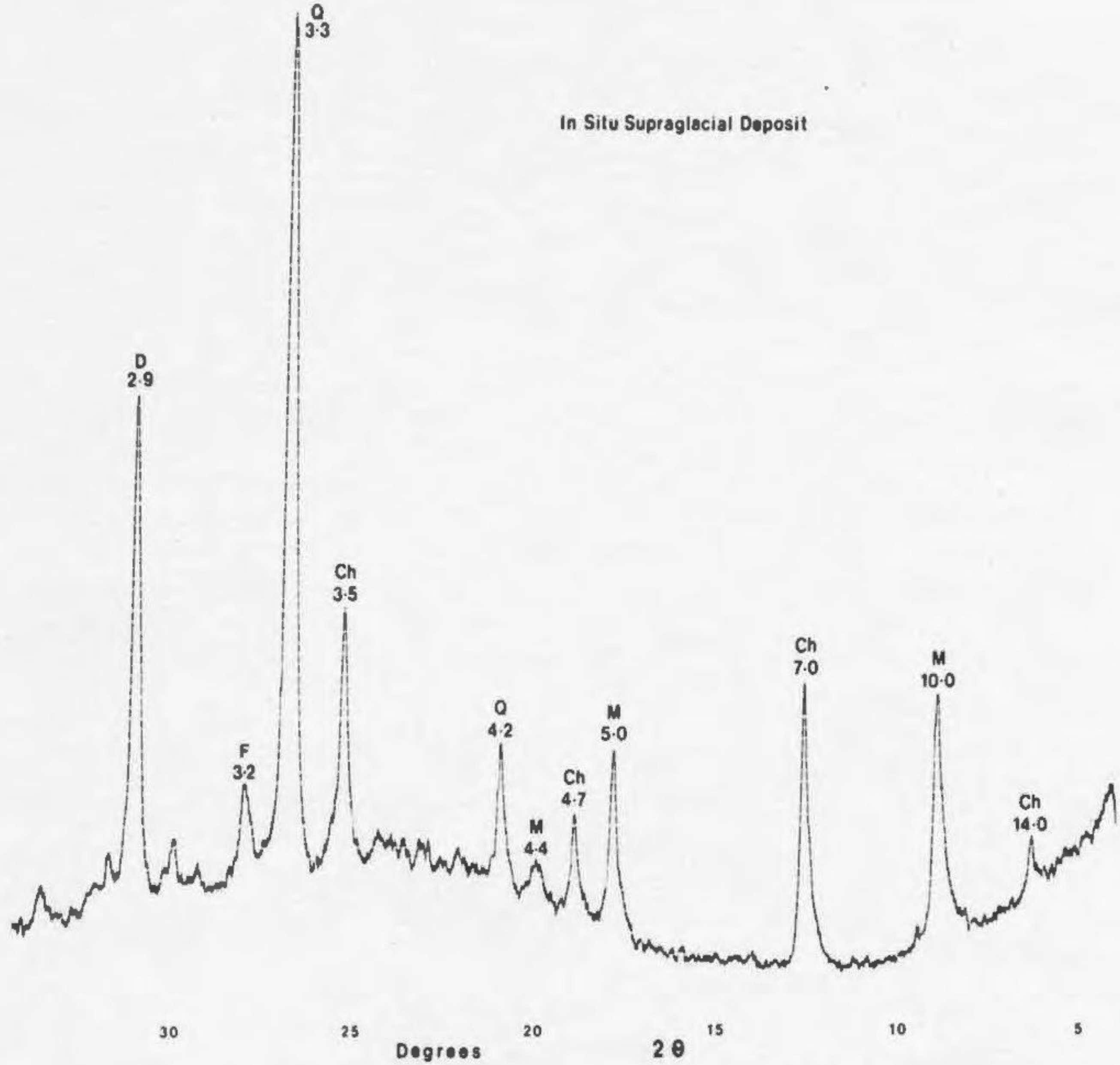

Figure 4-16 X-ray diffraction pattern of in situ supraglacial debris from Emerald Glacier (Ch = Chlorite, M = Mica, Q = Quartzite, F = Feldspar, Ca = Calcite, D = Dolomite).

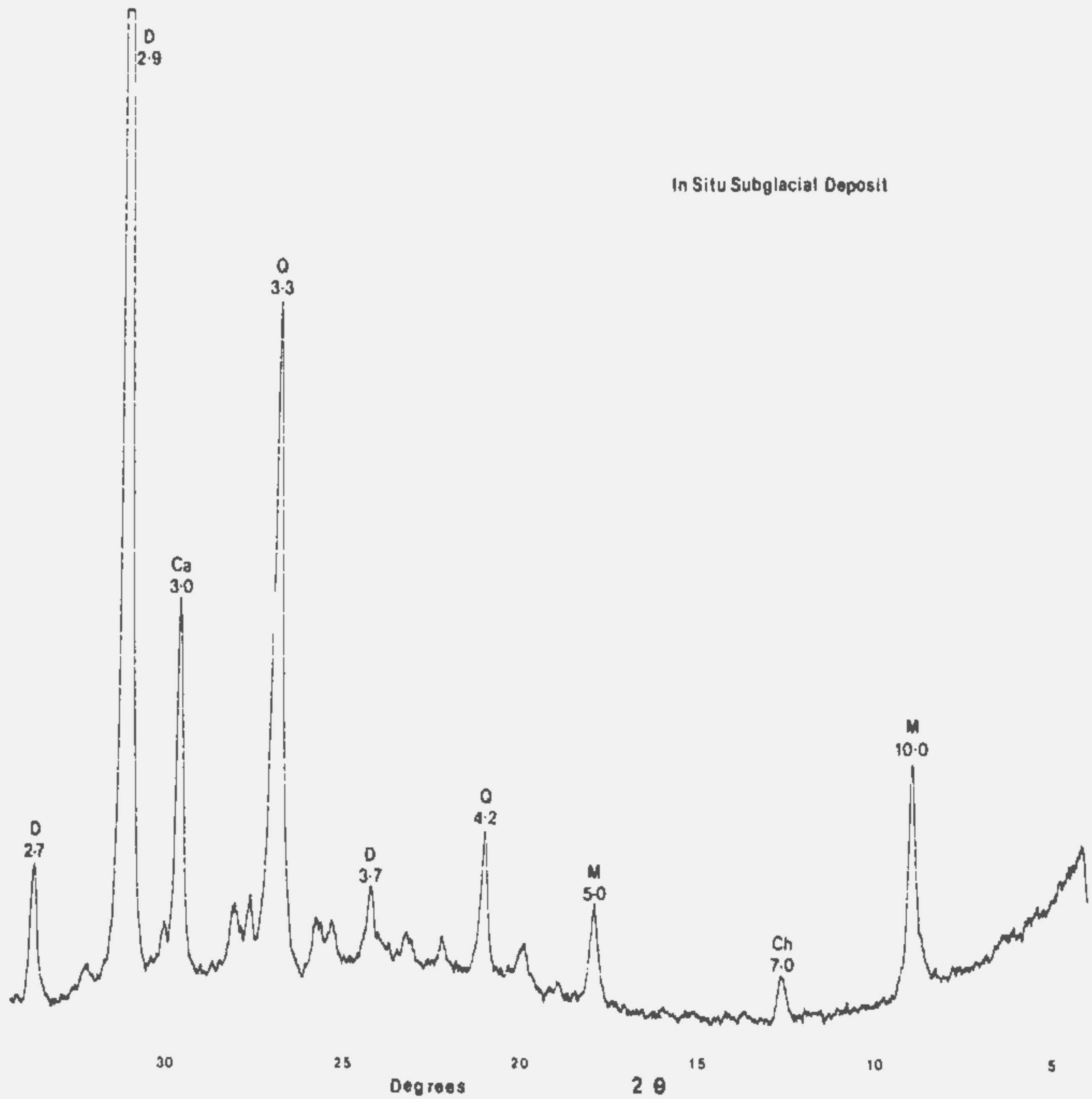

Figure 4-17 X-ray diffraction patterns of in situ subglacial debris from Emerald Glacier (Ch = Chlorite, M = Mica, Q = Quartzite, F = Feldspar, Ca = Calcite, D = Dolomite).

4.6 Atterberg Limits

Several authors have discussed the geotechnical properties of tills (e.g. Easterbrook 1964, Kazi and Knill 1969, Boulton 1975b, Boulton and Paul 1976, Millegan 1976). Most used these properties to characterise tills of different ages or merely as an aid to description.

Boulton and Paul (1976) however, noted the possibility of utilising geotechnical parameters to distinguish between glacier sub-environments. They proposed the 'T-line', which represented a direct relationship between the liquid limit and the plasticity index. It is located to the left of Casagrande's 'A' line, which was constructed from values derived from sedimentary clay deposits, and thus takes into account the poorer sorting of glacial tills. Cohesionless deposits possess a liquid limit less than 10.5 on the T-line, compared to 20 for the 'A' line (Terzaghi and Peck 1967). The percentage of clay-minerals and thus cohesion, increases from left to right along the T-line, and therefore position reflects the nature of the grain-size distribution (Boulton and Paul 1976).

The liquid limit versus plasticity index plot for Emerald Glacier deposits is presented in Fig. 4-18. It reveals a distinct separation of Moraine A and Moraine B samples. Moraine A deposits are located to the right of those from Moraine B, and therefore reflect a higher average clay content. Facies b samples are close to the T-line, as is one in situ

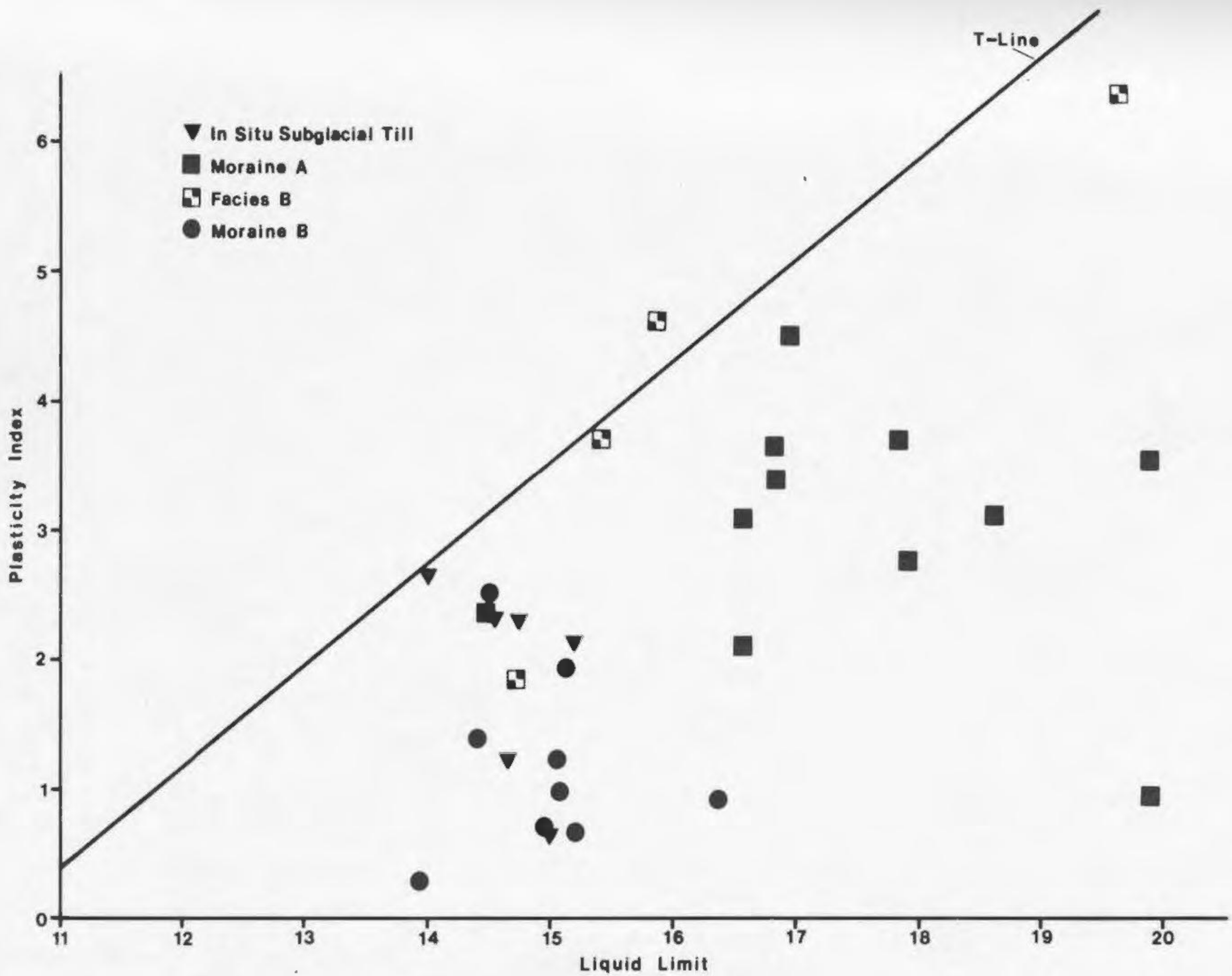

Figure 4-18 Plot of liquid limit against plasticity index values for Emerald Glacier deposits.

subglacial deposit (#33). Samples from other facies units, plus subglacial deposits are located to the right of the T-line. These correspond to deposits that have been modified by post-depositional processes (Boulton and Paul 1976).

A wide range of factors influences the Atterberg limits of any sample. The liquid limit for instance, is dependent upon the mineral constituents of the parent deposit, which affect the intensity of surface charges and the thickness of attached water layers (Means and Parcher 1964). Increases in the non-clay mineral fraction will have the effect of reducing liquid limits.

The plasticity index depends on the cohesive properties of clay minerals, related to physiochemical forces acting between particles (Yong and Warkentin 1966; Baver, Gardner and Gardner 1972; Carson and Kirkby 1972). Cohesion is a concept restricted to clay minerals (Carson and Kirkby 1972), although some confusion in its usage is evident. For instance, when Boulton (1975) notes that ".... if the clay concentration is high the till will tend to develop high cohesion", a clay-size or clay-mineral meaning is not apparent, although one infers clay-mineral concentration. Similarly, cohesion will vary considerably with the type of clay mineral involved. Direct comparisons between areas of widely differing spatial location are therefore dubious, unless similar lithology and weathering processes can be proven.

Erosion of igneous or metamorphic rock produces rock-flour of a clay-size range, whilst some sedimentary rocks break down largely into clay-minerals. The relationship between clay-size percentages and plasticity will vary considerably with lithology and may be expressed by the 'activity' of a deposit.

Activity is defined as the "relationship between water sorption on one hand and mineralogy and size distribution on the other" (Boswell 1961). Means and Parcher (1964) suggest that values less than 0.75 denote relatively inactive soils, 0.75 to 1.50 normally active, and greater than 1.50 active soils Swelling type clays (e.g. montmorillonite) possess high activities. Of those samples analysed 81% were relatively inactive and 19% normally active (Table B-1). In general, the highest activity values were recorded for in situ supra-glacial, facies c and facies d deposits.

Relative inactivity may be the result of increases in the non-clay mineral proportions (Dumbleton and West 1966a, 1966b). They have the effect of 'diluting' Atterberg limit values (Boulton and Paul 1976) by lowering the liquid limit and plasticity index. A decrease of values down the T-line from right to left is expected.

4.7 Fabric

Three dimensional till fabric results are presented in Fig. 4-19. They indicate moderate to strong upglacier imbrication in three samples and weak downglacier imbrication in two. All fabrics were significant at the 95% level according to the Poisson test against randomness.

The strongest fabric was at site 3, within facies d_1. Azimuth was normal to the moraine crest and imbrication was distinctly upglacier (average 27^o).

Facies b at site 2 revealed a strong fabric with both upglacier dip and an aximuth normal to the moraine crest.

Fabric has been widely used as an indicator of formative processes (e.g. Holmes 1941, Andrews and Smithson 1961, Kruger 1970, Mills 1977). Price (1970) used fabric as a basis for his squeeze hypothesis for moraines in Iceland. Similarly, Walker (1973) used fabric as a tool to differentiate between possible origins of ridges in the Bow Valley, Alberta.

At Emerald Glacier, fabric contrasts exist between facies a and facies a_1, b and d_1. Facies a possesses a weak fabric and downglacier imbrication while a_1, b and d_1, possess a stronger, upglacier fabric. Similarly, facies a_1 and d_1 exhibit a mean azimuth orientation normal to the moraine crest, while facies a and b record an azimuth value oblique to the moraine axis. On this basis a contrast in formative processes may be suggested.

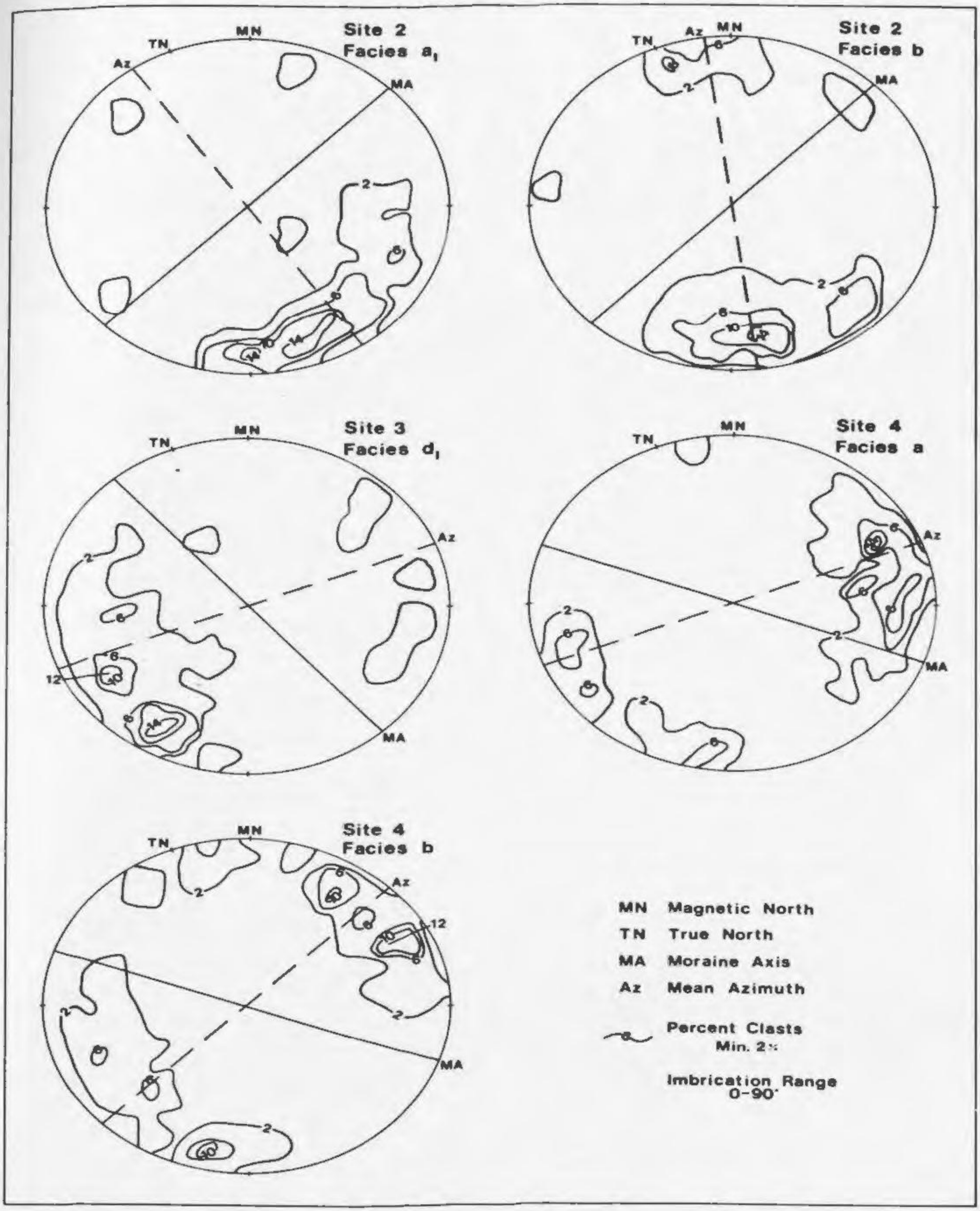

Figure 4-19 Fabric diagrams from facies a, a_1, b and d.

4.8 Summary

The main focus of this study is the contrasts between subglacially and supraglacially derived tills. A number of different techniques have been used to distinguish between them.

Particle size has been the main differentiating criterion and has confirmed the contrast between a passively transported supraglacial till element where little or no comminution occurs along a flow line and subglacial till which is characterised by rapid comminution and a decrease in mean particle size along a flow line.

These environmental differences manifested themselves in terms of contrasting sorting and kurtosis values within the grain size distribution curves. Similar contrasts were noted from the mean size values of debris, ranging from -0.8 Ø in the supraglacial to 1.5 Ø in the subglacial environment. This is a function of the low silt-clay content of in situ supraglacial deposits (0.7% to 3.5%) compared to in situ subglacial tills (18.4% to 44.9%).

Deposits composing the moraines also revealed distinct grain-size distributions. Facies units within Moraine A illustrated between facies variation, with distributions encompassing those values recorded by both in situ supraglacial and in situ subglacial deposits. Moraine B samples however, were confined to a narrow range of values, although resembling the in situ subglacial elements in character rather than their supraglacial counterparts. Such a differentiation between moraine deposits may point to a contrast in formative processes.

These contrasts between in situ subglacial and in situ supraglacial deposits, and between Moraine A and Moraine B units was confirmed by X-ray diffraction analysis. Although no differentiation between glacier sub-environments was observed on the basis of 7Å/10Å peak ratios as Reich et al. (1974) had found, contrasts were noted on the basis of non-clay mineral: clay mineral ratio values. In situ supraglacial and in situ subglacial deposits again revealed discrete distributions (average 0.22 and 0.42 respectively). Similarly, Moraine A deposits were widely scattered, encompassing both in situ supraglacial and in situ subglacial values, whilst the distribution of Moraine B samples was confined within a narrow range located adjacent to in situ subglacial till values. The use of X-ray diffraction analysis as a method of differentiating between glacier sedimentary systems is a novel one, and its potential for revealing diagnostic characteristics of glacier sub-environments can only be evaluated on the basis of further research.

Further evidence of contrasts between Emerald Glacier sedimentary environments was provided by the assessment of Atterberg Limit values. Moraine A samples recorded a wide scatter for liquid limit and plasticity index values, compared with a narrow distribution of Moraine B deposits. No in situ supraglacial deposit values were recorded because of the insufficiency of fine sized material.

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The distinction between in situ supraglacial and in situ subglacial tills which has been demonstrated by several sedimentological and geotechnical techniques is entirely in keeping with the contrasting manner in which the material has been transported and deposited. The difference between samples derived from Moraine A and Moraine B may however, be of greater significance to this research thesis. The wide scatter of Moraine A values recorded by each technique encompass elements of both the discrete in situ subglacial and in situ supraglacial distributions, and may suggest a complexity of origin. Moraine B deposits on the other hand exhibit a narrow scatter of values which may point towards a common genesis for these deposits.

Chapter 5

Discussion and Synthesis

5.1 Characteristics of the near terminal area: a discussion

Distinct morphological and sedimentological differences are evident between Moraine A and Moraine B, and are the direct result of contrasting proximal ice, debris-cover conditions.

Supraglacial debris retards glacier-melt proximal to Moraine A. In consequence, the glacier is coupled to the moraine throughout the ablation season. Compressive strain should therefore increase towards the terminus; and this is confirmed by a decrease in both ice and sediment movement rates. The increased strain culminates in a forward movement of the moraine itself, at a rate of between 0.8 and 0.9 cm/day during July/August. Downglacier decrease in glacier velocity is related to the distance from the glacier margin, which similarly reflects increases in debris thickness (Fig 5-1).

An effect of increasingly compressive flow downglacier can perhaps be seen in the changing imbrication of surface clasts. They evidence a gradual divergence away from the

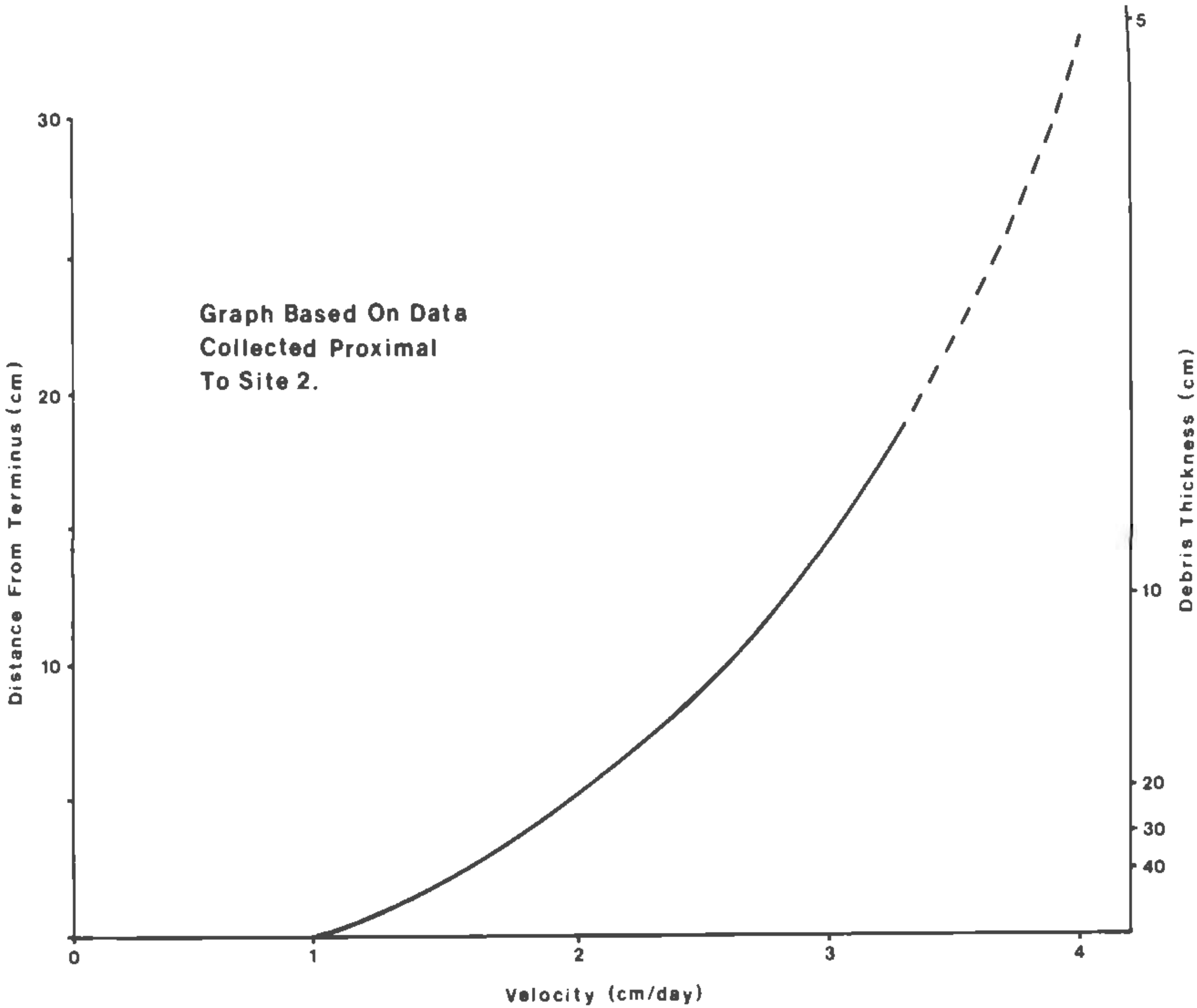

Figure 5-1 The relationship between near terminus glacier movement, debris thickness and upglacier distance at Emerald Glacier.

surface slope, until clasts are almost vertical immediately proximal to the moraine (Fig. 5-2). It appears the condition arises from the greater movement rates of glacier ice compared to the sliding velocity of surface clasts. The effects are progressively accentuated downglacier as the platy clasts are 'rotated' by the faster moving ice (Fig. 5-3). As far as the author is aware no report of this phenomenon exists within the glaciological literature, and it may therefore represent a further characteristic of the effects of compressive strain at a glacier terminus.

The source of supraglacial debris has been inferred as cirque headwalls with subsequent englacial transportation. The similarity of avalanche slope and near terminal supraglacial debris grain-size distributions adds weight to this argument. However, ice surface observations in late August 1979 (Fig. 5-4) suggest that simple incorporation of material into the glacier system on a yearly basis may not be an accurate characterisation. The existence of crevasses suggests that surface accumulation of debris is unlikely, and leads to the conclusion that material may be collected in crevasses to be exposed below the firn line by ablation (Fig. 5-5).

However, factors such as the finer texture of crevasse fill deposits compared to terminal supraglacial debris and the predominance of primary stratification within ice cliffs (Fig. 5-6), suggests that a purely crevasse-fill origin for

Figure 5-2 The changing imbrication of surface clasts in the near terminal area (Moraine A). Almost vertical clasts are located immediately proximal of the moraine. Spade shows scale.

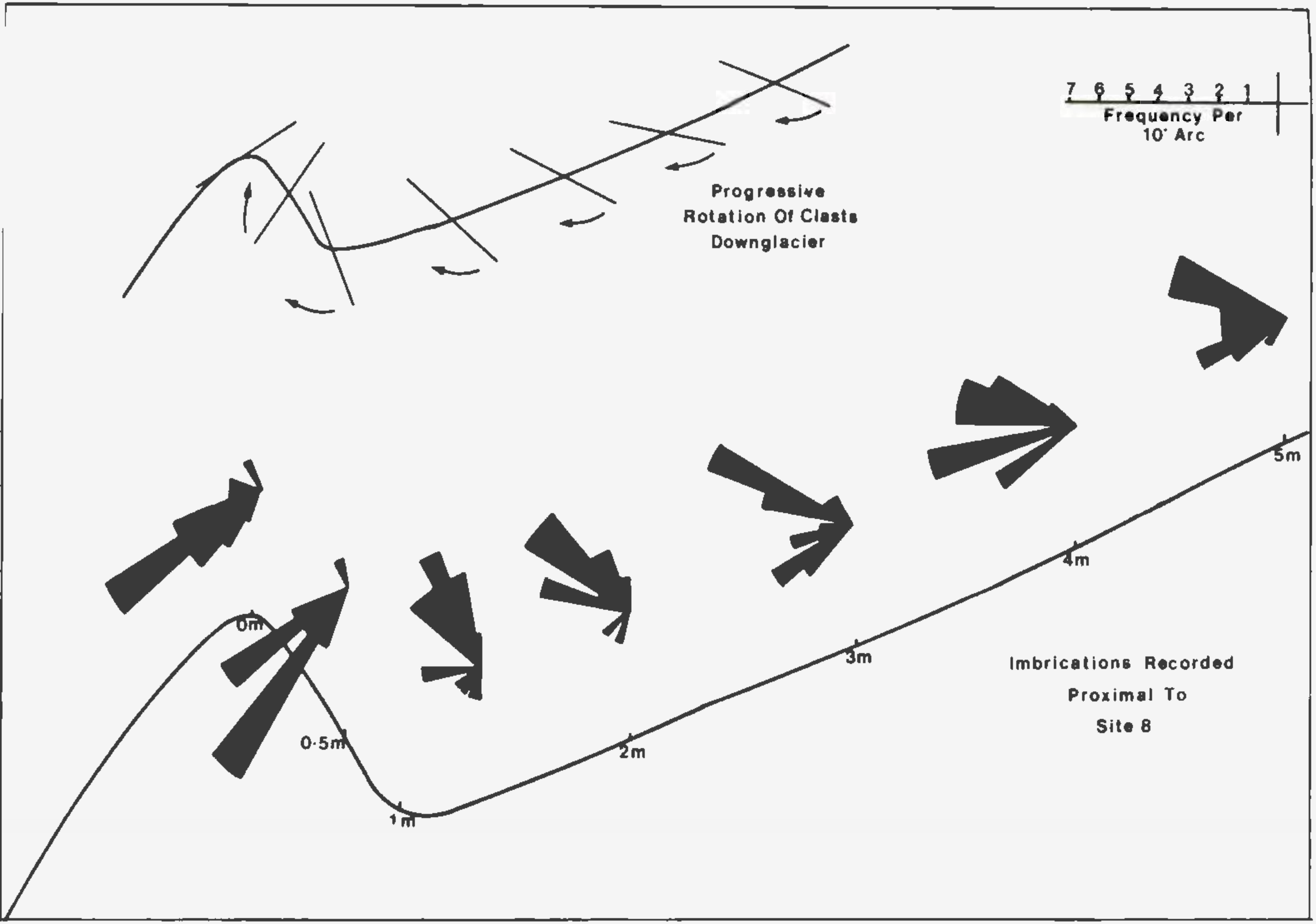

Figure 5-? Downglacier variations in surface clast imbrication patterns.

Figure 5-4 The ice surface in late August 1979.

Figure 5-5 Crevasse fill located upglacier of the area of heavy surface debris cover.

Figure 5-6 Terminal ice cliff showing predominantly primary stratification. Minor secondary features are also evident.

supraglacial debris is a dubious conclusion.

The existence of abundant supraglacial material along the north-west cirque walls suggests this may be a potential source area. Material from this area probably contributed to the supraglacial debris suite through mass-wasting processes and was subsequently transported to the glacier terminus.

Supraglacial debris is reflected in the Moraine A stratigraphies by facies a/a_1 and c. This conclusion is based upon the similarity of both sedimentology and X-ray diffraction patterns with the in situ supraglacial debris samples. Eyles (1979) however, considers that a skeletal soil (i.e. facies c) ".... never survives final till deposition.. often being mechanically remixed with the till ...". Nevertheless, the stratigraphic position of facies c immediately below a coarse textured supraglacial deposit, and the textural similarity with the in situ skeletal soil, suggests a common genetic environment. Facies a contains no such skeletal soil.

In contrast, the sedimentological characteristics of facies b are far removed from those of a supraglacially derived till, and as such a subglacial origin is proposed. Facies b however is finer than the in situ subglacial material, which is proximal to, and which appears to comprise Moraine B. It may thus be suggested that facies b is in a less disturbed state.

The area proximal to Moraine B was observed as an actively glaciofluvial environment, with a series of small outwash streams evident. In such an environment washing of material occurs, resulting in the removal of fine grained material. The washed debris collects immediately behind the moraine barrier or migrates through it, in suspension within melt-water channels. Comparison of facies d and d_1 deposits reveal that the latter is considerably richer in fines, although not in the relative proportions of clay minerals which are easily removed by glaciofluvial action. The role of lithology in controlling the clay-mineral content must also be considered however.

Facies d may represent a concentration of washed debris, and its emplacement upon the moraine itself would represent a more recent depositional phase.

On the whole, Moraine B possesses a greater percentage of fine grained material than Moraine A; a reflection of the contrasting dominant genetic environments. Although all facies units within Moraine A have proportionally greater clay mineral content per unit weight than Moraine B, this is offset by the lower percentage of clay-sized material within Moraine A as a whole, dominated as it is by supra-glacially derived debris.

The concretion effect of fines being washed over the moraine surface may explain the steeper moraine slopes of Moraine B. The process of interparticle quartz cementation outlined by Whalley (1974b) may be of relevance here. Moraine A on the other hand, is consolidated solely by the existence of the facies b unit, and the steepness of slopes here can be considered a reflection of the active nature of the moraine itself.

A washing process may also explain the anomalous site 2 facies a, and the skeletal soil which develops within Moraine A, both of which reveal relatively large proportions of clay minerals as well as clay-sized material. It is suggested that whereas depletion of fines occurs proximal to Moraine B, washing processes in the supraglacial environment have the effect of concentrating fines, and therefore clay minerals immediately above the glacier surface, which acts as an impermeable layer. Any movement of material is downglacier within the unit itself, to be concentrated still further proximal to the glacier terminus. The existence of the skeletal soil layer is a result of, and is dependent upon, glaciofluvial activity. Proximal to site 2, debris thicknesses were less than further east. Glacier melt was therefore pronounced, which enhanced the washing process. This may explain the poor upglacier extension of the skeletal soil; washing transporting fine material downglacier more rapidly, compared to the less glaciofluvially active site 4 area.

The downward migration of fines through the moraine due to washing, especially at site 2 where glaciofluvial activity was considerably more intense than further round the ice margin, may explain the anomalous grain-size distributions. The downwashed material may have been either supraglacially derived or have been washed from the facies b lens. Similarly, the lateral migration of fines through the moraine itself (Fig. 5-7) may be cited as a further explanation.

Moraine A is thus characterised by complexity, resulting from the interstratification of supraglacial and subglacial till elements, which would thus explain the wide scatter of results derived from grain-size and X-ray diffraction analyses. The influence of the supraglacial cover is to induce glacier-moraine contact throughout the ablation season, thus increasing compressive strain and generally providing the moraine with forward momentum.

Moraine B has no proximal supraglacial debris cover and a subglacial till element predominates, although units of different depositional phases may be identified. The lack of supraglacial cover results in ice decoupling from the moraine during the summer. In 1979, decoupling had occured by July 24 and was continuing on August 25. Marginal retreat would presumably be reversed with the first significant fall of snow, which stops ablation and allows the ice to advance to the moraine. Longitudinal compressive strain will thus probably increase to a level similar to that experienced behind Moraine A.

Figure 5-7 The washing of fine material from beneath Moraine A.

The Emerald Glacier moraine is a contemporary feature of the ice margin, with initial development occurring within the last decade. This is suggested by a series of factors.

Firstly, Moraine B is within 4 metres of a gently sloping (approximately 10^{o}-15^{o}) bedrock surface exposed in the subglacial environment. There was no vertical height change at its termination within the proglacial debris proximal to Moraine B. Water from summer ice-melt was observed to wash this surface clear of debris deposited during the subglacial melt-out process. Material from this washing process collected proximal to Moraine B. No ridging of this washed debris was noted and therefore glacial rather than fluvioglacial activity seems likely to have constructed the moraine. The rate of ice movement and the length of the winter accumulation season when ablation is halted and readvance of the ice occurs, suggest a short time period for a maximum moraine advance of 4 metres. Indeed, Moraine A advanced at 0.8 to 0.9 cm/day during the summer 1979 and sections of it were 2-3 metres further forward in the summer of 1980 (Rogerson, personal communication). Winter snowfall would act as a supraglacial debris cover in the case of Moraine B.

Secondly, within Moraine A no development of illuviated or eluviated horizons was noted in any of the facies units in which detailed sampling was undertaken. Washing of fines from facies b down through facies a units may be expected, and the fact that no such process has occurredmay point to the recent development of the moraine stratigraphies.

The discussion on the characteristics of the near termnal area is summarised by Rogerson and Batterson (1980).

5.2 Formation of the Emerald Glacier moraines: A synthesis

Possible modes of formation were discussed in section 1.4. Several mechanisms were considered plausible to account for small moraines existent along the margins of temperate glaciers i.e. a dump (e.g. Boulton and Eyles 1979), push (e.g. Rabassa et al. 1979) or squeeze (e.g. Price 1970) process.

The applicability of these models to Emerald Glacier can be reviewed on the basis of the preceding discussion.

a) Moraine A

The complex stratigraphy and existence of supra-glacial debris, suggests a simple squeeze mechanism is not a satisfactory explanation for the formation of this moraine. Indeed, the continual forward motion of the moraine is more in keeping with a pushing process. However, the emplacement of a subglacial deposit above its usual stratigraphic position (Boulton 1976, Sugden and John 1976) does not correspond to a simple push either. As such, a composite genesis is suggested (Fig. 5-8).

Initial stable conditions resulted in the formation of a talus slope at the ice margin, as material was 'conveyor-belted' off the ice front (Fig. 5-8a). This corresponds to a minor dump moraine (Hewitt 1967; Whalley 1974a; Boulton and Eyles 1979, facies A_1). Material sliding off the glacier

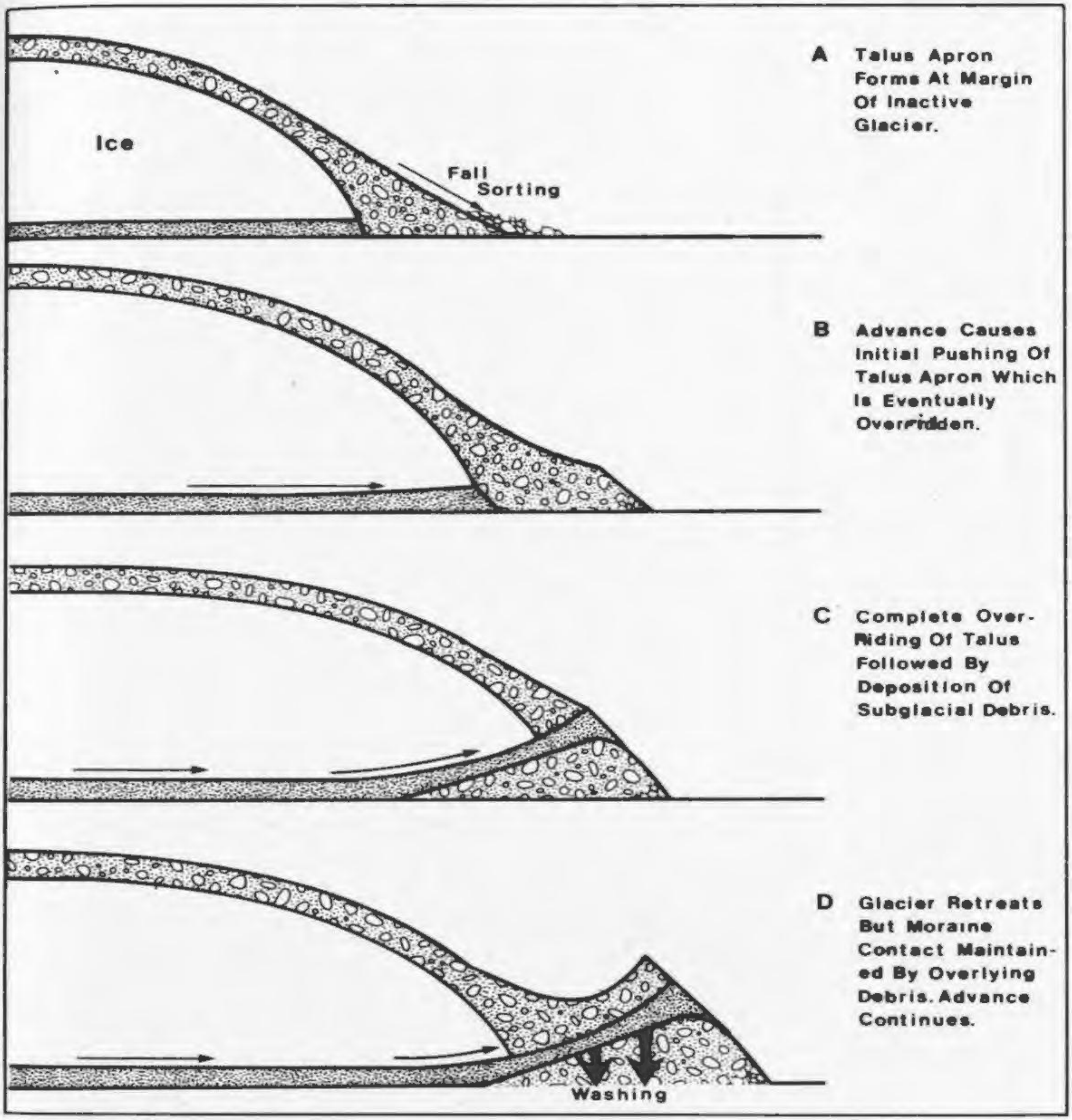

Figure 5-8 Model of the formation of Moraine A.

surface in this manner will possess downglacier imbrication i.e. the fabric possessed by facies a. Predominance of large clasts at the base of the talus apron due to fall-sorting will be expected (Whalley 1974a, Eyles 1979).

A change in glacier equilibrium leading to a reactivation of the ice front, will have the initial affect of ice pushing against the talus slope. This will continue until the shear strength of the pushed debris is greater than 1 bar (Whalley 1974a, Haeberli 1979); the approximate yield strength of ice. At such time the talus slope will be overridden (Fig. 5-8b). Compressive strain in the terminal area is a component factor in this process.

Subglacial deposition occurs as overriding takes place (Fig. 5-8c). The emplacement of subglacial debris as a result of overriding rather than for instance, intrusion through the moraine is suggested by deposition well above the present glacier sole. The fabrics recorded from facies b also concur with this process. The thickness of the facies b unit ranges from 25-80 cm, considerably thicker than expected (Boulton 1975), especially since the moraine has only been activated within the last 10 years. A squeezing out of material due to pressure release as the subglacial unit is exposed to a subaerial environment, is suggested to account for this.

Whilst the overriding process was active, supraglacial debris was continually being supplied to the glacier terminus. Sorting through washing is an ongoing process, and it is suggested that upon glacier recession a sorted stratigraphic unit was deposited over the subglacial, facies b lens (Fig. 5-8d). This is further evidence for the rejection of a subglacier lens intrusion hypothesis.

The distinct junction between the subglacial unit, and over and underlying deposits testifies to its plasticity. If this were not the case, a general downwashing of fines from the wastage of basal ice would be expected, nullifying the effects of overriding.

Although ice receded from a supra-morainal position as a result of ablation, contact with the moraine is maintained. Indeed, pushing continues. Subsequent overriding is possible if reactivation continues, but unlikely due to the relative height of the moraine above the glacier sole and the constraining effect of an extensive supraglacial debris cover.

b) Moraine B

Conditions proximal to Moraine B differ from those adjacent to Moraine A, and it may thus be presumed that the mode of formation will differ too.

A supraglacial debris cover is largely absent and therefore the 'dump' mechanism of moraine formation may be rejected.

However, fabric recorded at site 3 corresponds to that recorded within squeeze moraines (Price 1970) i.e. normal to the moraine crest.

Moraine B deposits are also fine textured; 20% finer than 4 Ø, compared to 5-10% for Fjallsjökull moraine sediments (Price 1970). Similarly, the immediate proglacial area of Emerald Glacier is actively fluvioglacial, conducive to the development of a saturated till body for subsequent squeezing. Nevertheless, a push mechanism to account for Moraine B is suggested.

This conclusion is based upon a series of observations in the terminal area. On July 12, at least two distinct ridges were noted in the vicinity of site 3 (Fig. 5-9), with ice in contact with the ridges. By July 24 the ice had decoupled from the moraine and only one ridge was evident (Fig. 5-10). Similar multi-constructional units have been described for lateral and neoglacial moraines (Humlum 1978, Osborn 1978, Ahmed 1979, Luckman and Osborn 1979).

Squeeze moraines are developed by the flowage of water soaked till towards the ice margin (Worsley 1974). However, Moraine B is within 4 metres of a subglacial bedrock surface that was observed to be washed clear of debris during the field season. Thus, the only available material potentially suitable for squeezing is located within this 4 metre area. It therefore seems unlikely that moraines 2 metres in height could have resulted from the ice pressures and squeezing processes acting on such a limited debris cover, especially as it would have been considerably lesser extent during previous seasons.

As such, the squeezing hypothesis for the formation of Moraine B is not acceptable.

Figure 5-9 Two ridges evident at Moraine B on July 12, 1979. The ice axe divides the two deposits. Glacier ice is immediately proximal to the upper ridge.

Figure 5-10 One ridge evident at Moraine B on July 24, 1979. An overriding of the moraine is suggested. The junction noted within the moraine is the result of slumping rather than the boundary of 2 depositional phases. This latter boundary is indistinct.

The proposed push formation depends upon the production of debris in the summer months (July to September), to be remolded during the winter/spring period (Fig. 5-11).

Initial conditions involve the melt-out and washing from the subglacial environment of regelation debris (Fig. 5-12) and glacier sole material (Fig. 5-13), into the proglacial area (Fig. 5-11a).

During winter reactivity, the ice front which may be frozen to its bed in marginal areas, bulldozes material into an initial ridge, which may be partially overridden (Fig. 5-11b). Compressive strain in the terminal areas enhances this process.

The lack of a protecting mantle of supraglacial debris results in ablation rates exceeding forward movement in the summer, and the decoupling of glacier ice from the moraine occurs. Material is again derived from the melt-out of subglacial deposits, to collect proximal to the moraines (Fig. 5-11c)

Winter reactivity again bulldozes material, but the previously formed ridge impedes forward movement and causes overriding. Material is therefore accreted to the proximal side of the moraine (Fig. 5-11d and Fig. 5-14). The overriding process is similar to that suggested for Moraine A, but in this case may occur as an 'annual' process; a 'pulse' mechanism.

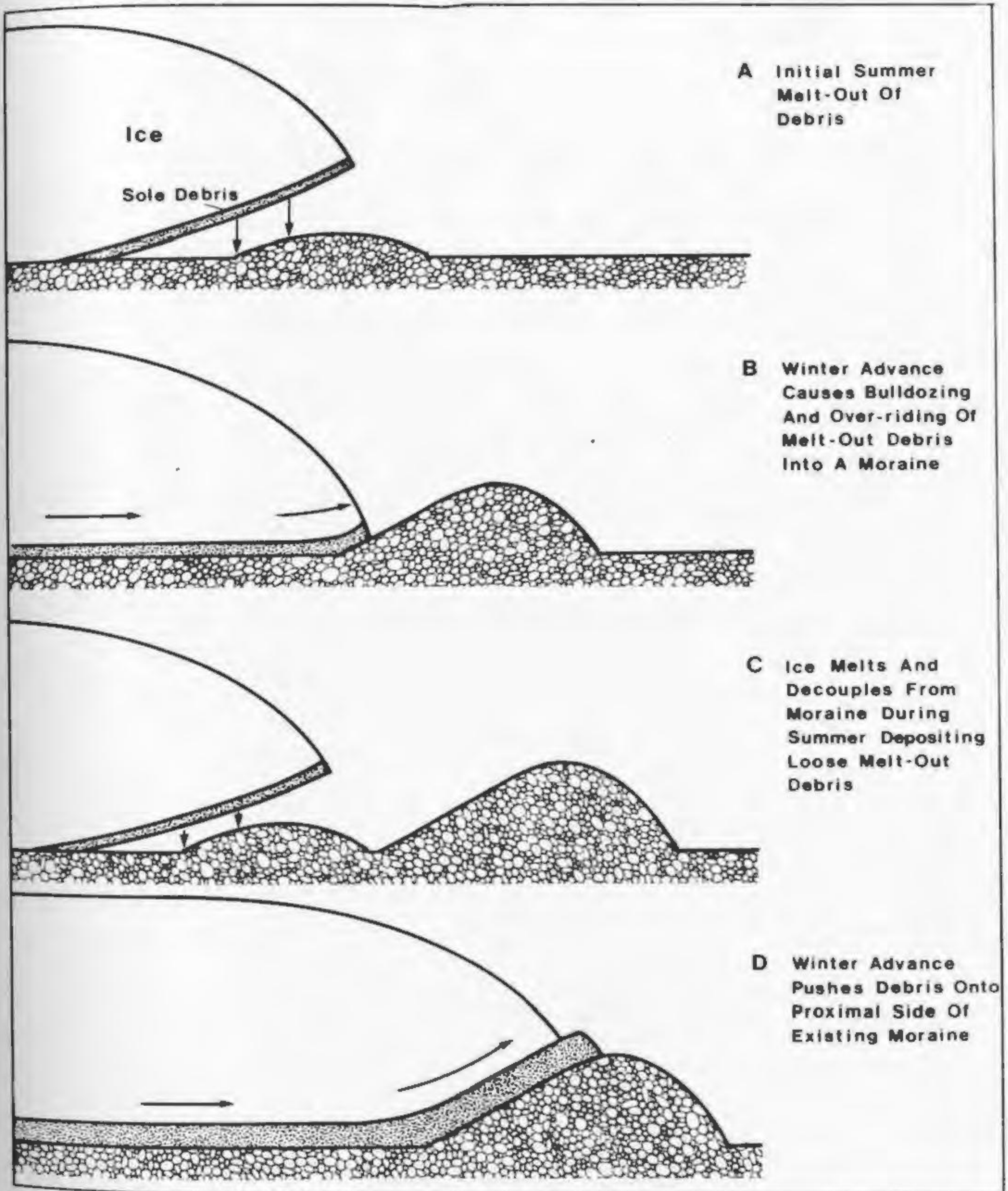

Figure 5-11 Model of the formation of Moraine B.

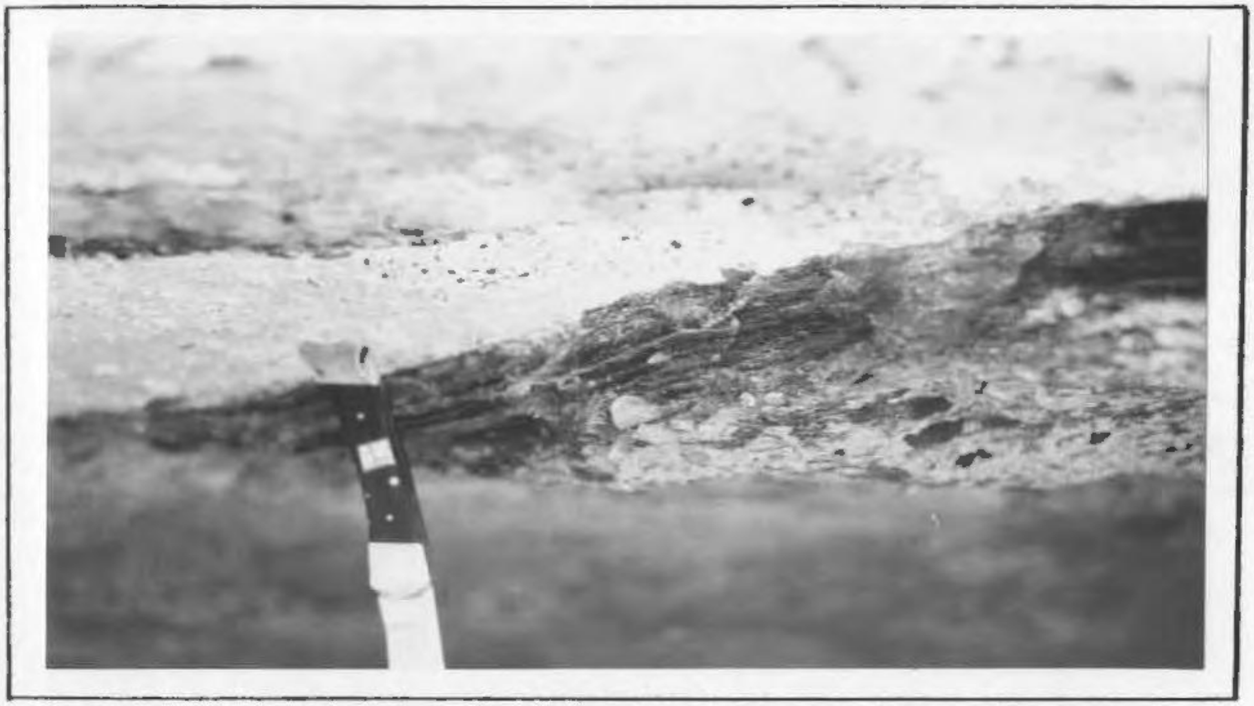

Figure 5-12 Regelation ice beneath Emerald Glacier. Banding within the deposit can be noted. This layer contributes to the melt-out material that subsequently comprises Moraine B.

Figure 5-13 The glacier sole. Melt-out debris of this kind collects proximal to Moraine B in the ablation season to be accreted to the moraine during winter advance of the terminus.

Figure 5-14 Deposition of till during overriding of Moraine B. The moraine is to the left of the photograph.

Under conditions of renewed winter activity, the moraine may be pushed forward prior to overriding, much in the same way as the continual pushing of Moraine A occurs. In this case however, the overburden is snow rather than supraglacial debris.

5.3 Summary

Moraine A is characterised by complexity of stratigraphy and origin. Initial deposition was in the form of an ice-front talus apron, whereas subsequent pushing and overriding moulded the moraine's present morphology. The squeezing/plastering of a plastic subglacial till over a short period as overriding occurred produced the distinct facies b unit.

In contrast, Moraine B exhibits a less complex stratigraphy and origin, that of winter pushing and overriding. Formative processes continued until July in 1979, although the precise time of winter advance to the moraine is uncertain. Birnie (1977) suggests the moraine forming period to be late in the accumulation season for Antarctic glaciers, and based upon glacier flow rates and the distance of glacier retreat, a similar situation may be expected for Emerald Glacier.

The contrast between the two areas stems from the existence of a heavy supraglacial debris cover proximal to Moraine A and its absence proximal to Moraine B.

Chapter 6

Conclusions

The type of winter reactivated push moraine characterising Moraine B is a not uncommon feature of temperate glaciers (e.g. Worsley 1974, Birnie 1977, Matthews et al. 1979, Rabassa et al. 1979). Similarly, minor dump moraines have been described at temperate glacier termini (Okko 1955, Boulton and Eyles 1979, Eyles 1979), although not exhibiting complexity in the form of an interstratified subglacial unit.

However, the existence of two moraine types, one of a possibly 'annual' nature around a single ice margin is a distinctly uncommon feature. It is one directly related to the existence or not of a supraglacial debris cover on proximal ice, and thus combines elements of both facies A_1 and B type moraines (Boulton and Eyles 1979).

Classically, complexity within moraines typifies subpolar or polar glaciers (Boulton 1967, 1970a, 1972), with a tripartite division of flow, melt-out and lodgement till elements representing a single depositional phase. In the case of Emerald Glacier however, complexity results from more than one formative process; the development of a talus apron and its subsequent pushing and overriding. This process has been described by Hewitt (1967) and Whalley (1974a), but in both cases only an essentially simple stratigraphy was evidenced.

Complexity at Emerald Glacier suggests several important facets of the depositional process. On the one hand a complex supraglacial stratigraphy observed within the in situ environment, is shown to have survived deposition within the moraine itself. Perhaps this testifies to the thickness of supraglacial debris, thus limiting washing and the potential removal of the fine layer. In any event, the existence of a supraglacial 'skeletal soil' within the moraine itsel , contradicts the assertion of Eyles (1979) that such a unit would never survive final deposition.

Similarly, the existence of a subglacial unit within the moraine suggests it possesses characteristics that have contributed to its survival as a distinct unit. The plastic nature of the lens is the major consideration and attests perhaps to a near terminal subglacial environment little influenced by glaciofluvial processes. The steep slopes surrounding the ice margin may result in a sub-glacial drainage system normal to the slope, rather than towards the terminus. Again, the protecting mantle of supraglacial debris limiting glacier melt may be a contributory factor in the maintenance of the subglacial unit, in that an active supraglaciofluvial environment would probably result in its removal by washing. Nevertheless, the emplacement of a subglacial unit well above its usual stratigraphic position is an unusual feature of moraine stratigraphy, but which confirms an active overriding process.

The study as a whole suggests that the moraines at Emerald Glacier cannot be simply characterised as dump or push type features. The predominance of one type of depositional process does not preclude others from occurring around a single ice margin. Similarly, the development of a supraglacial debris cover and a talus apron does not necessitate stagnation of the ice margin. Indeed, depositional processes may vary from year to year depending on glacier and sediment dynamics.

Although the depositional processes leading to the development of the Emerald Glacier moraines may be complex, the factors influencing a readvance of the ice front are equally complex.

The 1960's and early 1970's have been characterised by recession rates gradually moving toward a still-stand position, following the relatively warm trend of the previous 30 to 40 years (Gardner 1972). Minor oscillations of the glacier snout are the result of a direct response to annual climatic oscillations (Sugden and John 1976). Periods of retreat may be evident within one season and have a more profound affect on the morphology of the ice front than advances, in which a lag is involved (Young, in press).

Data from Peyto Glacier over the past 13 years suggests two periods of positive mass balances, superimposed on a general trend of glacier retreat. The years 1966-68, and 1973-4 and 1976 recorded positive balances, and corresponded to years of heavy winter snow accumulation. Similar trends may be expected in the Emerald Glacier situation.

Emerald Glacier is a considerably smaller glacier than Peyto (0.7 and 13.4 km^2 respectively). Similarly, Emerald Glacier has a less than 330 metre decrease in altitude between the backwall of the firn basin and the terminus. Based on the assumption that similar mass balance trends are evident between the two areas, it is suggested that positive changes in mass at Emerald Glacier will be reflected in a terminal advance within a short period of time. If the beginning of the most recent positive mass balance phase was 1973, it would be expected that it would affect the terminal area within a few seasons. This agrees with the recent initiation of moraine formative processes.

Similarly, Emerald Glacier has steep surface slopes (average 17.5° from backwall to terminus), and therefore negative mass balances will be poorly recorded by retreats of the glacier margin, compared to a glacier confined within a gently sloping valley situation e.g. Peyto Glacier. Also, a supraglacial debris cover protects the surface from large losses of mass through surface ice-melt.

In contrast, there will be a short response time for positive mass balances to be evidenced at the terminus, again as a result of steep slopes.

As such, small steep sloped cirque type glaciers like Emerald Glacier, especially with a supraglacial debris cover, may be sensitive indicators of periods of climatic deterioration. Valley glaciers on the other hand, are more sensitive to periods of climatic amelioration.

However, the unanswered question is whether readvance is climatically induced by deterioration, or whether past recession has established a more appropriate size/climate equilibrium. A long term measurement programme would be required to respond to such a question.

In any event, the conclusions reached at Emerald Glacier may be applied to areas of both present and past moraine formation, in two major respects. Firstly, more than one moraine forming process may be observed around a single ice margin, one of which may be 'annual' in nature; and secondly, complex moraines can be the result of temperate glacier deposition, and such a possibility must be considered when examining deposits of past glacier depositional phases.

Key to Abbreviations used in References

Am. J. Sci.	American Journal of Science.
Arctic and Alpine Res.	Arctic and Alpine Research.
Arctic Inst. N. Am. Res.	Arctic Institute of North America Research.
Bull. Can. Petroleum Geol.	Bulletin Canadian Petroleum Geology.
Bull. Geol. Soc. Am.	Bulletin Geological Society of America.
Can. J. Earth Sci.	Canadian Journal of Earth Science.
Clays and Clay Min.	Clays and Clay Minerals.
Ecol. Monographs	Ecological Monographs.
G.S.A.	Geological Society of America.
Geogr. Bull.	Geografiska Annaler.
Geogr. Polonica	Geographical Bulletin.
Geogr. Tidsskr.	Geografisk Tidsskrift.
Geolog. Foren. Forhand.	Geologiska Foreningens i Stockholm Forhandlingar.
Geol. Surv. Canada Pap.	Geological Survey of Canada Paper.
Int. Assoc. Sci. Hydrol.	International Association of Scientific Hydrology.
J. Geol.	Journal of Geology.
J. Geol. Soc. Lond.	Journal of the Geological Society of London.

J. Glaciol.	Journal of Glaciology.
J. Sed. Pet.	Journal of Sedimentary Petrology.
N.Y. Acad. Sci. Trans.	New York Academy of Science Transactions.
Norsk. Geogr. Tidsskr.	Norsk Geografisk Tidsskrift.
Quart. J. Eng. Geol.	Quarterly Journal of Engineering Geology.
Quat. Res.	Quaternary Research.
Scient. Amer.	Scientific American.
Trans. Inst. Br. Geogr.	Transactions of the Institute of British Geographers.

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Appendix A

Grain-Size Analysis

Sediment samples were collected from both constructional forms and in situ deposits, and analysed in the laboratory. Forty-nine field samples were split, leaving a residue of approximately 300-400 grams per sample. Each was soaked within a marked basin for a period ranging from 1 to 24 hours, depending on the amount of clay-size material present. With large amounts of clay a small quantity (5 ml.) of Calgon was added to aid dispersion, otherwise agitation by hand was sufficient. No organics were found within any of the samples and so no peroxide or HCl treatment was required.

Wet sieving was conducted through a 4 Ø sieve and the superlatent collected in a 4 litre beaker. This was floculated with between 5 and 20 ml. of a 0.5N $MgCl_2$ solution, depending on the amount of fines. After settling, the clean water was syphoned off. The residue was transferred to a 1 litre settling tube and dispersed in de-ionised water with 20-30 ml. of a 25 gram/litre Calgon solution.

The finer than 4 Ø fraction was analysed by the pipette method , whose recording intervals were determined by:

$$v = \frac{2}{9} (d_s - d_w) \frac{g}{n} r^2$$

where, v = velocity (cm/sec); ds = specific density ($\approx$ 2.65); d_w = density of water (0.99825); g = gravity (9.81m s^{-2}); n = atmospheric pressure (10.05 X 10^{-3} at 20°C); and r = radius of particle.

Readings were therefore taken at:

hr	m	s		hr	m	s	
0	0	28	5 Ø	0	28	31	8 Ø
0	1	47	6 Ø	1	54	05	9 Ø
0	7	08	7 Ø	7	36	20	Finer than 9 Ø

At each recording interval 20 ml of material was extracted at a 10 cm. depth, and placed in a pre-weighed weigh boat, dried and then reweighed on a chemical balance. Weight corrections were made for the admixture of flocculant and dispersant, and the decreasing percentage of 1 litre that each 20 ml. sample represented. Volume was corrected by a weighting of 1.0204 for the 6 Ø fraction, increasing to 1.1111 for finer than 9 Ø. Corrections ranged from 1.455 grams for a 20 ml. $MgCl_2$/20 ml. Calgon addition to 0.74 grams for 5 ml. $MgCl_2$/20 ml. Calgon.

The coarser than 4 Ø fraction was oven dried after wet sieving and sieved through a nest of sieves ranging from -4 Ø to 4 Ø, including half-phi intervals. Six sieves could be placed on the Rotap shaker simultaneously, and each nest of sieves were shaken for a period of 15 minutes. The separate fractions were weighed on a sartorius top-loading balance, which measured to one hundreth of a gram, and individual weights were logged on a data sheet. Results were plotted using the programme outlined in section 3-6, and the numeric values for characterising parameters are presented overleaf, in Tables A-2 to A-4.

Table A-1

Descriptive Terms for Skewness; Sorting and Kurtosis

Measured on the Phi scale

(from Briggs 1977)

Skewness (Sk)

Very Negatively skewed	-1.0 - -0.3
Negatively skewed	-0.3 - -0.1
Normal	-0.1 - 0.1
Positively skewed	0.1 - 0.3
Very Positively skewed	0.3 - 1.0

Sorting (σ)

Very well sorted	<0.35
Well sorted	0.35 - 0.50
Moderately well sorted	0.50 - 0.70
Moderately sorted	0.70 - 1.00
Poorly sorted	1.00 - 2.00
Very Poorly sorted	2.00 - 4.00
Extremely Poorly sorted	>4.00

Kurtosis (K)

Very Platykurtic	<0.67
Platykurtic	0.67 - 0.90
Mesokurtic	0.90 - 1.11
Leptokurtic	1.11 - 1.50
Very leptokurtic	1.50 - 3.00
Extremely leptokurtic	>3.00

Table A-2

Grain size Distribution of in situ supraglacial

and in situ subglacial samples

#	Location	Gravel	Sand	Silt	Clay	$\bar{m}$	sk	σ	k	Nk	C/A
16	Avalanche Slope	61.6	33.2	4.1	1.1	-1.08	.48	1.87	1.44	.59	6.24
17	Avalanche Slope	57.7	32.0	9.7	0.7	-0.51	.57	2.54	1.26	.56	3.60
18	Avalanche Slope	60.5	34.1	4.6	0.8	-0.95	.50	1.95	1.34	.57	5.20
19	Avalanche Slope	53.0	41.1	4.7	1.3	-0.78	.43	2.11	1.41	.59	4.77
25	Transect 1	36.9	51.9	8.9	2.4	0.39	.36	2.74	0.92	.48	1.69
26	Transect 1	68.3	27.3	3.3	1.2	-1.34	.50	1.75	1.47	.60	7.24
27	Transect 1	59.4	34.3	5.0	1.2	-0.82	.54	2.11	1.33	.57	4.54
28	Transect 2	62.7	31.0	4.8	1.5	-1.11	.57	2.13	1.31	.57	5.20
29	Transect 2	57.9	39.1	2.3	0.8	-1.01	.36	1.67	1.07	.52	6.28
39	Transect 3	74.2	23.8	1.3	0.7	-1.62	.49	1.34	1.21	.55	11.95
50	Crevasse Fill	55.4	26.4	14.7	3.5	0.24	.67	3.34	0.97	.49	2.50

Table A-2 (continued)

In Situ Subglacial samples

#	Location	Gravel	Sand	Silt	Clay	$\bar{m}$	sk	σ	k	Nk	C/A
33	Decoupling	37.2	41.6	14.0	7.3	1.11	.49	3.75	1.00	.50	1.53
35	Decoupling	37.4	38.4	16.4	7.7	1.26	.50	3.87	0.88	.47	1.46
37	Decoupling	36.3	34.9	22.2	6.6	1.31	.46	3.80	0.76	.43	1.33
38	Decoupling	46.9	34.8	12.9	5.5	0.49	.58	3.48	1.04	.51	2.11
40	Decoupling	44.3	45.0	8.1	2.7	-0.06	.46	2.71	1.15	.53	2.47
41	Decoupling	26.6	28.5	22.1	22.8	3.37	.23	5.16	0.71	.41	0.70
49	Regelation	33.5	39.9	26.3	0.3	1.66	.42	3.78	0.67	.40	1.79

Table A-3

Grain Size Distribution of

Moraine A Samples

#	Location		Gravel	Sand	Silt	Clay	$\bar{m}$	sk	σ	k	Nk	C/A
10	Site 2	a	52.8	27.5	10.2	9.5	0.68	.70	4.11	1.15	.54	2.27
11	Site 2	a	41.4	29.5	13.4	15.7	1.80	.59	4.00	0.78	.44	1.40
12	Site 2	a	45.1	31.0	12.2	11.7	1.19	.63	4.45	1.00	.50	1.82
13	Site 2	a	51.7	30.1	9.8	8.4	0.54	.67	3.04	1.25	.55	2.39
14	Site 2	b	25.6	30.4	20.3	24.0	3.36	.27	5.12	0.70	.41	0.72
15	Site 2	c	52.5	39.4	5.9	2.2	-0.53	.49	2.49	1.11	.53	3.13
20	Site 4	a	73.7	19.4	4.6	2.3	-1.50	.61	1.97	2.05	.67	7.01
21	Site 4	a	67.3	25.8	4.6	2.4	-1.21	.57	2.10	1.77	.64	5.83
22	Site 4	a	62.1	32.1	3.4	2.4	-1.07	.52	2.04	1.49	.60	5.57
23	Site 4	b	30.3	41.0	15.4	13.2	2.14	.45	4.51	0.86	.46	1.05
24	Site 4	c	43.8	46.2	7.4	2.7	0.01	.49	2.67	1.22	.55	2.70
30	Site 5	a	72.2	25.3	1.7	0.8	-1.54	.45	1.39	1.29	.56	10.89
31	Site 5	a	72.0	24.5	2.3	1.3	-1.50	.46	1.51	1.40	.58	9.97
32	Site 5	a	73.6	22.1	2.7	1.6	-1.57	.59	1.64	1.52	.60	8.84

Table A-3 (continued)

#	Location		Gravel	Sand	Silt	Clay	$\bar{m}$	sk	σ	k	Nk	C/A
43	Site 6	b	34.2	40.1	17.9	7.8	1.39	.41	3.83	0.84	.46	1.65
44	Site 7	b	42.6	33.0	12.6	11.8	1.30	.61	4.43	0.98	.50	1.69
45	Site 7	c	39.9	43.8	10.3	6.1	0.48	.47	3.36	1.25	.56	2.00
46	Site 7	a_1	62.1	28.0	5.4	4.4	-0.91	.62	2.61	1.71	.63	4.41
47	Site 8	b	42.9	34.1	13.3	9.7	1.09	.58	4.16	0.97	.49	1.61
48	Site 8	a	57.8	28.0	7.7	6.7	-0.26	.67	3.25	1.61	.62	3.28

Table 1-4

Grain Size Distribution of

Moraine B Samples

#	Location		Gravel	Sand	Silt	Clay	$\bar{m}$	sk	σ	k	Nk	C/A
1	Site 1	d_1	40.1	34.3	20.6	5.0	0.98	.46	3.58	0.74	0.43	1.46
2	Site 1	d_1	29.2	33.2	31.9	5.7	1.00	.20	3.65	0.68	0.40	0.90
3	Site 1	d_1	43.8	35.9	13.8	6.5	0.76	.56	3.70	0.98	0.50	1.87
4	Site 1	d	60.5	32.8	5.0	1.7	-0.97	.54	2.14	1.33	0.57	4.82
5	Site 1	d	51.8	36.7	9.1	2.4	-0.28	.53	2.71	1.13	0.53	2.94
6	Site 1	d	48.3	40.5	8.6	2.6	-0.23	.48	2.72	1.13	0.53	2.74
7	Site 1	d	45.9	37.5	13.1	3.5	0.34	.53	3.18	1.01	0.50	2.25
8	Site 1	d	48.8	38.1	9.6	3.5	-0.06	.53	2.95	1.11	0.53	2.53
9	Site 1	d	50.0	39.6	7.7	2.7	-0.25	.52	2.66	1.17	0.54	2.86
42	Site 3	d	44.2	36.7	13.7	5.5	0.64	.53	3.57	0.94	0.49	1.84

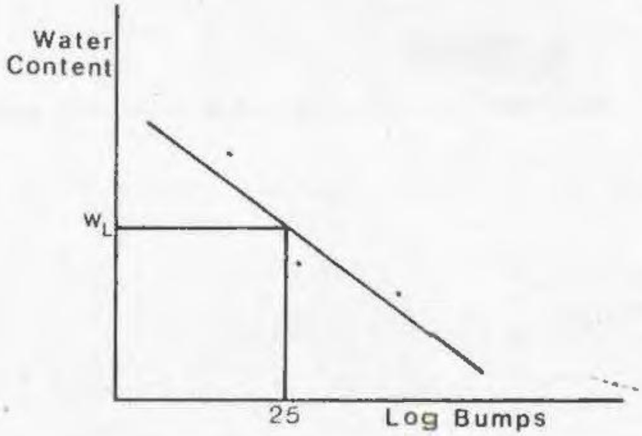
Appendix B

Atterberg Limits

Atterberg limits are conducted on samples which pass through the U.S. standard sieve No. 40 (about 1.25 Ø) i.e. medium-fine sand. Approximately 100 grams of sample is needed for the completion of the Atterberg tests, and therefore only 31 out of 49 samples were analysed. The major omissions were supraglacial samples which have a low silt-clay content.

Tests completed were the determination of the plastic limit and liquid limit. In the former a small quantity (≈20 grams) of sample is wetted and rolled, until the resultant thread crumbles at a diameter of 1/8 inch. The process is repeated three times, and after drying the average water content defines the plastic limit.

The liquid limit is arbitrarily defined by the water content at which 25 blows (a 1 cm. drop each) of a liquid limit device (Casagrande 1932) closes a standard (11 mm. long) groove cut in a soil pat. Experimentally, the position of the 25 bump water content is calculated from a plot of 3 separate determinations of water content ranging either side of 25 bumps.



Determination of Liquid Limit (W_L)

The plasticity index is defined as the difference between the liquid and plastic limits.

Results are presented in Table B overleaf. Values for the activity ratio are derived from the relationship between the plasticity index and the percentage of clay-size material (Means and Parcher 1964). Relatively inactive (RI) soils are characterised by values of 0-0.75, normally active (NA) soils by 0.75-1.50, and active (A) soils over 1.50.

Table B

Atterberg Limit Values from In Situ Subglacial,

Moraine A and Moraine B Deposits

In Situ Subglacial Deposits

#	Location	Liquid Limit	Plastic Limit	Plasticity Index	% Clay Size	Activity	Ratio
33	Decoupling	14.00	11.32	2.68	7.31	0.37	RI
35	Decoupling	15.20	13.04	2.16	7.74	0.28	RI
36	Washed Thro. Mor. A	14.80	12.50	2.30	24.16	0.10	RI
37	Decoupling	14.65	13.41	1.24	6.62	0.19	RI
38	Decoupling	14.55	12.19	2.36	5.45	0.43	RI
40	Decoupling	14.98	14.29	0.69	2.66	0.26	RI

Moraine A Deposits

#	Location/Facies	Liquid Limit	Plastic Limit	Plasticity Index	% Clay Size	Activity	Ratio
10	Site 2 / a	16.82	13.15	3.67	9.54	0.38	RI
11	Site 2 / a	16.85	13.42	3.43	15.73	0.22	RI
12	Site 2 / a	16.55	14.42	2.13	11.68	0.18	RI
13	Site 2 / a	16.58	13.48	3.10	8.41	0.37	RI
14	Site 2 / b	19.65	13.22	6.43	23.98	0.27	RI
15	Site 2 / c	18.62	15.48	3.14	2.18	1.44	NA
21	Site 4 / a	17.82	14.11	3.71	2.37	1.57	A
23	Site 4 / b	15.86	11.19	4.67	13.21	0.24	RI
24	Site 4 / c	17.90	15.14	2.76	2.68	1.03	NA

Table B (continued)

Moraine A Deposits

#	Location/Facies	Liquid Limit	Plastic Limit	Plasticity Index	% Clay Size	Activity	Ratio
30	Site 5 / a	19.90	16.24	3.56	1.26	1.80	A
31	Site 5 / a	19.89	18.93	0.96	0.82	1.17	NA
43	Site 6 / b	14.73	12.90	1.83	6.97	0.24	RI
44	Site 7 / b	15.42	11.69	3.73	11.79	0.32	RI
46	Site 7 / a_1	16.95	12.40	4.55	4.44	1.02	NA
48	Site 8 / b	14.48	12.09	2.39	6.65	0.36	RI

Moraine B Deposits

#	Location/Facies	Liquid Limit	Plastic Limit	Plasticity Index	% Clay Size	Activity	Ratio
1	Site 1 / d_1	13.45	13.67	0.28	5.02	0.06	RI
2	Site 1 / d_1	16.36	15.43	0.93	5.68	0.16	RI
3	Site 1 / d_1	14.51	11.96	2.55	6.54	0.39	RI
4	Site 1 / d_1	15.08	14.10	0.98	1.65	0.59	RI
5	Site 1 / d_1	14.43	13.03	1.40	2.43	0.58	RI
6	Site 1 / d_1	14.97	14.27	0.70	2.60	0.27	RI
7	Site 1 / d_1	15.07	13.84	1.23	3.54	0.35	RI
8	Site 1 / d_1	14.77	12.47	2.30	3.54	0.65	RI
9	Site 1 / d_1	15.22	14.55	0.67	2.71	0.25	RI
42	Site 3 / d	15.11	13.17	1.94	5.50	0.35	RI

Appendix C

X-Ray Diffraction

Analysis of the finer than 4 Ø fraction allows the relative abundance of clay mineral and silt fraction to be determined, as well as defining the type of clay minerals present.

Slides were prepared by making a smear of a solution of the silt/clay component and allowing to dry.

A Phillips diffractometer with a 2 k.w. normal focus copper anode tube was used, and results recorded by a proportional detector with a modified monochromator. Scan speed was 1° per minute between 4° and 35° of 2θ, with a 1° divergence slit and a 0.2 mm. receiving slit. Chart speed was 1 cm. per minute.

Three slides were prepared for each sample, one being run as an untreated sample. The presence of Kaolinite was tested by heat treating a second slide at $550^\circ C$ for 1 hour, analysing the resultant and comparing to the untreated. Collapse of the 7 Å peak, identifies kaolinite (Grim 1962, Carroll 1970, Brown 1972). The third slide is placed within a dessicator containing ethelene glycol at $80^\circ C$ for 1 hour.

A characteristic swelling from a peak of about 15 Å to one of about 17-18 Å is associated with montmorillonite-type clays.

After the random analysis of 15 samples using heat treatment and glycolation no kaolinite or montmorillonite-type clays were found. Subsequent analyses were therefore undertaken on untreated samples alone.

Results were presented in a chart form. Relative values for clay and non-clay mineral peaks were characterised on the basis of the height of the peak measured by a scale on the chart itself:

Very low	(L^-)	<10
Low	(L)	10 - 40
Moderate	(M)	40 - 70
High	(H)	70 - 100
Very High	(H^+)	>100

In the following tables major second and fourth order chlorite peaks and first and third order Illite/mica peaks are recorded, as well as the major non-clay mineral peaks of quartzite (3.3Å), calcite (3Å) and dolomite (2.9Å). Other identified diffraction peaks e.g. 14Å and 4.7Å chlorite, 4.4Å mica, 3.7Å and 2.7Å dolomite, 4.2Å quartz, 3.2Å feldspar consistently had very low or low peaks and are therefore not recorded.

The clay: non-clay ratio value is derived from a numeric representation of major chlorite and illite/mica peaks, (7Å and 10Å respectively) and the major quartz and dolomite peaks (3.3Å and 2.9Å respectively).

Table C

X-Ray Diffraction analysis results: In-Situ supraglacial and subglacial deposits.

In-Situ Supraglacial deposits

#	Location	Chlorite		Mica/Illite		Quartz	Dolomite	Calcite	Clay: Non-clay
		7Å	3.5Å	10Å	5Å	3.3Å	2.9Å	3Å	
16	Avalanche Slope	H^+	H^+	H	M	H^+	L	L	.79
17	Avalanche Slope	L	L	L	L	H^+	H^+	L	.13
18	Avalanche Slope	L	L	L	L	H^+	H^+	L	.16
19	Avalanche Slope	H	M	L	L	H^+	H^+	L	.41
25	Transect 1	L^-	L	M	L	H^+	H^+	L	.20
26	Transect 1	L	L	H	M	H^+	M	M	.42
27	Transect 1	L^-	L^-	M	L	H^+	L	M	.56
28	Transect 2	L	M	L	L	H	M	L	.42
29	Transect 2	M	M	M	L	H^+	H^+	L	.40
39	Transect 3	M	M	H	L	H^+	L	L	.78

In-Situ Subglacial deposits

#	Location	Chlorite		Mica/Illite		Quartz	Dolomite	Calcite	Clay: Non-clay
		7Å	3.5Å	10Å	5Å	3.3Å	2.9Å	3Å	
33	Decoupling	L	L	L	L^-	M	H	L	.30
35	Decoupling	L	L	M	L	M	H^+	L	.29
37	Decoupling	L	L	M	L	M	H^+	L	.30
38	Decoupling	L^-	L	L	L	H	H^+	M	.18
40	Decoupling	L^-	L^-	L	L^-	M	H^+	L	.14

Table C

X-Ray Diffraction Analysis results:

Moraine A

#	Location	Chlorite		Illite/Mica		Quartz	Dolomite	Calcite	Clay: Non-clay
		7 Å	3.5 Å	10 Å	5 Å	3.3 Å	2.9 Å	3 Å	
11	Site 2 a	M	M	M/H	L	H^+	M	M	.62
12	Site 2 a	L	L/M	L	L	H^+	H	M	.30
13	Site 2 a	M	M	H	L	H^+	H	M	.50
14	Site 2 b	L	L	L	L	H^+	M/H	M	.28
15	Site 2 c	M	M	M	L/M	H^+	H	L	.45
20	Site 4 a	L	L	H	M	H^+	H^+	M	.37
21	Site 4 a	L	L	M	L	H^+	H	L	.33
22	Site 4 a	L	L	M	L	H^+	M	H	.37
23	Site 4 b	L	L	H	L	H	H	L	.58
24	Site 4 c	L	L	H^+	H	H^+	H	L	.54
30	Site 5 a	M	M	H	M	H^+	–	L	.77
31	Site 5 a	H	M	H	M	H^+	–	L	.67
32	Site 5 a	M	M	H	L	H^+	L	L	.91
43	Site 6 b	L^-	L^-	L	L^-	M	H^+	L/M	.15
44	Site 7 b	L	L	L	L	H^+	H^+	M	.20
45	Site 7 c	L	L	L	L	H^+	H^+	M	.14
46	Site 7 a_1	M	M	M	L	H^+	H^+	M	.39
47	Site 8 b	M	M	H	M	H^+	H^+	L	.39
48	Site 8 a	M	L	M	L	H^+	H^+	L	.28

Table C

X-Ray Diffraction analysis results:

Moraine B

#	Location	Chlorite		Mica/Illite		Quartz	Dolomite	Calcite	Clay: Non-clay
		7 Å	3.5 Å	10 Å	5 Å	3.3 Å	2.9 Å	3 Å	
1	Site 1 d_1	L^-	L^-	L	L^-	L	H^+	L	.11
2	Site 1 d_1	L^-	L^-	L	L^-	L	H^+	H	.14
3	Site 1 d_1	L^-	L^-	L	L	H	H^+	L	.18
4	Site 1 d	L^-	L^-	L	L	H	H^+	L	.14
5	Site 1 d	L^-	L^-	L	L	M	H^+	L	.17
6	Site 1 d	L^-	L^-	L	L^-	L	H^+	M	.16
7	Site 1 d	L	L^-	L	L	H^+	H^+	L	.16
[illegible]	Site 1 d	L^-	L^-	L	L	H^+	H^+	L	.14
9	Site 1 d	L	L^-	L	L	H^+	H^+	L	.18
[illegible]	Site 3 d	L^-	L^-	L	L^-	M	H^+	M	.17

Appendix D

Sampling Procedure

The selection of sampling sites for both continuous measurement and stratigraphic sections was constrained by a series of limiting factors.

The major constraint was that of time. The field season was a short one, being further reduced by occasional adverse weather conditions. Only a limited amount of data could therefore be accurately and efficiently collected.

The continuous measurement programmes were further hampered by the need to find suitable datum points located away from the ice margin. A stratified random line sampling pattern (Cole and King 1968) was therefore employed, with the aim of distributing measurement sites around the ice margin as much as possible.

The field area was also sufficiently remote to warrant some selectivity in sediment sampling being employed. This factor is also considered on page 41, and partly explains the lack of sedimentological data for the coarser than -2.5 Ø fraction.

In general, the sampling procedure undertaken in the field was considered the most effective means of collecting data from the ice margin area within the limitations imposed by time and environmental conditions.

Reference: Cole, J.P. and King, C.A.M., 1968: Quantitative Geography , Wiley & Sons London