EROSION OF PALEOZOIC BEDROCK IN THE TERMINAL ZONE OF YOHO GLACIER, BRITISH COLUMBIA

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THE TERMINAL ZONE OF YOHO GLACIER,
BRITISH COLUMBIA

by

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Abstract

Recent research has revealed differing points of view as to whether the physical characteristics of glacial debris are determined by the lithology of eroded bedrock or are a function of the environment of transport and deposition. A gap exists in the literature concerning short distances of transportation where relatively soft sedimentary rocks predominate and the resultant eroded bedform morphologies.

In the proglacial zone of Yoho Glacier, British Columbia, there are seven main sedimentary lithologies which have been exposed by recent, post-Neoglacial recession. The beds lie across the former direction of ice movement. Both erosional and depositional surfaces are exposed and virtually unmodified by other geomorphic and subaerial processes. It is therefore possible to compare the bedrock erosion and comminution processes for varying adjacent sedimentary bedrock types under the same general ice conditions.

In order to gain an understanding of sub-glacial erosional processes from the examination of bedrock lithology, contents of related tills, and surface roughness, certain field and laboratory procedures were adopted and designed. In an attempt to combine glaciological and glacial-geological practices, certain procedures can be viewed as standard. Others have been adopted from the engineering and computer sciences disciplines to help demonstrate some physical and chemical characteristics of lithology, tills and surfaces (bedrock) that have in the past been given only passing examination.

Laboratory analyses conducted on till samples did not produce results that enable definite conclusions to be drawn about the relationship between

bedrock type and their volumetric abundance in till. Although bimodality is observed in the grain size distribution curves with apparent terminal grade modes, it is doubtful that these characteristics are solely the result of either bedrock type, distance of transport, or mode of transport, but rather a combination of these. The excavation of large boulders and their subsequent deposition down-valley from their source (1.5 km) clearly indicates that not all materials are greatly reduced in bulk. However, because bimodality is observed in all samples, the bedrock units were probably comminuted to some degree due to a combination of abrasion and plucking, in transport for up to 1.5 km from their source.

To indicate the roughness of a given bedrock unit, the extent of each physical and chemical characteristic of the bedrock unit must be specified. It is hypothesized that 'ideal plane' or 'ideal sliding surface' configurations (surface roughness) of bedrock units need not necessarily be horizontally level to enhance glacier flow. Based on physical and chemical characteristics of bedrock types, and of glacier ice conditions at the ice-rock interface, each bedrock unit tends toward its own configuration which enhances flow and retards erosion. The roughness of a bedrock unit can be viewed as its morphological variation from its ideal plane or ideal sliding surface configuration which may not necessarily be a smooth surface (horizontally level).

The results of bedrock unit strength tests (susceptibility to erosion by physical and chemical processes), and of the ranking of bedrock units with reference to slope frequency distribution, ideal plane configuration and degree of horizontal levelness clearly indicates that over a given time period, relatively weaker bedrock units will erode to a smoother surface than stronger bedrock units.

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Dedicated to Judy Ford of Port aux Basques.

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Chapter 1

Objectives and Introduction

The Symposium on Glacier Beds held in Ottawa in 1978, (Glen *et al.* 1979) provided a forum in which both glaciologists and glacial geologists presented and discussed related works and made recommendations concerning the coordination of future endeavours. The previous state of these sciences was summed up by Meier (1979) who stated: "One of the problems that we have noticed in the past is a separation of people who are interested in ice from those people who are interested in the effects of past ice movements." Lliboutry (1979) commented that more data on geology and microstructures of glacial beds was needed in order to advance the theoretical knowledge of sliding and bottom creep.

Comments made by Hallet (1979), Röthlisberger (1979) and Boulton (1979) brought into perspective several of the more popular concerns discussed at the symposium. Hallet (1979) states: "Glacial geologists could study the debris cover of recently deglaciated glacier beds to obtain information on the abundance of debris present within or at the base of retreating glaciers...if one sees a vast expanse of glaciated bedrock with little debris, perhaps bordered by distinct moraine, and if one can establish that much of the material was not removed by proglacial streams, the former debris load in and on the glacier must have been very low." This suggests perhaps limited subglacial erosion of bedrock.

Röthlisberger (1979) commented on the relative importance of the plucking and abrasion processes, and stated that because more fine particles than coarse were carried by glaciers, abrasion seems to be the dominant process. Boulton (1979) on the other hand suggests that given certain circumstances, plucking can be the more dominant process. Basically, Boulton believes that abrasion is brought about by "...rubbing by a tool which produces fine materials from breakdown both of the bed and the tool." The plucking process produces the tool.

The study of the process (abrasion, plucking, solution) - response (bedform, till content) system at an interdisciplinary level (glaciology: glacial geology) can enhance our understanding of this problem.

The basic concern of this study is to gain an understanding of subglacial erosional processes from the examination of bedrock lithology, contents of related tills, and surface roughness characteristics. The field area is recently deglaciated and consists of at least seven Paleozoic bedrock units with relatively little surface till cover. It has a distinct Neoglacial latero-terminal moraine complex (similar to that described by Hallet (1979) in his recommendations for field study), and shows few signs of post-glacial modification. The area also shows signs of widespread plucking as evidenced by roches moutonnée forms, and of abrasion which is manifested by striated surfaces and related erosional bedforms.

The objectives of this thesis can be itemized as follows:

i) To determine (through the use of three-dimensional graphics and related statistical parameters) if different Middle Cambrian rock units exhibit different glacially eroded bedforms (roughness) at a microscale.

ii) To demonstrate how various lithological characteristics (physical and chemical) can be used in the analysis of erosional processes.

iii) To examine the contents of tills and relate these to the erosional processes.

iv) To attempt a general assessment of the rates of erosion on different substrata.

1.1 The Field Area

Yoho Glacier (51°35'N:116°32'W) is located in the northern extremity of Yoho National Park in the Waputik Mountains (Continental Ranges, Rocky Mountains, British Columbia) approximately 18 kms north of Field, British Columbia. The glacier is part of the Wapta Icefield which straddles the British Columbia/Alberta border and the Yoho and Banff National Parks boundary (Figure 1). The elevation of Yoho Glacier ranges from 2130 m amsl (above mean sea level) at the terminus to over 2430 m amsl at its source in the Wapta Icefield.

The area of detailed study lies between .5 km and 2.0 km from the terminus of Yoho Glacier. The area is part of a bedrock valley floor which was covered by glacier ice less than 50 years ago. This freshly glaciated surface takes the form of a concave upwards rock bench which is obscured in a few areas by small pods of drift. The Yoho River to the east, lies in a gorge more than 20 m below the bench. Proglacial fluvial modification of the glaciated surface is therefore slight and confined to small tributary streams which generally flow along the bedding planes of the bedrock.

A continuous Neoglacial latero-terminal moraine, marking the Neoglacial maximum (c. 1844: Bray and Struik 1963) extends along the western edge of the glaciated valley for about 500 metres. The moraine is broken where the valley narrows and steep cliff-forming units of the

Yoho Glacier: Location of Study Area

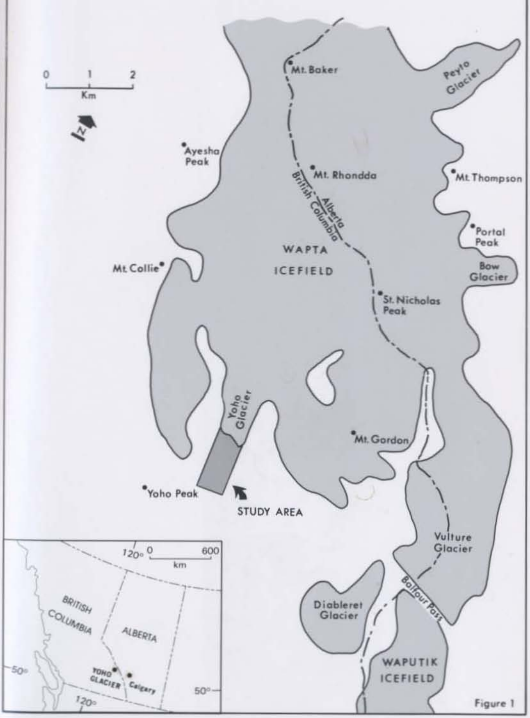


Figure 1

Sullivan Formation dominate. Three Neoglacial recessional moraines approximately 3 to 5 metres in height are located 2 km down-valley from the present glacier terminus. They are spaced approximately 50 metres apart and can be traced across the Yoho River. Wheeler (1931), and Bray and Struik (1963) date these moraines at about 1901, 1904 and 1910 respectively on the basis of dendrochronologic techniques.

1.2 Climate

The essential characteristics of climate in the Rocky Mountains result from a continental location, high relief and a north-south oriented topography, normal to prevailing westerly winds. Several short treatises referring to the climate of the Rocky Mountains have appeared but for the most part these publications have been extremely general or have dealt with specific areas or topics outside the Yoho Valley. The lack of data and the complexities of mountain climates have discouraged comprehensive description of climatic conditions (Janz and Storr, 1977). Generally, stations are scattered over large areas, almost all located in valleys and at best represent only the conditions at that elevation and site. Climatic elements are controlled to a large extent by such terrain characteristics as elevation and aspect. This obviously results in extreme spatial variability of climate within mountainous areas for any one period of time.

Because there is no meteorological station located in the immediate vicinity of the field area, this discussion necessitates the use of data and summaries from the Banff townsite; Field, British Columbia; and Peyto Glacier stations (Table 1). The application of these data to the field area is therefore suggestive and intended to give a realistic impression rather than describe precise climatic parameters.

	Field	Banff	Peyto Glacier
Location	Field, B.C. Lat 51° 24'N Lon 116° 29'W	Banff, Alta. Lat 31° 11'N Lon 115° 34'W	Banff National Park, Alta. Lat 51° 40'N Lon 116° 35'W
Glacier area			13.4km²
Max. elevation			3185 masl
Min. elevation			2125 masl
Mean elevation	1246 masl	1368 masl	2635 masl
Mean surface slope			12.9°
Mean azimith of surface/valley	c. 090°	c.160°	033°
Instrumentation elevation		1396 masl	2220 masl (100m from glacier margin)

Source: Young and Stanley (1976)

Table 1

General Features of the Field (British Columbia) and Banff (Alberta) Townsites, and Peyto Glacier Meteorological Station

1.3 Sunshine

Janz and Storr (1977) define the duration of bright sunshine as of sufficient intensity to burn or to scorch standard sunshine cards inserted in a Campbell-Stokes Sunshine Recorder (Environment Canada, Atmospheric Environment Service). The closest station continuously recording such data is located at the Banff townsite (Global radiation measurements are carried out sporadically at the Peyto Glacier: Goodison (1972); Young, per. comm.). The average number of hours with bright sunshine at Banff is given in Table 2.

Although the actual hours of sunshine at the field area are thought to differ from those recorded at Banff, relative changes from month to month will probably not vary drastically. Surrounding mountains may reduce the hours of potential sunshine at various sites on many days, especially during the winter months when the sun is low in the sky. Because the field area is positioned in a narrow north-south oriented valley, the valley sides greatly reduce the number of sunshine hours. It is estimated that during the field months of July and August only 150 to 210 hours of sunshine per month were experienced in the centre of the valley (based on observation, no instrumentation).

1.4 Temperature

Temperature regimes in the mountain regions are controlled mainly by the radiation cycle, the succession of airmasses, and elevation (Foessel 1974; Janz and Storr 1977). Average monthly temperatures during the summer (June to August) range from 12°C to 5°C at Field, British Columbia. Lapse rates of .7°C to .8°C/100 m reduce the temperature at the field area (Yoho Glacier and vicinity) to a range of approximately 6°C to 0°C. In the immediate glacier environment (and down valley from

Table 2

Average Number of Hours With Bright Sunshine at Banff, 1941-1970

Jan	59
Feb	93
Mar	126
Apr	160
May	200
Jun	207
Jul	255
Aug	214
Sept	171
Oct	130
Nov	82
Dec	92

Annual Total: 1739

Source: Atmospheric Environment Service Janz and Storr (1977)

Yoho Glacier), the influence of katabatic winds further reduces temperatures by 2°C to 3°C.

Temperature ranges are characteristic of continental mountain climates; 83°C in the Peyto Glacier area, 78°C at Field, British Columbia (Janz and Storr 1977). The proximity of the Wapta Icefield would affect the temperature range at the field area with winter extremes being more pronounced and summer extremes less pronounced than at Field, British Columbia. This condition is partly a function of elevation, resulting in temperature regimes in higher icefield areas which are less extreme than those in the valleys.

1.5 Precipitation

The continental nature of the climate is demonstrated in the year to year variations of precipitation (Janz and Storr 1977). Topographic influences probably cause rather complex variations of rainfall and snowfall over short distances. The estimated average precipitation at the field area is over 1000 mm, compared to 500 mm at the Banff townsite and 750 mm at Field, B. C. (Janz and Storr 1977). The principle that precipitation increases with increasing elevation is generally accepted.

1.6 Wind

Janz and Storr (1977) stress two very important conditions when dealing with wind data from mountainous regions: i) Winds vary greatly (in both direction and strength) and therefore where they are measured may only be representative of the place of measurement. ii) While broad scale pressure differences control the general wind conditions, topography exerts a modifying influence on both direction and speed, as does the presence of glaciers (A. Stenning 1980, unpublished).

Since the prevailing wind aloft over the field area is westerly, it follows that valleys with an east-west orientation are the high wind areas. However, this is not to say that the north-south oriented Yoho Valley does not experience strong winds. Glacier winds have been shown to have considerable influence on circulation patterns in glacial valleys (A. Stenning, 1980, unpublished). Janz and Storr (1977) and Stenning (1980, unpub.) suggest these influences are generally of relatively short duration. During winter months, easterly winds over Kicking Horse Pass combine with winds off the Wapta Icefield which are funnelled down the Yoho Valley and create the famous "Yoho Blows"; events of high storm conditions at Field, B. C. (R. A. Doherty, personal communication).

1.7 Vegetation

The field area lies in a transitional subalpine forest-alpine tundra vegetation zone or timberline area. Characteristic of this area are the spire-like trees of heavy snow country and stunted or dwarfed trees known as Krummholtz. This dwarfing is largely a result of wind action and inadequate soil development (Blood 1976).

Bray and Struik (1963) have carried out botanical studies in the Yoho Valley and described the major vegetational species. They report that *Picea engelmannii* is the dominant tree species with *Abies lasiocarpa* occasionally appearing in the forest stands. Important understorey plants are *Menziesia glabella*, *Rhododendron albiflorum*, *Vaccinium membranaceum*, *Vaccinium scoparium*, *Vaccinium caespitosum*, *Phyllodoce empetriformis*, *Cassiope mertensiana*, *Valeriana sitchensis*, *Lycopodium annotinum* and *Sphagnum*.

Forest stands occur on a variety of slope conditions at elevations ranging from 1750 to 2100 m amsl. All sites are reported to be mesic to mesic-dry (Bray and Struik 1963).

The immediate field area is largely unvegetated due to its recent date of deglaciation and poor skeletal soil cover. It does, however lie below the immediate tree line by 120 m.

1.8 History of the Yoho Glacier

Glaciers of various sizes and types are present in the Rocky Mountains. They include icefields, valley glaciers of the outlet type, valley glaciers of the alpine type, cirque glaciers and cliff mountain shelf or niche glaciers (Groom 1959; Gardner 1972; Ommanney 1976). These glaciers are largely remnants of the Neoglaciation, a period covering the last 400 to 500 years B. P., although the larger and higher icefields may have survived since the late Wisconsin and therefore be at least 30,000 years old (Heusser 1956; Porter and Denton 1967; Gardner 1972; and others).

Yoho Glacier, previously known as Wapta Glacier (Wheeler 1931), is the largest southerly outflow from the Wapta Icefield. The area of the icefield is approximately 50 km^2 and occupies a basin enclosed by mounts Gordon, Olive, Thompson, Baker, Collie, Des Poilou, McArthur, Agesta, Isolated and Yoho Peaks together with connecting ridges (Figure 1).

Scientific observation of the Yoho Glacier was initiated by W. Hillell under the auspices of the Smithsonian Institute in 1901 (Sherzer 1907). In 1906, the Alpine Club of Canada commenced an annual surveillance of Yoho Glacier with two objectives:

i) To ascertain retreat or advance, and

ii) To record surface flow through the force of gravity (Wheeler 1931).

Although these treks were mainly of an outing nature, they did provide valuable scientific data and photographs which are useful for comparison with present day glacier extent (Figures 2 a, b, c, and d).

Much of the glacial data from the Canadian Rocky Mountains concerns frontal or terminal recession. The condition of recession has prevailed for the last seventy years and likely since the Neoglacial maximum (Gardner 1972). It has occurred at a variable rate, probably reaching a maximum in the 1940's and 1950's. Sherzer (1905), Heusser (1956), and Bray and Struik (1963) place the Yoho Glacier Neoglacial maximum at around 1857 when the terminus was almost 2 km down-valley from its present position. Recession data for Yoho Glacier are given in Table 3.

The fluctuating rate of glacial recession has been explained in terms of climatic change (Heusser 1956; Brunger, Nelson and Ashwell 1967; Henoch 1971; Gardner 1972; and Slaymaker and McPherson 1972). Recession from c.1910 to c.1950 is explained with respect to a general increase in mean annual temperatures and a decrease in precipitation up to the early 1940's. Collier (1957) postulates that a decline in temperature and an increase in precipitation in the 1950's resulted in a glacial readvance in the Cordillera. Although this readvance did not occur in many parts of the Canadian Rockies, a marked decline in the rate of recession is noted (Hubley1956; West and Maki 1961; and Gardner 1972). Climatic conditions during the last decade may be responsible for present day readvance evident at the termini of Emerald and President glaciers in the Yoho Valley (M. Batterson 1980, unpublished; R. J. Rogerson, personal communication). Terminal readvance is not currently evident at Yoho Glacier.

1.9 Canadian Rocky Mountain Pleistocene Chronology

Intensive study in the valleys and passes east of the main range of the Canadian Rocky Mountains has produced some descriptive accounts of glaciation for the mountain region (Beach 1943; Bostock 1948; Belyea 1960;

Figure 2a, b c and d

This series of photographs was taken from approximately the same location on the Neoglacial latero-terminal moraine looking towards the terminus of Yoho Glacier. Figures 2a, b and c are reproduced with the permission of the Peter and Catharine Whyte Foundation, Archives of the Canadian Rockies, Banff, Alta. Figure 2d taken by R.J. Kodybka, 1980.

Figure 2a Terminus of Yoho Glacier, 1906.

Note lack of supra-glacial debris, especially along cliff wall (out of photo view) to the left of photo.

Figure 2b Terminus of Yoho Glacier, 1913.

Figure 2c Terminus of Yoho Glacier, 1931.

Figure 2d
View looking up-valley towards Yoho Glacier which is partly obscured from view by the bedrock formation on the eastern side of the Yoho River, 1980.

Year	Recession (m/yr)
1901-1904	11.0
1904-1906	---
1906-1907	06.0
1907-1908	11.4
1908-1909	11.8
1909-1910	14.1
1910-1912	14.9
1912-1914	09.4
1914-1916	29.8
1916-1917	07.9
1917-1918	14.5
1918-1919	15.0[1]
1919-1931	13.8[2]
1931-1980	20.4[3]

Total recession 1901-1980, 1412m

Source: Wheeler (1931); 1 and 2 from Slaymaker and McPherson (1972); 3 from Kodybka (1980 field season)

Table 3

Recession Rates at the Terminus of Yoho Glacier

Wagner 1966; Rutter 1966; Shaw 1972; Harris and Boydell 1972; Harris and Howell 1977; Roed 1975; Luckman and Osborn 1979). A variety of interpretations have been made including differing numbers of glacial advances within similar time periods. This implies that either glacial events cannot be correlated between adjacent valleys or that the evidence is equivocal (Shaw 1972; Luckman and Osborn 1979). Most attempts at evaluating chronology are based primarily on stratigraphic evidence. Although it is assumed that over long periods of time glacial responses will be synchronous, over short periods advances have been reported to produce contrasting responses in adjacent valleys. Therefore, Shaw (1972) and Rutter (1972) state that for any one glaciation, advances may appear as oscillations superimposed on larger period oscillations related to glacier stades and interstades.

Morphological evidence suggests that ice stagnation was an important feature in the process of retreat in Mountain glaciers (Shaw 1972). Wagner (1966) states that glacier erratics of Mountain provenance found in the Foothills of southwestern Alberta may be used to establish maximum cordilleran ice thickness during the Wisconsin at about 600 metres.

Attempts to correlate Mountain glacier episodes with Continental sequences have led to chronological deductions based on inconclusive evidence (Wagner 1966; Richmond 1965; Stalker and McPherson 1969; Shaw 1972; Röed 1975).

Holocene glacial events have been neglected until recently (Luckman and Osborn 1979; Bray 1964). Luckman and Osborn (1979) and Rutter (1976, 1977) state that Holocene glacier advances in the Banff-Yoho area of the Canadian Rockies have been of limited extent. Limiting ^{14}C dates within 1 km of contemporary glaciers indicate that the late Wisconsin Ice Sheet and valley glaciers disappeared prior to 9660 yrs. B. P. Luckman and Osborn (1979) identify two subsequent glacial advances: an early Crowfoot advance prior to 6600 yrs. B. P. which is identified as early Holocene or perhaps late Wisconsin in age, and a late Neoglacial Cavell advance consisting of several glacial advances of approximately equal extent. Because of the variation in number, age, and extent of Cavell advance moraines this advance is considered to be a model for the whole suite of Holocene glacial advances in the area. Luckman and Osborn (1979) conclude that Holocene glacial events are much more complex that previously thought, and that overriding of earlier glacial events by the Cavell advance complicates the chronological record.

1.10 Structural Geology

Rocks of the study area form part of a single thrust plate over the Simpson Pass Thrust. They are described structurally and stratigraphically as belonging to the Eastern Main Ranges (Cook 1975).

The structural evolution during the Lower Paleozoic in the southern Rockies probably resulted from several superimposed periods of tectonic activity, some of which may have occurred in the Lower Paleozoic and are possibly contemporaneous with sedimentation (Harrison 1969; Cook 1975). The previously held idea (Harrison 1969 and others) that a series of separate thrust plates containing internal imbrications and folds had developed in the Eastern Main Ranges has since been refuted (Cook 1975).

Harrison (1969) and Cook (1975) state the regional structure of the study area is dominated by broad, open concentric folds and a series of north and northwest trending normal faults downthrown to the west. These merge with the complex belt of thrusts and folds that occur along the zone of facies transition.

Thrust faults and folds occurring in the eastern Main Ranges have been identified by Aitken (1966, 1971, 1978), Harrison (1969), Balkwill (1972), Cook (1975) and others. The Stephen-Cathedral fault is in close proximity to the study area and marks the contact between the Pika and Sullivan Formations. It can be traced along the western margin of the upper Yoho Valley and is characterized by extremely steep, high cliffs.

1.11 Regional Geology

North and Henderson (1954) divide the southern Canadian Rocky Mountains into four physiographic-stratigraphic-structural sub-provinces: the Foothills, Front Ranges, Main Ranges and the Western Ranges. The area of concern in this study is the eastern sector of the Main Ranges consisting of Lower Paleozoic (Middle Cambrian) rocks which are predominantly carbonate.

The stratigraphic succession described by Aitken (1966, 1978) consists of sediments which formed on a stable, slowly subsiding shelf adjacent to the landmass of the Canadian Shield. Two factors dominated deposition on this shelf: a persistent supply of clastic sediments (crystalline rocks) from the Canadien Shield, and a transgressing continental shoreline on the Shield with temporary retreats. Throughout the Cambrian, the eastern edge of the shale basin fluctuated. Its position is marked by the eastern edges of the principal Lower Paleozoic carbonate formations. Palmer (1960) and Robison (1960) suggest that these factors gave rise to three discernable facies:

i) An inner detrital facies characterized by shales and silt stones with subordinate carbonate interbeds. This grades to sandstones as contact with the crystalline Precambrian basement is approached. In Lower Cambrian deposits, sandstones are the dominant lithology.

ii) A middle carbonate facies with a variety of carbonate rocks. Shales are virtually absent. Carbonates contain some clay and beds with quartz, silt and sand occur in some sequences.

iii) An outer detrital facies characterized in some locations by mudstones and thin-bedded, argillaceous and silty carbonate rocks, and in others by carbonate-shale couplets (Aitken 1978).

Aitken (1966, 1978) describes sedimentation in the uppermost Lower Cambrian, Middle and Upper Cambrian and the Lower Ordovician as being cyclic: Grand cycles spanning from one to three trilobite assemblage zones have been identified with the appearance of an inner detrital facies (Lower shale) passing gradually upwards into a carbonate facies (Upper carbonate) (Aitken 1978; Fritz 1975) (Table 4). Cyclicity within outer detrital deposits has not been observed (Aitken 1978) although Palmer (1971) has suggested restricted grand cycle sequences in some middle carbonate with outer detrital deposits. The lithologic sequences or sub-cycles found within the grand cycle scheme are interpreted as recording increases in water depth and increased supply of terrigenous sediment, followed by a gradual decrease in both water depth and sediment supply. These variations may indicate tidal events or changes in tidal range.

Aitken (1966), and others have proposed a tilting craton theory to explain the linked behaviour of source and depositional areas, recorded by cycles (grand cycles) involving inner detrital and middle carbonate deposits. It is suggested that the Lower Paleozoic transgression took place during tilting of the craton, resulting in uplift of the source area and subsidence of the depositional area. This occurred in combination with a continuous, slow subsidence of the axis of tilting. Tilting occurred through movements of short period, reflected in sedimentary rhythm which

Age	Formation	Max. Thickness (m)	Lithology	Facies	Grand Cycles	Fossil Zones (Aitken 1978)
Ordovician (Lower)	Survey Peak (Aitken and Norford, 1967, Cook 1975)	240	Slate, some limestone	Middle Carbonate, Inner Detrital		
Upper Cambrian	Mistaya (Aitken and Greggs 1967, Cook 1975)	152	Limestone, minor amounts of dolomite (interbeds)	Middle Carbonate	?	Saukia
	Bison Creek (Aitken and Greggs 1967, Cook 1975)	213	Shale, some limestone and calcarenite interbeds	Outer Detrital		Ptychaspis Prosaukia; Conaspis
	Lyell (Walcott 1908, Deiss 1939, Aitken 1966, Cook 1975)	370	Dolomite and limestone with chert zones	Middle Carbonate		Elvinia; Aphelaspis
	Sullivan (Walcott 1920, Aitkens and Greggs 1967, Cook 1975)	431	Shale; interbedded limestone and silt, massive limestone and dolomite.	Inner Detrital		Crepicephalus
Middle Cambrian	Waterfowl (Aitken and Greggs 1967, Cook 1975)	165	Dolomite, some shale	Middle Carbonate		Cedaria
	Arctomys (Walcott 1920, Aitkens and Greggs 1967, Cook 1975)	228	Shale; siltstone, some dolomite	Inner Detrital		?
	Pika (Deiss 1939, Cook 1975)	273	Limestone, dolomite	Middle Carbonate, Inner Detrital		Bolaspidella
	Eldon (Walcott 1908, Deiss 1939, Aitken 1966, Cook 1975)	333	Limestone, dolomite	Middle Carbonate		Bathyuriscus Elrathina
	Stephen (Walcott 1908, Rosetti 1951, Aitken 1966, Cook 1975)	137	Limestone, shale	Inner Detrital		Glossopleura
	Cathedral (Walcott 1908, Cook 1975)	364	Limestone, dolomite	Middle Carbonate		Albertella
	Mount Whyte (Walcott 1908, Deiss 1939, Rasetti 1951, Cook 1975)	137	Limestone, siltstone, shale	Inner Detrital		Plagiura Poliella; Bonnia-olenellus
Lower Cambrian	Gog Group (Deiss 1940, Okulitch 1956, Mountjoy 1962, Cook 1975)	2133	Quartzite, minor shale, sandstone and some limestone at top of unit.		?	?

Aitken (1978)

Table 4 Table of Formations

are superimposed on motions of a similar nature but extended period, typified in the sedimentary cycle.

1.12 Bedrock Stratigraphy

Bedrock in the field area belongs to the Middle Cambrian Eldon and Pika Formations. The Eldon Formation is characterized by distinctive thick-bedded dolomite and argillaceous limestone (Cook 1975). The Formation occurs throughout the area as a massive cliff-forming carbonate unit in which large, irregular, sharply bounded dolomitized zones do not follow specific stratigraphic horizons. The contact between the Eldon Formation and the conformably overlying Pika Formation is marked by a change from thick-bedded dolomite and limestone to the thin-bedded limestones with shaly partings. The Pika Formation is composed of thin-bedded to flaggy limestones with argillaceous partings. Beds of hard, dense dolomite often form a resistant rib at the top of this formation. Beneath this rib, the formation generally weathers as a recessive unit relative to the underlying Eldon Formation, so that the lower contact of these two formations is often marked by a distinct break in slope.

Figure 3 shows bedrock formations in and around the immediate field area.

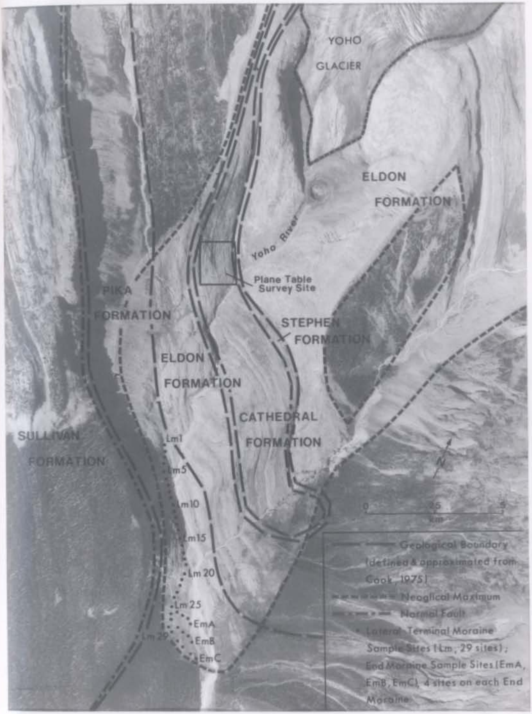


Figure 3 BEDROCK FORMATION in the VICINITY of YOHO GLACIER

Chapter 2

Previous Research

2.1 Bedrock Erosion

Bedrock erosion by glaciers may take the form of abrasion, plucking or solution by subglacial water. Boulton (1979) states that glacially eroded bedrock surfaces reflect at least two important processes: an abrading process which smooths surfaces and produces the characteristic streamlined bedforms; and a plucking process which produces locally roughened areas of bed (Matthes, 1930).

The fundamental requirements for abrasion are a supply of basal debris; sliding of basal ice; and the effective migration of debris down towards the bedrock surface due to net loss of surrounding ice through basal melting. Factors affecting the rate and type of abrasion include ice thickness, basal water pressure, relative hardness of rock particles and bedrock, rock particle characteristics and the efficiency of rock flour removal. Glacier abrasion has been observed directly in the field by Boulton (1974), experimentally by Boulton and Vivian (1973), Hope, et al. (1972), Brepson (1976), and Mathews (1979), and estimated from sediment discharge by Thorarinsson (1939), and from excavated rock volumes by Andrews (1972), and Kaszycki and Shilts (1979).

Various degrees and scales of plucking are said to take place under a variety of basal ice conditions (Boulton 1979; Metcalf 1979; Glen 1979; and others). Metcalf (1979), and Glen (1979) point out that at a micro-scale, individual grains may be either pulled off by glacier ice or pulled off by basal debris. Based on previous literature (Boulton 1972, and others), four mechanisms of plucking have been proposed. These are listed by Kemmis (1979) as follows: i) Plucking during glacier sliding related to enhanced basal creep where the ice is at the pressure melting point; ice envelops boulders, rocks or pebbles on the glacier bed and penetrates fissures. ii) Plucking by regelation where ice sliding over a fine grained bed (carbonate) may enable basal melt water to carry in suspension individual clay to sand size particles. These may then be frozen into the glacier bed during regelation. iii) Plucking is also related to a sequence of thermal regime zones. Basal melt water is produced in an up-glacier zone of ice at the pressure melting point where it flows outward under a hydrostatic head to an outer zone where the ice is below the pressure melting point. Plucking occurs as available basal debris is frozen in with the basal melt water in the cold outer zone. iv) Plucking where a sequence of basal thermal regime zones allows melt water flow in the glacier bed materials but not between the bed and basal ice. The melt water flow weakens the bed by increasing pore pressures in localized bed areas (Moran et al. 1980).

Clearly, different plucking mechanisms may depend on the thermal regime of the glacier. The recognition that several plucking mechanisms can occur under a glacier simultaneously may be important if the relative importance of abrasion and plucking is to be determined.

It has been demonstrated (Galibert 1962; Boulton 1978, 1979; and others) that the process of plucking may be enhanced by pre-existing joints in the rocks. Several processes have been proposed whereby entirely joint-bounded blocks may be pulled from the bed and incorporated into the glacier, or where partially jointed blocks may be further fractured and incorporated (Boulton 1979). Carol (1947) states that regelation melt water in lee-side positions may lead to the shattering of bedrock due to ice growth in the interstices. Lewis (1940) proposed that glacial unloading may cause dilation joints in bedrock thus weakening the rock. Boulton (1974) discovered that internal stresses induced in a sub-glacial hummock by glacier flow may lead to fracturing in lee-side positions. Boulton (1978) also observed that boulders in traction in basal ice produce fracturing on lee-side crests of hummocks where they are expelled from the ice under reduced basal pressure.

The relative importance of abrasion and plucking as erosional agents is difficult to assess. Plucking, though volumetrically important, is a highly localized process concentrated on the distal sides of bedrock hummocks and more profound on jointed rocks (as those found in the Yoho Valley study area). The streamlined forms of glaciated bedrock surfaces are largely a product of more widespread abrasional processes (Boulton 1979). Evident in the study area are the highly streamlined glacially eroded bedforms indicative of the abrasion process (Figure 4 a) and large quantities of block boulders down glacier (Figure 4 b) which by their provenance constitute evidence of substantial amounts of plucking (Figure 4 c).

2.2 Glacier Sliding at the Ice-Rock Interface

Glacier sliding at the ice-rock interface has been the topic of much recent glaciological literature (Andrews 1972; Kamb 1970; Nye 1969;

Figure 4a

Streamlined glacially eroded bedform indicative of the abrasion process. Note striated surface.

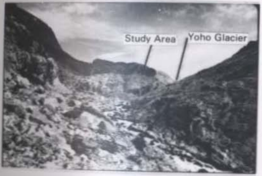


Figure 4b

Block boulders approximately 1 km down-valley from the bedrock units under study. Their provenance constitutes evidence of large-scale plucking.

Figure 4c

Plucked surfaces on lee sides of bedrock obstructions.

1970; Hallet 1976; Boulton 1978, 1979; Fowler 1979; Robin 1976; Morris 1979; Broster, et al. 1979; Goldthwait 1979; Budd, et al. 1979; Hambrey and Muller 1978; Lliboutry 1977, 1978; Weertman 1976; etc.). Andrews (1972) states that the erosive power of a glacier can be estimated from a determination of the power expended in sliding over its bed. He defines the total power (W_T) as equalling $T g\overline{U}$ where $\overline{U}$ is the average glacier speed and Tg is the shear stress at the glacier bed. The effective erosive power is therefore some fraction of W_T, which depends on the relationship between glacier sliding velocity and internal flow velocity. It might be construed from this that glacial erosion is effected by glacier ice alone. Boulton (1974) shows that although plucking may be caused by sliding ice alone, it is debris in basal ice that actually produces an abraded surface. The sliding/abrasion process is typical of many warm-base glaciers. Where cold-base glaciers are concerned, shearing in the glacier ice or beneath the bedrock/ice interface is a major process (Moran, et al. 1980) and erosion due to sliding is retarded.

Most theoretical concepts concerning glacier sliding were originated by Weertman (1957) and later modified and re-interpreted by Lliboutry (1965, 1968, 1976, 1979) as well as by Weertman (1962, 1968, 1979, 1972, 1974). Further contributions have been made by Boulton (1975), Robin (1976), Nye (1969), Kamb (1970), and Morland (1976).

Weertman (1957) suggests two mechanisms which allow the ice of a temperate or warm-base glacier to move over its bed. The first of these mechanisms, plastic flow, enables ice to flow over and around obstacles. The other, pressure melt results when ice melts under pressure on the upstream surface of the obstacle and refreezes on the downstream side where pressure is less. Weertman's (1957) treatment of glacier sliding assumes obstacles are all the same size and the remaining bed is perfectly smooth. Under a given shear stress, the plastic flow mechanism results in ice movement increasing with larger obstacle size, whereas in the pressure melt mechanism, ice movement decreases with increasing obstacle size. Weertman argues that the pressure melt mechanism enables ice to flow around smaller obstacles while the plastic flow mechanism enables ice to flow around larger obstacles. Hence, sliding speed is determined by obstacles of a size at which the sliding speeds due to the two mechanisms are equal: the controlling obstacle size.

Weertman (1964) refines his approach by taking account of the resistance offered by obstacles smaller and larger than the controlling obstacle size. This is accomplished by simplifying the distribution of obstacle sizes. Instead of looking at a continuous spectrum (Weertman 1957), three discrete sizes are examined. Weertman (1964) considers a bed to be a superimposition of a series of beds all having similar roughness parameters, but at a variety of scales. Thus he concludes that an appreciable part of the resistance to sliding comes from obstacles other than the controlling obstacle size.

Lliboutry (1968) models a sliding bed in respect to sine waves. The amplitude divided by the wavelength of a single sine wave is the roughness parameter. Lliboutry's (1976, 1979) subsequent treatment calls for the superimposition of several sine waves of the same roughness but differing wavelengths. As a result the wavelength and amplitude vary in geometric progression.

Both Weertman's and Lliboutry's theories rely on intuitively appropriate simplification, evident by their reference to discrete obstacle sizes applied to a bed which most likely contains a continuous spectrum of obstacle sizes.

2.3 Newtonian Viscous Flow

Early theoretical works on glacier sliding (Weertman 1957, 1962; Lliboutry 1958, 1965; Nye 1969) considered glacier ice as a Newtonian viscous material. This treatment of ice flow would be appropriate in the absence of regelation and presence of basal ice debris (Nye 1967, 1969; Vivian and Bocquet 1973; Boulton, et al. 1979; Morris 1979). The melting and refreezing of ice onto basal ice and the glacier bed; cavitation induced by bedrock obstructions; and clasts held in ice, all have an effect on altering the lower boundary conditions at the ice-rock interface, and thus flow in the ice.

Creep in ice is accomplished by slippage along crystal planes (Nye 1965, 1969, 1970; Glen, et al. 1957; Lliboutry 1965; and others). Individual grains have enough freedom to rotate and line up parallel with one another. Creep rate becomes noticeable when a critical stress is reached (in ice; $1 Kgf/cm^2$). Experimentally, creep rate is proportional to the cube of the stress (Glen's Law).

Barnes et al. (1971) studied the frictional and creep properties of polycrystalline ice. They concluded that the behaviour of sliding ice on bedrock under conditions of strong interfacial adhesion fall into three regimes. At very low sliding velocities (approximately $10^{-6} m.sec^{-1}$), recrystallization is produced in a thin zone of ice close to the interface with the basal planes being preferentially oriented in the direction of sliding. This situation was observed to favour easy creep in the sliding direction. At somewhat higher sliding velocities (approximately $10^{-4} m.sec^{-1}$), shear failure and brittle fracture limit the shear strength of the ice and lead to a coefficient of friction that does not vary significantly with sliding velocities. At still higher velocities (approximately $10^{-1} m.$

sec^{-1}), heating at the interface is sufficient to produce some melting and the friction coefficient drops drastically.

2.4 Ice-Bound Debris

Studies of basal ice sliding have generally dealt with clean ice (Weertman 1957; Lliboutry 1965, 1968; Nye 1969, 1970; Kamb 1970; etc.). Morris (1979) maintains that an understanding of the motion of a clast in the basal ice of a glacier lies at the heart of physically-based models of erosion and deposition.

Boulton et al. (1979) report that debris found in glacier ice near the bed may occur in concentrations of up to 30 to 40% by volume. Besides being an important erosive agent, debris is comminuted and retarded relative to the moving glacier, presumably because of drag at the debris-ice and debris-bed interface, rock-on-rock friction values being much higher than those for ice-on-rock. Where the frictional drag between debris and bed is large, and concentrations high, the presence of basal debris must be an important determinant of glacier sliding velocity and subsequently of ice dynamics. Ice must not only slide over bedrock but also around the mobile obstructions found in basal ice and at the rock-ice contact. Glen et al. (1977) present analytical solutions for the motion of isolated clasts in ice unbounded by other solid surfaces. Lewis (1960) and Boulton (1975) have examined the motion of clasts in basal ice and have described their effects on erosion and deposition. Lewis (1960) describes the maximum force that could be exerted by a rock held in ice and Boulton (1975) adapts Weertman's (1957) sliding theory to describe a condition for the onset of abrasion. Röthlisberger (1968) describes the processes which tend to bring clasts into contact with the bed.

Weertman (1957), Lliboutry (1968), Nye (1969, 1970) and Kamb (1970) assume when considering basal sliding theories with clean ice, a lubricating layer of water exists between ice and bedrock. Thus shear stress at the ice-rock interface is negligible. Morris (1979) suggests that when the bed undulates about a given base plane, the ice exerts a net force on the rock which can be interpreted as the result of an average shear stress acting over the base plane.

Boulton (1975) defines the conditions for the lodgement of a stone in basal ice by estimating the horizontal force exerted on the stone by the ice and equating this to the retarding friction or drag between the stone and the bed. If clasts are close to the bed, the existence of irregularities on the stone would allow normal and frictional forces to be transmitted without disturbing the pattern of flow of ice or altering the physical processes at the rock-ice contact.

2.5 Regelation and Precipitates

Morris (1979) observes that in the steady state, the classical regelation boundary condition as described by Weertman (1957, 1970, etc.) and Lliboutry (1977, 1978), cannot be obeyed both at the glacier bed and at the surface of clasts in the overlying ice. Temperature and stress distributions cannot ensure that both surfaces are at the pressure melting point (Nye 1967). Morris (1976, 1979) has argued that melting and refreezing within ice would produce an internal temperature distribution to be added to the regelation temperature distribution at the bed. This method of analysis depends on the assumption that any internal component of temperature is negligible compared to the regelation component at the ice-rock boundary. Morris (1979) shows that this is not always the case as melting and refreezing within the ice produces an internal temperature distribution such that the ice at the two boundaries and at the internal water inclusion (possibly in a cavity in the ice of clasts held in suspension in ice), is at the pressure melting point.

Vivian and Bocquet (1973) and Boulton, et al. (1979) observe that cavities frequently occur in the lee of large clasts. Some of the cavities are filled with water, while others develop spicules of ice. Clearly, the size and shape of these cavities will depend on the flow patterns of the ice andare observed to vary on a seasonal basis (Vivian and Bocquet 1973). Morris (1979) states that if clasts move towards a part of the bed with a different flow pattern, bulk melting or refreezing must take place as the cavities adjust to the new pressure and stress situations.

Hallet (1976) states that thin superficial chemical precipitates on bedrock are characteristic of many retreating temperate glaciers. Bauer (1961), Kers (1964), Ford, et al. (1970), and Hallet (1975, 1976) identify calcite as being the most common and best developed deposit generally formed on limestone and other carbonate-rich rocks. The calcite deposits are characteristically found in the lee of bedrock protuberances (Figure 5 a) where they are oriented parallel to the direction of the former ice flow or regelation water migration (Figure 5 b) (Hallet, 1975). The author has also identified calcite precipitates on the bottom surface of an overhanging bedrock form (President Glacier, Yoho National Park). This suggests a pressure release of the calcite deposits onto a surface rather than, as the term precipitate may imply, a gradual gravitational fall-out of the deposits.

Hallet (1975, 1976) suggests that the genesis of such forms appears to be closely related to the regelation-slip process (Weertman 1957; Lliboutry 1979), which Lliboutry (1965) and Hallet (1976) identify

Figure 5a

Calcite deposits in lee of bedrock protuberances.

Figure 5b

Calcite deposits (spicules) oriented in the direction of former ice flow or regelation water migration.

as being characteristic at the bed of temperate glaciers.

The role of adhesion during the sliding of the ice over bedrock has been studied by Roraty and Tabor (1958), Barnes, et al. (1971) and more recently by Robin (1976). Lliboutry (1968) has drawn attention to the lack of adequate theoretical treatment of variations in the flow properties of ice in the presence of melt water.

The assumption that basal ice is separated from bedrock by a continuous thin film of water plays a major role in theories of glacial sliding (Weertman 1957, 1964, 1972, for example). The implication of this assumption is that the pressure at the water film boundary is normal and no shearing takes place as a result of adhesion between ice and rock. However, it has been shown by Budd (1976) and Robin (1976) that adhesion is present when glaciers are sliding at relatively low velocities (presumably due to a decrease in friction). Cold patches may also occur where adhesion forms between ice and bedrock.

Robin (1976) has drawn attention to the role of impurities in basal ice which may be a major cause of cold patches. Souchez, et al. (1973) have described the complex changing physical and chemical properties of basal ice as it moves through variations in temperature and pressure conditions around bedrock irregularities. These phenomena may have important effects on the stick-slip motion often reported in temperate glaciers (Boulton 1976, 1979; Hallet 1979; and others).

2.6 Glacier-Streamlined Bedrock Surfaces

The investigation of glacier-streamlined bedrock surfaces has been pursued by both geomorphologists and glaciologists. The characteristics of roches moutonnées, riegeln, rock drumlins, flutes, grooves, striations and p-forms are well described in geomorphic and glaciological literature

(Ljunger 1924, 1925; Hjulstrom 1935; Johnsson 1956; Dahl 1965; Gjessing 1966; Embleton and King 1968; Boulton 1974; Sugden and John 1976; and others). Recent studies by Boulton (1979) and Weertman (1979) have emphasized their development with reference to glacier sliding. In search for a glacier sliding law, Weertman (1979) and Lliboutry (1979) have described the effects of these hummocks under glaciers.

Benoist and Lliboutry (1978) and Benoist (1979) have provided a spectral analysis of micro-relief from roches moutonnées and developed a shadowing function to describe the effects of bumps on lee-side cavitation beneath a glacier. The fine scale of roughness on synthetic slabs has been related to experimentally-induced ice sliding by Budd, et al. (1979). Benoist and Lliboutry (1978) suggest that the relief spectrum of sub-glacial bedrock hummocks is a response to the sliding law of the ice and the erodibility of the substrata. Lliboutry (1979) suggests that detailed maps of these forms on a micro-scale (of the order of centimetres in vertical height) could provide valuable information on the theory of ice sliding.

Chapter 3

Strategy and Approach

The strategy and approach adopted for this study was basically dictated by field conditions (time and accessibility) and laboratory facilities.

In order to gain an understanding of sub-glacial erosional processes from the examination of bedrock strength, surface roughness and the contents of related tills, certain field and laboratory procedures were designed and adopted (these are outlined in the discussion to follow). Certain of the glaciological and glacial-geological practices, can be viewed as standard (in that they are accepted practice). Others have been adopted from the engineering and computer sciences disciplines to help demonstrate some physical and chemical characteristics of lithology, tills and surfaces that have not been subject to detailed examination in the past.

This study does not claim that the procedures and techniques used here are the only or ideal methods. They have, however, been successful in providing data and analyses of these data which are of some value.

3.1 Field Methods: Sampling Procedures

Daugherty (1974) writes: "The essence of sampling lies in the fact that a large number of items, individuals or locations may, within specified limits of statistical probability, be represented by a smaller group of items (samples) selected from the larger group...If we carry out sampling correctly, a limited number of samples will be sufficient for making generalizations... sampling represents a more efficient use of our energy while still allowing us to make reliable statements about the whole population... The key to success in sampling lies in adopting a procedure which permits us to draw satisfactory conclusions about a parent population from a sample of minimal size."

Daugherty (1974) identifies three basic considerations before adopting a sampling procedure and before conclusions can be drawn from the samples collected:

i) The exact size and extent of what is being sampled must be determined.

ii) The most appropriate sampling procedure must be adopted, taking into account time constraints and field conditions.

iii) The minimum size of sample both in number and volume must be determined so that reliable representation of the feature sampled can be attained.

In attempting to solve most geochemical field problems (Klassen and Shilts 1977), and geological engineering and mechanical problems (Milner 1976; and others), geologists have conventionally tended to collect as many samples as time and laboratory facilities permit and to spot localities as evenly as possible over the area of study. A sampling design of this nature implies rather restrictive assumptions about the variability of element composition of the rock and/or sediment unit. First of all, it assumes both that the material sampled is fairly uniform on a local scale, as exempified by the taking of only one sample per locality, and that a more important variation in composition is exhibited on a regional scale as specimens from many parts of the parent unit are needed. Any rational decision as to the number of specimens or the distribution of sampling localities can only be based on prior knowledge of the natural variability in composition in the parent units, and on intuition. Visual assessment and interpretation of unit geochemical and physical features are stressed as being essential before sampling can be undertaken. Thus sampling in this thesis is stratified or systematic rather than random.

3.11 Scale

A consideration of the scale of approach one is taking is usually beneficial to make as it puts into proper perspective the magnitude of size of form one is studying. The generally accepted classification of large, medium, small or mega, meso, micro (Lliboutry 1979; and others) usually refers to forms such as glaciated valleys or cirques; roches moutonnées, whalebacks, etc.; and striations, precipitates, small relief forms, etc.; respectively. Of concern to this study are the meso and micro scale.

3.12 Lithologic Samples

Glacial geological studies require bedrock samples for refined lithological detail by microscopic examination and laboratory tests (chemical and physical). Krumbein and Sloss (1963) state that normally, samples should not be large (depending on laboratory analysis) as their evacuation from the field area may become a burden. However, each sample should fully represent, in relative volumes collected, the lithologic variations in each bedrock unit (Lamar and Thomson 1956).

In small scale studies, samples are generally collected within 1.5 to 3 metre intervals, rather than within natural subunits, since the regular intervals provide samples comparable to those derived from drilling. This sampling technique is useful in stratigraphic studies involving the comparison of surface and sub-surface data (Krumbein and Sloss 1963). It has been noted that systematic sampling procedures are suitable for the purposes of preparing a more complete description of stratigraphic sections through microscopic and laboratory analysis (Krumbein and Sloss 1963; Schenck and Adams 1943; Milner 1962; Lahee 1961; and others).

Because of the observed homogeneity exhibited in each bedrock unit, and because only loose surface boulders were allowed to be taken from the study area (Parks Canada, Collection Permit), the sampling of lithologic units consisted of a stratified sampling of boulders that were suitable in size and which resembled the physical and chemical characteristics of the parent bedrock units. Where possible, samples were taken from the parent units using systematic intervals.

3.13 Boulder Count

Approximately 1.5 km down valley from the present glacier terminus and .5 km to 1 km down valley from the bedrock bench, is a field of glacially transported boulders whose provenance is the study area (Figure 3b).

A boulder count on 500 bedrock blocks measuring approximately $0.25m^3$ was conducted to determine dominant rock type and the effects of large scale plucking on the bedrock units. Boulders supplied from the adjoining cliff face (Sullivan Formation) were few and were distinguished on the basis of their characteristic black colour, smaller size and shattered platey appearance.

3.14 Till Sampling

The sampling of till along moraines in the Yoho Valley generally followed the line sampling and systematic sampling prodecures described by Daugherty (1974), and others. Till samples were collected at a predetermined interval of 20 metres along the moraine. All samples were obtained from the proximal side of the moraine and were considered to be surface-subsurface samples (taken at a depth of approximately 10 cm) (Phillips 1955). This depth was chosen in order to reduce the probability of cliff-side debris being incorporated into the samples. Generally, from field observation, cliff-side debris on the surface of the lateral moraine was discontinuous and was less than 10 cm thick.

The problem of choosing a representative size of sample (volume) is always difficult. Ideally, the larger the sample size, the more likely it is to be representative of the till. In all probability, the only correct way to sample till without the sample being biased is to extract it by 'bull-dozing'. In this manner all size fractions at a variety of depths can be accounted for. This method however proves to be extremely expensive and destructive when a large number of samples is required. Suitability and convenience, therefore, dictate the choice of sample size and number.

Twenty-nine sub-surface till samples (approximately 400 to 600 grams each) were collected from the Neoglacial latero-terminal moraine (western side of the Yoho Valley) at a regular interval of 20 m. Twelve till samples were collected from the proximal side of 3 recessional moraines near the Neoglacial maximum moraine (four samples from each moraine).

3.2 Survey Methods: Bedrock Surface Morphology

3.21 Surface Area Determination: Plane Table Survey

The area of the seven bedrock units was surveyed with a plane-table and a self-reducing alidade. The purpose of the survey was to determine an approximate proportional surface area of contact between individual bedrock units and glacier ice, and to examine the three dimensional nature of the bedrock units at the medium scale. The same bedrock unit surface areas on the eastern side of the Yoho River were estimated from aerial photos.

3.22 Bedrock Unit Site Selection: Micro-Relief Sample Site

The concept of micro-relief is not new. The term was apparently coined by Le Conte (1877) to describe prairie mounds in California and Oregon, and was used to describe the configuration of a surface in which coalescent low mounds were 6 to 10 inches higher than the adjoining depressions. Subsequent use of the term has varied in application and scale (Dwornik et al., 1959; Strahler and Koons 1959; Van Lopik and Kolb 1958; Mabbutt 1963; and others).

In this study, the term micro-relief refers to a glacially eroded bedrock surface with the vertical relief generally not exceeding 30 cm and is represented by a series of $1m^2$ plots on each bedrock type. In its broader context, micro-relief refers to all bedrock surfaces eroded by glacial ice with vertical relief not exceeding 30 cm or those relief forms which generally do not fall into the elongated erosional or asymmetrical rock form category (i.e. whalebacks, roches moutonnées, hummocky bedrock, etc.) (Laverdière, et al., 1979).

3.23 Micro-Relief Survey

A three dimensional micro-relief contouring survey was performed on seven $1m^2$ bedrock plots, one on each bedrock type. The purpose of this survey was to gather relief information which could be displayed and analysed with the aid of computer mapping techniques. It is suggested that bedrock characteristics at the micro-scale can be assessed as indicators of bedrock erosion and can be utilized in identifying and describing erosive processes.

Each plot was on the crest of a whaleback form, characteristic of each bedrock unit (Figure 6). The survey was by centimetric ruler held perpendicular to a metric scale which was moved over a fully levelled wooden frame at 2 cm intervals. A total of 2601 relief readings was obtained from each micro-relief plot. Note was made of fractures, bedding planes and surface precipitates. Striations were mapped and measured.

3.3 Laboratory Analyses: Lithology

3.31 Bedrock Identification

A sample from each bedrock unit was thin-sectioned and examined microscopically for constituent minerals to aid in identification. The preparation and analysis of thin sections followed the standard procedure set forth by Dickson (1966).

3.32 X-Ray Diffrection Techniques

Analysis of the finer than 4Ø (0.0625 mm) fraction of artificially crushed bedrock unit samples using X-ray diffraction techniques was facilitated by a Phillips Diffractometer and followed the procedures outlined by Grim (1962), Carroll (1970), and Brown (1972).

The purpose of this analysis was to supplement the thin section identification of constituent minerals in each bedrock unit.

3.33 Bedrock Strength Characteristics: Total Carbonate Content

Since bedrock strength involves resistance to solution, abrasion and crushing, tests were designed to describe the solubility and mechanical strength of the bedrock units.

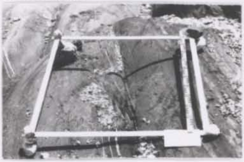
A comparison of the relative solubilities of the bedrock units was

Bedrock Unit A

Bedrock Unit B

Bedrock Unit C

Bedrock Unit D



Bedrock Unit E

Bedrock Unit F

Bedrock Unit G

Figure **6**

Bedrock Unit Plot Sites

obtained by measuring the percent (%) carbonate by weight in each unit. This was accomplished by powdering the sample to pass it through a 40 mesh sieve and adding 50 ml .4 normal H_2SO_4 and the required .45 normal N_aOH to one gram of the powdered sample. The weight loss after reaction (which takes place in a matter of minutes) indicates the % of carbonate in 1 gram of sample reacting with the acid.

3.34 Carbonic Acid Solubility

Since natural limestone solution is accomplished by H_2CO_3, a solubility test employing a weak solution of H_2CO_3 (pH. 4.0) was undertaken. It is the opinion of the author that the results from such a test should better reflect the natural conditions of limestone solubility in a glacial environment. A procedure was designed which enabled an efficient analysis under controlled conditions (Strong, per. comm.; Yoxall, per. comm.) (Appendix A).

3.35 Apparent Porosity

The porosity of a rock is defined as the ratio of the volume of pores to the bulk volume of the rock (Lama and Vutukurí 1978). If pore volume and bulk volume can be determined, then apparent porosity can be calculated. The pore volume measured was of interconnected pores, hence the value calculated is the apparent porosity (Lama and Vutukuri 1978; Vutukuri et al., 1974).

Apparent porosity calculations were performed on cylindrical specimens (diameter 2.5 cm, length 2 cm) with bedding planes perpendicular and horizontal to the elongated cylindrical configuration of the specimens. Vutukuri et al. (1974) state that the number of specimens to be tested should vary with the type of rock. Ideally, the greater the specimen number, the more accurate the results. For carbonate rocks, 5 to 10 samples are generally sufficient for testing purposes.

Apparent porosity calculations on the seven bedrock units enabled a relative comparison of apparent void space, hence susceptibility to water and solute intake, related erosive processes and compressive strength determinations.

3.36 Bedrock Abrasion

Tablets of each bedrock unit were tested for their susceptibility to abrasion using a custom-built 'abradometer' (Figures 7 a, b, and c). This test was not intended to simulate the abrasion of rock by sliding ice, but rather to compare and contrast abrasion strength (time) of each bedrock unit sample under similar, controlled conditions.

The abradometer consists of a 50 W, 100 rpm. geared-down motor which drives a replaceable 63 mm diameter, 6 mm wide, medium-fine grit grindstone. Each rock tablet was squared, levelled, and clamped in the vise, and oriented in the direction of the striations, with respect to the grindstone rotation. The tablets were held at an arbitrarily fixed constant pressure of 1.34×10^5 Pa against the grindstone. (The pressure was determined by the combined weights of the heaviest tablet and the moveable vise assembly. Weights were added onto the tray assembly of lighter tablets to equal the constant pressure.) The instrument stopped automatically when a volume of rock equivalent to 3.6 cm^3 was abraded. The time required for this process to complete itself was recorded on an electric trip-stop clock. Each bedrock unit was tested four times. Similar runs were conducted on a granite and talc tablet to determine an approximate upper and lower time boundary.

Figure 7a
Custom Built Abradometer

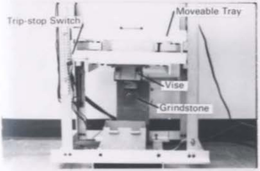


Figure 7b
Moveable tray with vise assembly and automatic trip-stop switch.

Figure 7c
Rock tablet in position to be tested. Each tablet was positioned in respect to the trip-stop switch mechanism so that exactly 3.6 cm^3 of rock was abraded

Smorodinov (1966) conducted laboratory experiments to determine an abrasiveness index on limestones which was dependent on the time of grinding. The pressures used in his experiments were also arbitrarily arrived at and generally did not exceed 3.00×10^{5} Pa. Mathews (1979) tried to simulate glacial abrasion by turning a grindstone made of ice and crushed quartz between two stone plates (limestone and feldspar) and found that the relatively quick abrasion of limestone (as compared to feldspar) was likely due to the ease of plucking grain from grain, or to the breaking of weak inter-crystalline bonds linking calcite grains.

3.37 Uniaxial Compression

Compressive strength or crushing strength is defined as the stress required to crush a cylindrical rock sample unconfined at its sides (Farmer 1968). The stress value at fracture (collapse of internal pores) is defined as the compressive strength of the specimen (Q_c) and is given by the relationship $Q_c = F/A$; where F is the applied force at failure and A is the initial cross sectional area transverse to the direction of force (Vutukuri et al., 1974).

Uniaxial compression testson saturated bedrock samples (apparent saturation) with bedding planes parallel and perpendicular to the elongated cylindrical configuration of the specimens were conducted to determine the relative strengths of each unit under conditions of applied compressive force. This test should give some insight into the magnitude of pressure required to:

i) Crush and pluck the bedrock units in situ.

ii) Comminute the individual bedrock boulders while they are in basal ice or traction.

Preparation and testing procedures followed the standard methods described by Vutukuri, et al. (1974); Mazonti and Sowers (1965); Yamaguchi (1970); Lama and Vutukuri (1978); and the I.S.R.M. Committee on Laboratory Tests (1972).

The ratio of length to diameter of the specimens was 2.5 to 3.0. The availability of coring bits dictated the specimen size at 2.0 cm X 2.5 cm. All samples were 'apparently' saturated ('apparently' refering to the apparent porosity) which resulted in compressive strengths being less than if the samples were oven dried (Vutukuri, et al., 1974). A Versa Tester 30M uniaxial compression device was employed to crush the specimens at a constant loading rate of 1.0 MPa/sec. The number of specimens tested corresponds to the number tested for apparent porosity.

3.4 Till Characteristics

The sampling of the moraine material was undertaken to demonstrate the textural change in moraine material progressively down glacier, and to postulate which bedrocks provided the greatest abundance of sediments through a comparison of till and bedrock minerology. This was accomplished through the analysis of constituent clay minerals in the till matrix, and through visual examination and identification of clasts.

3.41 X-ray Diffraction

Analysis of the finer than 4Ø (0.0625 mm) fraction of twelve recessional end moraine till samples using X-ray diffraction techniques followed the procedures outlined above.

Because individual bedrock units may be typified by peak values for certain minerals (Pettijohn 1975), the patterns observed from the bedrock units were compared to end moraine diffraction plots to identify similarities.

From this comparison suggestions can be made as to which units are more abundant in the recessional moraine samples.

3.42 Grain Size

All samples (split to approximately 200 grams each) were wet sieved using a 4ϕ (0.0625 mm) sieve following standard preparation to remove organic material. The coarser than 4ϕ fraction was oven-dried and sieved according to standard methods (Folk, 1968; Griffiths 1967; Bowles 1974). Sedigraph analysis of the less than 4ϕ fraction was facilitated by a Sedigraph 5000 Analyser and X-ray beam (Geological Survey of Canada, Ottawa).

Cumulative particle size, frequency curves and statistical parameters (four moments measures; mean size, standard deviation, skewness and kurtosis) were derived from raw weight data by computer. The computer program employed was that of Slatt and Press (1976) for a Hewlett-Packard desk top calculator model 9821and plotter model 98 62A in which textural statistical parameters were derived by the graphic method rather than the moments method.

3.5 Analysis of Bedrock Surface Roughness

3.51 Computer Techniques

Computer programs were used to:

i) Analyze the detailed slope characteristics (segment length) of each micro-relief plot.

ii) Conduct a three-dimensional, 3rd order polynomial trend surface analysis.

3.52 Slope Frequency Distribution

Slope characteristics (segment length, rise/run; Carson and Kirby

1972) on each micro-relief plot were computed to produce a frequency distribution of slopes at 1 mm intervals from 0 mm to 100 mm (Appendix B). Slope values were calculated between surveyed relief points 2 cm apart (2550 slopes/plot). A comparative study of the seven slope frequency distributions may indicate varying erosion processes and magnitudes.

3.53 Synagraphic Mapping System (Symap): Trend Surface Analysis

A Symap computer mapping program with a trend surface analysis elective was employed to plot and analyse data obtained from micro-relief surveys.

Trend surface analysis is a mathematical technique in which surfaces of increasing complexity are fitted to point observations. The positions of a best fit plane is such that the sum of the squares of the vertical distances between the points and the plane is reduced to a minimum. In this way, residual values which indicate local variations not predicted by the general trend are plotted. Their magnitude may be taken to provide a measure of variation from an 'ideal' surface, or roughness.

This method of surface fitting is related to regression analysis except that it takes place in three dimensions rather than two. As in 2-dimensional regression analysis, the fit of the curve through a series of points can be improved by the use of a high order polynomial function which changes the straight line regression to a curve. In trend surface analysis, higher order surfaces can be fitted to the scatter of points in the same manner to obtain a best fit plane. Figure 8 illustrates the theoretical application of the relationship between a third order or cubic two dimensional curve and the three dimensional counterpart.

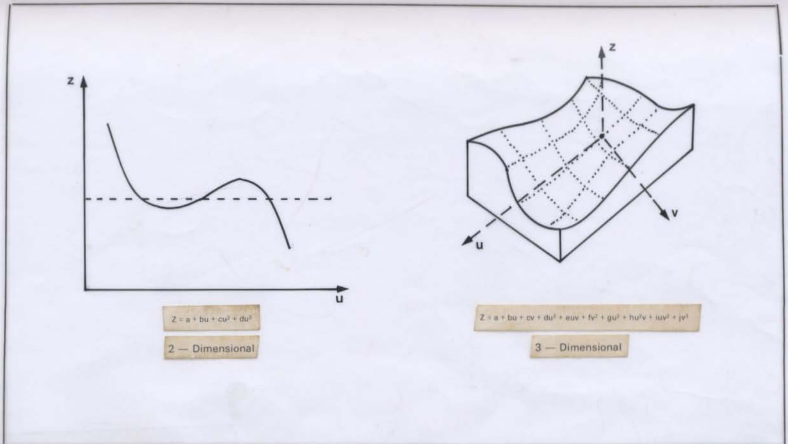


Figure 8

Relationship between a third order polynomial plane and equation in 2 and 3 dimensional view.

3.54 Three Dimensional Viewing: Symvu Program

Symvu is a computer graphics program written for the purpose of generating three-dimensional line-drawing displays of data. Symap used in conjunction with Symvu, generated accompanying statistical parameters of the third order polynomial analysis.

The Symvu program is written in Fortran IV and is operated on the 18 m 370/168 using 250K memory. A Cal-Comp plotter is used to sketch the illustrations (Schmidt 1975).

Chapter 4

Results and Interpretations

4.1 Bedrock Surface Morphology and Unit Description

Seven bedrock units were identified (Units A, B, C, D, E, F, G) and described in the field on the basis of strike-dip relations, bedding, unit thickness and drift cover.

Three levelled profiles measured parallel to striations are given in Figure 9. They show the two dimensional nature of the bedrock units at the medium or meso scale. Relative distances of bedrock units traversed by glacier ice can be seen from the profiles (areas of bedrock type in contact with glacier ice were obtained from the plane table survey, Figure 10; Table 5). The contact between the Eldon Formation (bedrock unit A) and the Pika Formation (bedrock unit B) (Figure 12) is described in detail on Figure 9, Section 1.

Section 1 from Figure 9 clearly shows bedrock unit A lying below the bedrock bench (units B, C, D, E, F and G). Angles of bedding contact are given in the profiles.

A qualitative assessment of the relative range of relief (levelness; smoothest to roughest) for the bedrock units in the Pika Formation indicates that unit D is the smoothest while unit E is the roughest. Bedrock units C, F, G, and B are ranked within this range respectively. Because bedrock unit A (Eldon Formation) lies below the bedrock bench, it is not included in this ranking.

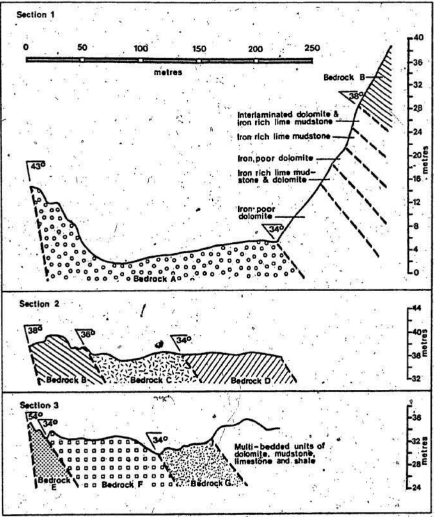
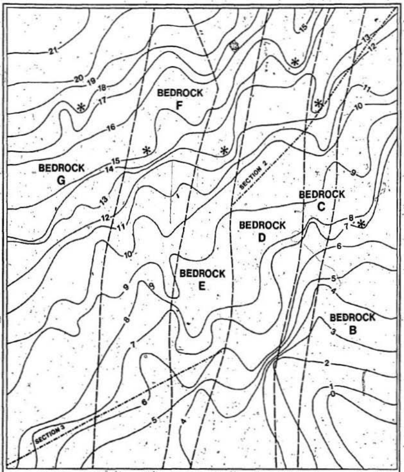


Figure 9

Levelled profiles measured along or parallel to glacial striations across the bedrock units



—·—·— Profiles measured parallel to or along the direction of former ice movement (striations) (Fig 9)

* Micro relief survey sites

0 —— 20 Meters

One meter contour interval

NOTE: THE PLANE TABLE SURVEY DID NOT INCLUDE BEDROCK A

PLANE TABLE SURVEY

Figure 10

Age	Formation	Bedrock Unit	Description	Strike/dip	Surface Area (m^2) West of Yoho River	East of Yoho River *	Comments
Middle Cambrian	Pika Formation	Bedrock Unit G	Lime mudstone: iron poor	N11°W 34°SW	36	39	Approximately 15 major beds per metre. Till cover less than 10%. Striations at 124° to 160°.
		Bedrock Unit F	Lime mudstone: partly dolomitized, spicular	N08°W 34°SW	45	52	Some fractures and cracks. Two major beds per metre. Till cover approximately 20%. Striations at 138° to 158°.
		Bedrock Unit E	Marble-lime mudstone, iron rich	N10°W 54°SW	38	46	Fractures and cracks widespread. Quartz veins dominate relief. They protrude from the unit up to 5 cm. Till cover approximately 30%. Striations at 136° to 153°.
		Bedrock Unit D	Shale: scattered relic calcite grains	N08°W 34°SW	25	32	Fractures, cracks widespread. Quartz veins throughout. Chatter marks abundant. Till cover approximately 10%. Striations at 130° to 142°.
		Bedrock Unit C	Dolomite — limestone: interbedded, iron poor, cleavage oblique to bedding	N08°W 36°SW	27	35	Isolated fractures and cracks. Quartz veins abundant. Till cover approximately 5%. Striations at 142° to 156°.
		Bedrock Unit B	Lime mudstone, iron poor, slightly fossiliferous, burrow mottled (iron rich)	N08°W 38°SW	35	42	Isolated fractures and cracks. Scattered precipitates. Till cover approximately 10%. Striations at 135° to 141°.
	Eldon Formation	Bedrock Unit A	Dolomite: iron poor, medium to coarse crystalline, hypidiotopic, scattered pyrite	N17°W 43°SW	>200	>250	Plucking widespread. Precipitates abundant. Quartz and pyrite evident. Till cover approximately 30%, boulders, gravel. Striations at 144° to 165°.
					* Estimated from photographs (aerial)		Note: All bedrock units are laterally bounded by valley walls, or as in the case of the units on the eastern side of the Yoho River, a distinct break in slope (fault).

Table 5 Bedrock Unit Descriptions

4.2 Micro-Relief Survey

The relief data obtained from each $1m^2$ bedrock plot were drafted to produce Figure11. Quartz veins, precipitates, till and striations were also identified on the plots. Bedding planes show up quite clearly on bedrock units B, F, and G where contours are densely grouped in linear fashion. Bedrock units C, D, and E exhibit elongated quartz veins (characteristic of those units as a whole). Bedrock unit plot E shows the deposition of minor till in the lee of some quartz veins (note the direction of ice flow) which were noted to protrude from the surface of the rock unit. Generally, till cover was sparse over all the bedrock plots as were precipitate deposits. A complete description of the bedrock units/plots appears in Table 5.

4.3 Bedrock Unit Description: Lithology

Seven bedrock units are examined, all but A are contiguous. Between unit A and B, a steep slope of alternating thin bedded dolomite and lime mudstone marks the stratigraphic boundary between the Eldon and Pika Formations (Figure12). A detailed description of the bedrock units is given in Table 5.

Bedrock unit descriptions and approximate surface area of contact between individual units and glacier ice are given in Table 5. Beds strike approximately north-south and dip steeply westward between 34^{o} and 54^{o}. On the western side of the Yoho River, they cross the Yoho Valley at an angle of approximately 40^{o} to the former direction of ice movement. They lie normal to ice movement on the east side of the river (Figures 13 a, b, c).

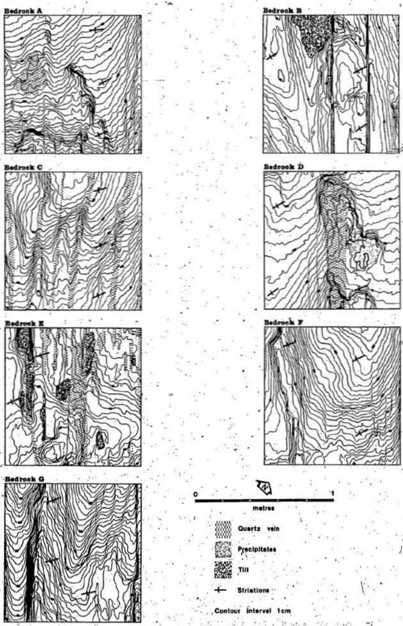
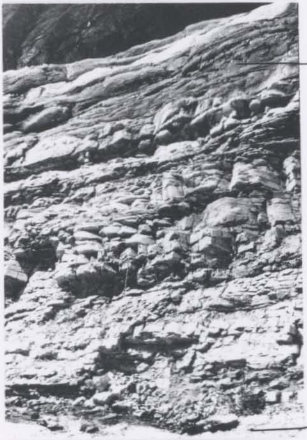


Figure 11 Contouring of bedrock unit plots



Contact Between the Eldon Formation (Bottom) and Pika Formation (Top)

Figure 12

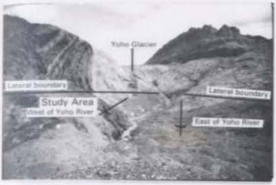


Figure 13a

Bedrock units in relation to former ice flow.

Figure 13b

Bedrock units on western side of Yoho River showing approximate limits on individual units and former ice flow. Note the units are laterally bounded by the valley wall.

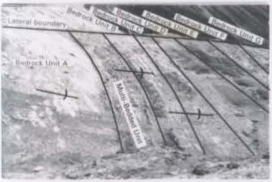


Figure 13c

Extension of the limits of bedrock units on the east side of Yoho River. Note the direction of former ice flow as evidenced by till deposits and striations, and the distinct break in the units slope (laterally bounded).

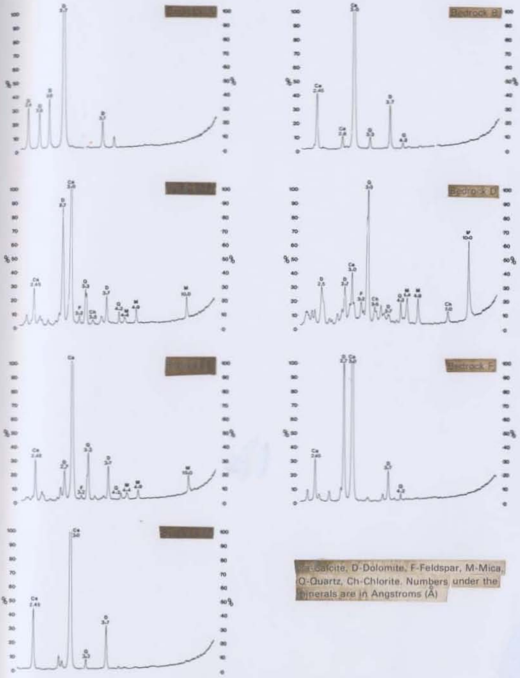
4.31 Bedrock Textures

Although no precise examination of the bedrock textures was made, generalities concerning the relative bedrock unit textures are appropriate. Dolomite, which is a major constituent of most of the bedrock units, generally occurs in rhombs and measures between 30 to 120 microns in size. Fine dolomite rhombs (less than 30 microns) are scattered through most limestone units. The grains of the calcite matrix, which were observed to be mainly nonferroan calcite, are 3 to 4 orders of magnitude smaller than dolomite grains.

A relative grouping of bedrock units based on textural characteristics can be made as follows: Bedrock unit A exhibits the coarsest textural size followed by bedrocks C and F respectively. Bedrocks B, E, G (grouped together), and bedrock unit D exhibited the finest textural characteristics.

4.32 X-Ray Diffraction Analysis

X-ray diffraction plots of the seven bedrock units are given in Figure 14. The presence of illite/mica is evidenced by a large peak between 9 to 10 Angstroms (Å), and generally a smaller peak at 4.4 Å. Quartz exhibits peaks at 4.2 Å and 3.3 Å while chlorite peaks at 14 Å, 7.0 Å, 4.7 Å, and 3.5 Å. Dolomite peaks at 2.5 Å, 2.6 Å, 2.7 Å, and 3.7 Å while calcite exhibits peaks at 2.45 Å and 3.0 Å. Most minerals in the montmorillonite group are identified by a peak between 12 to 15 Å. Interstratified clays and other mixed layer minerals are difficult to identify because of their complex structures. Generally, mixed layering of mica is indicated by a series of small peaks on the high Å side of the 10 Å peak. Mixed layering of the montmorillonite and chlorite are indicated by the small peaks on the lower Å side of the 14 Å peak (Weaver 1958; Carroll 1970; Beaumont 1972; Vahtra per. comm.).



Ca-Calcite, D-Dolomite, F-Feldspar, M-Mica, Q-Quartz, Ch-Chlorite. Numbers under the minerals are in Angstroms (Å)

Figure 14

X-ray Diffraction Plots of Bedrock Units

Glycolation and heat treatment showed that no kaolinite or montmorillonite was present in any of the bedrock or morainic samples. Batterson (1980, unpublished) came to a similar conclusion while studying morainic material in the Emerald Glacier area, 8 km southwest of Yoho Glacier.

4.33 Total Carbonate Content

Although solubility is only a marginally relevant test for erosional potential (kinetics of dissolution being more appropriate) it was felt that a relative comparison between the various bedrock solubilities would enable a qualitative assessment of total available carbonates and erosional potential.

Total carbonate content (% wt.) of the seven bedrock units is given in Table 6. In considering the kinetics of dissolution, the results suggest that all bedrock units except unit D show high susceptibility to this process of erosion if the subglacial environment were suitable (i. e. , continuous supply of melt water, Collins 1979).

4.34 Carbonic Acid (Ca CO_3) Solubility

Test results of the carbonic acid solubility analysis are presented in Table 7. Bedrock units B, C, D, E, F, and G quickly reached saturation. Bedrock unit A did not reach solution saturation until the 19th hour. The decreasing concentrations of Ca CO_3 exhibited in the 7th to 43rd hour readings in bedrock units B, F, and G were probably due to two factors:

i) change in pH level (acidity) of the solution; and

ii) experimental error in regards to the washing of the solution off the conduction probe after each reading.

Table 6

Total Carbonate Content (%)

	Percent carbonate by wt. in 1 gm. powdered sample
Bedrock Unit A	99.55
Bedrock Unit B	95.05
Bedrock Unit C	91.90
Bedrock Unit D'	28.20
Bedrock Unit E	84.90
Bedrock Unit F	99.70
Bedrock Unit G	97.75

(Acid-neutralization method; Allison and Moodie, 1965)

Table 7
Carbonic Acid Solubility

Time	Standard H_2CO_3		mg/L $CaCO_3$						
hr	pH	Temp °C	Bedrock Unit A	Bedrock Unit B	Bedrock Unit C	Bedrock Unit D	Bedrock Unit E	Bedrock Unit F	Bedrock Unit G
1	4.2	6.0	83	452*	321	278*	256*	365	387
2	4.2	5.5	138	452	365*	278	256	408*	408*
4	4.2	5.0	197	321	365	278	256	408	408
7	4.2	5.0	206	343	365	321	256	408	365
19	4.2	5.0	223*	300	321	278	256	332	321
31	4.2	5.0	223	273	321	321	256	332	321
43	4.5	5.0	223	234	321	278	256	278	234

*indicates solution saturation

The solution saturation concentrations (mg/L.) of the bedrock units are taken as those concentrations where stabilization of the readings has occurred. The highest solution saturation concentration levels were, in descending order, bedrock units B, F and G, C, D, E, and A. If solution play an important role in the erosion of the bedrock units, it is suggested that this process and associated erosional forms would be more pronounced on the surfaces of the bedrock units in the order given above.

There appears to be no relationship between the percent carbonate present in the bedrock units and the H_2CO_3 solution rates. Bedrock unit A, which exhibited one of the highest % carbonate values (99.55%), took the greatest amount of time to reach solution saturation (19 hours). Bedrock unit D, with only 28.2% carbonate by weight, exhibited immediate solution saturation of $CaCO_3$.

4.35 Apparent Porosity

The average apparent porosities of the bedrock units are given in Table 8. Microscopic examination of thin sections revealed that pore configuration is generally preferentially oriented in the direction of the bedding planes. The pore surface exposed on the exterior of the specimens will therefore vary with the orientation of bedding planes, hence apparent porosity will be expected to vary conformably. Table 8 gives the number of specimens tested for each bedrock type, and summarizes the resultant average apparent porosities. Because of the flaggy nature of bedrock unit D, cylindrical specimens could not be extracted, thus apparent porosity had to be calculated with the use of liquid displacement techniques (Lama and Vutukuri 1978). The total average apparent porosity for for bedrock unit D was calculated at 2.2% (5 samples tested).

All bedrock units except units C and E exhibited higher apparent

Table 8
Average Apparent Porosity of Bedrock Units

	Bedrock Unit A	Bedrock Unit B	Bedrock Unit C	Bedrock Unit D	Bedrock Unit E	Bedrock Unit F	Bedrock Unit G
Number of specimens with bedding parallel to the elongated cylindrical configuration of specimens.	10	10	13		14	19	16
Average Apparent Porosity (%)	2.33	0.59	0.95		0.91	0.82	0.52
Number of specimens with bedding perpendicular to the elongated cylindrical configuration of specimens.	18	16	17		8	16	17
Average Apparent Porosity (%)	1.51	0.54	0.95		2.12	0.61	0.48
Total Average Apparent Porosity	1.80	0.56	0.95		1.35	0.68	0.50

Bucause of the flaggy nature of bedrock unit D, cores could not be extracted from sample boulders.

porosities with beddings parallel to the elongated configuration of the specimens. The average apparent porosities of bedrock unit C with bedding perpendicular and parallel to the specimens elongated cylindrical configuration were identical, while bedrock unit E exhibited a higher apparent porosity with the bedding perpendicular to the elongated cylindrical specimen. The largest total average apparent porosity was exhibited by bedrock unit D (2.2%), the smallest was exhibited by bedrock unit G (0.5%). The large value exhibited by bedrock D is probably a result of its flaggy nature (laminated beds) and its high clay content which may have been responsible for the adsorption of water to the mineral surfaces (Brady 1974; Hillel 1971).

Bedrock unit A with a 1.8% total average porosity value was identified as having the largest textural size range of all the bedrock units. The large dolomite minerals resulted in large void spacings between minerals, and therefore a high apparent porosity value. Bedrock unit G, a lime mudstone, had a relatively small textural range with fewer and smaller void spaces.

High apparent porosities can influence a number of erosional mechanisms. Since subglacial water is usually acidic (Collins 1979) and because of the carbonate nature of the bedrock units, the water is able to penetrate into rocks (if pore pressures are suitable) leaving greater surface areas of rock susceptible to the solution process. If this process acts quickly, or if it persists for an extended period of time, a bedrock micro-relief could result that would resemble a coarse sandpaper appearance (Figure 15).

Another mechanism could result if subglacial water enters the rock pores and due to changing basal thermal regimes, freezes. The expanding, freezing water could cause a shatter effect on the rock surfaces, thereby reducing the surface relief of the bedrock after thaw takes place. (This

Figure 15

Pitted bedrock surface, possibly the result of solution at the ice-rock contact caused by abundance of sub-glacial water and high apparent porosity of the rock unit (Bedrock unit E).

shatter effect may resemble the more common 'frost shattering' of rocks when water freezes inside rock cracks or bedding planes).

Clearly, the greater the apparent porosity of a rock unit, the more likely it is that several erosional mechanisms can operate under various ice pressure and temperature conditions.

4.36 Bedrock Abrasion

The results of the abrasion test are presented in Table 9. The relative ease with which the bedrock units were abraded with respect to time taken for abrasion of 3.6 cm^3 of rock, can be ranked as follows: talc; bedrocks D, C, B, A, G, E, F and granite respectively.

Laboratory tests indicate, under given abrasion conditions, bedrock unit D will likely be eroded volumetrically the most whereas bedrock F will be eroded the least.

4.37 Uniaxial Compression

Compression tests on cylindrical bedrock samples have been used extensively by engineering geologists (Misterek 1970; Obert and Duvall 1967; and Jaeger 1972) mining engineers (Means 1976; Donaldson 1974; and others), and in related rock and soil mechanics (Lama and Vutukuri 1978; and Vutukuri, et al. 1974). Relatively little attention has been focused on this aspect of rock behaviour by glacial geologists.

In concept, the compressive strength of bedrocks is influenced not only by ice physics but by internal rock properties such as mineralogy, grain size, porosity, bedding and other inherent rock properties (Vutukuri, et al. 1974). Clearly, a compression test on bedrock units can provide valuable data to help describe the general theory of glacial erosion and comminution of rock debris.

Table 9
Abrasion Test Results (time required to abrade 3.6cm³ of rock sample)

	Bedrock Unit A		Bedrock Unit B		Bedrock Unit C
	16 minutes 05 seconds		12 minutes 25 seconds		06 minutes 41 seconds
	13 minutes 13 seconds		15 minutes 10 seconds		11 minutes 28 seconds
	15 minutes 04 seconds		16 minutes 09 seconds		07 minutes 28 seconds
	16 minutes 11 seconds		13 minutes 19 seconds		14 minutes 51 seconds
Average	15 minutes 11 seconds	Average	14 minutes 16 seconds	Average	09 minutes 35 seconds

	Bedrock Unit D		Bedrock Unit E		Bedrock Unit F
	03 minutes 30 seconds		30 minutes 14 seconds		43 minutes 23 seconds
	02 minutes 58 seconds		33 minutes 27 seconds		45 minutes 23 seconds
	05 minutes 27 seconds		31 minutes 53 seconds		43 minutes 08 seconds
	06 minutes 15 seconds		33 minutes 57 seconds		43 minutes 03 seconds
Average	04 minutes 33 seconds	Average	31 minutes 38 seconds	Average	43 minutes 17 seconds

	Bedrock Unit G		Granite		Talc
	19 minutes 27 seconds		93 minutes 56 seconds		01 minutes 35 seconds
	20 minutes 44 seconds		95 minutes 25 seconds		01 minutes 30 seconds
	15 minutes 38 seconds		96 minutes 10 seconds		01 minutes 25 seconds
	15 minutes 25 seconds		98 minutes 17 seconds		01 minutes 22 seconds
Average	17 minutes 38 seconds	Average	95 minutes 57 seconds	Average	01 minutes 27 seconds

Table 10 summarizes the average compressive strengths of the bedrock units with bedding planes perpendicular and parallel to the applied force.

Ruiz (1966) determined that the compressive strength of limestone and dolomites under apparent saturated conditions is up to 20% less than air dried samples. Vutukuri, et.al. (1974) report similar results. The results reported in Table 10 are therefore minimum strengths of bedrock types under apparent saturated conditions.

There is a great deal of literature regarding compressive strengths and orientation of bedding planes (Berenbaum and Brodie 1959; Dube and Singh 1969; Barron 1971; etc.). Lama and Vutukuri (1978) suggest that compressive strength is greatest as the angle of applied force reaches 0^o (parallel) and lowest as this angle approaches 90^o (perpendicular) to the bedding planes. Clearly, this would vary with rock types and inherent rock properties. Results from modelling of various angles of bedding planes and joints in rocks under compression tests generally agree with Lama's and Vutukuri's (1978) reported findings.

The results in Table 10 show that all bedrock units except A and E have higher compressive strengths with bedding at 0^o to the applied force. Bedrock unit A is exceptional probably because of the massive nature of the unit. Bedrock unit E on the other hand has a large number of quartz veins which probably lead to fracturing at the quartz-limestone interfaces regardless of bedding plane orientations.

Bedrock unit A exhibits the highest total average compressive strength (22.6 MPa.). The high compressive strength is probably due to the large dolomite content, which Hugman and Friedman (1979) state has a

Table 10

Summary of the Uniaxial Compression Test Results

	Bedrock Unit A	Bedrock Unit B	Bedrock Unit C	Bedrock Unit D	Bedrock Unit E	Bedrock Unit F	Bedrock Unit G
Number of specimens with bedding planes parallel to the applied force (0°)	10	10	13		14	19	16
Average Compressive Strength (MPa)	21.5	16.3	13.3		8.8	21.6	21.3
Number of specimens with bedding planes perpendicular to the applied force (90°).	18	16	17		8	16	17
Average Compressive Strength (MPa)	23.2	12.4	11.2		10.9	11.5	16.6
Total Average Compressive Strength (MPa)	22.6	13.9	12.1		9.5	17.0	18.9

Because of the flaggy nature of bedrock unit D, cores could not be extracted from sample boulders. Ringheim (1964) recorded compressive strengths of between 0.06 MPa and 0.63 MPa for soft shale units from southwest Saskatchewan which are believed to be similar in texture to some Alberta and eastern British Columbia shale units (Cook 1975). Bedding orientations during the compression tests were not stipulated.

significant effect on the ultimate strength of low porosity, carbonate rocks.

4.4 Boulder Count

The results of the boulder count are given in Table 11.

Recent literature emphasizes the causal relationship between jointed rocks and the ease with which they are plucked, crushed, and transported in glacial and other environments (Boulton 1978; Lama et al. 1978). Galibert (1962) and Boulton (1978, 1979) suggest that the plucking or the large scale excavation of a bedrock may be enhanced by pre-existing joints in rocks. Laboratory investigations have also resulted in similar findings (Chenevert and Gatlin 1965; Chappell 1974; 1975; Byerlee and Summers 1975; etc.).

The strike-dip relationship will also have an effect on the plucking process. With the beds striking approximately north-south and dipping steeply west between 34° and 54°, the resultant ice contact was at approximately 45° (Figures 9 and 13 a, b). From laboratory compression tests Vutukuri, et al. (1974), and Lama and Vutukuri (1978) have shown that similar orientations of rock cores enhance rock failure.

Although jointing and bedding planes are not widespread in bedrock unit A, a relatively large boulder count (100 or 20%) was recorded. This can best be explained with regard to the bedrock surface area exposed to glacial processes. Bedrock A extends from the raised bedrock bench (described previously), northward to beyond the Yoho Glacier into the Wapta Icefield (Eldon Formation; Cook 1975). The total area exposed far exceeds that of the combined bedrock unit areas of 580 square metres on the western side of the Yoho River. The extensive surface area available for glacier erosion would therefore suggest that a greater probability exists for the process of plucking to occur.

Table 11
Boulder Count Results

	# of Boulders	% of Total
Bedrock Unit A	100	20
Bedrock Unit B	116	23
Bedrock Unit C	51	10
Bedrock Unit D	15	3
Bedrock Unit E	45	9
Bedrock Unit F	117	24
Bedrock Unit G	20	4
Other	36	7

On bedrock units B and F where fractures and cracks were identified, the boulder count of 116 and 117 respectively may be related to the angle of contact between ice motion and bedding planes (dip). Lama and Vutukuri (1978) and Vutukuri, et al. (1974) showed that as the angle of applied force reaches 90° to bedding planes of experimentally compressed rocks, the compressive force required to crush the specimen increased.

Tests were conducted on the shear strength of specimens where similar results were reported (Vutukuri, et al. 1974). The force required to shear the specimen decreased as the angle of force approached 180° to the bedding planes. These findings therefore suggest that given a constant pressure, whether it be compressive, shear or frictional, bedrock units with bedding planes oriented parallel to the direction of the applied force, will tend to be less resistant to fracture than those oriented at right angles to the applied force.

This argument would also apply to bedrock units C, E, and G, with boulder counts of 51, 45 and 20 respectively. Only 15 boulders were identified as bedrock D. This is probably due to the nature of the rock which is clearly susceptible to rapid comminution of entrained clasts. The number of bedding planes (both major and minor) is clearly a controlling factor in the plucking and crushing of bedrock units.
The pressure-induced plucking of elongated erosional froms (Boulton 1979, and others) is probably enhanced when bedding planes are present.

4.5 Till Analyses

4.51 X-ray Diffraction

X-ray diffraction plots for the twelve recessional moraine samples are presented in Figures 16 a and b. Moraine samples 1-1, 1-2 and 1-4

Figure 16a

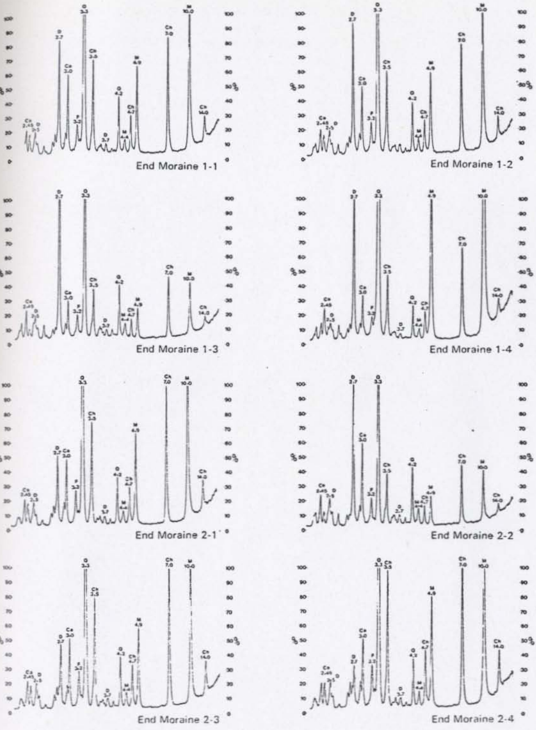
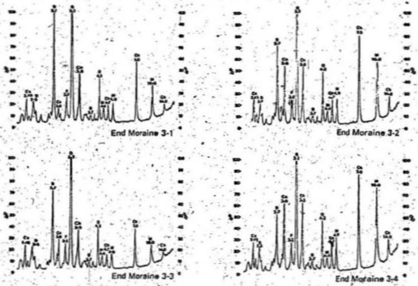
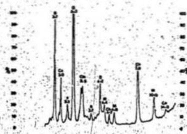


Figure 16a and b
Recessional End Moraine X-Ray Diffraction Plots

Figure 16b





End Moraine 2-2

(This sample was glycolated and heat treated to determine if kaolinite or montmorillonite were present. The minor split in the 3.5Å (Chlorite) peak indicates their absence.

show extremely high quantities of dolomite, quartz, chlorite and mica, with lesser amounts of calcite and feldspars. Moraine sample 1-3 exhibits high dolomite and quartz peaks and lesser calcite, chlorite, feldspar and mica peaks.

Moraine samples 2-1, 2-3 and 2-4 exhibit virtually the same patterns as those described above for moraine samples 1-1, 1-2 and 1-4. Moraine sample 2-2 demonstrates high dolomite and quartz peaks and lesser calcite, feldspars, chlorite and mica peaks (similar to end moraine 1-3).

Moraine samples 3-1 and 3-3 show extremely high dolomite and quartz peaks with chlorite, mica, calcite and feldspars following in order of magnitude. Moraine samples 3-2 and 3-4 demonstrate a high quartz peak with secondary high chlorite, dolomite, calcite and mica peaks and a low feldspar peak.

The origin of the tills based on the X-ray diffraction plots is difficult to assess. No clear relationship exists between bedrock unit plots and moraine plots. All bedrock units generally exhibit high peaks in dolomite and/or calcite. Only bedrock unit D shows a dominant quartz and mica peak. All end moraine samples exhibit peaks in quartz and mica which for all except samples 1-3, 2-2, 3-1, 3-3 are major peaks. By using transparent overlays of all bedrock unit and moraine samples, it was evident that the patterns and magnitudes of peaks exhibited in bedrock D closely resembled all of the moraine plots. This is in itself, however, not conclusive evidence that bedrock D is in greater abundance in the till matrix. Based on previous studies of various shale units (Batterson 1980, unpublished; Beaumont 1971), X-ray diffraction analysis showed high chlorite, quartz and mica peaks. These peaks were also observed in all the recessional moraine samples, although their magnitude varied. The cliff face adjoining the lateral moraine is part of the Sullivan Formation (Cook 1975):

a shale unit. However, as shown on Figures 2a and b, the 1903 and 1913 Yoho Glacier extent, very little supra-glacial debris is present at the ice-valley wall contact. Therefore, the volumetric component of the valley side debris in the tills is probably very low. It is suggested that valley side debis that did reach the recessional end moraines supra-glacially (approximately 100 m to 300 m from the 1913 glacier terminus), is unlikely to have been rapidly comminuted in such a short distance, and via supra-glacial transport. The observed peaks of chlorite, quartz and mica may well be of bedrock unit D provenance. The dearth of supra-glacial debris evidenced in photographic records (Figures 2a,b,c) shows that the debris supply to the moraines was mainly of subglacial origin (i.e. abraded and plucked debris).

4.52 Mineralogy: Frictional Properties of Minerals

In establishing the mineral content of the moraine samples, it is necessary to discuss and speculate on the affect these minerals may have on glacier sliding and upon bedrock morphology.

Horn, et al. (1962) studied the frictional properties of various minerals including calcite, quartz, chlorite and others. Their investigations revealed that for various moisture conditions kinetic friction is generally equal to or slightly less than static friction. The exception to this occurred in the case of quartz where a stick-slip phenomenon was observed. This is attributed to kinetic frictional resistance developed between the saturated quartz surfaces and the sliding medium. Dry quartz exhibited relatively little frictional resistance, as did calcite. The presence of water was noted to act as an anti-lubricant when it was applied to surfaces of minerals that had massive crystal structures such as quartz and calcite, whereas it lubricated surfaces of minerals such as chlorite and micas that had layer-lattice crystal structures.

The stick-slip process clearly varies with rock type and bed roughness, however, under given conditions there are definite trends with mineral content. Brace (1972) and Byerlee and Brace (1968) state that chlorite and micas have little influence on the stick-slip process. Increased quartz content is more conducive to stick-slip behaviour. Stick-slip occurs in carbonate rocks but only under suitable moisture conditions. Byerlee and Brace (1968) found that stick-slip behaviour occurs on limestone and dolomites less readily than on sandstones, granites or quartzites (Ohnaka 1973; Engelder 1974; Friedman, et al. 1974).

Clayton (1951) states that the frictional characteristics of many coefficients of quartz do not vary with the rate of sliding in both saturated and dry states (Horn, et al. 1962). Clay minerals on the other hand show increases in static frictional resistance under both wet and dry conditions.

All layer-lattice minerals (like those found in bedrock unit D) have perfect basal cleavage (Horn, et al. 1962). Atoms in a molecular sheet are held together by relatively strong covalent bonds. The molecular sheets on the other hand are held together by weak van der Waals forces. As a result, it is extremely easy to cleave clean sheets from crystals of these minerals. It is difficult, however, to rupture a crystal along a non-cleavage plane. This implies that the abrasion that results when boundary-lubricated surfaces of layer-lattice minerals are rubbed together generally exposes freshly cleaved surfaces. If no solution is present to neutralize the cohesive forces between the fresh cleavage surfaces, the frictional resistance will be high, thus sliding along the plane is reduced and abrasion is enhanced.

A number of conclusions can be drawn from this discussion:

1) Mineral type and content have a significant role to play in the sliding and/or abrasion process of the ice over bedrock.

ii) The presence or absence of a lubricating layer of water can enhance or retard those processes depending on mineral type.

iii) The relative presence of quartz in rock types is at least partially responsible for the stick-slip process often reported in glaciological studies.

All bedrock units studied were shown to be high in calcite (Figure 12). Therefore, frictional resistance to sliding ice should have been relatively small if the ice-rock interface were dry and greater if it were wet. The presence of precipitates over all the bedrock units clearly indicates that the ice was warm based and water was present at the ice-rock interface in most locations at some time.

Based on frictional properties of minerals, a water layer generally impedes glacier sliding over most of the bedrock units and encourages abrasion. An exception occurs with bedrock unit D. Having a high quartz and mica content (partial perfect basal cleavage), the unit should exhibit cleavage at sheet planes when moisture is in short supply, and enhanced sliding when water lubricates the bedrock surface.

This may be manifest in the surface morphology of this unit as resembling an ideal plane configuration that takes the form of the units dip (34° SW) or foliation (in this case, a fairly horizontal morphology at the meso and micro-scale).

4.53 Grain Size

The plots of particle size, frequency distribution, and statistical parameters for the 41 moraine samples are presented in Appendix C.

A bimodal size distribution is apparent in most of the till plots, having a clast size consisting predominantly of rock fragments (-2.5Ø) and a till matrix consisting mainly of mineral fragments at around 5Ø. This is in keeping with Dreimanis and Vagners (1971) finding on glacial erosion and transport of tills (predominantly of dolomite origin) in Southern Ontario. Buller and McManus (1973) found that tills from valley glaciers tend to have bimodal size distributions as well. Controversy exists as to whether till textures are a function of the environment of transport or dependent upon the nature of the bedrock source. Slatt (1971) examined the textures of sediments taken from terminal deposits of various Alaskan valley glaciers which eroded five different bedrock types and determined that the modal occurrance of proglacial sediments is probably a function of glacial transport and independent of the bedrock type.

Contrary to Slatt's (1971) findings, Mills (1975) working on temperate valley glaciers, and Dreimanis and Vagners (1971) studying sediments of continental glacier origin, determined that the texture of tills is dependent upon sedimentary bedrock lithology.

The bimodal trend exhibited by the sample analysis is probably a result of the bedrock types rather than the environment of transport. The dominance of dolomite and calcite in both bedrock and moraine samples (recessional end moraines) confirms this as the terminal grade of dolomite and calcite, and of the samples (till matrix mode), with size fractions at between 4Ø and 8Ø (Dreimanis and Vagners 1971). It is surprising to find terminal grades appearing after only 1 km or so of transport. This may reflect the high sliding-abrading energy of temperate alpine glaciers (Rogerson 1980, per. com.), but is probably more the result of the general bedrock type which in relative terms is fairly weak.

Textural changes in the morainic material (progressively down valley) along the latero-terminal moraine are demonstrated in Figure 17.

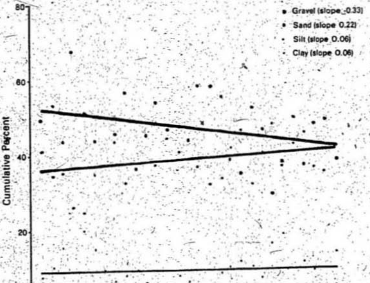


Figure 17
Textural changes in lateral moraine till progressively down valley

Percent gravel, sand, silt and clay fractions were taken from the particle size and frequency curves, and a simple regression analysis was performed. The slopes of the individual regression lines are thought to represent constant values or slight changes in the cumulative percent of the individual textural classes over the distance of the latero-terminal moraine (600 m.)

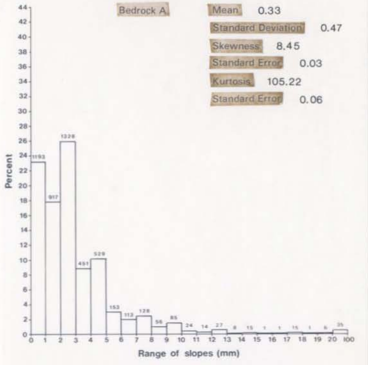
The gravel fraction exhibits the only negative slope (-0.33) indicating that the percent gravel is decreasing down valley as the sand (slope 0.22) increased very slightly. The silt and clay fractions remain relatively constant (silt slope 0.06, clay slope 0.06). Dreimanis and Vagners (1971) believe that textures of tills derived from sedimentary rocks depend not only upon the terminal grades of individual sedimentary minerals which are mainly located in the silt and clay fractions, but also upon the medium/hard sedimentary rocks composing the sand fraction. These are often aggregates of dolomites and calcites. This bimodality is said to take place anywhere from 0 to 3 km from bedrock sources in continental ice deposits (Dreimanis and Vagners 1971). Despite the alpine glacial environment, some similar trends in textures and terminal grade modes of sedimentary rocks are observed in the present study. There is some tendency towards increasing sand fraction down valley and a decreasing gravel fraction.

4.6 Roughness - Morphologies: Computer Analysis

4.61 Slope Frequencies

The graphs of the slope frequency distributions at the micro scale and accompanying statistical parameters for each bedrock unit plot are given in Figures 18 a, b, c, and d. The means of the slopes (segment

Figure 18a



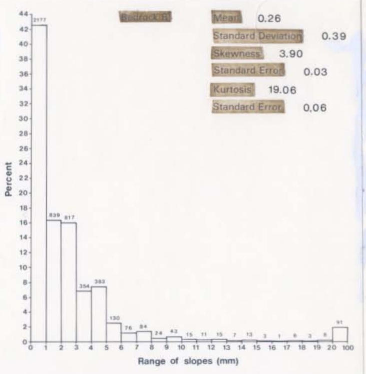
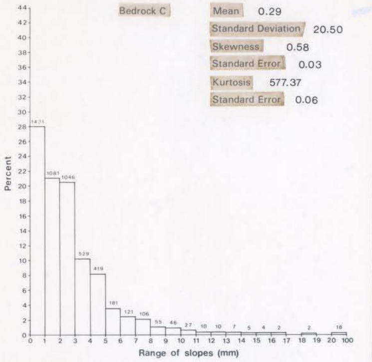


Figure 8 a, b, c and d Slope Frequency Distributions.


Bedrock C
Mean 0.29
Standard Deviation 20.50
Skewness 0.58
Standard Error 0.03
Kurtosis 577.37
Standard Error 0.06
Percent
Range of slopes (mm)
1081
1046
529
419
181
121
106
55
46
27
10
10
7
5
4
2
2
18

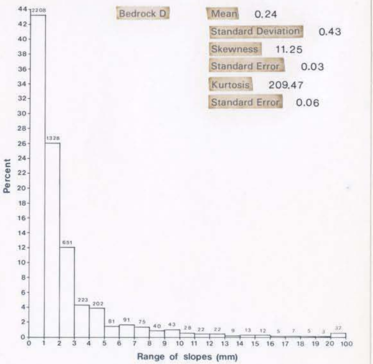
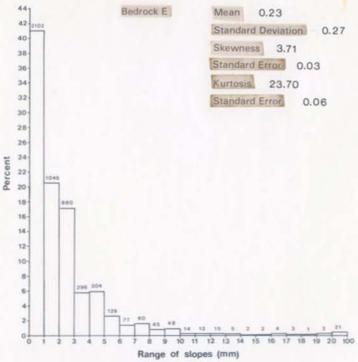

Bedrock D
Mean 0.24
Standard Deviation 0.43
Skewness 11.25
Standard Error 0.03
Kurtosis 209.47
Standard Error 0.06
Percent
Range of slopes (mm)
2208
1328
651
223
202
81
91
75
40
43
28
22
22
9
13
12
5
7
5
3
37

Figure 18b


Bedrock E
Mean 0.23
Standard Deviation 0.27
Skewness 3.71
Standard Error 0.03
Kurtosis 23.70
Standard Error 0.06
Percent
Range of slopes (mm)

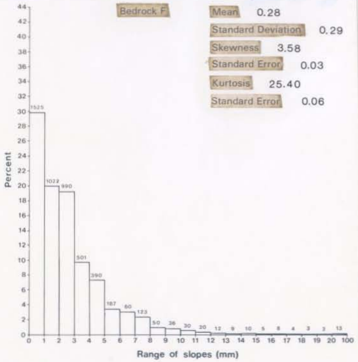

Bedrock F
Mean 0.28
Standard Deviation 0.29
Skewness 3.58
Standard Error 0.03
Kurtosis 25.40
Standard Error 0.06
Percent
Range of slopes (mm)

Figure 18

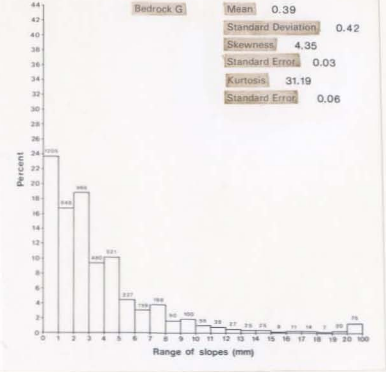

Bedrock G
Mean 0.39
Standard Deviation 0.42
Skewness 4.35
Standard Error 0.03
Kurtosis 31.19
Standard Error 0.06
Percent
Range of slopes (mm)

Figure 18d

lengths) range from 2.3 mm on bedrock E to 3.9 mm on bedrock G. The median is only influenced by the position of values in the ranking and not directly by the magnitude of these values. Therefore, it is a measure of central tendency (King 1969). This suggests that at this particular sampling interval, the tendency towards a more uniform, levelled relief is exhibited by units E, D, B, F, C, A, and G respectively where E is the most and G is the least uniform or levelled (horizontally). All units exhibit positive skewness with bedrock units D and A being strongly skewed (11.25 and 8.45 respectively), and bedrock C exhibiting the least positive skewness (0.50). Kurtosis values all indicate an extremely low concentration of slopes around the mean.

Of particular interest is the range of slope segments graphed at 10 mm to 100 mm. Considering slope sample points are 2 cm apart, the frequency distributions in this range may be a result of relatively weak bedding planes or fractures within each rock type. It is therefore suggested that these relatively large slope values can be explained in terms of 'micro-plucking' of individual units by overburden ice (similar to processes described by Kemmis 1979). An examination of the bedrock descriptions (Table 5) verifies that bedrock unit G with 12% of slopes between the 10 mm to 100 mm range is also very well bedded.

Other bedrock unit slope frequencies between 10 mm and 100 mm are as follows: bedrock unit A, 5.7%; bedrock unit B, 6.7%; bedrock unit C, 4.9%; bedrock unit D, 6.1%; bedrock unit E, 3.2%; bedrock unit F, 4.5%. A qualitative assessment of the extent of micro-plucking appears in the next section.

4.62 Trend Surface

Trend surface analysis enables both quantitative and qualitative interpretation of data. The Symvu 3-dimensional representation of bedrock unit form (residual surface) enables a visual assessment of bedform which provides some insight into the related glacial erosive processes.

A series of three diagrams were constructed for each bedrock unit; azimuth 045° views the surface in an up-glacier direction; azimuth 225° views the surface in a down glacier direction; azimuth 315° views the resultant glacially-eroded forms at right angles to glacier flow. The 045° azimuth view might be expected to highlight areas of micro-plucking on the lee of small, eroded forms while the 225° azimuth view illustrates the smoothed stoss side of these forms. The 315° azimuth view emphasizes the asymmetry in micro-relief. The altitude (the elevation of the viewing position above the horizontal plane) is kept constant at 45°, as is the viewing distance (25 cm).

The quantitative analysis includes the following (Appendix D):

1) Standard deviation (σ) or the second moments measure, which is a measure of dispersion. This statistical parameter indicates the spread of values on either side of the mean of normalized data. In this case the mean is represented by the mean plane defined by the 3rd order polynomial equation. Sixty-six and two thirds percent of the variation in the sample would be expected to lie within one standard deviation of the mean, and 95% within two standard deviations of the mean.
2) Total variation (σ^2) is the standard deviation squared. It is a measure of the spread of values in the individual bedrock unit plots.
3) Coefficient of determination (r^2) is the coefficient of correlation squared, expressing the proportion of variation in the dependent variables (observed data values) explained by the association with

the independent variables (expected data values) (Cole and King 1968; and Chorley and Kennedy 1971).

Recent studies on bed roughness and form have generally dealt with a 2-dimensional perspective on the scale of the order of metres (Boulton 1974; Kamb 1970; Hallet 1976; Fowler 1979; Lliboutry 1977, 1978; Weertman 1976; and others). Benoist (1979) examined longitudinal profiles of roches moutonnées measuring every centimetre over a total length of approximately 100 m, which he termed micro-relief, but which was a two-dimensional perspective. Both Lliboutry (1968, 1975, 1977, 1978) and Weertman (1966, 1967, 1969, 1976) have studied bed roughness parameters by introducing sine waves and relating the amplitudes to a roughness index. This treatment however only concerns itself with 2-dimensions. It is suggested that a 3-dimensional examination of bedform is a more realistic approach to the problem.

If one assumes that the erosion of a bedrock surface at the micro-scale (as in this study) decreases as the bedform reaches its 'ideal configuration' or 'ideal plane' (which allows ice to slide with minimum work) then, by inference, there must be a bedrock plane configuration which may enhance sliding but reduce erosion. This 'ideal plane' can be mathematically defined in terms of polynomials.

The equation describing the ideal plane was chosen at the 3rd order, or cubic (having a cubed as well as a squared linear term). This order was selected after a computer run of five polynomial orders was made for each bedrock unit plot and examined with reference to best fit (Dougenik and Shechan 1977). Peucher (1972) and Cole and King (1968) recommend polynomial equations of the 3rd order to describe most geological features.

4.621 Quantitative Analysis

King (1967), Cole and King (1968), and Chorley and Haggett (1965) suggest that the standard deviation of the plotted data can supply useful information as to where the values are in relation to the best fit polynomial surface. It is hypothesized that this statistical parameter can be used as an indication of how close the bedrock is to being eroded to its ideal plane.

The standard deviations of the bedrock units (ranging from 1.33 for bedrock G to 2.00 for bedrock B) suggest that from 66 2/3% to 95% of the data points lie within 1.33 and 2.00 standard deviations of the ideal plane. The high correlation coefficients, ranging from .72 (bedrock B) to .98 (bedrocks A and C) indicate that the plotted surfaces closely resemble the polynomially determined 3rd order surface. These values are perhaps not surprising considering the large sample number (2601) in each case, however, it must be remembered that residual values, which tend to accentuate local anomalies, were plotted. Therefore, the statistical values do have meaning.

If the ideal plane hypothesis is accepted, then all bedrock unit plots were very close to achieving the enhanced, non-erosive, flow configuration.

It is suggested that changes in ice-flow direction, basal pressure, shear pressure, etc. would likely necessitate an ideal plane of different configuration, and because these conditions are known to fluctuate (Paterson 1969; Lliboutry 1976; Weertman 1979), an ideal plane on a bedrock surface is probably never achieved. However, one may speculate that because the statistical parameters computed indicate the bedrock forms closely resemble the individual ideal planes, ice flows, pressures, and other ice-rock interface conditions must have been fairly constant over a period of time necessary to erode the bedrock units to their present configurations, or the bedrock units responded very quickly to glacial conditions.

4.622 Erosional Bedforms and Processes: A Qualitative Assessment

Bedrock Unit A

The residual plots of the trend surface analysis for bedrock unit A are presented in Figures 19 a, b, and c. The bedrock unit (dolomite) is, in respect to the other bedrock types studied, a fairly massive unit with few bedding planes. This is generally the observed case for most dolomites (Hugman and Friedman 1979; ~~Wanless 1979~~; etc.). No bedding planes were observed in the bedrock plot examined. Therefore, no inherent weakness in this particular plot contributed to its glacially eroded relief form.

Figures 19 a and c clearly show localized results of micro-plucking, whereas 19 b presents a smoother surface to the viewer. The micro-plucked surfaces are probably the result of ice pressure and subsequent fracture of the lee-side surfaces similar to the processes used to describe the plucked lee surfaces of elongated erosional forms of much larger dimensions (Boulton 1974, 1979; Carol 1947; Engelder and Scholz 1976). Figure 19 b illustrates the smooth stoss slopes of micro forms.

Bedrock Unit B

The residual plots of the three azimuth views of bedrock unit B are presented in Figures 20 a, b, and c. Two major bedding planes are evident in Figure 20 a. This view illustrates the erosive effects of sliding ice on the rocks with bedding planes. Figures 20 a and c show

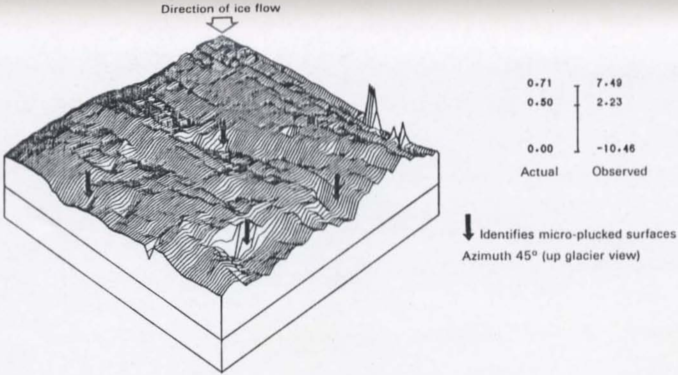


Figure 19a Residual plots of the trend surface Bedrock Unit A

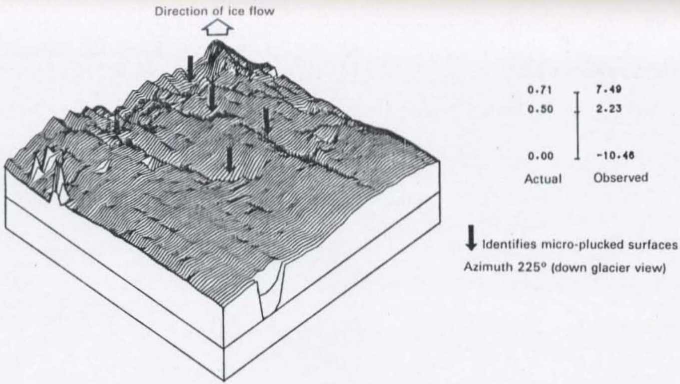


Figure 19b Residual plots of the trend surface, Bedrock Unit A.

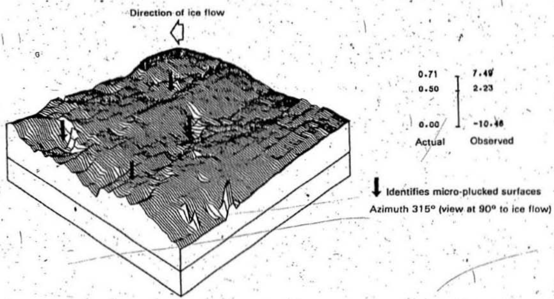


Figure 19c Residual plots of the trend surface Bedrock Unit A

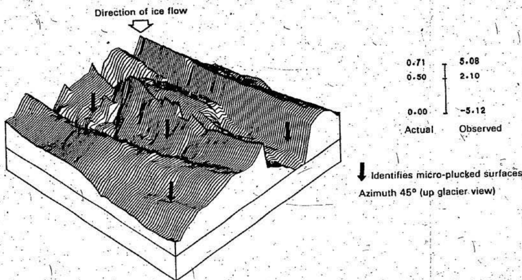


Figure 20a Residual plots of the trend surface Bedrock Unit B

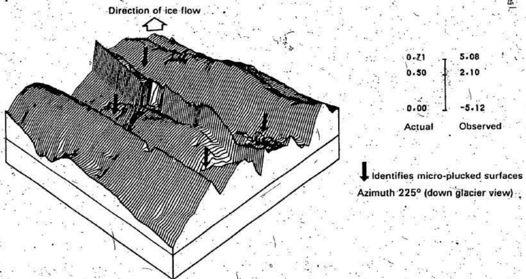


Figure 20b Residual plots of the trend surface Bedrock Unit B

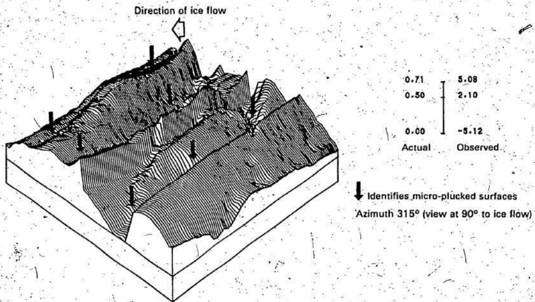


Figure 20 c   Residual plots of the trend surface   Bedrock Unit B the lee surfaces plucked at bedding plane contacts. This is evidenced by the jagged outlines and vertical slopes (analysis of slope frequency shows 6.7% of the slopes are greater than 10 mm).

Figure 20 b illustrates the smoothed stoss slopes, while Figure 20 c shows in profile, the combined effects of smoothed stoss and plucked lee surfaces. Micro-plucking is widespread over the entire plot and is accentuated along bedding planes. There is little evidence of solution erosion (solution hollows, etc.), however, the process may have had some effect along the bedding planes.

Bedrock Unit C

The three azimuth views of bedrock unit C are given in Figures 21 a, b, and c. The high peaked ridges exhibited in all three views can be attributed to quartz veins. Micro-plucking is evident in Figures 21 a and c. Relatively smoother stoss slopes are illustrated in Figure 21 b. Solution erosion appears to be minimal.

Bedrock Unit D

Fractures and cracks widespread in bedrock plot D are depicted in Figures 22 a, b, and c. The jagged appearance of the plot may be attributed to the rock type (shale). Bedding units may be easily picked out in Figures 22 b and c. The summits of the eroded forms (Figures 22 b and c) are attributed to the more resistant quartz veins. As in bedrock unit B, micro-plucking is accentuated along the bedrock planes (Figures 22 a and c). Figure 22 c indicates some smoothed stoss slopes and plucked lee slopes.

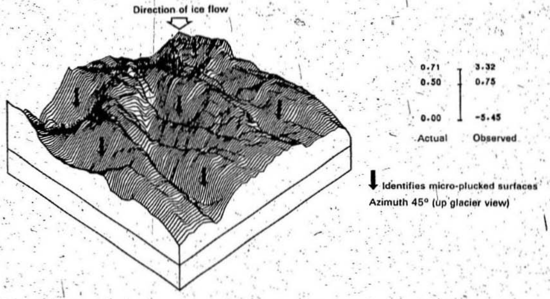


Figure 21 a Residual plots of the trend surface Bedrock Unit C

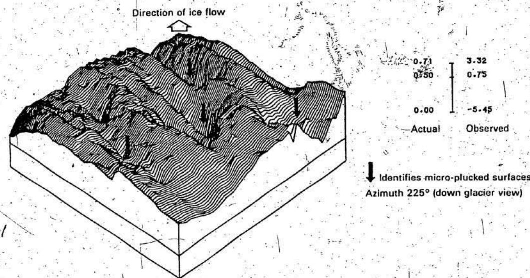


Figure 21 b   Residual plots of the trend surface   Bedrock Unit C

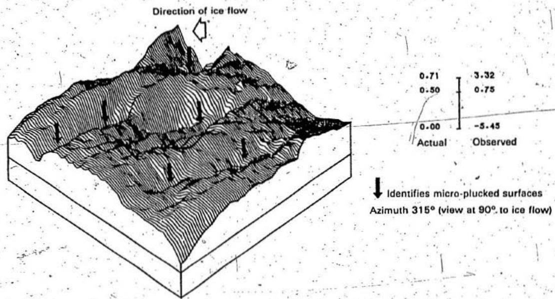


Figure 2f-c Residual plots of the trend surface Bedrock Unit C

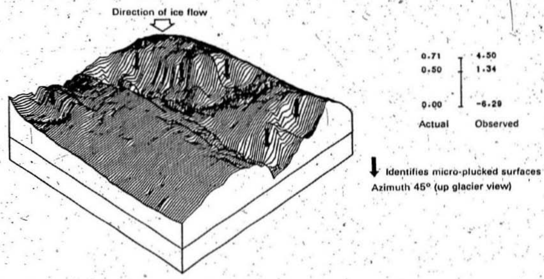


Figure 22a Residual plots of the trend surface Bedrock Unit D

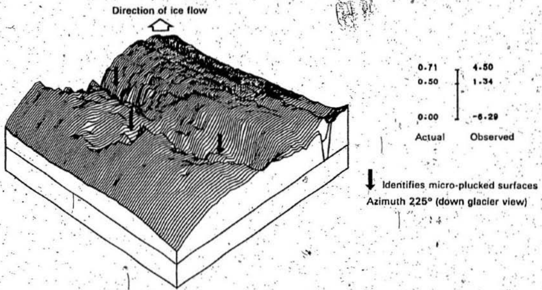


Figure 22b Residual plots of the trend surface Bedrock Unit D

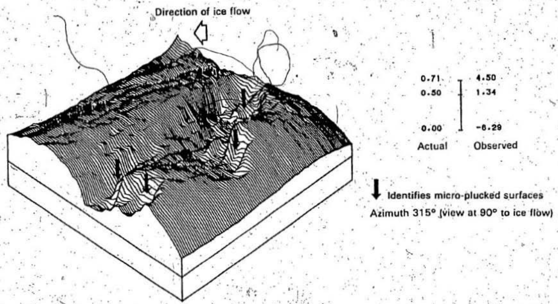


Figure 22c Residual plots of the trend surface Bedrock Unit D

Bedrock Unit E

Bedrock unit E was observed to have quartz veins dominating the relief. This is evident on the summits depicted in Figures 23 a, b, and c. Widespread fracturing and cracks were also observed which likely account for the apparent high degree of micro-plucking (Figures 23 a and c). The somewhat irregular stoss slopes shown in Figure 23 b do not appear to support the micro-plucking process, however, Figure 23 c does contain evidence of this phenomenon in the form of plucked lee surfaces.

Bedrock Unit F

Bedrock unit F (Figures 24 a, b, and c) shows evidence of at least one major bedding plane. As in other bedrock units, plucking along the bedding plane dominates the relief (Figure 24 a and c). Figure 24 b clearly demonstrates a smoothed stoss slope. The steep slopes in plucked areas around bedding planes is typical of other bedrock units which exhibit bedding planes. Solution effects in plucked cavities at the bedding plane (Figures 24 a and c) were probably responsible in part, for the resultant relief form.

Bedrock Unit G

The residual plot of the trend surface analysis for bedrock unit G is presented in Figures 25 a, b, and c. Bedding planes and partings were o bserved to be abundant on this bedrock unit. Figures 25 a and c show that at least two major bedding planes appear in the bedrock plot. Plucking is accentuated all along the bedding planes (Figures 25 a and c). Figure 25 b shows some resemblance to relatively smoothed stoss slopes. Solution effects appear to have been restricted to plucked cavities.

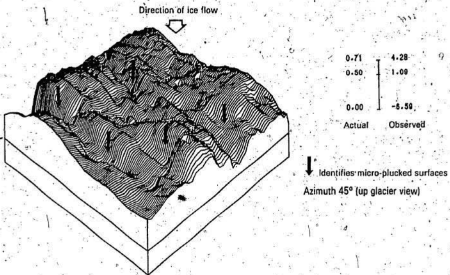


Figure 23a Residual plots of the trend surface Bedrock Unit E

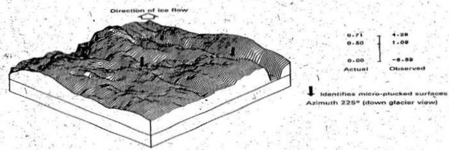


Figure 23b Residual plots of the trend surface Bedrock Unit E

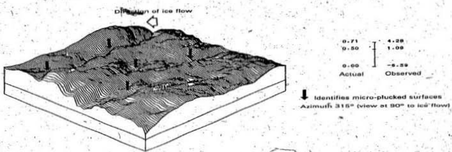


Figure 23c Residual plots of the trend surface Bedrock Unit E

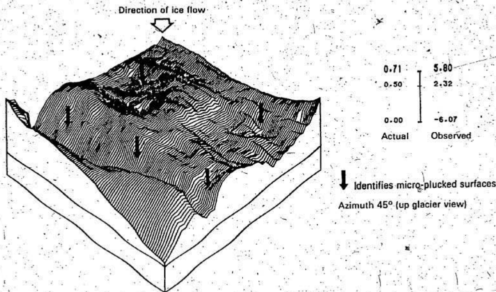


Figure 24a Residual plots of the trend surface Bedrock Unit F

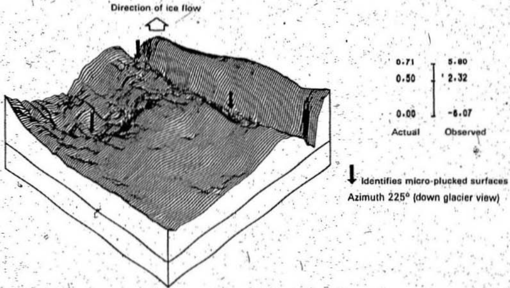


Figure 24b Residual plots of the trend surface Bedrock Unit F

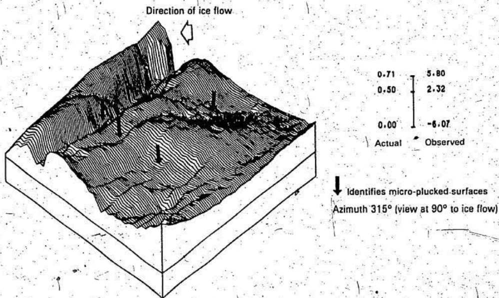


Figure 24c Residual plots of the trend surface Bedrock Unit F

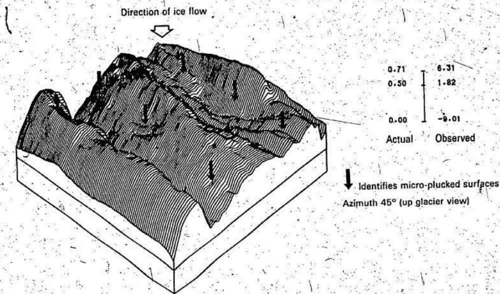


Figure 25a Residual plots of the trend surface Bedrock Unit G

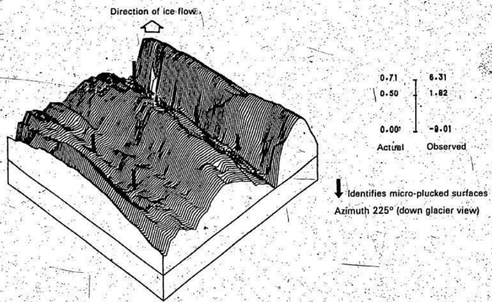


Figure 25b Residual plots of the trend surface Bedrock Unit G

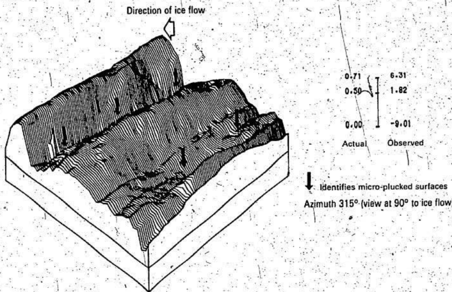


Figure 25c Residual plots of the trend surface Bedrock Unit G

4.623 Conclusions

The 3-dimensional trend surfaces reveal features which may be attributed to abrasion and micro-plucking at this scale. In addition, solution-induced erosional forms appear to be less pronounced. The micro-plucking process is widespread on all bedrock unit plots and appears to be accentuated in zones of fractures, cracks and particularly bedding planes. The classical smoothed stoss and plucked lee slopes often reported on meso-scale elongated erosional forms (whalebacks, roches moutonnées etc.) appear to occur also on erosional forms at the micro-scale.

The high degree to which the surfaces approximate an ideal sliding surface (plane) indicates that:

i) The bedrock units respond very quickly to glacial conditions, and/or;

ii) Ice flows, pressures and other ice-rock interface conditions must have been fairly constant over a period of time necessary to erode the units to their present configuration.

Chapter 5

Discussion and Conclusions

5.1 The Susceptibility of Bedrock Units to Erosion

The results from laboratory analysis performed on the lithologic units enable a ranking of relative bedrock strengths or susceptibility to erosion by physical and chemical processes. Table 12 is a ranking from 1 to 7 (corresponding to the number of bedrock units) whereby 1 represents the most susceptible to erosion or weakest bedrock unit and 7 the least susceptible to erosion or the strongest bedrock unit according to each laboratory test. The test scores are totalled so that the summed laboratory analysis scores are directly proportional to the net bedrock strength.

Bedrock unit D with the lowest net score (12) proves to be the weakest unit while bedrock unit G with the highest net score is the strongest. The other bedrock units, although showing some variations in the total scores, are very similar. All bedrock units generally show a wide variation of rank to various tests (i.e., no unit appears to be consistently low or high ranking in all laboratory tests). The implication of this is that in detail, different erosional processes may predominate on different bedrock units.

5.2 Abundance of Bedrock Units in Till Samples

The laboratory analyses conducted on till samples (X-ray diffraction, grain size) did not produce results that enable definite conclusions to be drawn about the relationship between bedrock type and their volumetric

Table 12

Ranking of Bedrock Strengths: Susceptibility to Erosion by Physical and Chemical Processes

	Bedrock Unit A	Bedrock Unit B	Bedrock Unit C	Bedrock Unit D	Bedrock Unit E	Bedrock Unit F	Bedrock Unit G
Total Carbonate Content	2	4	5	7	6	1	3
Carbonic Acid Solubility	7	1	6	2	3	4	5
Apparent Porosity	2	6	4	1	3	5	7
Bedrock Abrasion	4	3	2	1	6	7	5
Uniaxial Compression	7	4	3	1	2	5	6
Total Score	22	18	20	12	20	22	26

For Bedrock Abrasion and Uniaxial Compression, the rank of 1 to 7 refers to lowest and highest readings respectively. For the Total Carbonate Content, Carbonic Acid Solubility, Apparent Porosity and Boulder Count, the rank of 1 to 7 refers to the highest and lowest content respectively. Thus for this discussion, the rank of 1 implies most susceptible to erosion and 7 least susceptible."

abundance in till. No precise examination of bedrock textures was made, thus the relationship of these to till textures remains tentative.

Although bimodality is observed in the grain size distribution curves with apparent terminal grade modes, it is doubtful that these characteristics are solely the result of either rock type, distance of transport, or mode of transport but rather a combination of the three. Samples from the latero-frontal moraine were taken over a distance of approximately 500 m (1 to 1.5 km down valley from the bedrock source). All samples show a trend towards bimodality. Therefore, blocks plucked from the bedrock units were probably comminuted, due to a combination of abrasion and crushing for up to 1.5 km from their source.

Since all bedrock units have dolomite and calcite as a major mineral constituent, no distinction can be made as to which rock unit was dominant in the tills. Bedrock unit D which was observed to be high in chlorite, quartz, and mica content, generally dominates the till mineral constituent, according to X-ray plots. It is concluded that bedrock unit D (the weakest unit), supplies the major fine matrix constituent of the tills.

The excavation of large boulders and their subsequent deposition down-valley 0.5 to 2.0 km from their source clearly indicates that not all materials are greatly reduced in bulk. The mode of transport of these boulders were probably englacial and/or basal according to photographic records of the glacier terminus (Figures 2 a, b, c, and d). Boulders may have had a significant role to play in the erosion of bedrock units as evidenced by the abundance of striations, chatter marks and deep gouges over virtually all bedrock units. However, there is no way to tell from these erosional forms the size of eroding sediment.

The most influential bedrock characteristics in determining the number, and size of excavated boulders appear to be bedding joints. The strike-dip relationship also influences the degree to which bedrock units are plucked and excavated. With beds striking approximately north-south and dipping steeply west between 34° and 54°, overlying ice movement was at approximately 45° to strike. The situation has been reported to enhance rock fracture (Lama and Vutukuri 1978).

5.3 Bedrock Surface Morphology: Bedrock Roughness

Bedrock roughness was examined quantitatively and qualitatively at the micro-scale, and qualitatively at the meso-scale. From the slope-frequency distributions and trend surface analysis of individual bedrock unit plots, bedrock roughness can be ranked at the micro-scale.

Table 13 summarizes the ranking of bedrock unit plots with reference to slope frequency distribution and the measure of 'ideal plane' configuration (correlation coefficient of the trend surface analysis; 3rd order polynomial). The ranking ranges from 1 to 7. In the slope frequency ranking, the range represents a tendency towards a rough or non-uniform surface. A similar ranking for the trend surface ideal plane refers to 1 as being closest to resembling an ideal plane and 7 the least similar. In this discussion the ranking is a measure of relative roughness. At the meso-scale, a qualitative visual assessment was used in the ranking of bedrock unit profiles.

At the micro-scale, the slope frequency ranking has bedrock unit E as the smoothest closely followed by bedrock unit D, and unit G as the roughest. The 'ideal plane' ranking identifies bedrock units A and C as closely resembling the ideal plane configuration (surface) with unit B having the least resemblance.

Table 13

Ranking of Bedrock Units with Reference to Slope Frequency Distributions and Ideal Plane Configuration at the Micro-Scale and a Qualitative Assessment of the Degree of Horizontal Levelness at the Meso-Scale.

	micro-scale		meso-scale
	Slope Frequency Distribution	Correlation Coefficient (Ideal Plane Configuration)	Qualitative Assessment
Bedrock Unit A	6	1	3
Bedrock Unit B	3	6	6
Bedrock Unit C	5	1	2
Bedrock Unit D	2	4	1
Bedrock Unit E	1	2	7
Bedrock Unit F	4	3	4
Bedrock Unit G	7	5	5

Slope frequency ranking of 1 to 7 indicates a tendency towards a more uniform, levelled relief (surface) and least uniform levelled relief (smoothest and roughest) respectively. A similar ranking for the trend surface 'ideal plane' configuration (correlation coefficient) refers to 1 as being closest to resembling an 'ideal plane' and 7 the least similar. This table therefore infers that the ranking of 1 to 7 is a measure (relative) of smoothest to roughest respectively.

The slope frequency ranking ranks the units as to their resemblance to a levelled, horizontal surface. This, however, does not imply that a surface such as this is the smoothest in terms of ice sliding. As discussed in the trend surface analysis, a surface that resembles a 3rd order polynomial plane best typifies an ideal sliding surface at the micro-scale. Therefore, it would not be appropriate that the two rankings be added to produce an overall roughness ranking. The slope frequency rating may best be interpreted as providing an insight as to whether a bedrock unit was eroded evenly (i.e. was there preferential erosion on a particular bedrock unit, and was this the result of glacier ice characteristics or the actual physical characteristics of the bedrock unit?).

At the meso-scale, the ranking of a levelled horizontal surface has bedrock unit D as the smoothest with bedrock unit E the roughest. Again this does not imply that this qualitative measure of roughness can be interpreted as a measure of an ideal sliding surface at this scale.

A visual interpretation of bedrock smoothness at the meso-scale and a quantitative assessment of slope frequencies at the micro-scale show no apparent similarities in bedrock roughness or degree of levelness at these two scales. At the micro-scale, bedrock unit E is the smoothest (most horizontally level) and G is the roughest (least horizontally level); at the meso-scale, bedrock unit D is the smoothest and E is the roughest.

5.4 Bedrock Unit Strength and Morphology

The results presented in Tables 12 and 13 clearly indicate that bedrock unit D is the weakest and one of the smoothest bedrock units, while bedrock unit G is the strongest and roughest unit. The significance of this relationship is that over a given time period, weaker bedrock units erode to a smoother bedform as compared to stronger units which erode to a relatively rougher bedform.

Although roughness/smoothness or degree of horizontal levelness may be an indication of erosion (preferential) on bedrock units, they probably do not adequately describe a bedform in which sliding is enhanced and erosion retarded. It is more probable that each bedrock unit has its own unique bed configuration which enhances sliding and retards erosion. The morphology of this bed appears to be influenced by physical and chemical characteristics inherent in the bedrock (such as those identified in Table 12) as well as variations in basal ice conditions.

In summation, no comprehensive description of roughness and erosion can be expressed in simple numerical form. It is concluded that roughness and erosion are, in mathematical terminology, vectors rather than scalar quantities (Stone and Dungundji 1965). In any assessment of erodibility of a given bedrock unit, the extent of each physical and chemical characteristic of the unit must be specified. Through the qualitative and quantitative processes, some understanding of the resultant bedrock roughness may be achieved. Basically, bedrock units are more erodible or rougher only with respect to some characteristics. Both erodibility and roughness are in fact relative terms.

5.5 Bedrock Morphology and Glacial Flow: Concluding Remarks

It has been hypothesized in this thesis that ideal plane (surface) configurations of bedrock units need not necessarily be horizontally level to enhance glacial flow. Based on physical and chemical characteristics of bedrock types, and of glacier ice conditions at the ice-rock interface, each bedrock type will seek out its own particular configuration (ideal sliding surface) which enhances flow and retards erosion. The erodibility of a bedrock unit may be reflected in its morphological variation from an ideal plane or surface, although in this study all bedrocks appear to be of the similar magnitude of erodibility. Any differences between them appear to be compensated for by each unit assuming distinct micro surface characteristics.

Thus, each bedrock type probably has an affect on basal ice conditions and ice flow. However, where various bedrock types are in close proximity to one another, individual bedrock affects on ice flow may be masked by collective properties. Bedrock morphologies (prior to and during glacial activity) have a major influence in determining basal ice conditions and glacial flow.

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Appendix A.

Carbonic Acid Solubility

Since natural limestone solution is accomplished by H_2CO_3, a solubility test employing a weak solution of H_2CO_3 (pH 4.0) was undertaken. It was reasoned that this procedure would reflect the natural conditions of limestone solution. No reference could be found for similar tests on powdered rock samples. A procedure was designed which enabled an efficient analysis under controlled conditions (Yoxall, per. comm., Strong, per. comm.).

The procedure consists of adding a known amount of less than 40 powdered bedrock sample (5 gm) to a solution of H_2CO_3 (50 ml); pH 4.0. Samples and solutions are controlled by placing them in a freezer at 4.0 to 5.0° C. A standard solution of H_2CO_3 was made and monitered (pH and temp.).

Temperature, pH and concentration of $CACO_3$ (mg/L) readings were taken at 1 hr., 2 hrs., 7 hrs., 19 hrs., 31 hrs., and 43 hrs. A conductivity metre facilitated the determination of mg/L of $CACO_3$. Care was taken to assure samples were covered and kept at a constant temperature of 4.0 to 5.0° C at all times.

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Appendix B

Slope Frequency Distribution Computer List

```
type vecmak.for

C
C
C
C
C
C
C
      DIMENSION R(51,51),S(51,50),T(50,51)
C
      OPEN(UNIT=1,NAME='TOTH.DAT',ERR=999)
      OPEN(UNIT=3,NAME='H.DAT',TYPE='OLD',READONLY,ERR=999)
      OPEN(UNIT=5,NAME='kb:')
C
      READ(3,20,END=999) ((R(I,J),I=1,51),J=1,51)
 20   FORMAT(18F4.2)
C     WRITE(5,111) ((R(I,J),I=1,51),J=1,51)
C 111 FORMAT(18(' ',F5.2,' '))
      D=2.0
      DO 100 I =1,51
      DO 100 J=1,50
 100  S(I,J)=ABS((R(I,J)-R(I,J+1))/D)
      DO 200 J=1,51
      DO 200 I=1,50
 200  T(I,J)=ABS((R(I,J)-R(I+1,J))/D)
C
C
C
      BIG=-9999.9
      DO 30 I=1,51
      DO 40 J=1,50
      WRITE(1,135) S(I,J)
      IF ( S(I,J) .GT. BIG) BIG = S(I,J)
 135  FORMAT(F5.2)
 40   CONTINUE
 30   CONTINUE
C
C
C
      BIG2 = -9999.99
      DO 60 I=1,50
      DO 70 J=1,51
      WRITE(1,135)T(I,J)
      IF ( T(I,J) .GT. BIG2) BIG2 = T(I,J)
 70   CONTINUE
 60   CONTINUE
 999  CLOSE(UNIT=1)
      CLOSE(UNIT=3)
      WRITE(5,145) BIG
 145  FORMAT(' THE BIGGEST NUMBER IN SVEC.DAT IS ',F5.2)
      WRITE(5,155) BIG2
 155  FORMAT(' THE BIGGEST NUMBER IN TVEC.DAT IS ',F5.2)
      RETURN
      END

Ready
```

```
type hcpy.for

c                 *************
c                 * HISTOGRAM *
c                 *************
      logical*1 stars(80),title(40),iparm,ismfil,iend
      integer ihist(300)
      virtual array(5500)
      data stars/80*'*'/
      implicit real*8 (a-h,o-z)
c
c
      type 20
 20   format(' file or kb: ?')
      accept 30,iparm
 30   format(a1)
      if(iparm .eq. 'K') goto 12
      type 21
 21   format(' output file ?'/)
      call assign(9,'out.dat',-1)
      goto 13
 12   call assign(9,'kb:',0)
 13   iparm='N'
c
 1    type 22
 22   format(' input file ?'/)
      call assign(8,'in.dat',-1)
      if(iparm .eq. 'Y') goto 2
 11   type 23
 23   format(' # of intvls')
      accept *, intvls
      type 24
 24   format(' lo - high range ?')
      accept *, ilo,ihi
c     type 25
c 25  format(' variable ?')
c     accept *,ivar
      type 26
 26   format(' conversion ?')
      accept *,icvr
c     type 328
c 328 format(' IB8 status ?')
c     accept *,ibnotib
      top=ihi
      offset=ilo
      call cvr(top,icvr)
      call cvr(offset,icvr)
      sizint=(top-offset)/float(intvls)
c
 2    type 27
 27   format(' title ?')
      accept 28, title
 28   format(40a1)
```

```
c
c       initialize variables
c
        do 29 i=1,300
        ihist(i)=0
   29   continue
        sum=0
        num=1
        izero=0
        smisv=0
        imisv=0
        ismall=0
        large=0
        z=0.0
        sqr=0.0
        cube=0.0
        forth=0.0
        sx2=0.0
        sx3=0.0
        sx4=0.0
c
c
    3   read(8,*,end=99) array(num)
c   10  format(f5.2)
c       type 329,  ib8,(array(num,j),j=1,3)
c  329  format(3x,i1,3f6.0)
        x = array(num)
        if(x .gt. 0.0) goto 6
        if(x .ne. 0.0) goto 7
        izero=izero+1
c
c       disable the following 'goto 5' stmt when includinu zero
c       readings in the calculations.
c
c       goto 5
        if(ilo .gt. 0)ismall=ismall+1
        if(ilo .eq. 0)ihist(1)=ihist(1)+1
    7   if(x .eq. -99.0) imisv=imisv+1
        goto 5
c
    6   call cvr(x,icvr)
        array(num)=x
        sum = sum + x
         x2=x*x
         sx2=sx2+x2
         x3=x2*x
         sx3=sx3+x3
         x4=x3*x
         sx4=sx4+x4
        if((x.ge.offset).and.(x.le.top)) goto 4
        if(x.lt.offset) ismall=ismall+1
        if(x.gt.top) large=large+1
        goto 5
```

```
c
      do 60 i=1,intvls
      fhi=offset+(sizint*i)
      flo=fhi-sizint+0.01
      percnt=(ihist(i)/fn)*100.0
      ilnth = int(percnt*50.0/100.0)
c
      if(ihist(i) .GT. 0.0) goto 31
      write (9,53) ihist(i),percnt,flo,fhi
53    format(5x,i4,' readings (',f6.2,'% )  in the intvl ',
     !       f7.2,' to ',f7.2,'  : ',80A1)
      goto 60
31    write (9,53) ihist(i),percnt,flo,fhi,(stars(j),j=1,ilnth)
60    continue
c
c     Display Statistics
c
      write (9,65)
65    format(//20x,'STATISTICS'//)
      write (9,66) nt,sum,skewa,skew
      write (9,67) n,xm,sesk
      write (9,68) imisv,var2,xkurta,xkurt
      write (9,69) sd,seku
c
c
66    format(6x,'total entries : ',i5,5x,'sum  : ',f8.3,5x,
     !       ' skew  : ',2f9.4)
67    format(2x,'valid test scores : ',i5,5x,'mean : ',f8.3,5x,
     !       ' se    : ',f8.4)
68    format(5x,'missing values : ',i5,5x,'var  : ',f8.3,5x,
     !       ' kurt  : ',2f9.4)
69    format(32x,'s.d. : ',f8.3,5x,' se    : ',f8.4)
c
c
      write (9,70)
70    format(5x,'Statistics include readings above and below ',
     1       'the said limits.'/)
      write (9,71) ismall,offset
      write (9,72) large,top
      write (9,76) izero
71    format(5x,i5,' readings below limit of ',f8.2)
72    format(5x,i5,' readings above limit of ',f8.2)
76    format(5x,i5,' actual zero value/s encountered ( included in ',
     1       'statistics)',6(/))
c
c
      type 73
73    format(//' end??')
      accept 30, iend
      if(iend .eq. 'Y') goto 199
      rewind 8
      type 74
74    format(' Same parameters ?')
      accept 30,iparm
      type 75
```

hänsyn till glaciala riktnings - element och periglaciala frostfenomen."
Medd. fr. Lunds Univ. Geogr. Inst., Auh. 30.


```
    4  n = int(((x-offset)/sizint) + 0.99)
       ihist(n)=ihist(n)+1
    5  num=num+1
       goto 3
c
c
c      All data read in. Set up to calculate statistics
c
   99  nt=num-1
       n=num-1-imisv
       fn=float(n)
       sum2=sum**2.0
       sum3=sum**3.0
       sum4=sum**4.0
       xm=sum/fn
       xm2=xm**2.0
       xm3=xm**3.0
       xm4=xm**4.0
       var2=(sx2 - fn*xm2)/(fn -1.0)
       var = sqr/(fn-1.0)
       sd  = dsqrt(var2)
c
c      loop to calculate y minus ybar
c
       do 40 i=1,nt
       x=array(i)
         z=(x-xm)
         z2=z**2
         sqr=sqr+z2
         sz=z/sd
         z3=sz**3
         cube=cube+z3
         z4=sz**4
         forth=forth+z4
   40  continue
c
c      calculate statistics
c
       skewa=(cube*fn)/((fn-1.0)*(fn-2.0))
       xkurta=forth*((fn*(fn+1.0))/((fn-1.0)*(fn-2.0)*(fn-3.0))) -
    !        (3*(fn-1.0)*(fn-1.0))/((fn-2.0)*(fn-3.0))
       skew1=(sx3-(3.0*xm*sx2)+(3.0*xm2*sum))/fn-xm3
       skew2=((sx2-(fn*xm2))/(fn-1.0))**1.5
       skew=skew1/skew2
       sesk = dsqrt((6.0*fn*(fn-1.0))/((fn-2.0)*(fn+1.0)*(fn+3.0)))
       xkurt1=((sx4-4.0*xm*sx3+6.0*xm2*sx2)-4.0*xm3*sum)/fn+xm4
       xkurt2=((sx2-fn*xm2)/(fn-1.0))**2
       xkurt=xkurt1/xkurt2 - 3
       sekul = (24*fn*((fn-1.0)**2))/((fn-3.0)*(fn-2.0))
       seku = dsqrt(seku1/((fn+3)*(fn+5)))
c
c      Display Histogram
c
       write (9,50) title
   50  format(//10x,40a1///)
```

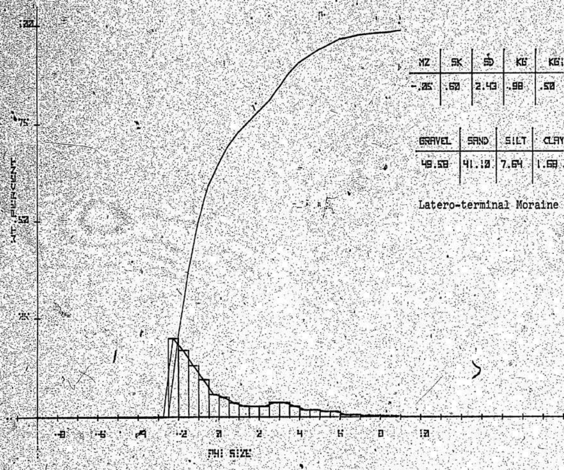
```
75   format(' Same file ?')
     accept 30,ismfil
     if((ismfil .eq. 'Y') .and. (iparm .eq. 'Y')) goto 2
     if(ismfil .eq. 'Y') goto 11
     call close(8)
     goto 1
c
c
199  call close(8)
     call close(9)
     stop
     end
```

1

Ready

Appendix C

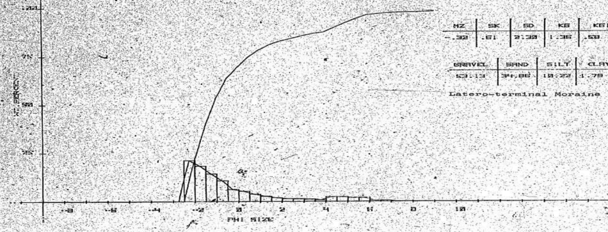
Latero-Terminal Moraine Grain Size Distribution Curves



MZ	SK	SD	KG	KG'
-.05	.60	2.43	.98	.50

GRAVEL	SAND	SILT	CLAY
49.58	41.10	7.64	1.68

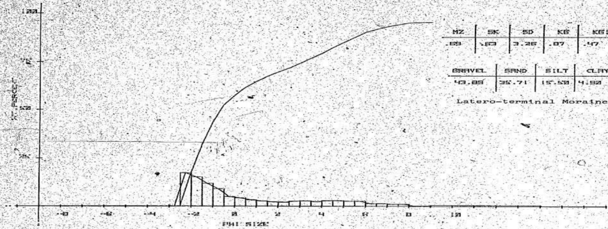
Latero-terminal Moraine 1



MZ	SK	SD	KG	KG'
-.32	.61	2.30	1.36	.58

GRAVEL	SAND	SILT	CLAY
53.13	34.86	10.22	1.79

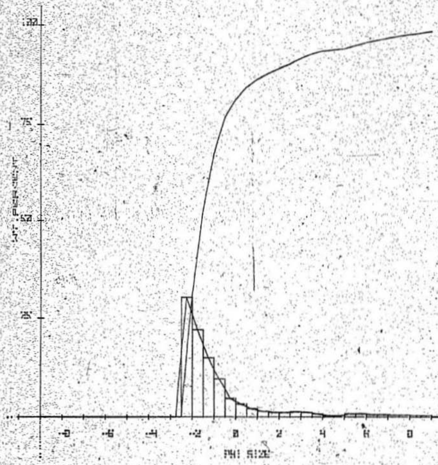
Latero-terminal Moraine 2



MZ	SK	SD	KG	KG'
.69	.83	3.26	.87	.47

GRAVEL	SAND	SILT	CLAY
43.89	35.71	15.50	4.90

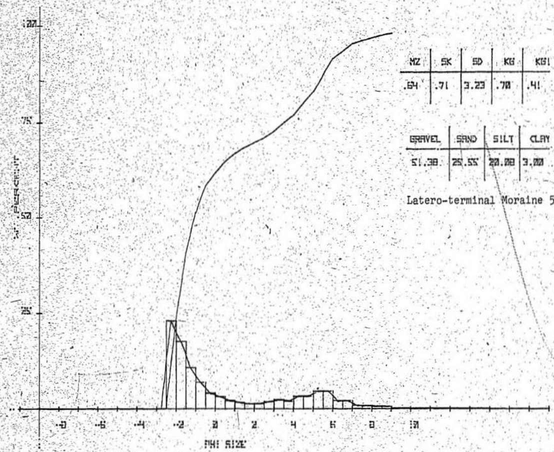
Latero-terminal Moraine 3



MZ	SK	SD	KG	KG'
-1.13	.64	1.86	2.16	.68

GRAVEL	SAND	SILT	CLAY
67.35	26.25	4.29	2.11

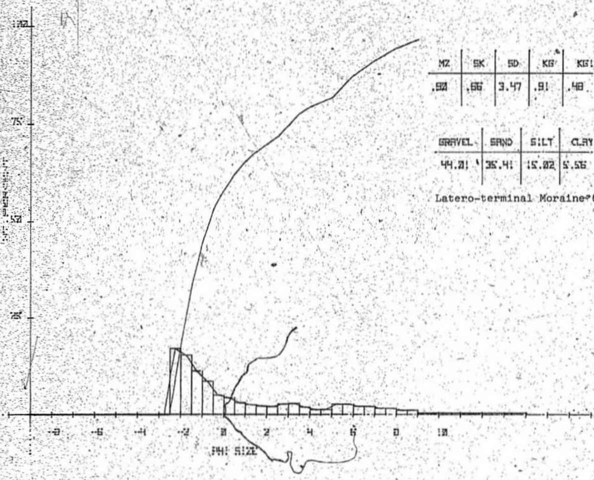
Latero-terminal Moraine 4



MZ	SK	SD	KG	KG'
.64	.71	3.23	.70	.41

GRAVEL	SAND	SILT	CLAY
51.38	25.55	20.08	3.00

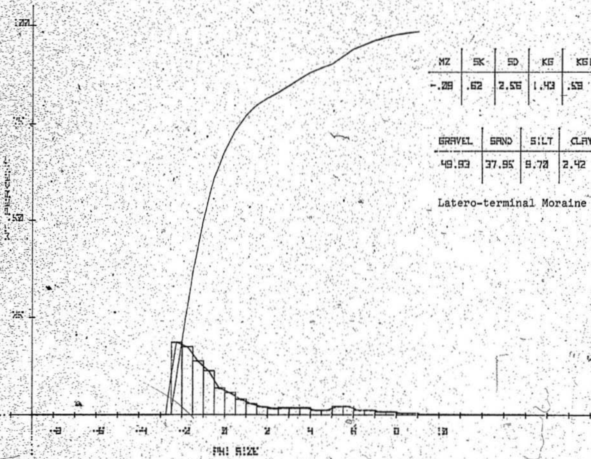
Latero-terminal Moraine 5



MZ	SK	SD	KG'	KG!
.50	.66	3.47	.91	.48

GRAVEL	SAND	SILT	CLAY
44.01	35.41	15.02	5.56

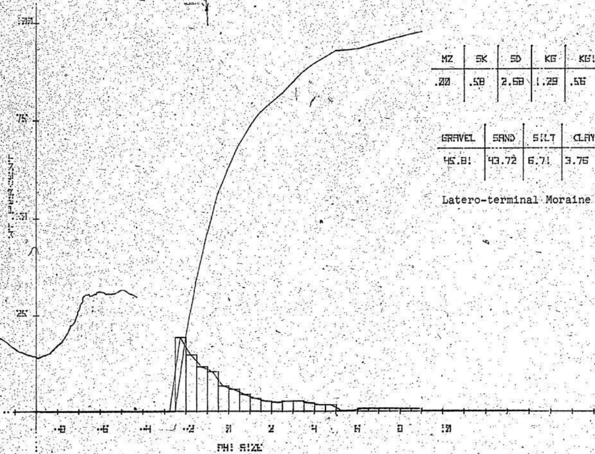
Latero-terminal Moraine 6



MZ	SK	SD	KG	KG1
-.09	.62	2.56	1.43	.59

GRAVEL	SAND	SILT	CLAY
49.93	37.95	9.70	2.42

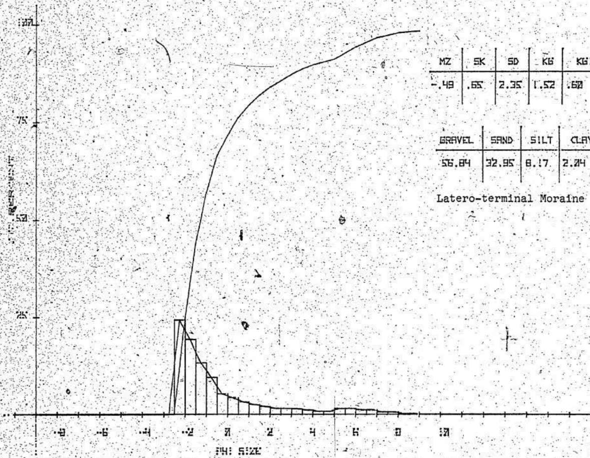
Latero-terminal Moraine 7



MZ	SK	SD	KG	KG'
.00	.58	2.68	1.29	.56

GRAVEL	SAND	SILT	CLAY
45.81	43.72	6.71	3.76

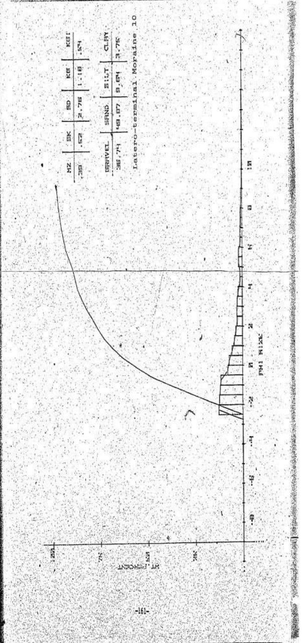
Latero-terminal Moraine 8

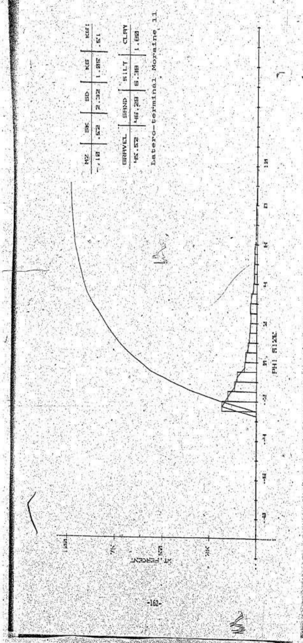


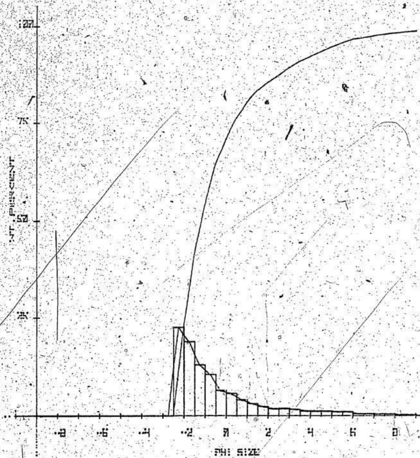
MZ	SK	SD	KG	KG1
-.49	.65	2.35	1.52	.60

GRAVEL	SAND	SILT	CLAY
56.84	32.95	8.17	2.04

Latero-terminal Moraine 9


MZ | SK | SD | KG | KGI
.39 | .52 | 2.76 | 1.18 | .54
GRAVEL | SAND | SILT | CLAY
36.74 | 49.87 | 9.64 | 3.75
Latero-terminal Moraine 10
PHI SIZE
WT. PERCENT
100
75
50
25
-6
-4
-2
0
2
4
6
8
10

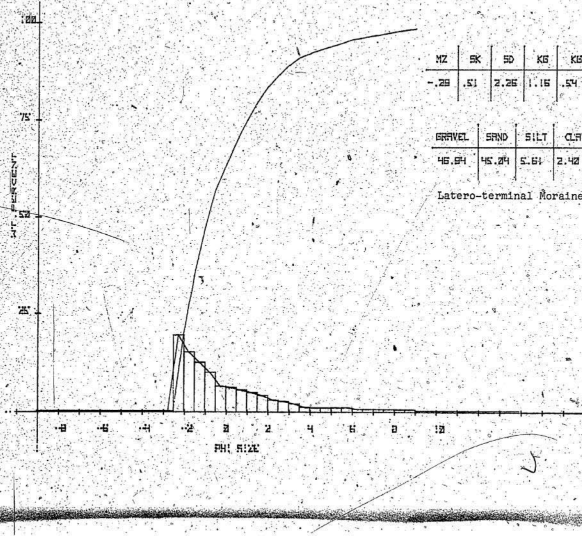

MZ | SK | SD | KG | KGI
-.10 | .52 | 2.32 | 1.38 | .51
GRAVEL | SAND | SILT | CLAY
45.52 | 46.29 | 6.39 | 1.80
Latero-terminal Moraine 11
PHI SIZE
WT. PERCENT
100
75
50
25
-6
-4
-2
0
2
4
6
8
10



MZ	SK	SD	KG	KG1
-.56	.58	2.09	1.35	.58

GRAVEL	SAND	SILT	CLAY
54.18	38.04	6.07	1.71

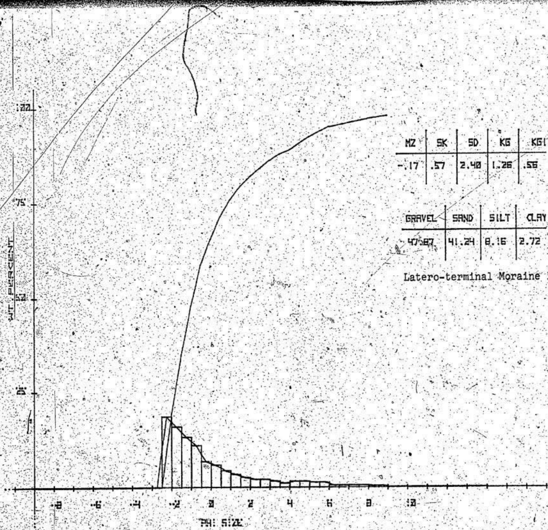
Latero-terminal Moraine 12



MZ	SK	SD	KG	KG'
-.29	.51	2.26	1.16	.54

GRAVEL	SAND	SILT	CLAY
46.94	45.04	5.61	2.40

Latero-terminal Moraine 13



MZ	SK	SD	KG	KGI
-.17	.57	2.40	1.26	.56

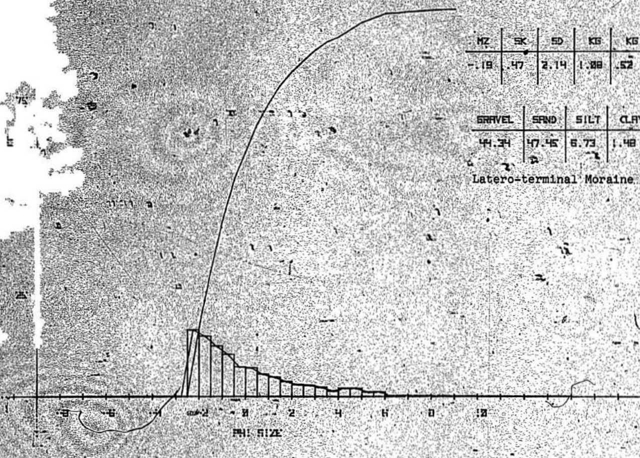
GRAVEL	SAND	SILT	CLAY
47.87	41.24	8.16	2.72

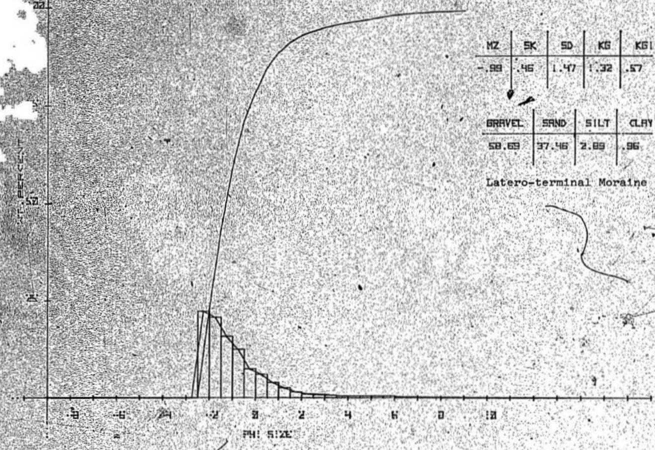
Latero-terminal Moraine 14

MZ	SK	SD	KG	KGI
-.19	.47	2.14	1.08	.52

GRAVEL	SAND	SILT	CLAY
44.34	47.45	6.73	1.48

Latero-terminal Moraine 15

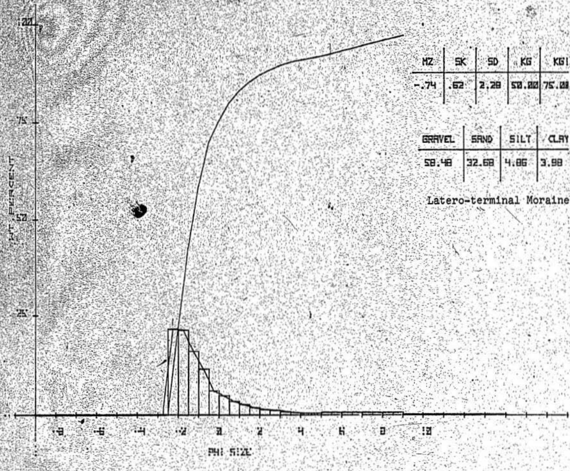




MZ	SK	SD	KG	KG1
-.99	.46	1.47	1.32	.57

GRAVEL	SAND	SILT	CLAY
58.69	37.46	2.89	.96

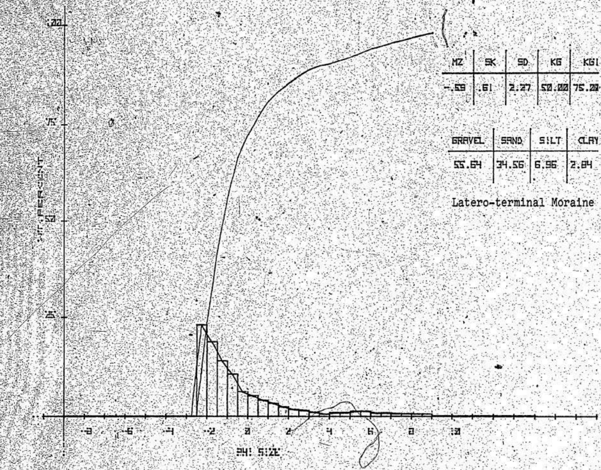
Latero-terminal Moraine 16



MZ	SK	SD	KG	KGI
-.74	.62	2.28	50.00	75.00

GRAVEL	SAND	SILT	CLAY
58.48	32.68	4.86	3.98

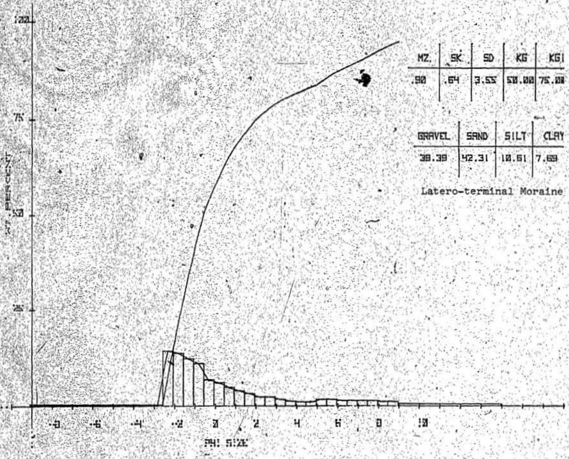
Latero-terminal Moraine 17



MZ	SK	SD	KG	KGI
-.59	.61	2.27	50.00	75.20

GRAVEL	SAND	SILT	CLAY
55.64	34.56	6.96	2.84

Latero-terminal Moraine 18

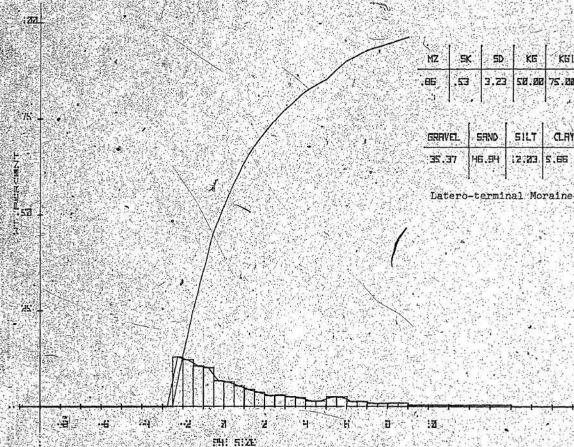


MZ.	SK	SD	KG	KGI
.90	.64	3.55	50.00	75.00

GRAVEL	SAND	SILT	CLAY
39.39	42.31	10.61	7.69

Latero-terminal Moraine 19

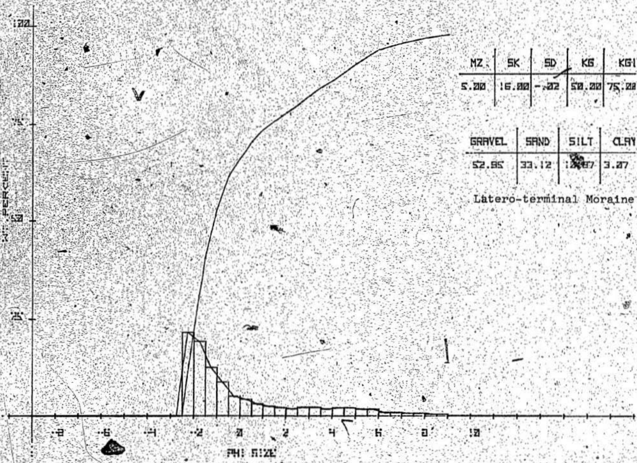
-156-
WT. PERCENT



MZ	SK	SD	KG	KG1
.86	.53	3.23	50.00	75.00

GRAVEL	SAND	SILT	CLAY
35.37	46.94	12.03	5.66

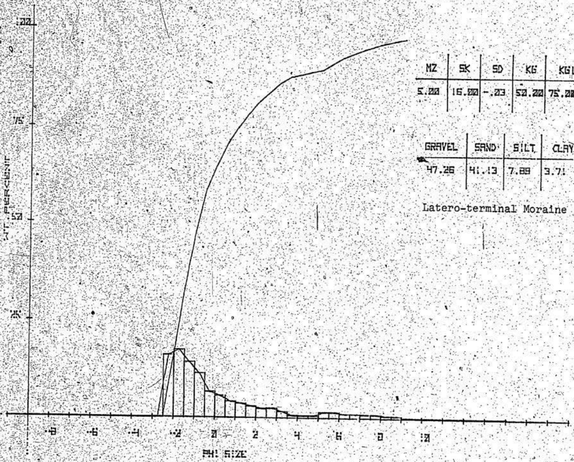
Latero-terminal Moraine 20



MZ	SK	SD	KG	KG1
5.00	16.00	-.02	50.00	75.00

GRAVEL	SAND	SILT	CLAY
52.95	33.12	[illegible]	3.07

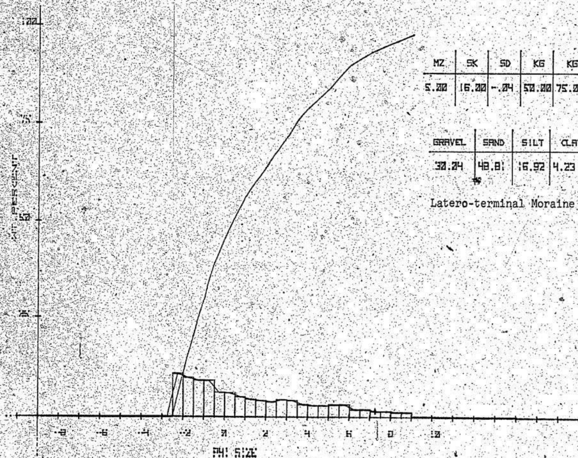
Latero-terminal Moraine 21



MZ	SK	SD	KG	KGI
5.00	16.00	-.03	50.00	75.00

GRAVEL	SAND	SILT	CLAY
47.26	41.13	7.89	3.71

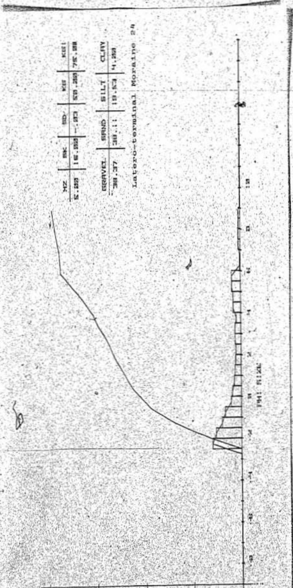
Latero-terminal Moraine 22



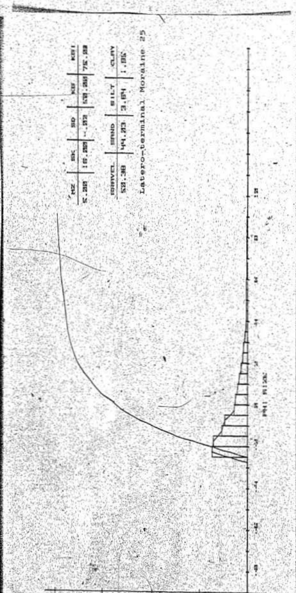
MZ	SK	SD	KG	KG'
5.00	16.00	-.04	50.00	75.00

GRAVEL	SAND	SILT	CLAY
30.04	48.81	16.92	4.23

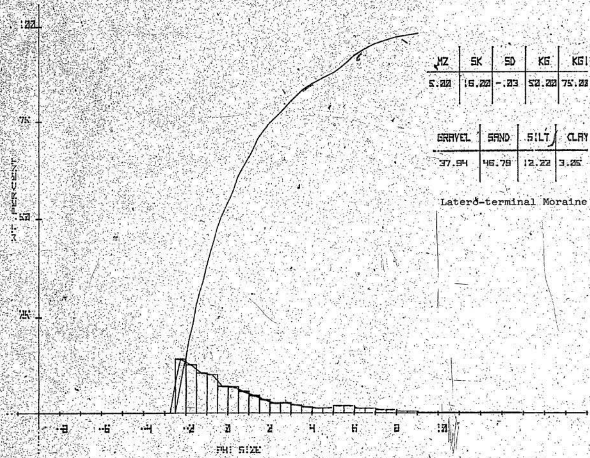
Latero-terminal Moraine 23



MZ | SK | SD | KG | KGI
5.00 | 16.00 | -.03 | 50.00 | 76.00
GRAVEL | SAND | SILT | CLAY
38.37 | 38.11 | 18.53 | 4.28
Latero-terminal Moraine 24
PHI SIZE
WT.PERCENT
100
75
50
25



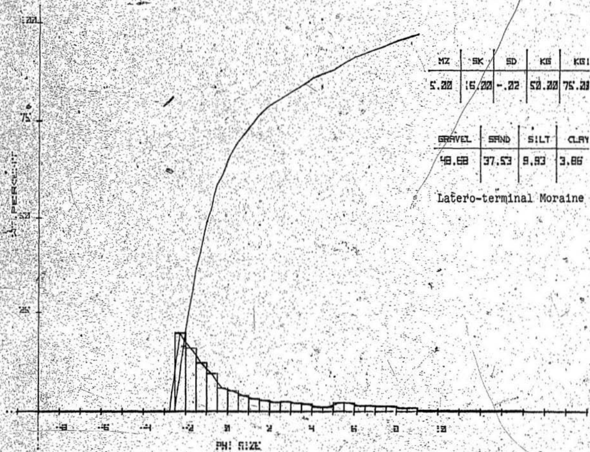
MZ | SK | SD | KG | KGI
5.00 | 16.00 | -.02 | 50.00 | 76.00
GRAVEL | SAND | SILT | CLAY
50.38 | 44.23 | 3.64 | 1.85
Latero-terminal Moraine 25
PHI SIZE
WT.PERCENT
100
75
50
25



MZ	SK	SD	KG	KGI
5.00	16.00	-.03	50.00	75.00

GRAVEL	SAND	SILT	CLAY
37.94	46.79	12.22	3.05

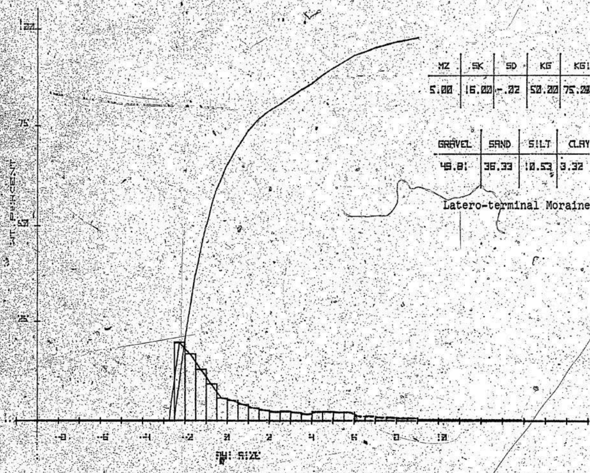
Latero-terminal Moraine 26



MZ	SK	SD	KG	KG1
5.00	16.00	-.02	50.00	75.00

GRAVEL	SAND	SILT	CLAY
48.68	37.53	9.93	3.86

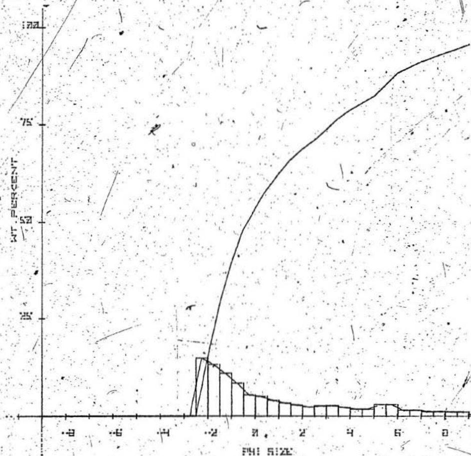
Latero-terminal Moraine 27



MZ	SK	SD	KG	KG'
5.00	16.00	-.02	50.00	75.20

GRAVEL	SAND	SILT	CLAY
49.81	36.33	10.53	3.32

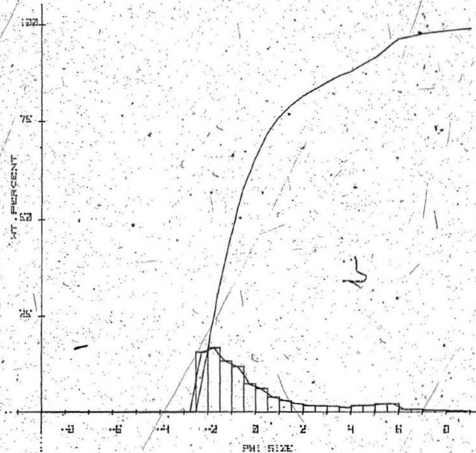
Latero-terminal Moraine 28



MZ	SK	SD	KG	KG1
5.00	16.00	-.03	50.00	75.00

GRAVEL	SAND	SILT	CLAY
39.38	39.44	14.82	6.35

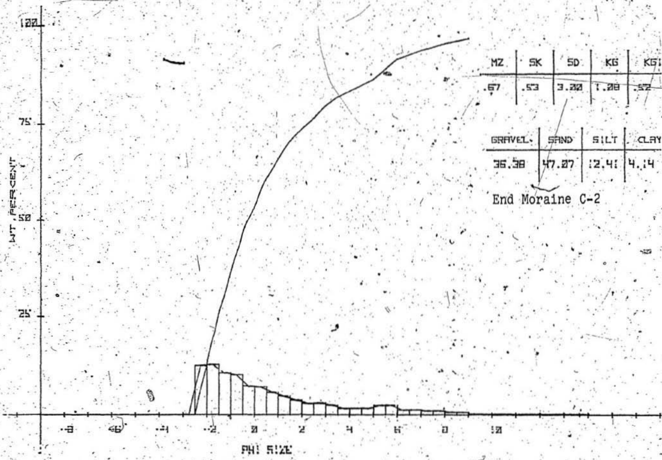
Latero-terminal Moraine 29



MZ	SK	SD	KG	KG'
-.05	.57	2.40	1.28	.56

GRAVEL	SAND	SILT	CLAY
46.15	41.73	10.55	1.57

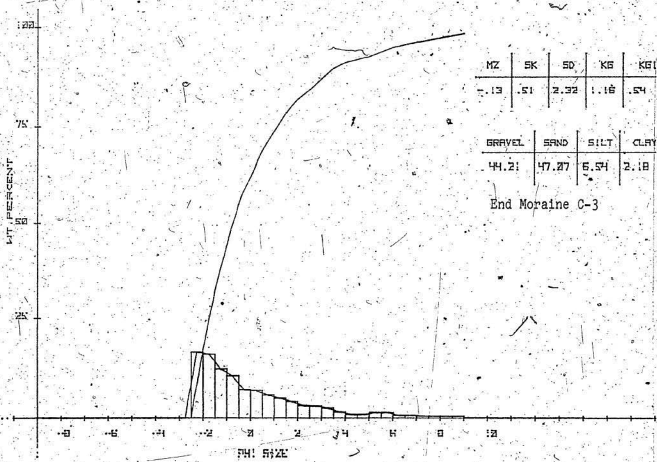
End Moraine C-1



MZ	SK	SD	KG	KG'
.67	.53	3.00	1.08	.52

GRAVEL	SAND	SILT	CLAY
36.38	47.07	12.41	4.14

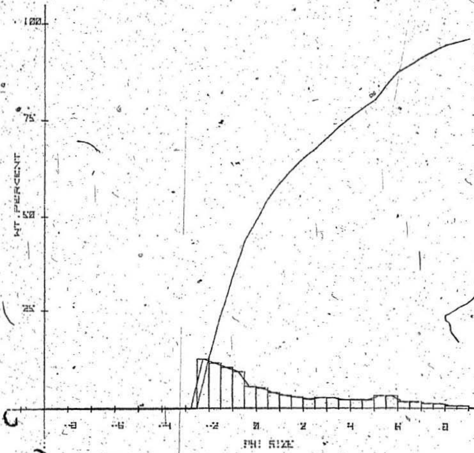
End Moraine C-2



MZ	SK	SD	KG	KGI
-.13	.51	2.32	1.16	.54

GRAVEL	SAND	SILT	CLAY
44.21	47.07	6.54	2.18

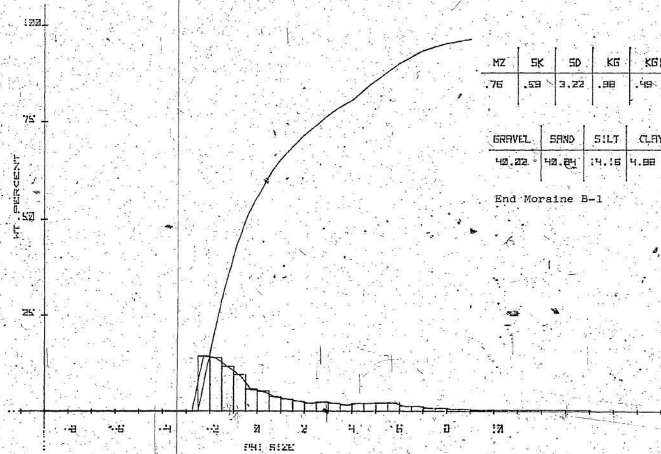
End Moraine C-3



MZ	SK	SD	KG	KG1
1.22	.51	3.41	.82	.45

GRAVEL	SAND	SILT	CLAY
34.13	41.93	18.68	5.27

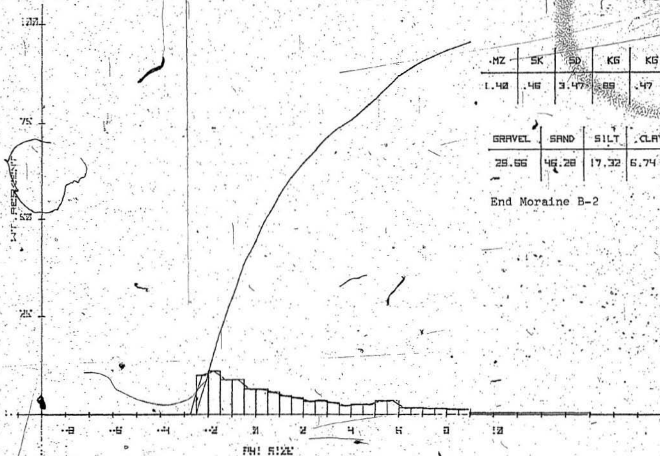
End Moraine C-4



MZ	SK	SD	KG	KG'
.76	.59	3.22	.98	.49

GRAVEL	SAND	SILT	CLAY
40.02	40.84	14.16	4.98

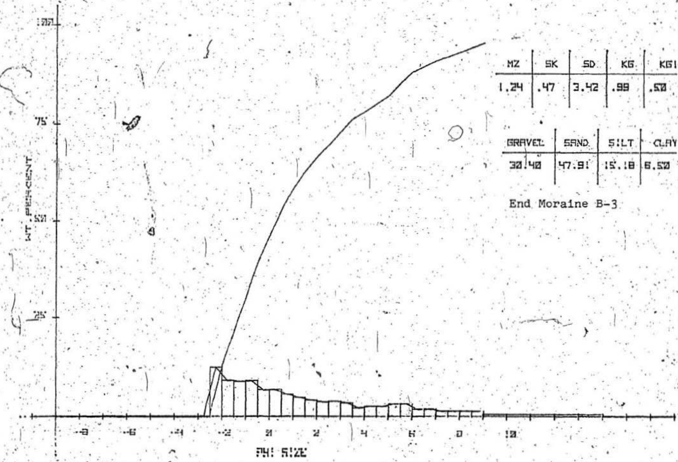
End Moraine B-1



MZ	SK	SD	KG	KGI
1.40	.46	3.47	.89	.47

GRAVEL	SAND	SILT	CLAY
29.66	46.28	17.32	6.74

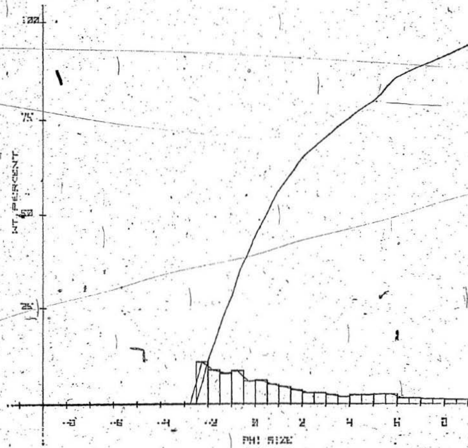
End Moraine B-2



MZ	SK	SD	KG	KG1
1.24	.47	3.42	.99	.50

GRAVEL	SAND	SILT	CLAY
30.40	47.91	15.18	6.50

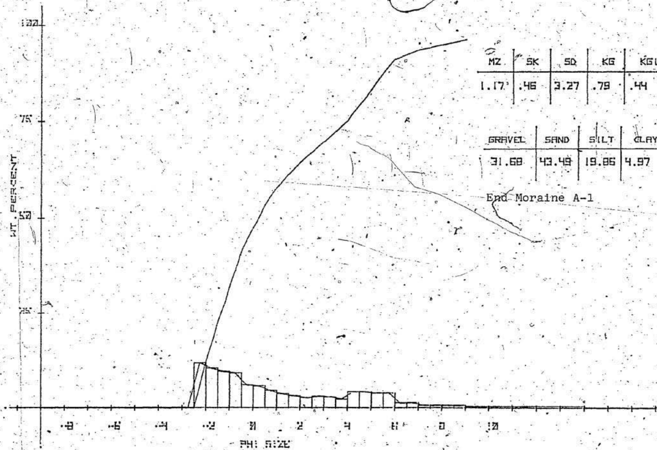
End Moraine B-3



MZ	SK	SD	KG	KGI
1.49	.48	3.69	.95	.49

GRAVEL	SAND	SILT	CLAY
28.44	45.62	16.46	8.48

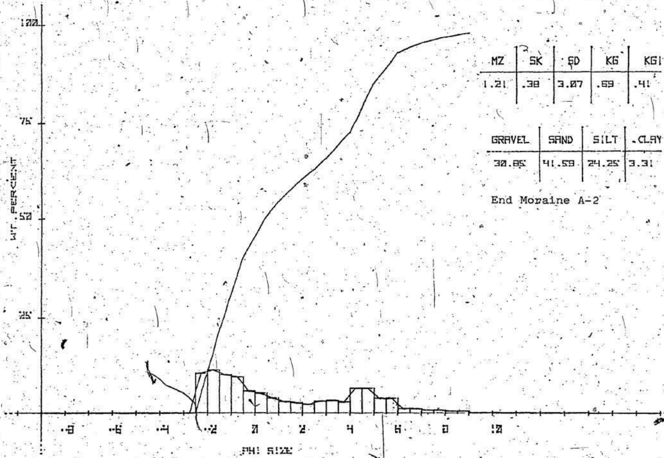
End Moraine B-4



MZ	SK	SD	KG	KGI
1.17	.46	3.27	.79	.44

GRAVEL	SAND	SILT	CLAY
31.68	43.49	19.86	4.97

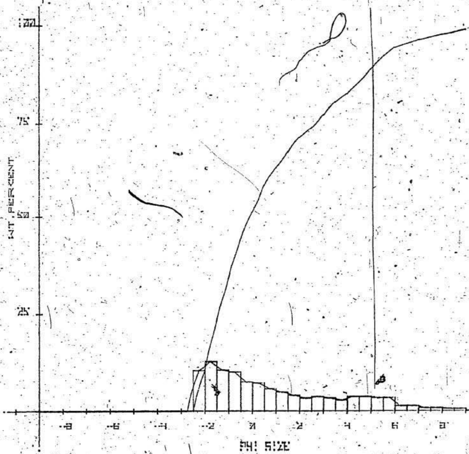
End Moraine A-1



MZ	SK	SD	KG	KGI
1.21	.38	3.07	.69	.41

GRAVEL	SAND	SILT	CLAY
30.85	41.59	24.25	3.31

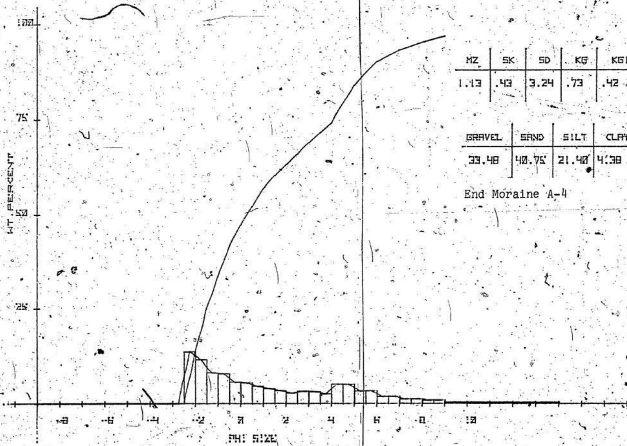
End Moraine A-2



MZ	SK	SD	KG	KGI
.81	.47	2.84	.86	.46

GRAVEL	SAND	SILT	CLAY
34.05	47.58	15.81	2.16

End Moraine A-3



MZ	SK	SD	KG	KGI
1.13	.43	3.24	.73	.42

GRAVEL	SAND	SILT	CLAY
33.48	40.75	21.40	4.38

End Moraine A-4

Appendix D

Statistics on SYNVU Trend Surface by Bedrock Type

	Bedrock Unit						
	A	B	C	D	E	F	G
Standard Deviation (σ)	0.97	0.52	0.96	0.90	0.94	0.93	0.83
Total Variation ($3\sigma^2$)	0.94	0.27	0.92	0.81	0.88	0.86	0.69
Correlation Coefficient (r)	0.98	0.72	0.98	0.95	0.97	0.96	0.91
Coefficient of Determination (r^2)	1.63	2.00	1.41	1.34	1.61	2.00	1.33