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CONDITIONS ASSOCIATED WITH FROST ACTION IN ROCKS:

A FIELD AND LABORATORY INVESTIGATION

by

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B.A.(Hons.), Durham University

A thesis submitted to the
Faculty of Graduate Studies and Research

in partial fulfilment of
the requirements for the degree of
Master of Arts

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September, 1985

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The undersigned recommend to the Faculty of Graduate Studies and Research acceptance of the thesis "Conditions Associated with Frost Action in Rocks: a Field and Laboratory Investigation" submitted by Michael John Hare, B.A., in partial fulfilment of the requirements for the degree of Master of Arts (Geography).

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ABSTRACT

Rock breakdown by frost action is largely controlled by temperature conditions, moisture availability and rock type. Time Domain Reflectometry (TDR) was used in the laboratory to examine the freezing of moisture in samples of Nisling Alaskite and Caen Limestone, and the role of climate was examined at a field site in the Ruby Range, Yukon Territory and through the use of a computer simulation.

TDR results indicated that over 40% of the initial moisture content of both rock types froze between 0°C and -2°C. Breakdown of the rock would occur when stresses developed by the freezing moisture exceed the tensile strength of the rock.

The field studies showed that although the general macroclimatic factors of latitude, altitude and aspect can be of considerable importance in determining the thermal regime of a rock, micro-factors may on occasion override them. Specifically, the moisture conditions and rock thermal properties are shown to affect frost penetration into the rock.

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TABLE OF CONTENTS

	page
ACCEPTANCE SHEET.....	ii
ABSTRACT.....	iii
ACKNOWLEDGEMENTS.....	iv
TABLE OF CONTENTS.....	v
LIST OF TABLES.....	vii
LIST OF FIGURES.....	ix
LIST OF PLATES.....	xi
 CHAPTER I: BASIS OF STUDY	
1.1 Introduction.....	1
1.2 Temperature-Moisture Relationships.....	3
1.3 Objectives of Study.....	4
1.4 Organisation of Thesis.....	6
 CHAPTER II: REVIEW OF LITERATURE	
2.1 Introduction.....	7
2.2 The Freezing of Moisture in Porous Media.....	8
2.3 The Role of Temperature in Frost Shattering.....	11
2.4 Investigations into the Role of Rock Type.....	22
2.5 Mechanics of Rock Shattering by Frost Action.....	25
2.6 Summary.....	34
 CHAPTER III: FIELD AREA DESCRIPTION	
3.1 Introduction.....	35
3.2 Geology of the Ruby Range.....	35
3.2.1 Mechanical Properties of Alaskite..	40
3.3 Quaternary History of the Ruby Range	44
3.4 Description of Field Site and Instrumentation.....	49
3.4.1 Field Site 1.....	51
3.4.2 Field Site 2.....	54
3.4.3 Field Site 3.....	58
3.5 Data Collection.....	62
 CHAPTER IV: THE FREEZING OF MOISTURE IN ROCKS	
4.1 Introduction.....	64
4.2 Freezing of Moisture in Rocks.....	64
4.2.1 Volumetric Expansion on Freezing...	65
4.2.2 Thermodynamic Aspects of Soil Freezing.....	67
4.3 Effects of Moisture and Freezing on Rock Strength.....	71
4.4 Measurement of Moisture in Rocks...	79

4.4.1	Field Measurement Techniques of Total Moisture Content.....	80
4.4.1.1	Neutron Moderation.....	80
4.4.1.2	Oven Drying of Field Samples.....	80
4.4.2	Laboratory Techniques for Measuring Unfrozen Moisture Content.....	83
4.4.2.1	Dilatometry.....	83
4.4.2.2	Differential Thermal Analysis.....	83
4.4.2.3	Suction-Moisture Content Tests....	84
4.4.2.4	Use of Ultrasound.....	85
4.5	Time Domain Reflectometry.....	85
4.5.1	The Dielectric Property and Water Content.....	86
4.6	Use of TDR on Freezing Rocks.....	90
4.7	Determination of the Freezing Characteristic Curve.....	93
4.8	Discussion of Results.....	100
4.9	Summary.....	104
CHAPTER V: CLIMATIC CONTROLS ON FROST SHATTERING		
5.1	Introduction.....	106
5.2	Factors Affecting the Thermal Regime of Rock.....	107
5.3	Field Observations of Rock Temperatures.....	108
5.3.1	General Climate.....	111
5.3.1.1	Altitude.....	112
5.3.1.2	Aspect.....	114
5.4	Computer Simulation Model.....	124
5.5	Computer Simulation of the Effects of Aspect.....	126
5.6	Effects of Surface Conditions.....	131
5.6.1	Surface Roughness and Presence of Cracks and Fissures.....	131
5.6.2	Moisture Conditions.....	134
5.6.3	Albedo.....	134
5.7	Thermal Properties of Rocks.....	135
5.7.1	Computer Simulation of Different Thermal Properties.....	140
5.7.1.1	Effects of Moisture Conditions...	140
5.7.1.2	Effects of Rock Type.....	143
5.8	Summary.....	147
CHAPTER VI: SUMARY AND CONCLUSIONS		
6.1	Introduction.....	148
6.2	Results.....	149
6.3	Other Considerations.....	154
6.4	Conclusions.....	155
BIBLIOGRAPHY.....		157

LIST OF TABLES

Table	Title	Page
2.1	Some Critical Freezing Temperatures and Measurement Methods.....	12
2.2	Field Observations of Susceptibility to Frost Shattering according to Rock Type.....	23
3.1	Mineral Composition of Alaskite Units	41
3.2	Mechanical Properties of Alaskite Units.....	41
3.3	Engineering Classification of Rocks...	43
3.4	Ranking of Alaskite Units Strength: Weathering Susceptibility.....	43
3.5	Late Pleistocene Glacial Chronology of the Northwestern St. Elias Mountains..	46
3.6	Thermistor Locations at Field Site 1..	53
3.7	Thermistor Locations at Field Site 2..	56
3.8	Thermistor Locations at Field Site 3..	60
4.1	Examples of the Hydraulic Conductivity of some Soils and Rocks.....	69
4.2	Field Rock Saturation Levels Determined by Oven Drying.....	82
4.3	Maximum Effective Temperature for Rock Breakdown.....	102
4.4	Thermal Contraction of Some Common Materials.....	102
5.1	Mean Rates of Cooling at Ruby Range Sites.....	115

- 5.2 Freeze-Thaw Oscillations on Different Aspects during Spring and Fall Months. 130
- 5.3 Typical Heat Capacities and Thermal Conductivity Measurements for Rocks and Minerals..... 137

LIST OF FIGURES

Figure	Title	Page
2.1	Freezing Characteristic Curves for Four Fine-grained Soils.....	9
2.2	Freezing Cycles Used in Experimental Studies.....	18
2.3	Hydraulic Pressure Theory of Rock Breakdown..	33
3.1	The Ruby Range Field Area.....	37
3.2	Geological Map of the Ruby Range.....	38
3.3	Glacial Limits in the Ruby Range.....	48
3.4	Morphological Map of Site 1.....	52
3.5	Morphological Map of Site 2.....	55
3.6	Morphological Map of Site 3.....	59
4.1	Reduction of Hydraulic Conductivity with Temperature.....	70
4.2	Tensile Strength as a Function of Water Content for Berea Sandstone and Indiana Limestone.....	73
4.3	Tensile Strength of Indiana Limestone with Temperature.....	74
4.4	Tensile Strength of Barre Granite with Temperature.....	75
4.5	Temperature-Time Dependence of Uniaxial Compressive Strength for Ottawa Sand.....	78
4.6	Typical TDR trace in Wet Soil.....	87
4.7	Probe Configuration of Coaxial Line for TDR Measurements.....	91
4.8	Suction-Moisture Content Curve for Caen Limestone.....	95
4.9	Freezing Characteristic Curve for Caen Limestone.....	96

4.10	Freezing Characteristic Curve for Nisling Alaskite.....	97
4.11	Freezing Characteristic Curves for Caen Limestone Exhibiting Hysteresis.....	98
5.1	Air and Rock Temperatures, Site 1.....	109
5.2	Air and Rock Temperatures, Site 2.....	110
5.3	Mean Daily Temperatures at Ruby Range Sites..	113
5.4	Mean Rates of Cooling Observed at Ruby Range Sites.....	116
5.5	Differences between Rock Aspect and Slope Aspect that may cause Complications in Predicting the Thermal Regime.....	117
5.6	Radiation as a Function of Slope and Aspect..	119
5.7	Site 1 Temperatures with Respect to Aspect...	120
5.8	Site 2 Temperatures with Respect to Aspect...	121
5.9	Site 3 Temperatures with Respect to Aspect...	122
5.10	Simplified Computer Program Flow Chart.....	125
5.11a	Temperature Profiles in Alaskite, April.....	127
5.11b	Temperature Profiles in Alaskite, September..	127
5.11c	Temperature Profiles in Alaskite, January....	128
5.12	Hypothetical Situation to Illustrate Possible Complications in Interpretation of Crack Temperature Regimes.....	132
5.13	April Frostline Penetration, East Face, Wet and Dry Alaskite.....	141
5.14	September Frostline Penetration, East Face, Wet and Dry Alaskite.....	142
5.15	April Frostline Penetration, East Face, Alaskite and Limestone.....	145
5.16	September Frostline Penetration, East Face, Alaskite and Limestone.....	146

LIST OF PLATES

Plate	Title	Page
1	Field Site 2	57
2	Field Site 3	61

CHAPTER 1

BASIS OF THE STUDY

1.1 Introduction

The frost shattering of rocks is a dominant process in the low-temperature regimes of alpine and polar environments. The visible evidence of frost shattering is the production of blockfields (or felsenmeer) on relatively flat surfaces, whilst on steeper slopes extensive talus or stratified screes may develop; the formation of loess material has also been attributed to this process (Konishchev, 1982). In the literature, the term frost shattering is often used synonymously with congelifraction (Bryan, 1946), gelifraction, frost riving, frost splitting and frost wedging (Washburn, 1979). However, frost wedging usually refers to the freezing of bulk water held within surface cracks to cause rock breakdown, whereas frost shattering is more normally associated with the freezing of pore water held within the rock to effect rock

disintegration. This thesis will focus primarily on the latter.

In 1973, Ives identified "...the efficacy of freeze-thaw in the role of bedrock disintegration..." as a "...major gap in geomorphological knowledge". It appears that this is still the case more than ten years later (see, for example, McGreevy and Whalley, 1982), for although recent papers show a growing awareness of the complexity of the frost shattering process, there are inherent difficulties in measuring the freezing of moisture within rocks, which have hindered research in this area. Thus, while there is an appreciation of the role of moisture and the effects of salts, there is a paucity of comprehensive field data (McGreevy, 1982; McGreevy and Whalley, 1982; Thorn and Hall, 1980). The lack of suitable instrumentation has frustrated attempts to clarify the precise role of temperature and moisture within freezing rocks, so that field studies have tended to emphasise the temperature aspects of frost shattering and have been unable to quantitatively evaluate the role of moisture. In laboratory experiments, work done by both Tricart (1956) and Potts (1970) has shown that the amount of disintegration in rocks supplied with abundant moisture was much greater than in similar rocks with less water, but there has been little progress beyond this qualitative distinction. As McGreevy and Whalley (1982) point out, the temperature-moisture relationship should indicate that any study into frost shattering in which there

is a preoccupation with temperature "...is both restrictive and misleading in that it threatens a neglect of the roles of moisture and rock type (and properties) in determining the nature and efficacy of rock breakdown." The result of this has been that whilst the main factors associated with frost shattering (rock type, temperature and moisture availability) are generally recognised, there appears to be little agreement on the precise role of each factor in cryogenic weathering (McGreevy, 1981).

1.2 Temperature-Moisture Relationships

Volume expansion, which accompanies a phase change of water to ice, has traditionally been considered the operative process causing rock breakdown in cold climates (e.g., Rapp, 1960; Ollier, 1969; Luckman, 1976). As early as 1936, however, Grawe recognised that there were a number of limiting conditions on this mechanism, and it is now well known that not all water freezes at 0°C; appreciable amounts of water can exist at sub-zero temperatures within soils and rocks. The amount of unfrozen water is related to temperature and pressure, as well as being affected by the chemical properties of the water and soil or rock. In this respect, work done in the field of building research on the freezing of concrete, glass and ceramics, as well as research into frozen soils, may be more useful than has yet been envisaged (e.g., see Hallet, 1984). Techniques developed in these fields have helped to clarify the intricate

temperature -moisture relationship at temperatures below freezing within porous media. McGreevy (1981) has recognised the relevance of such work to the study of rock weathering by freeze-thaw:

It is important that concepts developed in other areas of study are incorporated into the geomorphological literature and adapted to the interpretation of various landscape process-form relationships (for example, rockfall and scree slope development, blockfield development, nivation and nivation hollows, rock breakdown in bergschrunds and corrie development, and rock breakdown in subglacial environments).

Lautridou and Ozouf (1982) recognised the numerous difficulties associated with measuring rock moisture levels in situ, and this still remains an area for further research. In the laboratory, however, the technique of Time Domain Reflectometry (TDR), previously used for determining unfrozen water content in freezing soils (Patterson and Smith, 1981) has been developed in the present study to provide some insight into the phase transition of water in freezing rock masses. Using the dielectric constant to determine volumetric water content, this technique provides an indication of the temperatures at which water freezes within rocks.

1.3 Objectives of Study

It is evident that there are a number of areas in the understanding of frost shattering that can be usefully investigated to provide a more coherent understanding of the process as it occurs in nature. Since the amount of moisture

that is frozen within a rock varies with temperature, it is important to know the nature of the temperature-moisture relationship, and, also to examine the thermal regime of the rock to understand when shattering may occur. The objectives of this thesis are therefore twofold:

- (i) to elucidate the significance of the temperature-moisture relationship within the frost shattering process by examining the freezing of moisture within rock samples through the use of TDR, and
- (ii) to examine the role of specific microclimatic factors which affect the thermal regime of a rock face in an environment in which frost shattering is known to take place.

The field aspect of this study was carried out in the Ruby Range, Yukon Territory. Bedrock temperatures of three outcrops of a granodiorite known as alaskite were monitored continuously for 50 days at different altitudes. These readings were related to the microclimate of each outcrop. In addition, a computer program was used to simulate year round climatic conditions to investigate further those factors affecting bedrock temperatures. In the laboratory, samples of the bedrock were subjected to sub-zero temperatures and the freezing characteristic curves were examined; also, samples of a porous limestone were subjected to similar tests for comparison.

1.4 Organisation of the Thesis

The next chapter will provide a review of the literature on frost shattering and will analyse the present status of research within this field. A fuller discussion of the field study area, and the techniques used can be found in chapter 3; chapter 4 discusses the freezing of moisture within rocks and reports on results obtained by TDR, and chapter 5 focuses on the climatic controls of frost shattering, and discusses the results from the field and from the computer simulation. The final chapter draws the investigation to a conclusion, summarising the findings and recommending areas for further research.

CHAPTER 2

REVIEW OF LITERATURE

2.1 Introduction

This chapter provides a review of some of the approaches that have been taken to investigate the phenomenon of frost shattering and the theories that have been developed to explain it. One of the main problems that has plagued studies into frost shattering is the difficulty of analysing what is a truly polygenetic process. Whilst temperature, rock type and moisture content are almost universally recognised as the three main controlling factors in frost shattering, practical difficulties have led most investigators to concentrate on determining the role of a single factor (see, for example, Goudie, 1974; Thorn, 1979; 1980; and Williams and Robinson, 1981). It is evident, however, that a complete understanding of frost weathering requires an examination of how these three factors interact. From investigations into the behaviour of freezing soils it has been realised that by focusing attention on the moisture content (which is dependent on both the thermal

regime and the soil type) a better modelling of the processes within the frozen material can be accomplished. Such an approach would therefore seem appropriate when considering the freezing of moisture in rocks.

This chapter begins by outlining basic concepts in the freezing of moisture in porous media, and then a review of previous investigations into frost shattering is given. Possible mechanisms of rock breakdown by frost action are discussed and the chapter concludes by outlining the framework of the present investigation.

2.2 The Freezing of Moisture in Porous Media

The fact that water can exist at several degrees below 0°C in porous media is now well established, and can be attributed to the depressed free-energy status of moisture held in pores. Since the free-energy states of water and ice within the system must be equal, there is a freezing point depression (Edlefson and Anderson, 1943). The water that remains unfrozen is in thermodynamic equilibrium until there is a change in pressure or temperature. In soils, the relationship between temperature and unfrozen water content is described by the freezing characteristic curve, as shown in Figure 2.1. It can be seen that the finer the soil pores the greater the proportion of water that is able to exist in the unfrozen state at any given negative temperature. This is due to the effects of capillarity and adsorption: also, at these temperatures, coulombic forces associated with the

Freezing Characteristic Curves for Four Fine-Grained Soils
(From Burt and Williams, 1978)

Figure 2.1

soil particle surface maintain a double layer of charged solutes around the soil particle preventing water molecules from forming ice. The thickness of the double layer will depend on the constituent minerals of the soil (Mitchell, 1976). The amount of water held in this way in any soil will be dependent on the particle size distribution of the soil, the large surface area of clay minerals being able to hold more water than, for example, a sand (see Figure 2.1).

It would seem reasonable to assume that similar conditions apply to rocks, and a freezing characteristic curve can be derived for any particular rock type. The curve will likewise be dependent on the pore size distribution of the rock as well as its mineralogy. There is, however, some equivocation as to the precise role of pore-water chemistry in rock breakdown. Although the presence of salts is known to lower the osmotic potential of the soil or rock, studies by Goudie (1974) and Williams and Robinson (1981) have shown that the presence of salts in freezing rocks may indeed enhance breakdown, possibly due to the effects of salt crystallization. A recent study by McGreevy (1982), however, contradicts these results; he concluded that the efficacy of frost and salt weathering was still uncertain, requiring further field data concerning temperature regimes and salt availability at and below rock surfaces in cold regions. The role of salts in frost shattering is beyond the scope of the present study but it is evident that this is an area in

which further research is still required.

In soils the freezing characteristic curve has been established to be virtually independent of the total water content (Williams, 1962). If the curve is initially determined for saturation, then the same curve applies to lower moisture contents except that it will be truncated near 0°C . In unsaturated rocks this would imply that lower temperatures are required to freeze an equivalent percentage of the initial moisture content than in saturated rocks. This is confirmed by laboratory tests by Tricart (1956) and Potts (1970) in which the disintegration of rocks supplied with abundant moisture was greater than in similar rocks supplied with less water.

2.3 The Role of Temperature in Frost Shattering.

It is important to note that the freezing characteristic curve implies freezing occurs over a range of temperatures. In many previous studies of frost shattering, attempts have been made to define a critical temperature at which moisture freezes and frost shattering becomes effective. The temperature parameters that are often considered important are: (i) intensity of freezing (extent of temperature decrease below 0°C), (ii) rate of freezing, and (iii) duration of freezing.

Table 2.1 shows some of the critical freezing intensities that have been suggested, and the methods of determining these temperatures. Fukuda (1971) measured the elastic wave

Rock Type	Critical Freezing Temperature	Method	Reference
Sandstone	Between -2°C and -3°C	Linear Expansion	Davison & Sereda(1978)
Silty Carbonates	Between -2°C and -7°C	Differential Thermal Analysis	Dunn & Hudec(1966)
Chalky Limestone	-5°C for at least 10 hours	Repeated Observations	Lautridou (1971)
Sandstone	-3.3°C	Temperature Monitoring	McGreevy & Whalley(1982)
	-4°C	Elastic Wave Velocity	Fukuda (1971)
Uncited Rock Type	-5°C	Flexure Vibration	Fukuda (1972)

Some Critical Freezing Temperatures and Measurement Methods.
(See text for discussion).

TABLE 2.1

velocity and electrical resistance of porous rocks, both of which change when moisture freezes within the rock, and determined that freezing occurred at -4°C . In a separate experiment, Fukuda (1972) also measured the internal angle of friction ($\tan \delta$) of an uncited rock type by a flexure vibration method. This value ($\tan \delta$) is sensitive to rock structure and texture, especially porosity, pore structure, and water content. A sudden drop in the value of $\tan \delta$ at -5°C was taken to indicate the freezing of water within the rock pores, and therefore the critical temperature for frost shattering. Various other critical temperatures have also been proposed: Davison and Sereda (1978) used a specially constructed dilatometer to measure linear expansion of a sandstone on freezing, and observed expansion at "...between -2°C and -3°C ." Dunn and Hudec (1966) used differential thermal analysis to determine the freezing of moisture within silty carbonates between -2°C and -7°C . Using repeated laboratory observations, Lautridou (1971) proposed that a temperature of -5°C over a ten hour period was critical for breakdown to occur. McGreevy and Whalley (1982) observed "...the evolution of heat to give a small temperature rise at -3.3°C indicating that water froze at this temperature." It is proposed by the individual workers that for the specified rock type, the temperature at which water was observed to freeze is the threshold temperature and once freezing has been initiated at this critical

temperature frost shattering may occur. Such an approach may be useful in defining climatic zones for frost shattering, but it has two major drawbacks. First, it is evident from the freezing characteristic curve that the freezing of moisture occurs over a range of temperatures. A more important threshold is the temperature at which the pressure generated by ice growth overcomes the tensile strength of the rock, and cracking occurs. The point where this occurs on the freezing characteristic curve depends not only on the rock type but also on the integrity of the rock sample. Furthermore, it is important to ensure that these critical temperatures do not refer to a supercooling of the liquid within the rock (i.e., the water in the system in disequilibrium with the temperature and pressure states of the system) as opposed to freezing point depression (moisture in equilibrium with temperature and pressure states of the system). As noted earlier, saturation levels will also affect the temperature at which freezing begins.

The rate of freezing is considered by some investigators to be an important criterion in frost shattering. Battle (1960) proposed that the rate of freezing should equal or exceed $6.0^{\circ}\text{C}.\text{hr}^{-1}$, basing this on his own laboratory experiments. However, in the field such rates are seldom found: Battle and Lewis (1951) and Battle (1960) collected data from three bergschrunds in different environments. In the Tverrabreen glacier bergschrund in Northern Norway air

temperatures fluctuated minimally, with a maximum range of 0°C to -2°C in two field seasons of observations. The range of temperatures was slightly greater in the bergschrund of the Jungfrauoch glacier in Switzerland (-2 to -5°C in March 1951) and in the bergschrund of "glacier 32" on Baffin Island ($+0.5^{\circ}\text{C}$ to -3.7°C in June-July 1953). However, at all locations the rates of cooling were lower than Battle's proposed rate, being between $0.2^{\circ}\text{C}\cdot\text{hr}^{-1}$ and $0.02^{\circ}\text{C}\cdot\text{hr}^{-1}$. These results led Battle and Lewis (1951) to believe that frost shattering was not operative under present periglacial conditions but had occurred previously.

Gardner (1969) studied temperature fluctuations at the surface of two rock faces in the Canadian Rocky mountains, and found mean rates similar to those of Battle and Lewis (1951) and Battle (1960). In addition, during the observation period, temperatures within cracks that were in rockwalls above snowpatches rarely fell below freezing. Gardner (1969) noted that the "...magnitude and rate of freezing did not reach the minimum levels for frost shattering that have been presented in the literature..." and found this difficult to reconcile with the presence of shattered rock debris. Field evidence from this study (see chapter 5) as well as other investigations (e.g., see Hall, 1975; Thorn, 1979; 1980; McGreevy and Whalley, 1982) suggest that application of Battle's criterion to rock temperature data would eliminate the possibility of frost shattering in

most, if not all instances. Lautridou and Ozouf (1978) , noted that between 7-15% of the initial water content of a sample can be lost by a slow freezing rate of less than $2.9^{\circ}\text{C}.\text{hr}^{-1}$. Moisture loss by evaporation and "cryosuction" (movement of the ice-water mixture from the interior of the specimen towards its surface) would reduce the potential of frost damage. In this instance the freezing characteristic curve would be truncated as described earlier, reducing the potential for rock breakdown to occur. It should be noted that Lautridou and Ozouf refer specifically to the shattering of small rock particles that may be found on a scree slope, outcropping in solifluction lobes or in stream bank deposits, as opposed to the "macrogelivation" that operates on a fissured rock face.

Aguirre-Puente (1975) also argued on a theoretical basis that rapid freezing would seal off the fine pores that might otherwise conduct water to the surface, so that a closed system is created which would enhance pressure effects. Mellor (1970) found that the bulk freezing strain of rock increases not only with increased water content, but also, for any given water content, with increased rate of freezing, thereby lessening the opportunity for redistribution or extrusion of porewater. However, Mellor's rapid freezing rate was between $1.5^{\circ}\text{C}.\text{min}^{-1}$ and $0.9^{\circ}\text{C}.\text{min}^{-1}$ above -41°C , which is considerably higher than those reported in field tests. Nonetheless, McGreevy and Whalley

(1982) note that the question as to whether or not freezing occurs rapidly or slowly is probably less decisive than whether the water within the rock actually changes phase.

The third parameter that has been regarded as a necessary threshold for frost shattering to occur is duration of freezing. In this instance, it is suggested that frost damage may occur only if the sub-zero temperatures are sufficiently prolonged. Lautridou (1971) indicated that freezing periods of at least 10 hours are needed to effect breakdown at temperatures between 0°C and -5°C ; this may be reduced at temperatures lower than -5°C . The importance of freezing duration will depend on the mode of rock breakdown (see section 2.5); if there is moisture migration towards a freezing front within the rock, it is suggested that given sufficient time, the pressures exerted due to ice accretion at the freezing front will eventually overcome the tensile strength of the rock. Rock type (especially in terms of permeability and integrity) and intensity of freezing are therefore also important in this respect.

The relationship between the intensity and duration of freeze-thaw cycles in frost shattering has also been investigated, although the results have been far from conclusive (Frénch, 1976). Two types of freeze-thaw cycle have been identified (Tricart, 1956), and are shown in Figure 2.2: the Icelandic cycle refers to a maritime type of climate, with an amplitude of $+7^{\circ}\text{C}$ to -8°C over a 24 hour

Freezing Cycles used in Experimental Studies

Figure 2.2

period. A continental type of climate is described by the Siberian cycle which has a wider temperature range of $+15^{\circ}\text{C}$ to -30°C over a 3-4 day period. The results of laboratory tests were somewhat equivocal: Wiman (1963) and Potts (1970) concluded that a greater amount of weathering occurred from the Icelandic cycle, while Tricart (1956) concluded that the Siberian cycle was more effective. Furthermore, McGreevy (1982) believed that the temperature fluctuations under consideration are seldom found naturally, and Lautridou and Ozouf (1982) also showed that relative frost susceptibility is the same under both temperature regimes.

Freeze-thaw frequency is defined by Fraser (1959) as the number of times the recorded temperature "...falls below the point of effective freeze following a period when the temperature was at or above the point of effective thaw." A freeze-thaw cycle is thought to be necessary for two reasons: first, as water within the rock freezes, fissures may be enlarged by the presence of ice (the exact mechanism is uncertain -see section 2.5). When the ice thaws water may then penetrate further into the rock, eventually leading to rock breakdown. However, Whalley et al. (1982) note that:

the degree of capillarity within a crack will be continually changing as weathering proceeds. If, due to ice formation, crack widening occurs, capillarity may be reduced, possibly rendering the crack temporarily immune from further ice wedging because of a reduced ability to absorb moisture.

Thus, even though the cracks within the rock may be widened, this does not necessarily mean that water can penetrate

deeper into the rock; indeed water movement may be restricted as cracking occurs.

The second reason why freeze-thaw frequencies are assumed to be important is that rarely will the rock physically breakdown when in the frozen state. The ice within the rock serves as a cementing agent, even though this ice may cause further cracking within the rock. Once the ice melts, the shattered rock is no longer held together and physical disintegration may occur. Hence there is usually a marked seasonal variation in rockfall activity with a spring maximum and a subsidiary peak in the fall (Luckman, 1976).

A study by Fraser (1959) into the importance of freeze-thaw cycles in mechanical weathering in Canada indicated that freeze-thaw frequency (derived from meteorological observations) was greater in the south of the country than at higher latitudes. He suggested that the lack of snow and vegetation to conceal and insulate bare rock surfaces led to a greater observable abundance of shattered rocks in Arctic regions. He therefore concluded that the climatic factors of low precipitation and low mean annual temperature should be considered as secondary factors in the frost shattering process. Whilst surface conditions are certainly important controls on the ground thermal regime, the general climatic conditions are also relevant and caution should be exercised before relegating them to a secondary role. In section 5.2 the factors of climate, surface conditions and rock thermal

properties are shown to interact as the important elements determining the ground thermal regime of the rock mass.

The use of "perceived" data (i.e., expected/assumed intensity, frequency and rates of temperature fluctuations) in experimental work is the unfortunate result of a scarcity of field data on bedrock temperatures. Goudie (1981) suggests that this has probably had a greater influence on the understanding of physical weathering than on almost any other branch of geomorphology. Failure to be able to reconcile laboratory derived criteria for frost-shattering occurrence with the few field observations available has led some workers to call for a re-appraisal of the frost shattering process (e.g., Battle and Lewis, 1951; White, 1976; Thorn, 1979). Whilst it is important to review continually our understanding of the frost-shattering process, it is also important not to disregard it completely simply because laboratory data and field data do not agree. Rather it would seem more pertinent to use field data as the basis of an investigation, and from there examine what conditions could lead to rock breakdown, stemming from an understanding of the freezing of moisture within porous media. For example, the freezing characteristic curve describes the change in unfrozen water content with temperature. A temperature fluctuation below 0°C involves a shift up or down the curve indicating the freezing or thawing of some moisture within the rock. To refer to a

freeze-thaw cycle, therefore, has much wider implications than crossing the 0°C boundary. Rapid and frequent temperature oscillations across the 0°C line to a threshold temperature may favour rock breakdown, especially in the presence of moisture, but these conditions may not be the only ones in which frost shattering takes place and low amplitude temperature oscillations below 0°C may also be important.

2.4 Investigations into the Role of Rock Type

The importance of rock type and structure has been recognised by a number of workers, and Table 2.2 shows a ranking of certain rock types in order of susceptibility to frost action and mechanical disintegration based on field observations. Laboratory experiments have led to a similar analysis of the role of rock type (e.g. Potts, 1970).

Varying rates of shatter have been related to lithological and structural characteristics of rocks. Thus, sedimentary rocks which contain mica and illitic clays display planes of fissility through which water migrates preferentially so that shales and siltstones, for example, breakdown more readily than igneous rocks (Potts, 1970; Washburn, 1979). Porosity and permeability are also important controls in the frost shattering process determining the ability of the rock to hold moisture, and the ability of that moisture to move through the rock (Lautridou and Ozouf, 1982). In addition, contraction of minerals in cold environments may enhance breakdown

Location	Ellef Ringes Isle. NWT.	Ardennes, Belgium.	Dartmoor SW England
	Gabbro	Quartzite	Coarse/Medium Granite
Most Resistant	Sandstone	Grits/Limestone Limestones	Diabase Fine grained Granite
Least Resistant	Shale	Schists	Metamorphosed Sediments
Reference	St.Onge (1969)	Alexandre (1958)	Waters(1964)

Field observations of susceptibility to frost
shattering according to rock type (after French, 1976)

TABLE 2.2

on the physical and chemical characteristics of the rock, particularly rock structure and texture. In rocks that are characterised by large grains of different sizes (e.g., granites, conglomerate sandstones), the diversity of minerals and their different coefficients of contraction produce internal stresses reducing the cohesion of the rock, as cracks and crevices penetrate the rock. On the other hand, rocks with micro-crystalline and mono-mineral structure react as a whole unit (Martini, 1967). As the internal rock structure is disrupted, this provides channels through which moisture can migrate and in which freezing can occur leading to further crack propagation. The rock surface micro-relief will also be important in channeling water seepage, so that horizontal surfaces may be more susceptible to frost action due to the collection of moisture. Thus, Martini (1967) recognises five lithological parameters which would affect frost shattering:

- (i) rock structure (development of mineral grains and their interaction)
- (ii) rock texture (nature of mineral arrangement)
- (iii) the system of cracks and fissures
- (iv) the nature of the cementing material (in sedimentary rocks)
- (v) the nature and form of the rock surface (influencing porosity).

Also important is the resistance of individual minerals

to chemical weathering, which may prepare rocks for, or operate in conjunction with, frost action. All of these factors will contribute to the tensile strength of the rock mass, which in turn controls the effective pressures that must be generated within the rock to effect breakdown. In this respect, planes of weakness may be exploited by the freezing moisture, so that in some places lower pressures may be required for cracking to occur than in other parts of the rock mass.

2.5 Mechanics of Rock Shattering by Frost Action

Initial views of frost shattering relied on the 9% volumetric expansion of water on freezing as the principal explanation for the disintegration of rocks in cold regions. In an early study of the freezing of water (in a one component system), Bridgman (1912) showed that the highest pressures (207mPa) were attained at -22°C . Grawe (1936) suggested that such pressures would rarely be obtained in the field because of a number of limiting conditions; these were (i) only pure water being present and this being completely confined, (ii) the extreme temperature of -22°C and (iii) the ability of the rock to withstand such pressures. Although there is little doubt that such conditions do not have to be fulfilled for shattering to occur, Grawe's paper does offer much pertinent advice for investigations into the action of frost within rocks. His proposal that "...the force exerted when water freezes in

rock cannot be stated as a definitive value unless the conditions under which the process takes place are also stated..." should not go unheeded.

The fact that volumetric expansion occurs when there is a phase change from water to ice cannot be denied, though, and there must be some accompanying exertion of pressure if the system is not completely open. It is possible to envisage pressures generated by volumetric expansion being dissipated through the pore system within the rock, particularly if it is not completely saturated. Taber (1929) also pointed out that in soil freezing, water contained within the porous medium usually contains air which would initially be expelled or compressed on freezing, thereby lowering the maximum attainable pressures.

As mentioned earlier, the rate of freezing is considered by some workers to be an important parameter; it was first considered by Grawe (1936) and later Battle (1960) who suggested that if rate of freezing is fast enough then the rock face would become sealed with ice. This is supported by Lautridou (1971) who argued that rapid freezing negates the effect of water loss by evaporation and by extrusion of ice from discontinuities. Such high rates of freezing as those proposed by Battle are seldom found in the field (see section 2.4 and chapter 5), and it is still questionable whether or not the pressures generated on freezing would be dissipated through the inter-connecting cracks and fissures

within the rock mass if the surface is sealed with ice. This will depend on the nature of the rock in terms of porosity and degree of saturation. Nonetheless, although the volumetric expansion of moisture on freezing may not by itself account for the disintegration of rock, the pressures generated may induce some cracking within the rock (McGreevy, 1982).

Several other theories have also been developed to account for the pressures required to cause frost shattering. St. Onge (1969) considered that rock breakdown was effected by the crystallization pressure of ice crystal growth, and Evans (1970) suggested that freezing of water within a rock is analogous to solute crystallization in arid environments. In a laboratory experiment by Connell and Tombs (1971) it was shown that polycrystalline ice growth from melt can "...exert a pressure against an extended obstacle, even though it is free to grow in other directions." Although the maximum pressures measured in the experiment were not large (20 kPa), Connell and Tombs noted that "...no upper limit was found for the total force developed." Thus crystallization pressures could theoretically be quite large, although those realised in practice would be severely limited by the supply of material to the growth face in contact with the large obstacle. Nonetheless, Whalley (1976) points out that even these pressures may be important in concentrating stresses at the crack tip.

Both of the above theories must rest on the premise that the total force exerted during freezing exceeds the tensile strength of the particular rock. The temperature at which this occurs is dependent on the freezing characteristic curve (indicating the total ice volume) and the integrity of the rock sample/mass. As the ice volume increases, so the pressures acting on the internal structure of the rock will increase, whilst the individual particles of the rock will be less susceptible to any sort of deformation as they are cemented by the ice.

The process by which ice forms in smaller soil pores as temperature decreases is described by the capillary theory. Initially proposed by Taber (1929; 1930), the theory was subsequently developed on a more theoretical basis by Everett (1961); Taber later extended his ideas to explain the breakdown of limestone on lake shores (Taber, 1950). Whilst it is realised that the theory does not constitute a fully comprehensive theory of soil freezing but rather outlines "...the thermodynamic requirements for ice growth..." (Smith, 1985), it may nonetheless be a useful starting point for a consideration of ice growth within porous rocks.

Williams (1967) showed that a suction will develop in a frozen soil between the ice and the water phases as the temperature falls due to the changing differential pressure of ice and water. From Everett's (1961) work, the

differential pressure between ice and water is dependent on the interfacial surface tension between ice and water, and the radius of interfacial curvature; these relationships can be summarised in the following equation (Smith, 1985):

$$r_{iw} = \frac{2S_{iw}.V_w.T_o}{-(T-T_o)L} \quad [2.1]$$

where: r_{iw} = radius of interfacial curvature between ice and water

S_{iw} = interfacial surface tension ($3.05 \times 10^{-2} \text{ Nm}^{-1}$) between ice and water

V_w = specific volume of water ($0.001 \text{ m}^3 \cdot \text{kg}^{-1}$)

T = temperature of the system

T_o = freezing temperature of pure free water (273.15K)

L = latent heat of freezing ($0.334 \text{ MJ} \cdot \text{Kg}^{-1}$).

This describes the situation where, as T decreases in the frozen soil, r_{iw} becomes smaller, allowing the ice to penetrate smaller and smaller pore spaces. If r_{iw} is smaller than the minimum pore size in the soil (r_p), the ice invades and fills all the soil pores and the freezing front advances into the soil. If, on the other hand, the soil contains pore spaces that are too small to be penetrated at any particular temperature, a suction gradient is established, since the pressure of the water at the freezing front is less than the pressure of the unfrozen water. Water movement due to the suction gradient will lead to a rise in water pressure at

the freezing front, and therefore some water must freeze. In soils, this process is believed to lead to the formation of segregated ice and subsequent frost heaving.

In most rocks, however, it seems unlikely that segregated ice will occur except in large fissures, because the initial porosity is considerably lower than in soils, although Mellor (1970) pointed out that the presence of unfrozen water in rocks of a fairly high porosity (he quotes sandstone as an example but gives no specific porosity value) would enable the migration of water towards centres of nucleation. Furthermore, he showed that small residual quantities of unfrozen water below -10°C are electrically continuous and suggested that this water would be hydraulically continuous also, and perhaps therefore be mobile under potential gradients (mechanical or thermal). If this is the case then the movement of moisture towards a freezing front would lead to considerable pressures being generated at these centres of nucleation.

It has been proposed that it is, in fact, the unfrozen, adsorbed water which may be the effective force in frost shattering (Dunn and Hudec, 1966; White, 1976). Dunn and Hudec (1966) and Falconer (1969) noticed that in certain carbonate rocks some water remained unfrozen at temperatures as low as -50°C , and at such low temperatures the adsorbed water molecules may become ordered (oriented in a certain fashion). As Falconer (1969) explains:

the dipoles (of the water molecule) are attracted to the wall of the pore space, but their charges at the free ends repel each other at the centre of the pore space. As the temperature is lowered the alignment of the molecules becomes more rigid hence the repulsive force becomes greater but the water does not freeze. Variations in this force accompany temperature changes.

Dunn and Hudec (1966) noted a strong correlation between percentage of adsorbed water and rock disintegration in a freeze-thaw environment. White (1976) argued that since field temperature data and saturation levels found within rocks did not meet apparent threshold criteria established from laboratory experiments, then hydration would be the principle means of rock weathering in cold environments. Washburn (1979) recognised the possibility of rock disintegration by hydration, but called for more experimentation along these lines, arguing that:

the issue is whether these effects can really account for the angular fragments of sound rock such as those commonly ascribed to frost wedging. Even if ordering of water is capable of producing such fragmentation, additional evidence would be required to prove that frost wedging is unimportant.

Studies carried out in the field of building research on the freezing of concrete, glass and ceramics may also contribute to an understanding of frost action in rocks. Arni (1966) states that damage from freezing and thawing is the most common and the most studied type of physical deterioration in the weathering of buildings. A number of papers by Powers (1945; 1956; 1965, amongst others) have been foremost in examining the breakdown of concrete by freezing, as well as in developing methods to test the

durability of concrete. Powers (1945) noted that rocks susceptible to frost action are (i) likely to be saturated near the surface, and (ii) of pore sizes such that frictional resistance will resist hydraulic movement. He thus considered that water freezing from the outside of a block of stone, after causing an initial surface seal, would penetrate the rock body, displacing the region of lower water content (Figure 2.3). This movement would be resisted by the fine-grained porous material, causing gradients of hydraulic pressure. If this resistance against the force displacing the water inwards is sufficiently high, then, Powers believed, it would be capable of damaging the rock specimen.

Powers' hypothesis appears to rest on the competence of the rock, since any cracks or fissures would lead to a dissipation of the hydraulic pressure (as with the volumetric expansion of water on freezing). The size and shape of the rock sample being examined would also be important, determining whether freezing was from one or more sides, and as Lautridou (1971) noted, rate of freezing would also be important if there is to be surface sealing of the pores (see section 2.4). Powers (1965) is particularly concerned that laboratory experiments are considered with caution before being applied to the field situation:

My unorthodox belief is that durability is not a measurable property; rate of disintegration in the laboratory cannot have a consistent relation to rate of disintegration in the field; one can tell by test whether or not freezing produces internal stresses

EXPLANATION

A is the saturated area near the surface, and B is the region of lower water content. Freezing will occur in A before it does in B, because of the direction of the temperature gradient and because A has a higher water content. If the water displaced into region B was free to move without resistance, no hydraulic pressure would develop. However, since it is moving through a fine-grained porous medium, the force causing movement will result in a corresponding frictional resistance and, therefore, gradients of hydraulic pressure. If the reaction against the force displacing the water inwards is sufficiently high, then it can be regarded as being capable of damaging the rock.

Hydraulic Pressure Theory of Rock Breakdown
(After McGreevy, 1982)

Figure 2.3

even when the stress originates in just a few pebbles; design and control of frostproof concrete should be based on comprehensive studies of moisture in concrete in relation to environmental factors, not on the results of freezing and thawing tests, at least not as those tests are now interpreted.

Geomorphologists would do well to extend these comments to any consideration of frost shattering; too often laboratory tests are not related to natural conditions under which frost shattering is known to take place with the result that they are of limited value.

2.6 Summary

This chapter has provided an overview of the present state of understanding of the frost shattering process, with a discussion of some of the methods that have been used to define the role of individual factors affecting frost action in rocks. It is evident that there is no clear consensus as to how frost-shattering actually occurs, although there is general agreement that the primary factors in the process are temperature, moisture content, and rock type. It is suggested that, as in soil freezing, it is the interaction of these factors which is of prime importance.

The rest of this thesis examines the relationship between temperature and moisture content within freezing rocks, and attempts to relate an understanding of this relationship to rock temperatures observed during a short field season in the Ruby Range, Yukon Territory. The next chapter provides a brief outline of the field setting of this investigation.

CHAPTER 3

FIELD AREA DESCRIPTION.

3.1 Introduction

The field work aspect of this investigation was conducted in the Ruby Range, Yukon Territory during the summer of 1983. The aim of the field work was twofold: (i) to investigate the physical environment in which frost shattering is known to take place in terms of saturation levels of massive rock and surface conditions, and (ii) to record air and rock temperatures at a number of locations to analyse the factors which may affect the thermal regime of a rock face.

This chapter provides a brief outline of the geology and Quaternary history of the Ruby Range. The three field sites monitored during the study period are described in detail, together with a description of techniques used in the field.

3.2 Geology of the Ruby Range.

Located in the southwestern corner of the Yukon Territory, the Ruby Range is a small, primarily igneous, mountain group oriented northwest-southeast, and situated

along the western edge of the Yukon Plateau, east of Kluane Lake (see Figures 3.1 and 3.2). The Range parallels the Shikwak trench, which acts as a major divide between the complex geological formation of the St. Elias mountains to the southwest and the older Yukon Plateau to the northeast. The plateau is for a large part underlain by granitic rocks and associated migmatites and gneisses (Muller, 1967). The broad ridge of the Ruby Range forms a distinctive undulating upland surface in the Yukon Plateau. The higher peaks of the range average 2,130m.(a.m.s.l.), and there is a distinct flattening of the upland surface that may be an old erosion surface of the Yukon Plateau (Bostock, 1948). The range is cut by many glacial valleys containing past or present finger lakes.

The Ruby Range batholith is the largest intrusive body within the Yukon Plateau, and underlies most of the range east of its intersection with the Shikwak Trench. The main components of the batholith are hornblende-biotite, granodiorite and quartz diorite, although gabbro and diorite occur in places, and it is considered by some (e.g., Muller, 1958) to be a northern extension of the coastal intrusions.

In the study area the main rock type is alaskite, an alkali granite containing varying amounts of quartz, feldspar and some biotite. The rock is described by Muller (1967) as a light coloured granitic rock, which he differentiates from granodiorite primarily on the basis of

The Ruby Range Field Area

Figure 3.1

Geological Map of the Ruby Range (After Muller, 1967)

Figure 3.2

coarser mineral grains which are less firmly bonded in the alaskite. The alaskite can be subdivided into three groups according to grain size (Kodybka, 1984):

- (i) coarse alaskite -dominated by feldspars in a matrix consisting mainly of quartz and plagioclase with biotite and chlorite as minor constituents,
- (ii) medium alaskite -with approximately equal proportions of quartz and feldspars and minor amounts of biotite,
- (iii) fine alaskite -with approximately equal amounts of quartz and feldspars and minor amounts of plagioclase and biotite.

Further data on these alaskite types is summarised in Tables 3.1 and 3.2, and is analysed in detail in section 3.2.1.

The weathered surfaces of each alaskite type showed noticeable differences: the coarse grained alaskite weathers to a rough crystalline surface from which individual or small agglomerations of crystals fall. The medium grained alaskite commonly has a weathered crust, not unlike the desert varnish described by other writers (e.g., Dorn and Oberlander, 1982; Cooke and Warren, 1973) and characterised by a smooth sheen on the rock face. Some of this is almost certainly ventifacted as well, for such a polish can be produced by wind-driven fines and snow. In sheltered areas where the weathered crust does not develop on the medium grained alaskite, the rock face is generally rough and angular, but not of the crystalline nature of the coarse

alaskite. The fine alaskite also shows signs of a weathered crust, although much less distinctive and more angular than on the medium alaskite. Again, in the areas of fine alaskite where the weathered crust is not evident, the rock faces are rough and angular with a fine crystalline texture.

3.2.1 Mechanical Properties of Alaskite

Table 3.2 provides a summary of some of the mechanical properties of the three types of alaskite. From this data a ranking of the relative strength of each type can be compiled, as shown in Table 3.3. The table assumes that as porosity and apparent porosity increase, so does weatherability; the ability of water to invade and fill the rock pores is of fundamental importance in frost shattering.

The total porosity of a rock is defined as the ratio of the volume of the pores to the bulk volume of the rock (Jumikis, 1979). Pore volume and bulk volume can be easily calculated by weighing samples in a completely dried state, and then reweighing them after complete saturation, (under pressure if necessary). Since this gives a measure of the volume of interconnected surface pores, the term apparent porosity is preferable. Apparent porosity is important in frost shattering since it determines the amount of water that the rock can hold and which can be subsequently frozen; generally the more porous the rock the greater the susceptibility to frost shattering (Washburn, 1979). The total percent void space is the total porosity of a rock and

	Fine Alaskite	Medium Alaskite	Coarse Alaskite
Quartz	50%	50%	15%
K-feldspar	40%	40%	70%
Plagioclase	10%	5%	5%
Biotite	5%	5%	5%
Chlorite	-	5%	5%
Other	5%	5%	5%

Mineral Composition of Alaskite Units, by Point Count
(Approximately 200 points counted per section:
Results rounded to nearest 5%).

TABLE 3.1

	Fine Alaskite	Medium Alaskite	Coarse Alaskite
Total Porosity	18.66%	5.98%	12.87%
Apparent Porosity	1.53%	0.83%	0.81%
Compressive Strength	61.1MPa	152.2MPa	81.1MPa
Approx. Tensile Strength	6.11MPa	15.22MPa	8.11MPa

Mechanical Properties of Alaskite Units

TABLE 3.2

Data for both tables from Kodybka (1984); all samples from same source area as those used in present study.

there are a number of techniques for calculating this, although the majority involve long and complicated procedures (Lama and Vutukari, 1978). The method used for Table 3.2 involved submerging a totally crushed ($<2\mu\text{m}$) powdered specimen with a known bulk volume into a graduated cylinder of water. The displaced volume of water is equivalent to the grain volume. Since there is the possibility of some moisture evaporating from the cylinders or of air bubbles being trapped in the powdered sample, total porosity values should only be regarded as a good approximation. Total porosity may affect a number of other rock properties, such as the thermal conductivity and rock strength.

In the medium grained alaskite the low porosity is due to the tightly packed structure of the rock. The relatively large number of small grains in the fine grained alaskite increases its porosity, whereas in the coarse grained alaskite the increased porosity is accounted for by the larger crystals not packing as tightly (as in the medium grain alaskite).

The compressive strength of the alaskite was used as an indicator of the strength of each rock type (Attewell and Farmer, 1976). Although it is the tensile strength of the rock that is of primary importance in frost shattering, tensile strength tests on rocks are difficult to perform (Jumikis, 1979). Rock is relatively weak in tension because of discontinuities that occur within it in its natural

Compressive Strength (uniaxial)	Classification	Usual Rock Types
5-20 MPa	Very Weak	weathered and weakly compacted sedimentary
20-40 MPa	Weak	weakly-cemented sedimentary; schists
40-80 MPa	Medium Strength	competent sedimentary some low density, coarse igneous
80-160 MPa	Strong	competent igneous and metamorphic, & some fine grained sandstones
160-320 MPa	Very Strong	quartzites; dense, fine grained igneous

Engineering classification of Rocks
(after Attewell and Farmer, 1976)

TABLE 3.3

	Fine Alaskite	Medium Alaskite	Coarse Alaskite
Total Porosity	1	3	2
Apparent Porosity	1	3	2
Uniaxial Strength	1	3	2
Total Score	3	9	6

Ranking of Alaskite Units Strength: weathering susceptibility

For total and apparent porosity, 1 refers to highest values, 3 for lowest; for uniaxial strength, 1 refers to lowest strength, 3 for highest. Thus, the lower the total score, the more susceptible to mechanical weathering. (After Kodybka, 1984).

TABLE 3.4

state; thus the tensile strength of a rock is usually only about 10% of its compressive strength. It is nonetheless influenced by similar lithological factors as the compressive strength, including mineralogy, porosity and grain size (Jumikis, 1979). Table 3.3 shows an engineering classification of rock strength, after Attewell and Farmer (1976); under this classification, the fine grained alaskite units are considered medium strength rocks and the coarse and medium grained alaskite units both fall into the strong rock category, although they are the lower and upper extremes of this category respectively. In the weathered state it is to be expected that the packing of the grains will be altered in some way, so that generally porosity will increase and strength will decrease as weathering occurs.

3.3 Quaternary History of the Ruby Range

Since present day geomorphological processes depend on both present and past conditions, it is appropriate to give a brief overview of the Quaternary history of the Ruby Range. In the case of frost shattering the glacial history of an area will be important for two major reasons: first, glacial erosion and plucking may expose areas in which frost action may occur. Cirque headwalls, truncated spurs on valley sides, notches, trough heads and sub-glacial meltwater channels may all provide steep, bare rock faces in which frost may penetrate and subsequent weathering take place. Second, glacial erosion often involves the loosening

and weakening of blocks of rock beneath the ice, as well as producing dilatation jointing parallel to the rock surface.

These discontinuities within the rock can act as planes of weakness, lowering its tensile strength, as well as

providing channels in which water can be held and through which it can penetrate into the rock. Although it is,

generally accepted in the literature that most of the Ruby Range was glaciated at some time during the course of the Pleistocene (see Bostock, 1948; Kindle, 1953; Muller,

1967), attempts to interpret the glacial chronology of the area are complicated by its transitional nature (from the currently glaciated St. Elias mountains to the west-central Yukon which was unglaciated throughout the Pleistocene (Johnson, 1983)).

Table 3.5 shows the late glacial chronology of the ~~northeast~~ St. Elias (after Denton and Stuiver, 1967), with a tentative attempt to correlate it to the terminology of Muller, 1967 which is more appropriate for the Ruby Range.

Three progressively less extensive ice sheets can be identified (Figure 3.3), each terminating in numerous valley glaciers: the upper limit of the ice surface of each younger ice sheet was about 300-500 metres below the preceding one

(Muller, 1967). These advances are associated with the Cordilleran ice-sheet that flowed off the Rocky and St. Elias mountains during the Wisconsin glacial period.

A late-Wisconsin advance into the lower valleys of the

GLACIAL AND NEOGLACIAL EVENTS		C DATES
After Muller (1967)	After Denton & Stuiver (1967)	(Years B.P.)
	Neoglaciation	< 100
	Volcanic Ash	• 1460(± 70)
		• 2640(± 80)
St. Elias Advance	Slims Nonglacial Interval	• 9770 ± 200
	Kluane Glaciation	• 12500 ± 200
	Boutellier Nonglacial Interval	• 30100 ± 600
Ruby Ice Sheet		• 37700 ± 1400
	Icefield Glaciation	49000
	Silver Nonglacial Interval	• > 49000
Nisling Ice Sheet	Shakwak Glaciation	• > 46400

LATE PLEISTOCENE GLACIAL CHRONOLOGY
OF THE NORTHWESTERN ST. ELIAS MOUNTAINS

(Modified after Denton & Stuiver, 1967)

Table 3.5

47

Kluane Hills is marked by well preserved moraines. In the northern part of the Kluane Hills, as well as in the Ruby Range and in the Nisling Range, local alpine glaciers occurred "...that left sharp cirques with a very fresh appearance and with well defined end-moraines." (Muller, 1967). The upper limit of glaciation indicated by these cirques is approximately 1500m., although the upper level of the ice sheet surface would be about 900m. This advance is termed the St. Elias advance by Muller (1967).

A more extensive early Wisconsin advance (the Ruby advance) is indicated by subdued moraines, which penetrate the lower sections of the valleys from the Kluane Lake basin into the Ruby Range. End moraines, kame terraces and well developed marginal channels indicate that the ice surface was at about 1200m. in the northwest of the Range, a little higher in the southeast.

Evidence of a pre-Wisconsin advance is poorly preserved except by the presence of erratic material scattered on many plateau surfaces above the level and beyond the extent of the relatively well marked Ruby ice sheet (Muller, (1967). There is little further topographic evidence for this Nisling advance, and it is suggested that the plateau surface of the Ruby Range would be covered by ice upto an upper level of about 1500m., leaving a small number of nunataks, which would be subjected to severe para-glacial conditions.

Glacial Limits in the Ruby Range (After Muller, 1967)

Figure 3.3

Subsequent to the retreat of the most recent glaciation (St. Elias), southwest Yukon has experienced fluctuating climatic conditions similar to the rest of the northern hemisphere. Two major intervals of Holocene glacier expansion have been identified in the St. Elias mountains (Denton and Stuiver, 1967): the older advance occurred between 3,300 and 2,400 B.P., and the more recent advance during the little ice age of the last several centuries (460 to 100 B.P.). During these periods of climatic cooling there may also have been corresponding periods of local glaciation in other parts of the southwest Yukon. Denton and Stuiver (1967) note that in the Ruby Range cirques may have been repeatedly occupied by glacier ice and may therefore represent the effects of multiple glaciations at the local level.

3.4 Description of the Field Site and Instrumentation

The field camp was established in the southeastern section of the Ruby Range, within a col between the heads of Alaskite Creek and Talbot creek (see Figure 3.1), at an altitude of approximately 1700 metres (a.m.s.l.). The area was probably overridden by both regional and local ice (Johnson, 1983), with the general direction of flow from Alaskite to Talbot Creeks.

A number of practical considerations had to be taken into account in selecting suitable sites where rock temperatures could be monitored. It was initially decided that it would

be useful to monitor the thermal regime of a single rock type at a variety of aspects and altitudes. Time constraints and the scope of the study did not make it feasible to look at more than one rock type; as it is, within the single alaskite rock type, there are variations of grain size and mineral composition. The sites necessarily had to be accessible (only one automatic "Rustrak" recorder was available), and each site had to have a variety of aspects. Three sites were chosen as illustrated in Figures 3.4, 3.5 and 3.6; it should be noted that although at each site the rock exposures afforded a number of different aspects, they are all on the south-facing slope of the hillside. Whilst it is realised that work on exposures in directions other than south would have been desirable, they were inaccessible because of steepness and lack of technical equipment.

Thermistors were inserted into the rock faces at a variety of depths: some problems were encountered with the drilling procedure, but most thermistors were placed at depths of approximately 5cm (see next sections for specific details). The drilled holes were backfilled with rock debris mixed with a silicon sealant to produce a colour similar to that of the original rock surface. Leads from the thermistors were gathered at a central point to facilitate quick and easy reading at the two manual sites, as well as at the automated site.

At each site a modified Stevenson screen was erected so that air temperature and relative humidity could be

monitored. Two of the screens (sites 2 and 3) contained hygrothermographs for recording temperature and relative humidity continuously, while the screen at site 1 contained a maximum-minimum thermometer which was read at the same time as the thermistor readings were taken. Precipitation was also recorded at each site, and samples were taken for analysis in the laboratory.

3.4.1 Field Site 1

Site 1 was located on an area of exposed bedrock just above a stream draining the small lake in the col into Talbot Creek. The stream was also fed by local snowpatches. This site was the lowest of the three, at an altitude of approximately 1650m. The rock face was essentially of an angular, blocky structure, with the appearance of failure along the horizontal and vertical axes. On either side of the rock face, rockslides had occurred, possibly through undercutting by the stream. Much of the weathered material from the rock face was carried away from the base of the outcrop by the stream. Figure 3.4 gives a rough morphological field map of the site.

Samples taken from the site revealed the alaskite to be predominantly medium grained. Within the rock face, however, a small gulley area had been eroded away; the alaskite around the gulley area was fine-grained, and it is probable that preferential weathering in this area led to a removal of the material. The gulley area was darker than the rest of

Morphological Map of Site 1

Figure 3.4

Thermistor Number	Depth (cm)	Height (cm)*	Aspect	Nature	Surface Colour	Grain Size	Ground Water Avail.	Lichen Type & Cover	Vegetation in Immed. Vicinity
1A	5	186	110°	Med. smooth	Brown-grey; Speckled	M/F	N	0	None
1B(a)	5.5	240	170°	Some crust	Dark Grey	F	Y	0	Some moss
1C	6.5	205	245°	Very Pitted	Speckled	F	N	1; C	Grasses below
1D	5	220	335°	Rough, flaking	Light Grey; Speckled	M	N	0	None
1E	6	265	135°	Some Pitting	Grey; Speckled	F	Y	1; B	None
1F	4	240	160°	Rough, new: Some Flaking	Brown-grey; Speckled	M/F	N	0	None
1G	3.5	410	250°	Flaking	Light Grey; Speckled	M	N	1; C	None

KEY:

Grain Size:

C : Coarse
M : Medium
F : Fine

Lichen Cover:

0 : None
1 : ↑
2 : ↑
3 : Complete

Lichen Type:

A : *Acarospora Chloropharia*
B : *Xanthoria Elegans*
C : *Lecidello Stigmathea*

NOTES:

* Heights are above ground level
(a) 1B situated within gulley recess

THERMISTOR LOCATIONS AT FIELD SITE 1
TABLE 3.6

the rock face, but this may have been due to the moisture on the rock face since the recess acted as a channel for groundwater flow, at least during the early summer when snow was still melting.

Table 3.6 outlines the specific characteristics of each thermistor micro-environment; it is evident that the gulley / recess provided a protected area where vegetation can become established on the weathered debris that collects on the small ledges. This vegetation acts as an important retainer of moisture, providing at least some supply of water after exposed parts of the rock face have dried out.

3.4.2. Field Site 2

The second site was located on glacial till on the ridge between the mountain slope to the north and the stream to the south (Figure 3.3). This site was the middle of the three sites at an altitude of approximately 1750 metres (a.m.s.l.). The bedrock at this site had a distinct blocky structure, and the rock face was significantly stepped or terraced (Figure 3.5 and plate 1).

Two sides of a wide shallow gulley feature (west facing and east facing) were monitored; down the centre of the gulley was a boulder stream of what appeared to be collapsed bedrock. The two gulley sides were the exposed bedrock and they varied significantly in a number of respects. Below the west face was an accumulation of large boulders of collapsed bedrock. This provided some protection for the rock face,

MORPHOLOGICAL MAP OF SITE 2

Figure 3.5

Thermistor Number	Depth (cm)	Height (cm)*	Aspect	Nature	Surface Colour	Grain Size	Ground Water Avail.	Lichen Type & Cover	Vegetation in Immed. Vicinity
2A		200	245°	Rough: Micro-Fissures	Light-grey; Speckled	C	Little	3;C/B	Moss & Grass
2B	10	310	260°	Smooth, crust	Light-grey Speckled	M/C	Little	2;B/C	Some Moss
2C(a)	7	280	255°	Some Flaking	Light Grey; Speckled	M/L	N	2; B	Some Moss
2D			050°	Rough, Crystalline	Grey; Speckled	M	N	3;A/C	Some Moss
2E	5.5	170	085°	Med. Rough; Fissured	Light-grey	M	N	3; C	None
2F(b)	5	150	340°	Med, Rough; Crust	Light Grey; Speckled	M/C	N	2;B/C	None

KEY:

Grain Size:

C : Coarse
M : Medium
F : Fine

Lichen Cover:

0 : None
1 : ↓
2 :
3 : Complete

Lichen Type:

A : Acarospora Chloropharia
B : Xanthoria Elegans
C : Lecidello Stigmatea

NOTES:

- * Heights are above ground level
(a) 2C sheltered from above by overhang
(b) 2F sheltered in part by overhang; newly exposed face

THERMISTOR LOCATIONS AT FIELD SITE 2

TABLE 3.7

Field Site 2, Looking Eastward

Plate 1

providing shading from the sun as well as probably acting as a snow trap in the winter, so that the overall effect is to reduce the range of temperatures to which the face is subjected, and reducing the potential for weathering. The rock face itself was considerably rounded and of a "pillowy" nature, and was composed predominantly of medium grain alaskite, showing the ventifacted rock varnish described previously. In comparison, the east face had a block-like structure and was composed primarily of coarse grained alaskite. This gave the weathered areas of the east face a rough-crystalline surface from which particles of rock were easily dislodged. The relatively large amount and wide variety of lichen on the west face also suggest that this surface was subjected to less severe weathering than the east face (Rejmanek, 1971).

The details of each thermistor micro-environment are outlined in Table 3.7. Particularly noticeable is the lack of readily available groundwater at the site and the limited amount of vegetation compared to site 1. Although some moss was established on a number of the less exposed rock ledges, there was not the variety of vegetation present at site 1.

3.4.3. Field Site 3

The third site was the highest site monitored and was at an altitude of approximately 2,050m. (a.m.s.l.). A tor-like outcrop within an area of steeply sloping weathered bedrock was instrumented (Figure 3.6 and plate 2). The rock mass was

Morphological Map of Site 3

Figure 3.6

Thermistor Number	Depth (cm)	Height (cm)*	Aspect	Nature	Surface Colour	Grain Size	Ground Water Avail.	Lichen Type & Cover	Vegetation in Immed. Vicinity
3A	8	400	115°	Med. Rough Some crust	Dark Grey	M	N	2; B/C	None
3B	6	80	050°	Rough, crust Some Flaking	Dark Grey; Speckled	M/F	N	1; B/C	Some Ferns
3C	8	215	185°	Smooth; crust Flissured	Dark Grey; Speckled	F	N	0	None
3D	6.5	150	260°	Rough; Flaking	Light-grey Speckled	M	N	1; B	None
3E	6	205	080°	Rough; Some Flaking	Dark Grey; Speckled	M	N	2; A/C	Some Moss
3F	6.2	270	315°	Rough; Flaking	Light Grey; Speckled	M/F	N	2; C	Some Moss
3G	5.5	70	075°	Med. Smooth; Crust	Dark Grey; Speckled	M	Y	3; B/C	None

KEY:

Grain Size:

C : Coarse
M : Medium
F : Fine

Lichen Cover:

0 : None
1 :
2 :
3 : Complete

Lichen Type:

A : Acarospora Chloropharia
B : Xanthoria Elegans
C : Lecidella Stigmatina

NOTES:

* Heights are above ground level

THERMISTOR LOCATIONS AT FIELD SITE 3

TABLE 3.0

1. [illegible] [illegible] [illegible]

Page 2

ideal in offering rock faces at four aspects which were suitable for monitoring. Because of the distance to this site, and the difficulty of access, it was monitored using the Rustrack recorder and a hygrothermograph inside the Stevenson screen.

This site exhibited less jointing and faulting than either of the other two sites. The alaskite type was primarily medium-grained, although there were areas of fine grained rock within the outcrop. Exposed areas of the top showed excellent examples of the ventifacted rock varnish. Like the west face of Site 2, this outcrop did not have a blocky structure but was again rounded and pillowy in structure. Below the outcrop, long and steep scree slopes ran down to Talbot Creek, indicating that a considerable amount of weathering of the rock face has occurred at sometime. Specific details of each thermistor location can be found in Table 3.8. Little vegetation was found at this high altitude exposed site, except in the most sheltered areas; also, little surface water was available at the rock outcrop, although a considerable amount of moisture flowed over the weathered backwall as snowpatches melted during the summer season.

3.5 Data Collection

The field season lasted from June 5th until July 12th. Temperature readings at sites 1 and 2 (manual sites) were usually taken on a thrice daily basis. In order to provide

an indication of the diurnal temperature pattern, an intensive 48 hour period of study was undertaken at least once a week, with readings being taken every two hours. At site three there were some initial problems with the Rustrak recorder, so that the data set is considerably shortened. However, the Rustrak reads all the channels (thermistors) every 90 minutes, so that the temperatures obtained provide a continuous record. The raw data are analysed more fully in chapter 5. Rock samples were recovered from each site to examine the mineralogy and grain size, and some were preserved in wax to test for moisture content (see next chapter).

CHAPTER 4

THE FREEZING OF MOISTURE IN ROCKS

4.1 Introduction

The first two chapters of this thesis stressed the importance of considering the frost shattering process in terms of the interaction between temperature, rock type and moisture. This chapter focuses on the freezing of moisture within porous media and its possible effects on rock structure and strength. The chapter briefly discusses the processes by which pressures within frozen media are generated. Methods for determining the moisture content of rocks in the field, and for determining freezing characteristic curves in the laboratory, are reviewed. A description of the technique of Time Domain Reflectometry (TDR), used to determine the freezing characteristic curve of two rock samples Caen Limestone and Nisling Alaskite, is also presented.

4.2 Freezing of Moisture in Rocks

It is evident that moisture is the "sine qua non" of

frost shattering (Washburn, 1979); frost action without a moisture supply is impossible, and the amount of moisture will determine the effectiveness of the process (e.g., see Potts, 1970). In soils, frost heaving is caused by the volume change resulting from freezing of the in situ pore water and ~~also~~ by the ice segregation process. Freezing of the pore water in soils results in a volume expansion of 9%, but the largest part of the heave that occurs is attributable to the growth of ice lenses due to water movement under temperature-induced gradients of potential.

Within rocks, pressures will be generated by both volumetric expansion, and if the rock is of sufficient permeability, by ice segregation. There may also be pressures produced by crystallisation, although the manner in which crystal growth causes frost shattering is not yet fully understood (Whalley, 1976).

4.2.1 Volumetric Expansion on Freezing

In section 2.5 it was noted that the "classic" accounts of frost shattering are based on the 9% volume expansion of water on freezing. Grawe (1936) cited a number of limiting conditions on this mechanism, based on a critical review of the early work of Bridgman (1912). It is now evident that some of these critical conditions do not need to be fulfilled for shattering to occur, although other criteria have been proposed more recently. Mellor (1970) proposed that a minimum saturation level of 50% is required for

volumetric expansion to cause significant strains and internal cracking of the rock. Powers (1965) noted that theoretically the lowest degree of saturation in concrete is 91.7% for freezing to be effective. However, Fagerlund (1975) comments that any single critical degree of saturation refers only to a specific set of conditions and is peculiar to the composition of the material in question. Frost damage will only occur if, at any point within the material, the actual degree of saturation exceeds the critical degree of saturation for the same point. Since the water content in a massive rock will undoubtedly vary from point to point, identification of a single critical value for the rock mass as a whole can be misleading.

White (1976) used the 50% saturation presented by Mellor (1970) to suggest that frost action may not be the significant means of rock breakdown in cold regions. He questions how often rocks are above 50% saturation and then subjected to freezing. However, field results from this study (Table 4.2), as well as from a study conducted in the Scottish Cairngorms by McGreevy (1982), indicate that over half of the samples measured attained the degree of saturation proposed by Mellor (1970).

If the rock is not fully saturated, then theoretically expansion should be in the direction of least resistance, which would be into the unsaturated pores. However, Connell and Tombs (1971) conducted an experiment in which ice crystals were observed to grow in directions other than

those of least resistance.

4.2.2 Thermodynamic Aspects of Soil Freezing

That water can remain unfrozen at temperatures below 0°C in porous media is of fundamental importance to an understanding of the freezing process and the properties of frozen materials. The presence of water at negative temperatures is due to matric and osmotic potentials created by capillarity, adsorption and solution. Changes in temperature and pressure can alter these potentials, and thus alter the phase composition of the frozen material. The potentials thus created within the frozen medium will also lead to water movements. A comprehensive analysis of this moisture movement requires an understanding of the relationship between coupled heat and moisture flows. The hydrodynamic model for soil freezing assumes that the two transport equations, one for heat and the other for water, are coupled by the relationship between pore-water pressure and temperature below 0°C (Taylor and Luthin, 1978). The Clausius-Clapeyron equation defines how the pressure difference that is maintained between the ice and adjacent water increases as the temperature decreases. If the content of free solutes is negligible, then the equation takes the form (Fukuda, 1984) :

$$dT = -T/L (V_w dP_w - V_i dP_i) \quad [4.1]$$

where T = temperature, Kelvin

L = latent heat of fusion

P_i, P_w = pressure of ice and water, respectively

V_i, V_w = specific volume of ice and water, respectively

If P_i is atmospheric and P_w falls then $dP_w/dT = 1229 \text{ kPa}^\circ\text{C}$; if P_w is atmospheric and P_i rises (if the ice is constrained by the rock structure, which will normally be the case) then $dP_i/dT = 1127 \text{ kPa}^\circ\text{C}$. If the permeability of the rock prevents movement of water in or out of the material during freezing (a closed system where $P_i = P_w$) then $dP/dT = 13543 \text{ kPa}^\circ\text{C}$. This final situation is that anticipated by Lautridou and Ozouf (1978) where they consider that the rate of freezing will determine whether the system is open or closed. If these pressures are compared to the tensile strength of most rocks (Table 3.3), then it is evident that only "very strong" rocks could resist the pressures generated by a closed system freezing. The potential gradients that are established enable moisture to be redistributed and drawn towards the freezing front; in soils this results in frost heave and ice segregation. The degree of moisture movement and subsequent ice segregation depends on a number of different factors including the rock/soil type, permeability, stress history and rate of freezing (Johnston, 1981). Table 4.1 shows the permeability of a number of unfrozen rocks and soils, and Figure 4.1 shows the reduction of hydraulic conductivity of frozen soils as temperature decreases below 0°C . Fukuda (1984) showed that in a rock of sufficient porosity (he used a Neogene Tuff,

Soil/Rock Type	Bulk Density (gm/cm)	Hydraulic Conductivity (cm/sec)	Reference
Leda Clay	1.82	1.88×10^{-8}	A
Slims Valley Silt	2.05	1.04×10^{-4}	A
177-125 μ m Sand	2.00	9.44×10^{-3}	A
Carleton Silt	1.95	1.36×10^{-4}	A
Fithian Illite	1.80	4.81×10^{-6}	A
Oneida Clayey Silt	1.75	8.14×10^{-5}	A
Limestone (1)	2.42	1.60×10^{-8}	B
Limestone (2)	n.g.	1.00×10^{-6}	B
Dolomite	2.44	3.79×10^{-3}	B
Siltstone	n.g.	7.26×10^{-5}	B
Fine Sandstone	n.g.	5.00×10^{-7}	B
Coarse Sandstone	n.g.	8.00×10^{-5}	B

Key: (1): Right angles to bedding plane
 (2): Parallel to bedding plane

References: A: Burt (1974)
 B: Saniņa and Zaleskii (1964)

Examples of the Hydraulic Conductivity of some Soils and Rocks

TABLE 4.1

REDUCTION OF HYDRAULIC CONDUCTIVITY
WITH TEMPERATURE
(After Williams, 1982)

Figure 4.1

porosity 36%), a similar reduction in conductivity would also occur. As the freezing process establishes a potential gradient, water will migrate toward centres of ice nucleation within the rock. In a saturated rock, these centres of ice nucleation would be the largest pores and micro-fissures. If water is drawn into an already weakened plane, crack propagation may occur which could eventually lead to rock breakdown. In an unsaturated rock, moisture migration will be reduced because the water will be contained in smaller pores and the conductivity is lower. If water is restricted from moving toward centres of ice nucleation, then the stresses developed will be less with a subsequent reduction in the degree of frost damage.

As noted above, if the ice is constrained by the rock matrix, P_i may be raised above atmospheric pressure; given sufficient overburden or hydrostatic pressure the negative pore water pressure may be reduced, and even be positive. Thus the potential gradient is reduced and moisture migration is similarly reduced.

4.3 Effects of Moisture and Freezing on Rock Strength

An important consideration in frost shattering is the strength (both tensile and compressive) of the rock in the saturated and frozen states. The ability of the rock to withstand the pressures generated by the phase change of water to ice determines the resistance of the rock to frost weathering. The tensile strength of the rock will determine

its ability to resist cracking, whilst the compressive strength of the rock will determine the ability of the pores to resist deformation when stress is applied. As noted in section 3.2.1, the tensile strength of rock is considerably lower than the compressive strength, and so it is this parameter that is usually the important one. Although rock strength will vary between rock types, there are some general principles regarding the effects of moisture and freezing on rock strength that should be noted.

Mellor (1971) showed that the strength (compressive and tensile) of a saturated rock may be up to 20% less than in air-dried samples (Figure 4.2); and up to 60% less than in oven-dried samples. Thus he observed that the strain for a given stress was greater for saturated samples than for dry specimens. The precise nature of the dependence of rock strength on water content is not yet fully understood, but it is usually attributed to the deterioration of mechanical properties by adsorption or absorption of fluids (Rehbinder effect) (Mellor, 1971).

At temperatures below 0°C , the compressive and tensile strengths of rocks are strongly temperature-dependent (see Figures 4.3 and 4.4). If the rock is saturated, there is an abrupt increase of strength as the first freezing of pore-water takes place, and thereafter strength increases more gradually as temperature decreases. In nominally dry rocks, strength increases as temperature decreases, although the

Tensile Strength as a Function of Water Content
for Berea Sandstone and Indiana Limestone

(After Mellor, 1971)

Figure 4.2

Tensile Strength of Indiana Limestone with Temperature.
(After Mellor, 1971)

Figure 4.3

Tensile Strength of Barre Granite with Temperature

(After Mellor, 1971)

Figure 4.4

relationship is not linear; Mellor (1971) explains this temperature dependence in terms of

- (i) the effect of temperature on the bulk rock materials (thermal strain and thermal activation are both reduced)
- (ii) The effect of temperature on the rock surfaces and their adsorbates (change of intrinsic surface properties of the rock, such as surface energy, and change of properties and processes in the adsorbed water which influences static fatigue).

The explanation for the increase of rock strength with decreasing temperature under saturated conditions is best summarised in Mellor's (1971) own words:

The most obvious explanation for the dramatic increase of strength below 0°C is that pore water freezes and fills cracks and pores of the rock with solid material, effectively grouting the rock. In saturated rocks of moderately high porosity, the bulk of the pore water nucleates abruptly at a temperature slightly below 0°C (Mellor, 1970). The first water to freeze is that occupying the largest pore spaces, which are probably the defects that control strength in the unfrozen state. The water which remains unfrozen after the initial nucleation occupies the smaller pore spaces, as well as a thin interfacial layer between the ice and the rock material. Thus, after initial freezing, the size of the controlling defect structures, or "Griffith cracks", is reduced and the strength of the bulk material is increased. As temperature is lowered further progressive freezing occurs, the size of the unplugged cavities is reduced, and bulk strength increases.

Fatigue may also be an important consideration in rock breakdown by frost action (Washburn, 1979). The long term

application of pressure, or a continued rhythmic cycle of applied pressure, may be important in reducing the threshold strength of the rock. Figure 4.5 shows the temperature-time dependence of compressive strength for a fine sand, and a similar relationship can be expected to exist in rocks; yet Mellor's (1970) call for more research in this area seems to have gone unanswered.

It should be noted that the above considerations refer primarily to the strength of rock in the frozen state. Although the frozen rock may increase in strength, it is likely that on subsequent thawing the strength of the rock will be reduced below its original value because of the formation of micro-cracks, fissures and other planes of weakness. Le Tavernier (1984) used ultrasound measurement to determine rock sample degradation over a number of freeze-thaw cycles; the results showed that sample alteration could occur in some cases after a single freeze-thaw cycle.

The presence of fissures, cleavage planes, grain boundaries and other flaws within the rock mass can be an important limiting factor on rock strength. The Griffith criterion for crack failure states that there is a critical length of crack, greater than which there will be spontaneous extension of the crack. This can be determined by:

$$a_m = \frac{2\gamma E}{\pi \sigma^2}$$

[4.2]

where: σ_m = maximum applied tensile stress

TEMPERATURE - TIME DEPENDENCE OF UNIAXIAL
COMPRESSIVE STRENGTH FOR OTTAWA SAND

(After Sayles, 1968)

Figure 4.5

γ = free surface energy per unit area

E = Young's modulus of the material

a = crack length (Whalley et al., 1982)

Critical crack length for any rock type may be determined if the appropriate values in [4.2] are known. Whalley et al. (1982) suggest approximate values for a granite of $\gamma = 20 \text{ J m}^{-2}$, $E = 4 \text{ MN m}^{-2}$ and $\nu = 70 \text{ GN}$, giving a critical crack length of 5cm. However, because the Griffith criterion requires a homogenous, brittle material and an extremely narrow crack tip, this value should be viewed with caution. Rock breakdown may only occur if there is a profusion of such cracks (Brace and Bombolkis, 1963), although frost action may well be important in initiating some of these cracks.

4.4 Measurement of Moisture in Rocks

It is evident that both the initial rock water content and the amount of moisture that freezes at any given temperature (the freezing characteristic curve) are important considerations in any analysis of frost shattering. Determining these conditions has been difficult, however, and has been a major limitation on frost shattering studies. In situ measurements of moisture content are beset by a number of problems (Lautridou and Ozouf, 1982), although advances in building research may prove useful in the future. Laboratory techniques have been developed for determining the freezing characteristic curve with varying degrees of success; these will now be discussed briefly with

a fuller discussion of the TDR method used in this study. 80

4.4.1 Field Measurement Techniques of Total Moisture Content

4.4.1.1 Neutron Moderation

This is a promising method developed for measuring moisture content in concrete (Malhotra, 1976; Bungey, 1982). This non-destructive technique is based on the bombardment of the concrete by a neutron emitting isotope, on the principle that hydrogen rapidly decreases the energy of fast or high energy neutrons. The interaction of fast neutrons with a matrix containing hydrogen results in retarded or scattered neutrons, which are counted and give a measurement of the number of hydrogen ions, and hence the moisture content of the concrete. Bungey (1982) points out that although the field studies have yet to be standardised the potential for the technique is considerable. When used on concrete structures, the fact that the equipment is usually calibrated for samples of semi-finite volume is undesirable, although this would be a useful attribute if the technique were used on rock masses.

4.4.1.2 Oven Drying of Field Samples

Another method of determining rock moisture levels in the field is by collection of samples from the field site, and sealing them in wax to maintain their original moisture content. In the laboratory they are weighed (with wax removed), oven-dried for 24 hours and reweighed (McGreevy,

1982). Using the sample grain size as an indicator of the apparent porosity, the percentage saturation can be determined.

In this study, 20 rock samples were collected from all three field sites, dipped in a sealing paraffin wax and transported back to the laboratory for analysis. The results are shown in Table 4.2; degrees of saturation range from 22% to 92%. Although the small sample size does not allow any statistical significance to be placed on the results, the measurements do indicate that the range of saturation levels found in the field is considerable. It is also evident that the fine-grained and coarse grained alaskite units are generally more saturated than the medium grained units, which may be due to the increased roughness and fissuring on these surfaces, compared to the smooth, ventifacted surface of the medium grained alaskite. McGreevy (1982) carried out a similar study of quartzite and schist samples from the Scottish Cairngorms and his results led him to question the validity of White's (1976) contention that few rocks in alpine areas ever become over 50% water saturated. In McGreevy's study, 22 out of 26 samples had saturation levels of over 50%, and in the present study 14 of the 20 samples were above this level of saturation. Both studies suggest that White's critical value (derived from Mellor, 1970) is frequently attained, and thus frost action may still be considered an important process in cold regions.

Rock Grain Size	Dry Weight (g)	Weight Loss (g)	% Moisture Content	% Saturation
Coarse	19.37	0.07	0.36	43.37
"	171.14	1.22	0.71	88.0
"	142.92	0.89	0.62	76.9
"	60.04	0.45	0.75	92.47
Medium	607.73	1.12	0.18	22.75
"	98.13	0.45	0.46	57.87
"	16.56	0.05	0.42	52.87
"	14.31	0.07	0.49	61.25
"	65.58	0.13	0.19	24.75
Fine	157.36	1.22	0.77	50.33
"	375.57	2.97	0.79	51.63
"	20.01	0.12	0.59	39.19
"	38.57	0.48	1.25	81.37
"	23.7	0.23	0.97	63.39
"	10.61	0.04	0.37	24.18
"	25.66	0.2	0.78	50.91
"	34.27	0.3	0.88	57.18

Field Rock Saturation Levels Determined by Oven Drying

TABLE 4.2

4.4.2 Laboratory Techniques for Measuring Unfrozen Moisture Content

4.4.2.1 Dilatometry

Dilatometry involves the measurement of the volume change of the material on freezing and thawing, from which the proportions of ice and unfrozen water can be calculated (El Khoraibi, 1974). The technique has been used successfully with soils, but it is difficult (Williams, 1976). In order to determine the unfrozen water content the sample must be completely de-aired which can be a difficult procedure with rock samples. Davison and Sereda (1978) devised a similar type of experiment in which an extensometer measured directly the linear expansion of a brick on freezing.

4.4.2.2 Differential Thermal Analysis

This method is a type of calorimetry and has been used to study the freezing of pore water in rocks by Dunn and Hudec (1965) and Mellor (1970). The method involves heating and cooling a sample of rock and an inert reference material at the same rate, and recording the temperature difference between sample and reference. A phase change in a component of the test sample produces a perturbation in the record of temperature difference against time. From this the freezing point depression can be determined, and the amount of latent heat liberated can also be calculated.

4.4.2.3 Suction-Moisture Content Tests

This method enables indirect determination of unfrozen water content with temperature to be made through measurement of the moisture content within a rock sample when an external pressure is applied.

Keune and Hoekstra (1967) derived a relationship between soil water tension (or air penetration pressure) and partial free molar energy, and since the partial free energy of water and ice in a frozen medium is a function of temperature, a relationship also exists between freezing point depression and pressure differential. At high pressures mercury penetration can also be used (Mellor, 1970).

Suction-moisture content tests were carried out on 4 samples of Caen limestone and 2 samples of Nisling alaskite. The method used was air intrusion, in which saturated samples are placed under pressure for at least a day, after which they are weighed, dried and reweighed. There are a number of inherent problems of conducting such tests on rock samples (Mellor, 1970); a major one is obtaining a sample thin enough (and still intact) to use effectively on the apparatus (a thick sample will prolong the time for equilibrium to be reached). Also the roughness of the samples can lead to a scratching of the membrane surfaces used and therefore leakage of air. (Polishing the sample is not effective in that it can seal the pores). From the

2 OF / DE 2

original samples, only 2 tests were completed successfully, both on the Caen limestone: the results are shown in Figure 4.8, and are compared to those obtained by TDR in section 4.7.

4.4.2.4 Use of Ultrasound

This technique has been examined extensively at the Centre de Geomorphologie du CNRS at Caen, France. Ultra high frequency sound pulses have been used to determine the freezing point of moisture within the rock (Tourenq, 1970) and to determine the deterioration of rock during and after freezing (Le Tavernier, 1984).

The velocity of propagation of an ultrasound pulse depends on the characteristics of the medium through which the pulse is travelling. By knowing initial values of propagation velocity for air, water, ice and the dry rock sample, one can determine the freezing point depression and the amount of rock breakdown during a number of freezing cycles. The technique is obviously of great use in the laboratory, and its continued development is encouraging.

4.5 Time Domain Reflectometry

Time Domain Reflectometry (TDR) has previously been used to determine volumetric water content in frozen and unfrozen soils from the dielectric constant (see Topp et al., 1980; Patterson and Smith, 1981; Smith and Patterson, 1984). TDR has not previously been used on rocks, but it was felt that the method could be suitable for determining the freezing

TYPICAL TDR TRACE IN WET SOIL
(From Patterson and Smith, 1981)

Figure 4.6

relative permittivity (K') of the dielectric (Krause and Carver, 1973); relative permittivity can be defined as:

$$K' = E'/E_0 \quad [4.1]$$

where K' = relative permittivity of the dielectric
(dimensionless)

E' = permittivity of the dielectric (F/m),

E_0 = permittivity of a vacuum (F/m) (Patterson, 1980)

A soil or rock will behave electrically like a complex series of parallel resistance and capacitance circuits (Keller and Licastro, 1959; Selig, 1975). The dielectric constant (K') is thus often expressed in complex notation which involves capacitance and conductance. The "real" part (K') of the complex dielectric constant (K') is not strongly frequency dependent over the frequency range of 1 MHz to 1 GHz (Davis and Annan, 1977), the effective operating range of TDR. Above these frequencies, however, the dielectric loss (i.e. the total amount of energy dissipated as heat) becomes important (Olhoeft, 1976); this electrical loss term can be defined:

$$\tan \delta = [K'' + (\sigma_{dc} / \omega E)] / K'$$

where: K' = relative permittivity

K'' = dielectric loss

σ_{dc} = d.c. conductivity

ω = angular frequency ($= 2\pi f$)

E = free space permittivity (Davis and Annan, 1977)

If $\tan \delta$ is significantly less than 1.0 (as it is in low loss, homogeneous materials), then K' is approximately equal

characteristic curve for rocks.

The TDR technique is a form of pulse-reflection measurement. A broadband pulse travels down the transmission line of the TDR, and will remain unaltered as long as the characteristics of the line remain the same. Any discontinuities in the line will result in the partial reflection of the electrical pulse back toward the signal generator. Figure 4.6 shows a typical trace displayed by the TDR's cathode ray tube for a parallel transmission line in a wet soil. By determining the position of the electrical start and end points of the transmission line, the horizontal trace length can be measured. From this time of the pulse in the sample, the dielectric constant of the material can be determined.

The Tektronix 1502 TDR unit used in this study is a commercially available cable tester which provides the cable connector with a 200 mV step-voltage with an incident rise time of less than 0.1 nanoseconds.

4.5.1 The Dielectric Property and Water Content

The electrical permittivity of a substance is a measure of the degree to which the medium resists a flow of charge (relative to that of a vacuum), and is defined as the ratio of the electric displacement to the intensity of the electrical pulse that produces it (Gray and Isaacs, 1975). Since the permittivity of a dielectric is always greater than that of a vacuum, it is often expressed as being the

to K' ; the measurable portion of the complex dielectric constant (K') can be referred to as the apparent dielectric constant (K_a). At a given frequency, the magnitude of the loss is dependent on moisture content, mineralisation, and chemical composition.

The dielectric property of a rock is dependent on a number of factors: constituent minerals, their structure and density, the presence of water and ice and salt content, as well as temperature and measurement frequency (Parkhomenko, 1967). The dielectric constants of most rock forming minerals lie in the range 2 to 12 (Chernyak, 1964; Keller, 1971); for water, the dielectric constant is much higher (80.1 at 20 °C, 87.7 at 0°C). Thus the measurement of K' can provide a reliable indication of volumetric water content, and the use of TDR for such measurements in soils has been described by Davis and Annan (1977), Davis et al. (1977) and Topp et al. (1980).

Based on results from several mineral soils, Topp et al. (1980) determined an empirical relationship between K_a and volumetric water content (θ_v):

$$K_a = 3.03 + 93 \theta_v + 146 \theta_v^2 - 76.7 \theta_v^3 \quad [4.5]$$

This relationship was found to be only weakly sensitive to soil type, dry density and soluble salt content, and θ_v could be determined with a probable minimum error of $\pm 0.025 \text{ cm}^3 \text{ cm}^{-3}$. Sources of error are both in the TDR technique itself (0.015 in θ according to Patterson, 1980) and variations in

the electrical properties of different soils.

4.6 Use of TDR on Freezing Rocks

One of the major advantages of the TDR technique for soils is that it can be used, non-destructively, in the field and in the laboratory. At its present state of development for rock samples the TDR technique has only been used in the laboratory.

A coaxial transmission line was constructed to contain rock cores collected from the field (Figure 4.7). There are practical limits to the length of the coaxial line: Patterson and Smith (1981) calculated that minimum length for a medium with $K_a=3$ was 7 cm. If the line is too long (a maximum has yet to be determined, although Topp *et al.* (1980) used a 100cm line) there can be significant signal attenuation from the long travel time. In this study, the line length was determined by the size of core samples available, usually being about 10cm.

Laboratory results of K_a on two rock types (Nisling alaskite and Caen Limestone) by TDR have shown promise. Because of the very low porosity of the alaskite, initial TDR tests were carried out on a coarse grained limestone with a porosity of approximately 24%. At room temperature, the dried sample gave a K_a value of 3.51. This value is slightly higher than the dielectric constant for a completely dry soil ($K_a=3.03$), and may be due to the fact that in rock the percentage of minerals is higher and the

PROBE CONFIGURATION OF COAXIAL LINE FOR
TDR MEASUREMENTS

Figure 4.7

percentage of air filled voids lower than in a soil. The sample was saturated by boiling in water for 5 hours (taken to be maximum water content; McBurney, 1929) and K_a was determined by TDR, at room temperature, to be 11.56. Using the empirical relationship of equation [4.5], the saturated sample has a volumetric water content of $0.225 \text{ cm}^3 \text{ cm}^{-3}$, which compares well with the porosity. These results appear encouraging, although no other non-destructive method of measuring rock moisture contents was available for comparison. In this instance, then, a preferable term is apparent liquid water content.

The low porosity of the alaskite meant that the use of the TDR was somewhat limited; it was noted in the previous section that the probable maximum error in the TDR technique is $0.025 \text{ cm}^3 \text{ cm}^{-3}$ and so theoretically the technique should not be used in rocks of porosity less than 2.5%. Thus it was necessary that a sample of sufficient porosity be selected for testing by TDR. Because of its high apparent porosity, the fine-grained alaskite was selected for testing. A number of dry samples were weighed, and then boiled in water until saturated, and reweighed. The sample with the highest porosity (5.9%) was chosen for use in the initial experiments.

The oven dried alaskite sample gave a K_a value of 3.08 at room temperature, very close to the value for dry soil; the saturated sample, at room temperature, gave a K_a value of

4.04, or an apparent water content of $0.058 \text{ cm}^3 \text{ cm}^{-3}$, which again compares well with the measured porosity. The results for both rock types proved encouraging, and so freezing tests were then carried out.

4.7 Determination of the Freezing Characteristic Curve.

To obtain the freezing characteristic curve of the two rock types, the coaxial lines containing the samples were sealed in a water tight container and placed in a circulating bath of ethylene-glycol mixture. To avoid supercooling, freezing was initiated by dropping the sample to about -4.0°C ; when there was a visible change in the TDR trace length (indicating that the water was beginning to freeze) the sample was warmed to approximately -0.1°C and then left for at least 6 hours to equilibrate before any further TDR traces were recorded. This method should allow for freezing to be initiated without causing significant damage to the sample. Rate of cooling of the circulating bath mixture was approximately $0.2^\circ\text{C min}^{-1}$; the rate of cooling within the sample would obviously be reduced, affected by the thermal properties of the sample itself, the coaxial transmission line and the sealing container. Unfortunately it was not possible to monitor the rate of temperature change within the sample, nor to control the rate of its cooling.

The bath temperature was lowered by progressive steps: at temperatures between 0°C and -2°C , the temperature

differences between readings were approximately 0.3°C , since the unfrozen water content changes most rapidly within this range. Thereafter TDR traces were recorded at 1°C intervals. The samples were given at least 6 hours to equilibrate after a temperature change.

The freezing characteristic curves determined by measuring K_a using TDR for both the rock types are shown in Figures 4.9, 4.10 and 4.11. The initial tests on each rock type involved the freezing of the sample to approximately -10°C . It can be seen that in both cases the initial unfrozen water content is reduced by at least 40% between 0°C and -2°C . Beyond this temperature, the decline in unfrozen water content with temperature appears to be less. In the limestone, even at -8.75°C there was still over $0.11\text{cm}^3\text{cm}^{-3}$ of unfrozen water left in the sample according to the TDR results. In the second limestone sample, the temperature was lowered to -15°C but there was little further reduction in the unfrozen water content.

These results were compared to results obtained by suction-moisture content tests, since the relationship of suction to water content has been shown to be related to the unfrozen water content relationship for the same material (Williams, 1967; 1976). Figure 4.9 shows the temperature-water content curve for Caen limestone calculated from the suction-moisture content tests and equation [4.1], compared to the TDR results. It is evident that the TDR readings are somewhat higher than those produced by the suction-

SUCTION - MOISTURE CONTENT TEST CURVE FOR CAEN LIMESTONE

Figure 4.8

Figure 4.9

FREEZING CHARACTERISTIC CURVE FOR NISLING ALASKITE

Figure 4.10

Figure 4.11

moisture content tests, and again this may be due to the relatively high abundance of minerals within rock. The results from both techniques do indicate, however, that the majority of freezing takes place within the range 0 to -2°C . The TDR, then, provides a useful indication of the nature of the relationship between temperature and unfrozen water content, even though the values obtained should not be considered as absolute until further tests have been carried out.

In the alaskite sample similar conditions existed, although the initial water content was much lower than in the limestone (Figure 4.10). In this instance the apparent unfrozen water content was reduced by approximately 43% by the time the temperature reached -2°C . Again, even at -9°C , the unfrozen water content was reduced by only 56% of the original water content, to $0.025\text{cm}^3\text{cm}^{-3}$ according to the TDR: since this is approximately the error of the TDR in soils (Topp et al., 1980), these values at the lower end of the volumetric water content should be treated with caution.

Finally, the effects of repeated freeze-thaw on the freezing characteristic curve were investigated. The results for the limestone are shown in Figure 4.11. The unfrozen water content-temperature relationship appears to exhibit hysteresis, so that for the same temperature, the unfrozen water content of the sample is slightly lower on thawing

than on freezing. A similar effect is found during the wetting and drying of soils. The phenomenon is partly due to the shape of the pores; if the pores are assumed to be pear-shaped, with a narrower neck than body, then as freezing occurs, ice will only invade the pore when the temperature falls to the critical value for the neck, determined by equation [2.1]. Thus the unfrozen water content will be higher than the theoretical value calculated from pore size. On thawing, at the same temperature, the ice in the neck of the pore will melt, whilst the ice in the body of the pore remains frozen so that the unfrozen water content is slightly lower. Hysteresis may also be due to variations in the contact angle between the water and the pore surface (assumed to be 0°) on freezing and thawing. The implications of hysteresis are that the freezing-thawing (or wetting-drying) history of the rock sample may affect the freezing characteristic curve of any given rock type or sample.

4.8 Discussion of Results

The results from the TDR and the suction-moisture content tests indicate that as freezing occurs, considerable suctions are generated within the rock. However, it is apparent that most of the phase change occurs within the range of 0° to -2°C , suggesting that frost action should be most important at these temperatures. Hence temperature oscillations near 0°C may be sufficient to effect some rock breakdown. At lower temperatures, the rate of ice formation

appears to be considerably reduced. From the freezing characteristic curve, the moisture content of the limestone determined by TDR remains above $0.11 \text{ cm}^3 \text{ cm}^{-3}$; similarly at suctions of up to 10^{-7} N m^{-2} , the limestone still maintains a water content of just under 10% (dry weight); this water is obviously held under considerable suctions and may be the adsorbed water noted by Dunn and Hudec (1966). Mellor (1970) showed that rock strength increases as temperatures decrease; the effect is especially noticeable in saturated porous rocks where the ice can act as a cementing agent (Figures 4.3 and 4.4). In soils, the high suctions developed at low temperatures (or low water contents) can lead to an increase in soil strength as the soil particles are held together under increased force. In rocks, there is less likelihood of compaction, although these suctions may lead to considerable stresses being set up within the rock. This would be especially the case in the zone adjacent to the freezing front, which is weaker than the frozen area but may experience the suctions developed in the frozen zone. If these suctions are high enough they may overcome the tensile strength of the rock and cause cracking. Table 4.3 shows the highest temperatures required by rocks of different strengths to effect breakdown, based on the theoretical relationship between temperature and suction (equation [4.1]), and the rock strength classification of Attewell and Farmer (1976). The table is derived from Table 3.3 and the tensile strength is calculated as 10% of the compressive strength. No account is

ROCK CLASSIFICATION	TENSILE STRENGTH	MAXIMUM EFFECTIVE TEMPERATURE
Very Weak	0.5-2.0 MPa	-0.4 °C
Weak	2.0-4.0 MPa	-2.6 °C
Med. Strength	4.0-8.0 MPa	-3.3 °C
Strong	8.0-16.0 MPa	-6.6 °C
Very Strong	16.0-32.0 MPa	-13.3 °C

Maximum Effective Temperatures for Rock Breakdown.
(For explanation see text)

TABLE 4.3

Material	Temperature	dL/Lo %	$\alpha \cdot 10^{-6} \text{ K}^{-1}$
Ice	0 to -23 °C	57 - 50	0 to -0.125
Quartz	+20 to -23 °C	10.3 - 9.4	0 to -0.044
Silicon	+20 to -23 °C	2.6 - 2.2	0 to -0.01
Iron Oxide	+20 to -23 °C	12 - 1.9	0 to -0.05
Calcium	+20 to -23 °C	23.3 - 20.6	0 to -0.2

α = Coefficient of Thermal Linear Expansion

$\alpha = d/dT (dL/Lo)$ where: T = Temperature
dL = Length Change
Lo = Original Length.

Thermal Contraction of Some Common Materials
(Data from Touloukian et al., 1970)

TABLE 4.4

taken of the increase of strength on freezing, since this varies significantly between rock types and depends on the condition of the rock (i.e., whether it is saturated or not), and the values given therefore represent maximum temperatures. In some instances, the increase of tensile strength on freezing can be over 100% (Mellor, 1971). The tensile strength of a rock mass will vary from point to point, as well as being gradually reduced over time by fatigue and increased fissuring, and so critical values must necessarily relate only to a specified rock sample in time and space. The table does nonetheless give an indication of the relative temperatures required for rock breakdown to occur.

At low temperatures (below approximately -10°C) the stresses produced by the suction may be compounded by the thermal contraction of the constituent minerals. As the amount of water that freezes is reduced at lower temperatures, so the expansion caused by freezing (through movement of water to the freezing front and/or volumetric expansion) becomes less important and thermal contraction of the minerals may become the dominant process. Table 4.4 gives the coefficient of thermal expansion for ice and a few minerals at negative temperatures. The variation of thermal contraction between minerals can lead to stresses being developed between mineral particles and subsequently granular disintegration may occur.

4.9 Summary

The stresses generated within a rock on freezing are instrumental in causing rock breakdown. Given sufficient moisture availability and low temperatures, the mechanism by which these stresses are produced will depend on the rock type. In relatively impermeable rocks, the 9% volumetric expansion may exert enough pressure to cause significant strains and internal cracking (Mellor, 1970). In a permeable rock, ice segregation may be a more important mechanism in rock breakdown.

Time Domain Reflectometry appears to be a useful method for determining the freezing characteristic curve for two rock types. The curves determined for the limestone were compared to suction moisture content tests and indicated that considerable suctions are generated within the rock as the temperature falls. It is these large negative pore water pressures which are the driving force for moisture migration towards a freezing front, providing the rock is of sufficient permeability. Under such conditions it is the addition of moisture at the freezing front that causes rock breakdown.

It appears that most of the moisture freezes between 0 and -2 C in the two cases examined, and hence it is at these temperatures that most of the freeze-thaw activity takes place. In this temperature range, stresses developed by 9% expansion and by ice segregation will be the primary

mechanism of breakdown. Below these temperatures, as the amount of moisture that freezes is reduced, and hence the associated volumetric expansion is reduced, other processes may contribute to the breakdown process. High suctions developed at low temperatures, as well as the thermal contraction of minerals, may lead to the generation of additional stresses, enhancing breakdown.

CHAPTER 5

CLIMATIC CONTROLS ON FROST SHATTERING

5.1 Introduction

A complete understanding of frost shattering requires a knowledge of the thermal regimes in which the process operates. Temperatures that allow pore water to freeze are obviously a prerequisite, and these will be lower as the pore size of the rock gets smaller, although in the previous chapter it was shown that most of the moisture apparently freezes in the temperature range 0° to -2°C . Other parameters may therefore be equally important, such as the duration and rate of freezing (see chapter 2; also McGreevy and Whalley, 1982). This chapter draws on field observations of air and rock temperatures from the Ruby Range to analyse the factors which can influence the thermal regime of a rock face. But, because of the brevity of the field season and thus the range of field data available, a computer program has been used ~~to~~ simulate year round climatic conditions. This enabled a fuller investigation into those factors affecting freezing intensity, duration and rate.

5.2 Factors Affecting the Thermal Regime of Rock

The thermal regime of a rock mass is affected by three primary factors - position, surface conditions, and the rock thermal properties. The general climatic conditions will depend on the position of the rock face, which includes latitude, altitude, aspect and slope angle. These all influence the amount of incoming solar radiation reaching the surface of the rock mass. The nature of the surface boundary affects the thermal regime below the surface, and may indeed override larger scale macroclimatic factors. Surface roughness, wetness, colour and albedo, as well as the effects of vegetation and micro-organisms (especially lichen) are all important in the ground heat flux. Finally, the propagation of surface temperature fluctuations through the rock depends on the thermal properties of the rock mass, namely thermal conductivity and heat capacity (i.e., diffusivity).

The influence of these three primary factors can be usefully examined through the surface energy balance equation:

$$Q^* = Q_h + Q_e + Q_g + dQ_s \quad [5.1]$$

where Q^* = net all-wave radiation flux density

Q_h = turbulent sensible heat flux density

Q_e = turbulent latent heat flux density

Q_g = sub-surface heat flux density

dQ_s = change of energy storage (Oke, 1978)

These factors will now be examined in detail below.

5.3 Field Observations of Rock Temperatures

There is an unfortunate lack of bedrock temperature records from areas where frost shattering is known to take place (Thorn (1979) cites only 2 studies besides his own: Hall (1975) and Gardner (1969)). This situation is not surprising given the problems of bedrock temperature measurement (see Goudie (1981) for a review of suitable techniques) as well as the difficulties of access to high altitude or latitude areas during late fall, winter or early spring, when frost shattering is likely to be greatest (Whalley et al., 1984). This has resulted in a tendency to use air temperatures as indicators of the thermal regime of the rock face. Such an approach is unfortunate and can be misleading since bedrock temperatures can vary considerably from the ambient air temperature. For example, Figures 5.1 and 5.2 show diurnal variations in air and bedrock temperatures for two different aspects at sites 1 and 2; it is evident that bedrock temperatures are influenced by factors such as aspect, altitude and other, site-specific, factors which can cause significant differences between air and rock temperatures. It is therefore important to analyse those factors that affect the thermal regime of rock and their influence on frost shattering conditions.

Temperature data were recorded for the field period at each of the thermistor sites within the 3 field sites

Figure 5.1 AIR AND ROCK TEMPERATURES, SITE 1

Figure 5.2 AIR AND ROCK TEMPERATURES, SITE 2

described in chapter 3. Tables 3.6, 3.7 and 3.8 give descriptions of each thermistor location. Although the field season lasted only 6 weeks from June 1st to mid-July, a period when the temperature rarely dropped below freezing, the data may be used to investigate specific climatic relationships and principles pertinent to this study. Specifically, the effects of altitude and aspect are analysed in relation to the magnitude and rate of diurnal temperature variations. The data also provide a useful basis for evaluating theoretical models that have been developed to characterise the thermal regime of earth materials.

5.3.1 General Climate

Although air temperatures may not necessarily provide a good indication of the ground thermal regime, it is true to say that the macro-factors which affect the energy balance at the ground surface are important in any consideration of the climate below the surface. The amount of solar radiation incident at the earth's surface will vary with latitude, so that generally there is a decrease in mean annual temperature with an increase in latitude. However, attempts to define areas of frost shattering based on temperature (and /or moisture) data on a continental scale have met with limited success (see, for example, Russell, 1943; Peltier, 1950; and Tricart and Cailleux, 1965), and it is evident that other factors such as altitude, aspect and slope angle may be equally important in influencing the climatic

conditions experienced on a rock face.

5.3.1.1 Altitude

Increasing altitude is associated with decreasing temperatures -the rate of decrease being controlled by the temperature lapse rate in the prevailing air mass. Figure 5.3 shows mean temperatures observed at the 3 field sites in the Ruby Range with respect to aspect. As expected the high altitude site (3) exhibits the lowest mean temperatures. However, sites 1 and 2 do not conform to the expected pattern (temperatures at site 2 are generally higher than those at site 1). There are a number of possible reasons for this: -site 2 was more exposed than site 1, or incoming solar radiation at site 2 may have been supplemented by reflected radiation from nearby snowpatches, or site 1 may have been in an area of cold air drainage. Whatever the specific factor(s), it is evident that at these two sites the altitude influence is over-ridden by other, site-specific, factors. Harding (1978) also showed that whilst the altitude gradient is a major control on mean maximum temperatures, local topography, vegetation and soil (or rock) type may be at least as important for minimum temperatures.

Rate of freezing may be also affected by altitude. Solar radiation increases with altitude (due to the corresponding increase in atmospheric clarity), so that the more intense radiation that occurs at higher altitudes may serve to

Figure 5.3 MEAN DAILY TEMPERATURES AT RUBY RANGE SITES

increase the diurnal range of temperatures experienced at rock surfaces (Whalley et al., 1984). Table 5.1 and Figure 5.4 show rates of cooling observed in the Ruby Range. Again, although rates are generally higher at site 3, the higher range of temperatures at site 1 means that the mean maximum rates of cooling are slightly higher than those at site 2. Note that the rates are for cooling above 0°C ; when the temperature drops below 0°C , rates may be slower because of the "zero-curtain effect" as freezing occurs (Johnston, 1981). Even so rates of cooling were significantly lower than the critical rate of $6.0^{\circ}\text{C}.\text{hr}^{-1}$ proposed by Battle (1960). Other field studies also indicate that this rate is seldom attained: Hall's (1975) work in Northern Norway recorded only 4 occasions during three years of spring and summer observations when the rate of freezing was as great as $0.1^{\circ}\text{C}.\text{min}^{-1}$, and on these occasions the temperature did not fall below -1.5°C . Similarly, Thorn (1979) measured maximum freezing rates in the Front Range, Colorado, and found that freezing rates were a full order of magnitude below Battle's critical value. McGreevy (1982) noted that Battle's criterion was based on undisclosed experimental conditions and may not be appropriate to field conditions.

5.3.1.2 Aspect

The influence of aspect is also important and has been well documented: Barsch (1977) observed a noticeable increase of apparently frost shattered material on north-

Thermistor	Aspect	Mean Cooling Rate °C.hr ⁻¹	Maximum Cooling Rate °C.hr ⁻¹
1A	ESE	0.32	0.80
1B	S	0.24	0.99
1C	WSW	0.61	1.17
1D	NNW	0.24	0.62
1E	SE	0.68	1.35
1F	SSE	0.37	1.16
1G	WSW	0.73	1.96
MEAN		0.46	1.15
2A(iv)	WSW	0.22	0.85
2B	W	0.38	0.46
2C	WSW	0.27	1.00
2D(i)	NE	0.29	0.33
2D(ii)	NE	0.53	0.62
2D(iii)	NE	0.33	0.38
2E	E	0.93	1.30
2F	NNW	0.56	0.95
MEAN		0.44	0.74
3A	ESE	0.55	2.18
3B	NE	0.35	1.36
3C	S	0.93	3.66
3D	W	0.62	1.88
3E	E	0.51	1.62
3F	NNW	0.52	1.85
3G	ENE	0.49	1.55
MEAN		0.57	2.01

Mean Rates of Cooling at Ruby Range Sites

TABLE 5.1

Figure 5.4 Mean Rates of Cooling Observed In the Ruby Range

Figure 5.5 DIFFERENCES OF ROCK ASPECT AND SLOPE ASPECT THAT MAY CAUSE COMPLICATIONS IN PREDICTING THE THERMAL REGIME

facing slopes compared to south-facing ones in the Swiss Alps, and Mathys (1974) and Price (1973) both showed that depth of freezing beneath the ground surface varied considerably with aspect. An important consideration in this respect is that the rock face may frequently differ from the aspect of the slope on which the outcrop occurs (Figure 5.5). Shading of the rock face will reduce the amount of incoming radiation at certain times of the day or year, so that seasonal and diurnal temperature variations may be affected. In this respect, the nature of the topographic horizon can also be important.

Temperature data collected in the Ruby Range were analysed to investigate the effect of aspect on bedrock temperatures. The data were somewhat limited in that the depth of the thermistors was not uniform, and the slope angle of the rock face into which the thermistors were placed varied considerably. The idealised relationship between rock temperatures and aspect is an inverted parabola, with the maximum at the south, slightly skewed due to greater warming in the afternoon when the sun is in the west (see Figure 5.6). Figures 5.7 to 5.9 show the daily mean, maximum and minimum temperatures from the 3 field sites, averaged from the days when more than 10 readings were taken. Lines were fitted to the data points by eye, and indicate that whilst at site 1 the general pattern expected is observed, at sites 2 and 3 the effect of aspect is either not apparent (site 2) or considerably dampened (site 3).

Figure 5.6 RADIATION AS A FUNCTION OF SLOPE AND ASPECT
Radiation index expressed as a percentage of radiation for horizontal surface.
(After Frank and Lee, 1985)

Figure 5.7 SITE 1 TEMPERATURES WITH RESPECT TO ASPECT

Figure 5.8 SITE 2 TEMPERATURES WITH RESPECT TO ASPECT

Figure 5.9 SITE 3 TEMPERATURES WITH RESPECT TO ASPECT

Again, overriding factors such as local shading, or the nature of the rock surface may be more important at these sites. Rates of cooling, however, showed a much more consistent pattern for all 3 sites (see Table 5.1 and Figure 5.4). In general, rates were highest on the south easterly, south and south westerly aspects, with lower rates for the north and east. This is particularly noticeable at sites 1 and 3. It is also evident that rates are slightly higher on the west facing slopes than on east facing ones; on easterly aspects the rock temperatures begin cooling while the ambient air temperature is still rising, thus reducing rates of cooling. On westerly facing slopes, the rock and air temperatures are both cooling together in the late afternoon and hence rates are usually higher. At site 2 the relationship between aspect and cooling rate is not so well defined; although the highest rate of cooling is found on the east facing slope, this thermistor gave repeatedly high readings and is thought to be erroneous. The other anomaly at this site is thermistor 2F, on a NNW face, which has a relatively high rate of cooling of $0.95^{\circ}\text{C}.\text{hr}^{-1}$. There is no apparent reason for this high rate, although the thermistor was located in a vertical face, sheltered in part by an overhang, and it can be speculated that this situation exaggerated the "normal" temperature range. Alternatively the thermistor may be erroneous.

Practical considerations limited the number of

thermistors and their location and it is realised that the field data are not comprehensive and further field work would be useful. However, the data recovered from the field indicated that although factors such as altitude and aspect can be important influences on the thermal regime, their precise role is often obscured by noise from other factors.

Because of this, a computer model was used to supplement the field data, and to clarify further the role of aspect, as well as the effects of saturation levels and rock thermal properties on rock temperature profiles.

5.4 Computer Simulation Model

The numerical microclimatic model used was a version of that presented by Smith (1977) to simulate active layer development in the High Arctic and also to predict frost penetration beneath highways in southern Canada (Smith and Tvede, 1977). By considering the surface energy balance, the model calculates an equilibrium surface temperature from given microclimatic and atmospheric data. A ground thermal profile is calculated using the specified thermal properties in the thermal diffusion equation, with the surface temperature as one of the boundary conditions. A simplified flow chart of the program is shown in Figure 5.10.

The model requires that the site-specific factors of surface albedo, surface roughness and a wetness factor are specified. The model uses the site wetness parameter described by Stewart and Rouse (1976) which is equal to 1.26

Figure 5.10

SIMPLIFIED COMPUTER PROGRAM FLOW CHART
(After Smith, 1977)

for saturated conditions. For the simulations comparing aspects and those comparing rock thermal properties these site-specific factors were kept constant: the site wetness factor was the mean of the values found in the field and reported in the previous chapter (i.e. $56.38\% = 0.716$). When the effects of saturation were investigated, the site wetness factor and the albedo were varied. Since the model was used to determine the thermal profile in a vertical rock face, snow accumulation was assumed to be negligible. The rock thermal properties for the two rock types used in the simulation (alaskite and limestone) were taken from Jessop *et al.* (1984) and Forsythe (1969).

The climatic data used in this study were taken from the climatological records for Kluane Lake (Environment Canada, 1982a; 1982b). A microcomputer model (Patterson, 1977) was used to compute incoming solar radiation values for a specified latitude, aspect and azimuth.

5.5 Computer Simulation of the Effect of Aspect

Values for incoming solar radiation on an hourly basis were entered into the program for north, south, east and west aspects. The temperature profiles from the east and west aspects were very similar, and the greatest variation between rock temperatures was between the north and south aspects, as would be expected. This was especially noticeable in the fall and spring months. Temperature profiles for April and September are shown in Figure 5.11.

Figure 5.11 Temperature Profiles in Alaska

c) January Taütochrome
Figure 5.11 Temperature Profiles in Alaskite

and can be compared to a mid-winter profile (Figure 5.11c) when the temperature difference between aspects is minimal. During September, when air temperatures are warmer than those in April and the rock mass as a whole is cooling from the summer warming, the greatest daily range in temperature is found on the cooler north face. In April when the rock mass is warming (but still below 0°C at depth), the greatest daily range of temperatures is found on the southerly aspect. Since rates of cooling will largely be determined by the diurnal range of temperatures, these are also greatest on the north face in September ($0.43^{\circ}\text{C}\cdot\text{hr}^{-1}$ at 5 cm) and on the south face during April ($0.59^{\circ}\text{C}\cdot\text{hr}^{-1}$ at 5 cm). Hence, the precise nature of the effect of aspect on the ground thermal regime will depend on the season, particularly with respect to the rate of freezing.

The effect of aspect on the frequency of freeze-thaw cycles (defined here in terms of oscillations around the 0°C isotherm) shows a similar relationship to season. Table 5.2 shows freeze-thaw incidence for north, east, west and south aspects for the months of April and May in the spring and September and October in the fall. These are the most active months in terms of freeze-thaw activity. It is noticeable that freeze-thaw oscillations are most frequent on the easterly aspect, whilst north and south aspects remain either completely frozen or unfrozen for at least two of these months. Thus, whilst the intensity and duration of

Aspect	April	May	September	October
North	Frozen	F/T Oscil. +4° - -4.5°C	F/T Oscil. +3° - -5°C	Frozen
East (West)	F/T Oscil. +3° - +6°C	F/T Oscil. +7° - -1.8°C	F/T Oscil. +6° - -2°C	F/T Oscil. +0.5° - -5°C
South	F/T Oscil. +8° - -2°C	Unfrozen	Unfrozen	F/T Oscil. +4° - -3°C

Key: F/T Oscil. = Diurnal Freeze-Thaw Oscillation across the 0°C Isotherm.

FREEZE-THAW OSCILLATIONS ON DIFFERENT ASPECTS
DURING SPRING AND FALL MONTHS.

TABLE 5.2

may be greatest on north facing slopes, freeze-thaw frequency is greatest on east and west facing slopes, and this may affect the process-by which the rock face is weathered.

5.6 Effects of Surface Conditions

The characteristics of the rock surface can significantly influence the thermal regime below the surface. Detailed empirical analysis of the surface boundary layer is complicated and was not carried out in this study: instead, this section will focus on the theoretical implications of various surface conditions.

5.6.1 Surface Roughness and Presence of Cracks and Fissures

Surface roughness is important for a number of reasons: a rough surface increases the surface area from which moisture can be evaporated and can also increase turbulence across the surface of the rock. Both processes serve to increase the flow of heat away from the surface, thus cooling it.

In rock faces where cracks and fissures occur (usually the norm rather than the exception) the thermal regime of the rock mass will be complicated further by localised shading and trapped air pockets. Figure 5.12 shows a hypothetical situation which illustrates the differences in incoming solar radiation that may occur in adjacent cracks on a rock face due solely to the angle of the crack. Given uniform heating and cooling conditions, McGreevy (1982) noted that crack width is probably the most important

INCIDENT RADIATION

SOUTH FACING
ROCK SURFACE

Crack A

Crack B

Figure 5.12 HYPOTHETICAL SITUATION TO ILLUSTRATE POSSIBLE COMPLICATIONS
IN INTERPRETATION OF CRACK TEMPERATURE REGIMES (After McGreevy, 1982)

controlling factor: narrow cracks would experience more rapid temperature changes and more extreme temperatures than wider cracks. This would be due to surface heating by direct surface radiation raising the temperature of the air adjacent to the rock surface, whereas air temperatures in wider cracks would be similar to the ambient air temperature.

McGreevy (1982) reports on a short experiment designed to investigate the freezing of bulk water within cracks. He showed that the freezing temperature in small holes (0.5, 0.4, 0.3, 0.2 cm diameter, 5 cm length) drilled in perspex was usually between -0.6°C and -2.5°C . Although the presence of clay minerals along crack walls in rocks will reduce the freezing point of water in these cracks (Anderson et al., 1974), McGreevy commented that:

...it may be inferred from the experiment and the value of -3.3°C (obtained as the freezing temperature of water in solid sandstone) that whilst minimum temperatures experienced in cracks are not as low as in solid rock, these may often be sufficient to cause freezing.

In the 5cm hole, freezing at the top of the hole began at 0°C , and at -2.2°C at the base of the crack suggesting that complete freezing of water in a crack may not be a requirement for frost action to occur (see also Branson, 1973). It may also be concluded that frost damage can also occur at the rock surface by the freezing of bulk water in cracks without the intensity of freezing that may be required for rock breakdown by phase change of pore water

within the rock.

5.6.2 Moisture Conditions

Rates of evaporation or sublimation will depend on the moisture conditions at the surface and within the rock, as well as the relative humidity. The evaporation term (Q_e) in the surface energy balance equation [5.1] may assume a dominant role if the surface of the frozen rock is "wet" (0°C , isothermal) whilst the air temperature is above freezing if water availability is low or restricted, the value of Q_e will drop and become zero for a completely dry rock surface (Oke, 1978). Any water that percolates into the rock through cracks and fissures before freezing may act as energy source (in just the same way as rain can if it percolates into a frozen medium) and dQ_s may become significant over short periods of time.

Moisture conditions within the rock mass also affect the thermal properties of the rock and are an important control on the propagation of surface temperatures into the rock mass. These effects are considered further in section 5.7.1.1

5.6.3 Albedo

The albedo of the rock surface and its immediate surroundings has an important influence on the surface net radiation. Light surfaces reflect much of the incoming solar radiation whereas darker surfaces absorb most. Rock surface colour is often altered significantly by moisture: usually

the rock surface becomes darker with increased wetness (Ahmad and Lockwood, 1980). This is noticeable at thermistor B at site 1, where the damp rock face produces a darker surface, although because the thermistor is sheltered in a gulley temperatures do not seem to be significantly affected. A more noticeable effect can be seen at thermistor F, site 3; this rock face was almost completely covered by lichen and remained moist throughout most of the field season due to melting snowpatches on the ridge above, resulting in a very dark north westerly face. Maximum temperatures were higher than any others recorded at site 3 (see Figure 5.11), and rates of cooling were also high.

Vegetation on the rock surface may also be important in this respect: lichen can affect the colour, moisture availability and the surface roughness; plants reduce incoming solar radiation by shading parts of the rock face, and may also utilise any surface moisture available. Such effects are obviously very localised but will influence thermal conditions in the rock mass. Also, reflected radiation from adjacent steep facing slopes, especially if covered by snow and ice, can serve to increase the incoming radiation on the rock face in question.

5.7. Thermal Properties of Rocks

Temperature fluctuations at the surface of a rock or soil are damped and lagged as they penetrate deeper into the solid material, and the extent of the damping and lagging

will depend primarily on the thermal properties of the rock or soil. The thermal diffusion equation describes the variation of temperature, in both space and time, within a medium:

$$dT/dt = D \cdot d^2T/dx^2 \quad [5.2]$$

where T = temperature

t = time

D = diffusivity

x = depth (Hillel, 1982)

This describes the rate of change of temperature with time in terms of the rate of change of the temperature gradient and the diffusivity of the medium. Diffusivity can be defined as the ratio of the rate of heat flow within a material (thermal conductivity, K) to the amount of heat required to raise the temperature of the material (heat capacity, C). Hence:

$$D = K/C. \quad [5.3]$$

The thermal conductivity of a material is defined as the amount of heat conducted through a unit area in time under a unit temperature gradient (Hillel, 1982). It is dependent on the constituent minerals, the rock porosity, moisture content, and temperature. Table 5.3 gives the thermal conductivities of some rocks and minerals, as well as other common materials. Thermal conductivity varies with the volumetric proportions of component minerals, the high conductivity of quartz (8.4 W.m⁻¹K⁻¹) being particularly

Rock Type	Volumetric Heat Capacity (C) $10^6 \text{ J.m}^{-3}.\text{C}^{-1}$	Thermal Conductivity $\text{W.m}^{-1}.\text{C}^{-1}$	Source
Granite	2.1	1.7-4.0	Johnston, 1981
Sandstone	2.2	1.8-4.2	"
Basalt	2.3	1.8-2.2	Forsythe, 1969
Limestone	2.5	0.9-2.5	"
Shale	1.5	1.6	"
Quartz	2.0	8.4	"
Mica	3.0	2.4	Johansen, 1975
Felspar	2.0	2.0	"
Ice	2.0	2.3	Kaye & Laby,
Water	4.2	0.56	1973

Typical Heat Capacities and Thermal Conductivity
Measurements for Rocks and Minerals.

TABLE 5.3

important. Since there is usually minimal compaction of rock structure, at least in the upper layers of the earth's crust, thermal conductivity does not normally vary significantly with depth. The thermal conductivity of water ($0.57 \text{ W.m}^{-1}\text{K}^{-1}$) is much higher than air, therefore moisture content is the main short term variable. As moisture content increases, conductivity also increases through increased thermal contact and air expulsion. The thermal conductivity of ice is approximately four times that of water ($2.2 \text{ W.m}^{-1}\text{K}^{-1}$), and so the thermal conductivity of saturated porous media will also increase as freezing occurs. The importance of understanding the nature of the freezing characteristic curve is obvious in that as the proportion of ice to water varies with temperature, so the thermal conductivity will also vary, altering the flow of heat within the material. In general, when the rock pores contain ice rather than water, K will be about 14% higher for rock at 10% porosity and 30% higher at 20% porosity (Johnston, 1981). At temperatures close to 0°C , the rapid change of ice and water within the rock with a small temperature change means that the thermal properties of the material also change significantly, making their determination difficult (see Riseborough et al., 1984).

Diffusivity is also dependent on the volumetric heat capacity of the material, which can be defined as the change in heat content of a unit bulk volume of material per unit change in temperature (Hillel, 1982). It is a measure of the

amount of energy required to raise the temperature of the rock. Again it is dependent on the mineral composition of the rock, bulk density and particularly the wetness, since in this instance the heat capacity of water is considerably higher than that of most rock forming minerals (see Table 5.3). When the addition or removal of heat to or from the rock mass involves a phase change, then some heat energy is used in the vapourization or fusion process as latent heat. In frozen materials, the apparent heat capacity term (C_a) includes the latent heat of fusion since it is part of the energy requirement to effect the temperature change. Because of the unfrozen water content relationship C_a is temperature dependent below 0°C , changing rapidly at temperatures near 0°C .

The total thermal response of a rock is therefore directly proportional to its conductivity (K), but inversely proportional to its volumetric heat capacity (C). The thermal diffusivity (D) can be regarded as a measure of the time required for temperature changes to travel through the rock mass. As previously stated water content is the important short term variable. Initially the addition of water to a dry rock produces a sharp increase in D ; however, beyond approximately 20% volumetric water content, D begins to decline because whereas K levels off with increased moisture content, the value of C continues to rise at high moisture levels (Oke, 1978). A rock with a high

diffusivity will allow the rapid penetration of surface temperature changes into the mass. On the other hand, low diffusivity will result in a concentration of thermal exchanges in the upper most layers of the rock and consequently it will experience relatively extreme diurnal temperature fluctuations.

5.7.1 Computer Simulation of Different Thermal Properties

5.7.1.1 Effects of Moisture Conditions

Since, undoubtedly, saturation levels in the field vary between sites, and moisture content is such an important element in the thermal properties of rock, the computer model was used to investigate the effects of different moisture conditions on the rock thermal regime. The same climatic conditions were imposed on both saturated (90% saturation) and dry (20% saturation) alaskite. The albedo of the saturated alaskite was altered to 0.16 to reflect the increase in water content, and the albedo of the dry rock was increased to 0.26 (values taken from Smith, 1977). The results are best described in terms of the penetration of the frost line shown in Figures 5.13 and 5.14. The effects of the higher apparent heat capacity of the wet rock can be clearly seen in retarding the onset of freezing and reducing the intensity of freezing. The increased conductivity of the wet rock is also reflected in the earlier thawing as the warm morning temperatures quickly penetrate into the rock. Although the degree of saturation seems to have little

Figure 5.13 APRIL FROSTLINE PENETRATION EAST FACE
WET AND DRY ALASKA

Figure 5.14 SEPTEMBER FROSTLINE PENETRATION NORTH FACE
WET AND DRY ALASKITE

effect on the daily maximum temperatures (the increased albedo of the wet surface appears to be offset by increased evaporation), it is noticeable that the dry alaskite experiences considerably colder minimum temperatures and the increased range of temperatures also means that rates of cooling are increased in the dry rock ($0.69^{\circ}\text{C hr}^{-1}$ compared to $0.6^{\circ}\text{C hr}^{-1}$ in the wet rock). Thus although increased saturation may be important to produce the required pressures to effect breakdown, higher moisture levels may indeed inhibit freezing and reduce susceptibility to frost shattering conditions.

5.7.1.2 Effects of Rock Type

The computer simulation model was also used to investigate the effects of different rock types (with varying thermal properties) on the ground temperature profile. This time the same climatic conditions were imposed on two different rock types - the Nisling alaskite and a limestone (since this rock type had already been used in this study). Thermal properties of the two rock types were determined from the literature (Jessop, 1984, and Forsythe, 1969, respectively). Moisture conditions, roughness and albedo were assumed the same for both rock types, although undoubtedly under natural conditions these would vary: this was done so that specifically the effects of the different thermal properties could be seen.

The influence of rock type is again best illustrated by

looking at the penetration of the frost line into the rock mass. Figures 5.15 and 5.16 show the 0°C isotherm for September (north face) and April (east face). A number of points should be noted. The higher conductivity of alaskite means that temperature fluctuations at the surface are dissipated throughout the body of the rock giving higher day temperatures (retarding the onset of freezing) and generally lower minimum temperatures at night (see September isotherm). In April this is not the case since the warmer morning temperatures also penetrate into the rock quickly, reducing the extent of freezing. The higher heat capacity of the limestone retards the thawing of the rock in the morning hours, especially noticeable in April, prolonging the duration of freezing. Thus, a higher thermal conductivity may increase the intensity of freezing, but may also reduce the freezing duration; on the other hand, a high heat capacity will reduce the depth of freezing, but may prolong the freezing duration. Rate of freezing will also be affected by both these thermal properties: generally, rate of freezing will decrease as heat capacity increases (since more heat energy is required to alter the temperature of the rock) but will increase as conductivity increases (since the range of temperatures experienced is greater). In the simulation this is clearly visible between the two rock types: freezing is effected more quickly in the alskite (mean rate for the two months = $0.66^{\circ}\text{C}\cdot\text{hr}^{-1}$) than in the limestone (mean rate = $0.53^{\circ}\text{C}\cdot\text{hr}^{-1}$).

Figure 5.15 APRIL FROSTLINE PENETRATION EAST FACE
ALASKITE AND LIMESTONE

Figure 5.18 SEPTEMBER FROSTLINE PENETRATION NORTH FACE
ALASKITE AND LIMESTONE

The effects of thermal properties on the freezing of rocks have received scant attention in the literature, but are evidently of considerable importance in controlling the depth, period and rate of freezing. The examples given above use mean values to represent "average conditions", and extreme conditions may exaggerate these effects.

5.8 Summary

This chapter has focused on those factors that can affect the climatic conditions in which frost shattering can occur. Three factors were identified as being primary controls on the ground thermal regime - general climatic conditions, nature of the rock surface and the thermal properties of the rock mass. Within each of these factors there are a number of elements that can affect the ground thermal regime significantly. It is also evident that although macro-factors are important controls on the ground thermal regime, they can on occasion be overridden by site specific microclimatic factors.

CHAPTER 6

SUMMARY AND CONCLUSIONS

6.1 Summary

This thesis has focussed on two aspects of frost shattering -the freezing of moisture within rocks, and the factors that affect the thermal regime of a rock mass. Time Domain Reflectometry (TDR) was used to examine the freezing of moisture in two rock types in the laboratory, and the role of climate was examined in the field and through the use of a computer simulation model.

The field aspect of the study was carried out in the Ruby Range, Yukon Territory where three exposures of a coarse grained granodiorite known as alaskite were instrumented to monitor bedrock and air temperatures at a variety of aspects and altitudes. Samples of the bedrock were taken from the field for laboratory determination of water content.

The climatic data were supplemented through use of a computer simulation model designed to predict the ground thermal profile for a given set of conditions. This

facilitated an analysis of the effects of aspect, levels of saturation and rock thermal properties on the ground thermal profile on a seasonal basis, particularly as they relate to the penetration of frost.

In the laboratory, freezing characteristic curves for two rock types, Nisling alaskite and Caen limestone, were determined by TDR, a method previously used to determine moisture content in soils. This method uses the difference in the dielectric constants of ice and water to determine the dielectric constant of the freezing rock. From this value the unfrozen volumetric water content can be calculated. TDR results were compared to suction-moisture content tests conducted on the limestone; attempts at similar tests on the alaskite failed.

6.2 Results

The TDR technique appears to be an effective method of determining the unfrozen water content in rock. Two samples of Caen limestone and Nisling alaskite were tested using the TDR, and the results for the limestone were compared to those determined by suction-moisture content test. These results showed that the unfrozen water content values obtained by TDR were slightly higher ($0.025 \text{ cm}^3 \cdot \text{cm}^{-3}$) than those determined for the equivalent suctions by the suction-moisture content tests. The TDR indicated that over 40% of the initial moisture content freezes at temperatures between 0 and -2°C in both samples. In the limestone, freeze-thaw

hysteresis was also observed. Even at -15°C almost half of the initial moisture content remains unfrozen and is under considerable suction, probably as adsorbed water. If the permeability of the frozen rock is high enough, then these suctions could cause the migration of moisture towards the freezing front, with subsequent ice segregation. If, on the other hand, the permeability of the rock prohibits movement of moisture, the pressures exerted are reduced and correspond to the 9% volumetric expansion of water on freezing. In either case, the majority of the pore water in the rock apparently freezes at between 0°C and -2°C , and it is at these temperatures that frost action will be most effective.

Volumetric expansion of the pore water, either by the 9% increase in volume during phase change or by ice segregation, will generate stresses within the rock that will cause breakdown if they exceed the tensile strength of the rock. At temperatures below this, the suctions developed within the freezing rock, as well as the effects of thermal contraction of the rock minerals and pore constituents will generate additional stresses. Thus the means by which rock breakdown occurs in cold climates may be temperature dependant, although the processes probably also act in conjunction with each other.

If the rock is less than 100% saturated, then the freezing characteristic curve is truncated at the warmer

end, and the pressures generated are subsequently reduced. However, critical degrees of saturation have not been identified, since these vary between rock types and depend on the tensile strength of the rock in question.

The amount of moisture that is frozen within a rock depends on its temperature and the freezing characteristic curve. Thus the thermal regime of the rock is of utmost importance in determining the amount of moisture that freezes and therefore the stresses generated within the rock mass. This thermal regime is affected primarily by three factors: the general climate, the microclimatic properties of the surface and the rock thermal properties.

From field studies in the Ruby Range it is apparent that although the general macroclimatic factors of latitude, altitude and aspect can be of considerable importance, site-specific micro-factors may on occasion override them. Temperatures generally decrease with increasing altitude, but the influence of aspect is less clear, and so was investigated further through use of a computer simulation model. North facing slopes generally experience more intense freezing and for the longest duration. However, east and west facing slopes have the most frequent oscillations across the 0°C isotherm, particularly in the fall and spring months. Rates of temperature change, which may be important in sealing the rock surface and creating a closed system, are related to the diurnal temperature range at the rock face. These change with the season, so that in the fall

highest rates of freezing are on the north-facing slopes; during the spring months, south-facing slopes experience the highest rates of freezing. However, rates of freezing were never as high as the value of $6^{\circ}\text{C}\cdot\text{hr}^{-1}$ proposed by Battle (1960) and widely quoted in the literature, suggesting that this value may not be appropriate.

The nature of the site boundary layer will affect the energy flows at the surface of the rocks. Thus the albedo, surface moisture conditions, and surface roughness are all important in determining the rock thermal regime, but can vary significantly within a small area and are therefore difficult to measure or predict. A moist rock surface enhances cooling of the surface through evaporation, and a rough rock surface also increases the surface area from which evaporation can take place. Cracks and fissures at the rock surface will further complicate the thermal regime by localised shading and trapped air pockets, and can also provide suitable openings where bulk water can collect and lead to surface frost wedging. The albedo of the rock surface may be altered by the moisture conditions at the rock surface. Dark rock surfaces absorb more incoming radiation and therefore should lead to higher rock surface temperatures. Lighter surfaces, especially those that are covered by snow or ice, reflect more of the incoming radiation and are less prone to extreme temperatures.

The rock thermal properties are the final controlling

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(Caen Limestone and Nisling Alaskite) with differing thermal properties but the same surface conditions, and these were subjected to the same climatic regime. The higher conductivity of alaskite meant that temperature fluctuations at the surface are dissipated throughout the body of the rock, leading to higher day time temperatures and lower minimum temperatures at depth. Thus, the duration of freezing is reduced while the intensity of freezing is increased. During the spring months, however, the fast penetration of warm temperatures into the body of the rock reduces the duration of freezing. The higher heat capacity of the limestone prolongs the freezing duration but reduces the intensity of freezing, as more heat energy is required to change the temperature of the rock.

Rates of freezing are also affected by these properties. Generally, the rate of freezing decreases as the heat capacity increases, but increases as the conductivity of the rock increases, since the range of temperatures experienced is higher. However, the rates of freezing determined by the computer simulation model, as well as those found in the field, never met the critical rate proposed by Battle (1960).

6.3 Other Considerations

This study has concentrated specifically on one facet of a geomorphic process in which there may be a number of other important elements contributing to the same final result.

There has been little consideration of the role of salts in frost shattering, or of the chemical weathering that may occur in cold climates. Similarly, the role of vegetation and micro-organisms has been almost completely ignored, in an effort to concentrate on the the dominant mechanism of rock breakdown by frost action. These processes may work in conjunction with frost shattering to effect rock breakdown, and so future work could usefully be done in examining how these processes interact.

Although the TDR technique proved useful in the laboratory, there are still problems of measuring in situ rock water contents in the field. Similarly, there is still a paucity of field data of bedrock temperatures in frost shattering environments. There is a need for a fully automated recording device to monitor bedrock temperatures and moisture contents throughout the year, and to allow full analysis of the factors affecting the rock thermal regime.

6. - Conclusions

This study has provided information on the freezing of moisture within rocks, and has indicated that the significance of the relationship between temperature and moisture is shown by the freezing characteristic curve. From these results, it is suggested that frost action may be most effective at temperatures close to freezing. At lower temperatures, effects of suction and thermal contraction may be important in generating additional stresses within

the rock mass. Thus, the means by which the rock breaks down is temperature dependent, although these processes almost certainly act in conjunction with each other. Breakdown of the rock will occur only when the stresses generated within the rock exceed its tensile strength, which itself depends on the integrity of the rock.

It has also been shown that the thermal regime of a rock mass is controlled by climate, surface conditions and the thermal properties of the rock mass. Since the latter two can vary significantly within a small area, defining areas of frost shattering can be complicated and full consideration must be made of site specific factors that control surface and subsurface temperatures.

These conclusions suggest that it may be difficult to identify frost shattering environments in nature, or to distinguish between "frost shattering" and other means of mechanical breakdown in cold climates. Nonetheless, the thesis has laid some ground work for future investigations in this area, and which may be able to clarify the processes causing rock breakdown in cold environments.

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