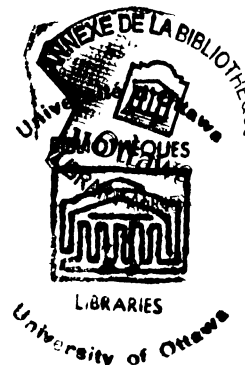


In Situ Creep of Perennially Cryotic Ground,
Rea Point, Eastern Melville Island
Northwest Territories

by
Lorne P. Bennett



Thesis presented to the School of Graduate Studies and
Research in partial fulfillment of the requirements for
the degree of Doctor of Philosophy in Geology.

UNIVERSITY OF OTTAWA
OTTAWA, CANADA, 1989



Lorne P. Bennett, Ottawa, Canada, 1989

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ABSTRACT

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At Rea Point, frozen soil creep is initiated by gravity, thermal contraction and expansion and secondary frost heave stresses. The rate of creep is affected by ground ice content, ambient temperature, slope angle and associated agents. While surface features such as thermal contraction cracks, earth hummocks and mudboils indicate frost action and suggest the possibility of frozen soil creep, downslope deformation of materials in the active layer (by frost creep and gelifluction) generally appear unrelated to the process of in situ frozen soil creep.

The mechanism of frozen soil creep has been used to partially explain phenomena such as valley bulges, cambers and gulls described in the Pleistocene periglacial and stratigraphic literature. It is believed that the data provided here and those of Savigny (1980) lend additional circumstantial support to the 'bulging' and 'cambering' mechanism.

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ACKNOWLEDGEMENTS

I wish to thank Dr. H. M. French for his support, comments and advice during the course of this thesis.

Discussions with Dr. W. Haeberli (Federal Institute of Technology, Zurich, Switzerland) and Dr. M. W. Smith (Carleton University), clarified some of the technical and theoretical aspects.

Support for this project was provided by Natural Sciences and Engineering Research Council of Canada, Grant A-8367 (H. M. French), the University of Ottawa Northern Research Group and the Polar Continental Shelf Project, Energy, Mines and Resources, Ottawa (Project 87-106, L. Bennett). I also owe a great debt to Panarctic Oils Ltd., Calgary who, between 1984-85, generously provided all logistics and transportation north of Edmonton. This support included accommodation and food at Rea Point, technical assistance by Rea Point personnel, seismic drilling operations in May 1985, and freight and other services on numerous occasions, in addition to air transport Edmonton-Ottawa on numerous occasions and surface (field) transportation at Rea Point. All this was made possible through the good offices of A. R. Rossiter, Chief

Administrator and Environmental Manager, Panarctic Oils Limited. Without his support and encouragement, this thesis could not have been undertaken. I also owe a great debt to G. D. Hobson, formerly Director, Polar Continental Shelf Project who in 1986, following the cessation of Panarctic operations at Rea Point, undertook to provide logistical support out of Resolute Bay for the 1987 summer field season. Appreciation is also expressed to EBA Engineering Ltd., and particularly to D. W. Hayley, for making available permafrost and ground ice data obtained during the 1981-82 ditching experiments at Rea Point.

Finally, I am sincerely grateful for field assistance provided by Geof Burbidge, Rob Graham, Gerry Butts and Ross and Tony Manik.

CHAPTER ONE

INTRODUCTION

1.1 Aims and Objectives

Frozen soil creep is defined as "the slow deformation that results from long-term application of a stress too small to produce failure in the frozen material" (ACGR, 1988, p. 19). This thesis attempts to understand the nature, magnitude and controls over the in situ creep of perennially cryotic ground (i.e. permafrost) on natural slopes at a number of localities in the vicinity of Rea Point (latitude 75°22'N; longitude 105°45'W), Eastern Melville Island, N.W.T. (Figure 1.1). For the purposes of this thesis, the terms 'frozen soil creep' and 'permafrost creep' are used interchangeably.

The aims of the thesis are as follows:

- 1) To determine the relationship between frozen soil (i.e. permafrost) creep and ground ice content, ground thermal regime and slope angle;
- 2) to determine the relationship between frozen soil (i.e. permafrost) creep and associated periglacial processes (thermal contraction cracking, frost heave, frost creep and solifluction/gelifluction); and,

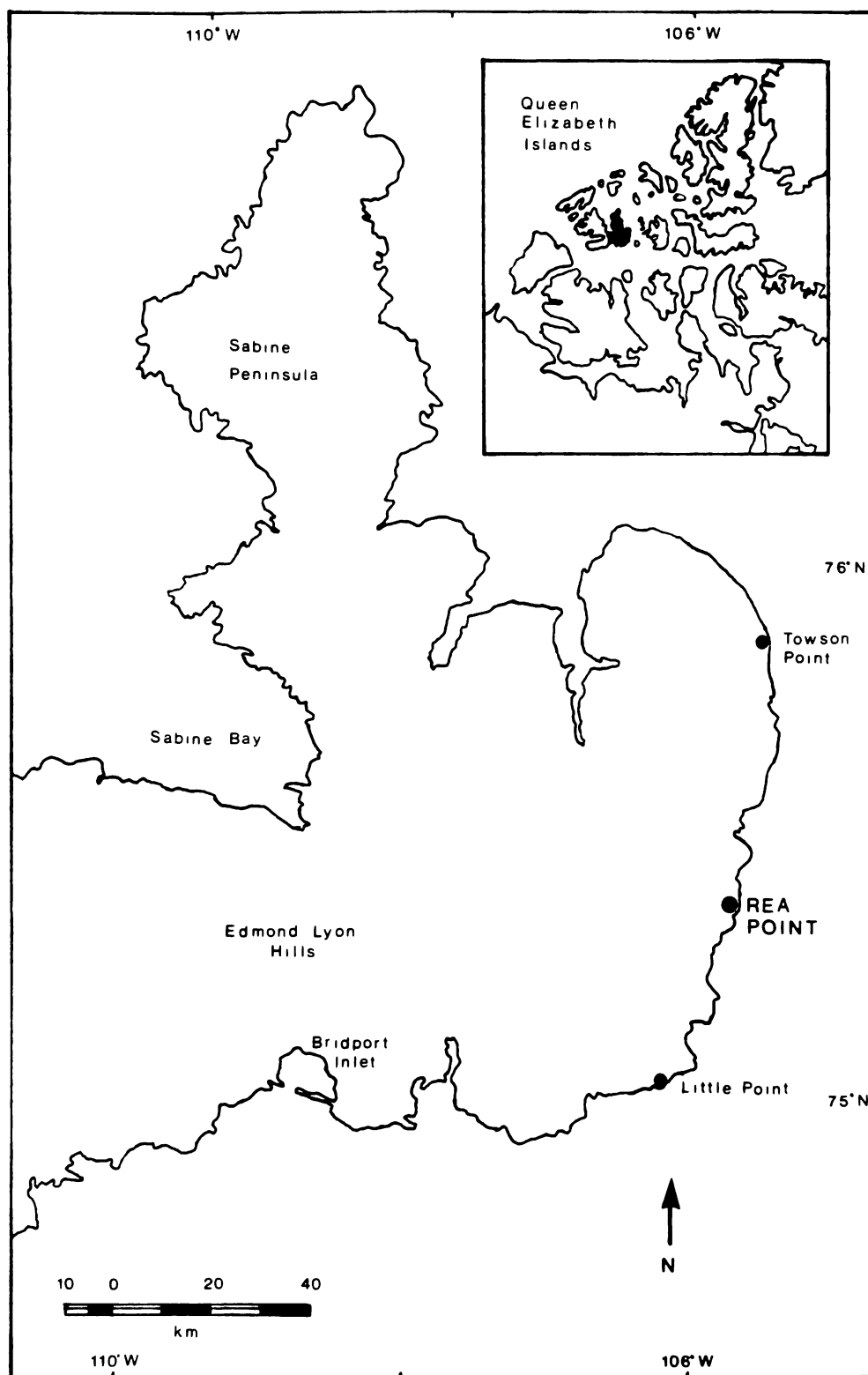


Figure 1.1 The study area, Rea Point, Eastern Melville Island.

- 3) to calculate threshold values for frozen soil (i.e. permafrost) creep, relating these to both primary creep regulators (ground ice content, ambient temperature, slope angle) and primary creep activators (gravity, thermal contraction and expansion, secondary frost heave).

1.2 Permafrost Creep

In North America, many field studies have been concerned with creep processes in both the seasonally cryotic zone and in unfrozen soils (e.g. Williams, 1957, 1966; Harris, 1980; Mackay, 1981; Mills, 1983; Egginton and French, 1985). However, apart from work on rock glaciers (e.g. Barsch, 1977; Giardino et al., 1984; Haeberli, 1985), only a few studies consider the creep of frozen sediments (e.g. Savigny, 1980; Savigny and Morgenstern, 1986a, b). Equally limited in number are laboratory tests which simulate natural frozen soil creep rather than induced frozen soil creep. The laboratory tests, in fact, provide most of the data used to develop theoretical models of creep. However, it is not known how closely these data compare with those of natural frozen soil creep (e.g. Man, 1983). Andersland and Anderson (1978) review many of the early laboratory studies of frozen soil creep while more

recent studies are reviewed by Vyalov (1986).

Laboratory tests are often conducted at temperatures and strain rates not usually experienced in nature (e.g. Sego and Morgenstern, 1983). This is due to the greater dependability of testing samples at -10° or -20°C than at pressure melting point. Critical changes in strength and ice structure occur as ground temperatures fluctuate below pressure melting point (Aguirre-Puente and Gruson, 1983). It is often difficult to maintain frozen samples at temperatures varying by less than a few tenths of a degree for the long periods of time required to simulate creep.

Strain rates used to simulate frozen soil creep in the laboratory are usually much greater than those measured in the field. As such, they create, or induce, unnatural stress. Such tests provide data demonstrating the bearing capacity of a frozen soil, but they shed little light on the natural process of in situ permafrost creep. While structures may be designed to settle at acceptable, pre-determined rates (i.e. 'induced creep'), one must conclude that the long term effects of natural frozen soil creep have been given little consideration to date (e.g. see Zaretsky et al., 1988).

In laboratory testing, the use of uniaxial (shear boxes) and triaxial machines often induce shear within test samples (e.g. Savigny and Morgenstern, 1986b). Although

this may be a component of the deformation of sediments in the field, it is not, by definition, creep. Test samples may be too small to allow the development of creep. In situ permafrost creep is most likely the result of the dispersal of natural stress among a number of ice lenses within a large soil block rather than the concentration of stresses within one or two lenses as occurs in test samples.

Laboratory tests would more closely simulate field conditions if samples could be prepared as scaled down models of a large soil block (e.g. scaled down ice lenses, soil particles, etc.). However, it is virtually impossible to reconstitute frozen soils containing excess ice (at any scale) (e.g. Weaver and Morgenstern, 1981).

Recent laboratory tests employing undisturbed frozen samples at field temperatures are closing the gap between laboratory test data and true creep processes. There are problems associated with the uneven distribution of ice in 'replicate' samples and results often vary widely (Morgenstern, 1981, p. 320; Maksimjak et al., 1983). However, the range of tests conducted allows further refinement of mathematical theories (e.g. Glen's Flow Law) for the effects of soil in ice (e.g. Savigny and Morgenstern, 1986b).

In Canada, one of the first theoretical studies of frozen soil creep was presented by McRoberts (1975). Based upon infinite slope analysis, subsequent studies showed that

a slope inclined at 15° and consisting of 10 m of ice-rich permafrost would creep at about 10 cm yr^{-1} (Morgenstern, 1981). Clearly this suggests that permafrost creep is a major contributor to mass wasting in northern environments and may pose considerable problems to engineering works on ice-rich soils.

One of the few attempts to measure the in situ creep of frozen sediments was undertaken in the Mackenzie Valley (Savigny, 1980; Savigny and Morgenstern, 1986a). It was found that the rate of creep varies according to the scale and spacing of ground ice structures at depth. Relatively uniform velocities occur over depth intervals that are several metres thick and which are ice rich (e.g. borehole GB1A at depth of 24 - 29m; Savigny and Morgenstern, 1986a). A second type of deformation, similar to shearing, occurs where large ice lenses are more widely spaced. All the movements measured took place in unconsolidated sediments, notably glaciolacustrine clay and silty clay.

If one excludes sedimentary sequences involving complex Quaternary stratigraphies and/or massive bodies of ice occurring at depth within permafrost, ground ice volumes are generally highest in the top 1 - 5 m of permafrost, being concentrated at or just below the base of the active layer (e.g. Pollard and French, 1980; Lawson, 1983). Such a concentration reflects the seasonal migration of water within the frozen ground in response to a temperature

gradient (e.g. Parmuzina, 1978; Cheng, 1983; Mackay, 1983). Accordingly, some of the highest rates of in situ frozen soil creep should occur within the upper 1 - 2 m of permafrost. In all probability, these movements should represent a more widespread process than the relatively deep-seated deformation described by Savigny (1980). On the other hand, the net downslope deformations occurring in areas underlain by cold permafrost might reasonably be expected to occur at much slower rates than the $0.15 - 0.30 \text{ cm yr}^{-1}$ reported by Savigny (Savigny and Morgenstern, 1986a, p. 513). This reflects the fact that the creep strength of soils is also strongly dependent upon temperature as stressed by Johnston (1981, p. 92). In a recent review, Lewkowicz (1988) suggests that permafrost creep may be an important denudational process in areas underlain by 'warm' permafrost and calculates mass wasting attributed to this process as equalling $800 - 1000 \text{ cm}^3 \text{ cm}^{-1} \text{ yr}^{-1}$.

1.3 Factors Affecting Permafrost Creep

Perennially cryotic ground is subject to a variety of time-dependent stresses and strains (e.g. Vyalov, 1959, 1986; Ladanyi, 1981). Gravitational, thermal and secondary frost heave stresses are of particular importance. These stresses are the primary activators of in situ frozen soil creep (Figure 1.2). Several factors govern or regulate

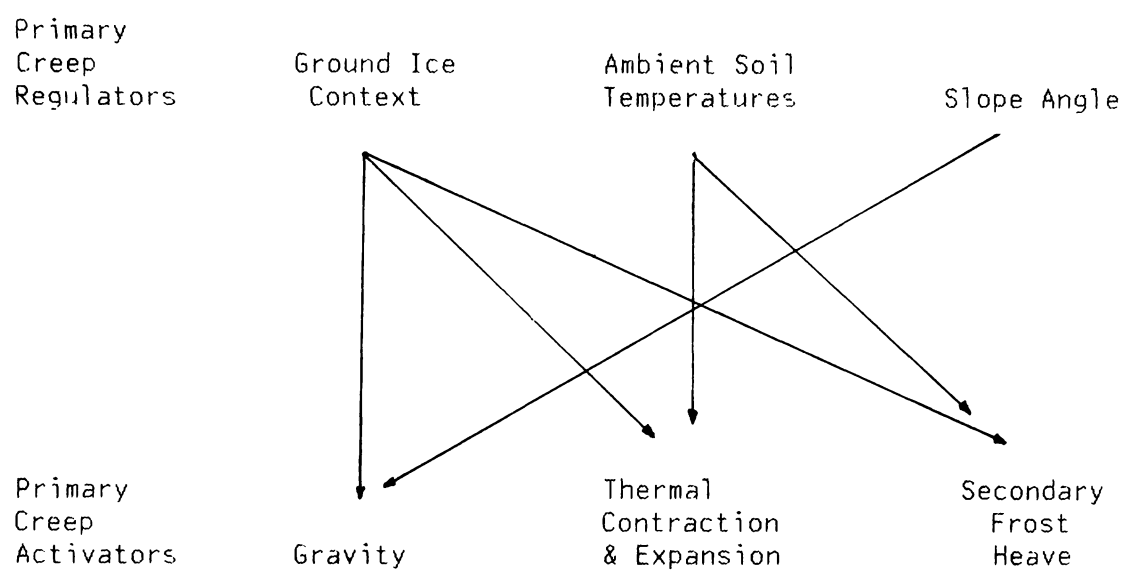


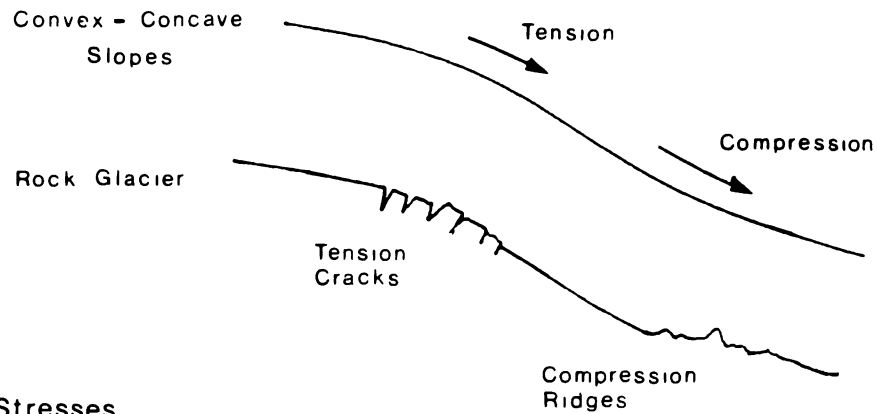
Figure 1.2 Inter-relations between the primary regulators and activators associated with in situ frozen soil creep.

these creep activators. The primary creep regulators are slope angle, ground ice content and ambient temperature. Secondary creep regulators, which may or may not affect creep by influencing the primary regulators, include slope aspect, insulation (e.g. snow cover) and depth.

Figure 1.3 illustrates how each of the primary creep activators places a soil in a state of tension or compression. It should be noted that the predominance of either tension or compression on a slope is a function of place (Figure 1.3a) while the predominance of either tension or compression resulting from either thermal contraction/expansion or frost heave stresses is a function of time (Figure 1.3b, c). In general, frozen soils are subject to an annual cycle of fall and winter tension followed by spring and summer compression.

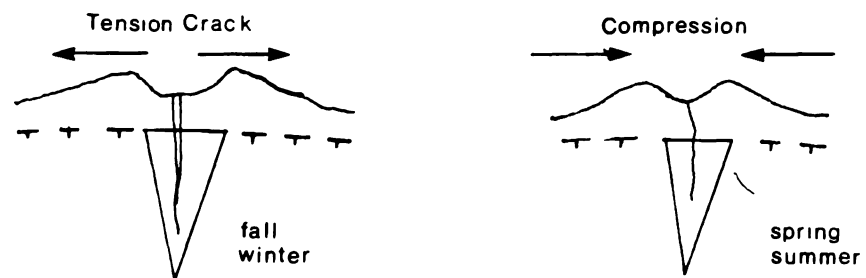
There are three primary modes of soil deformation; these are termed damped (or attenuating) creep, steady-state creep and progressive creep (Sayles, 1968). Figure 1.4a illustrates a selection of isothermal creep curves. The mode of creep in a frozen soil is governed by the magnitude of strain hardening (strengthening) and strain softening (weakening) processes. Examples of strain hardening include particle rearrangement and ice-rock bond creation. Particle rearrangement involves the squeezing out of pore ice leading to a net increase in particle surface contact and thus, internal friction. Where the volumetric ice content of a

A Gravity Stresses



B Thermal Stresses

Ice Wedges



C Secondary Frost Heave Stresses

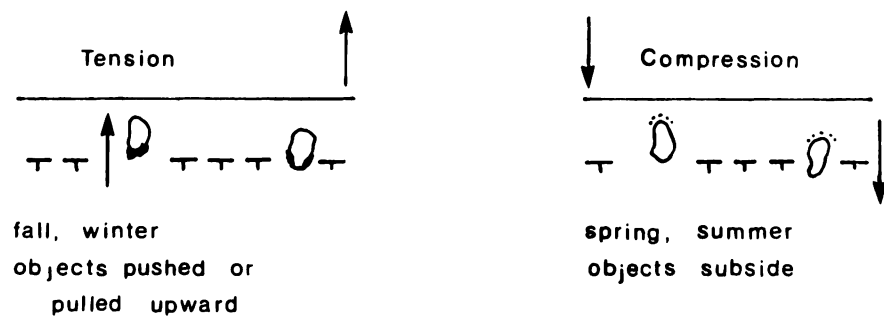


Figure 1.3 Examples of major stresses in frozen soils: A) Gravity stresses; B) Thermal stresses; C) Secondary frost heave stresses (after Haeberli, 1985; Mackay, 1974, 1981).

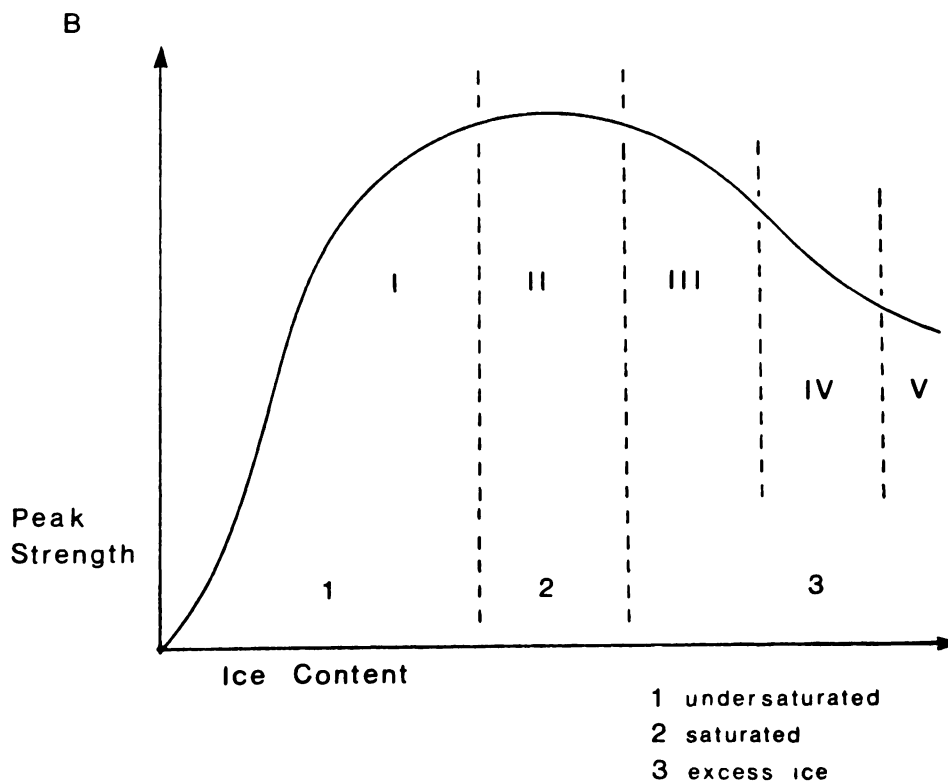
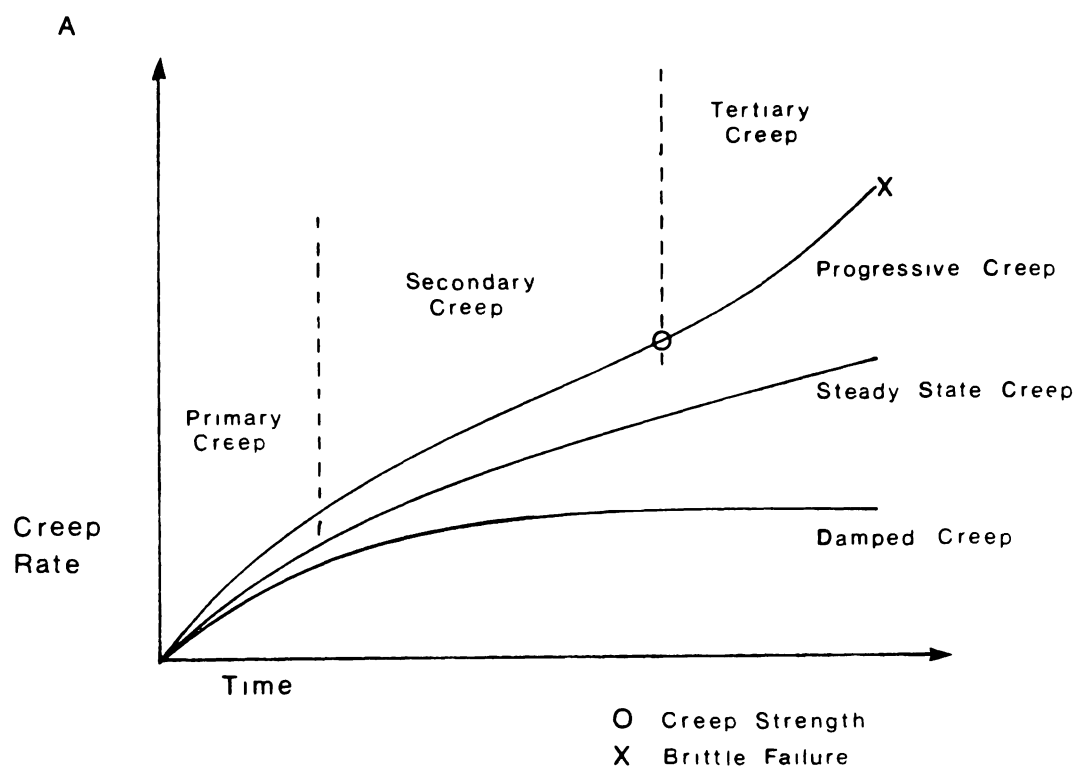


Figure 1.4 Typical creep performance of frozen soils: A) Isothermal conditions and creep curves, after Sayles (1968); B) Graph showing peak strength versus ice content. In B), the greater the peak strength, the greater the resistance to creep (after Shusharina and Bobkov, 1969).

soil is reduced in this manner, ice-rock bonds are created (ice lenses or ice-ice bonds are reduced). Examples of strain softening include ice crystal reorientation, ice-ice bond creation and micro-crack development. Ice may adapt to directional stress by crystal reorientation (c-axis parallel to stress direction) or crystal size reduction (an increase in ice-ice bonds). Alternatively, a cryogenic soil may develop micro-cracks in response to creep stress. The micro-cracks reduce the ability of the soil to withstand subsequent creep or shear stress.

The response to stress by strain hardening or strain softening is governed by cryostructure and internal bonding. A frozen soil may have either an orderly structure (e.g. the result of aligned platy soil grains), a random structure or a composite structure (e.g. one showing traits of both orderly and random structure such as a random mix of soil grains trapped between regular horizontal ice lenses). Ultimately, the cryostructure depends upon the geology, ice content and cryogenic history of the soil (see ACGR, 1988, p. 24).

Internal bonds may vary from weak to strong. Bonds between dry soil particles or those bounded by water are generally weak. The bonds are easily broken and easily re-formed (e.g. Vyalov, 1986). While ice-rock bonds are generally stronger than ice-ice bonds, these too are highly variable. Ultimately, the internal bonding depends upon the

ice content and temperature of the soil.

If the strengthening processes (strain hardening) exceed the weakening processes (strain softening), soil creep is damped (see Figure 1.4a). However, if the weakening processes exceed the strengthening processes, steady state or progressive creep develops. The possibility of the latter hinges upon soil creep strength ($\tilde{\sigma}$). The latter is defined as the largest stress that can be imposed without sustained deformation of the soil. When a stress is imposed which exceeds the creep strength of a frozen soil, viscoplastic creep results leading to progressive flow and shear or tensile failure (Vyalov, 1986; Chamberlain et al., 1972; Sayles, 1985). An example of shear failure is provided by multiple retrogressive thaw slides. In summary therefore, the creep rate of a frozen soil is highly variable since it is significantly influenced by a) ground ice distribution, b) ground thermal regime, and c) slope angle.

1.3.1 Ground ice content

Ice content plays a major role in regulating the creep deformation of cryogenic soils (e.g. Shusherina and Bobkov, 1969; Figure 1.4b). When a cryotic soil is totally devoid of ice, for example, it reacts to stresses in exactly the same manner as an unfrozen material. The shear strength

of this material may be determined with reasonable accuracy by means of the linear Mohr-Coulomb equation,

$$S = c + \sigma_n \tan \phi \quad (1.1)$$

where c is cohesion, σ_n is normal load and ϕ is the angle of internal friction. This equation, however, cannot be used to describe ice-rich soils as creep deformation affects the linearity of the relationship (Ladanyi, 1972).

The introduction of a small quantity of ice to a granular soil results in a substantial increase in soil strength (see Figure 1.4b). The strength increase is derived from the cohesive forces of pore ice. Initially, strength increases only at points of contact between soil particles or where individual ice crystals dominate. In soils of very low ice content, failure planes may form in areas lacking ice. As the pore ice increases (Figure 1.4b), particles progressively further from contact are cemented together. In conjunction with the increased area of ice-rock bonds, the angles of internal friction and effective normal stress increase when dilation of soil particles during shear is restricted by ice. This adds to soil shear strength.

In Figure 1.4b, 'I' represents a zone of increased cohesion (c), increased internal friction (ϕ) and increased effective normal stresses (σ_n). As the soil approaches

saturation (II), small changes in ice content do not create dramatic changes in c , ϕ and σ_h and the frozen soil reaches a point of maximum resistance to stress.

When excess ice is present, creep strength is reduced (III; Figure 1.4b). Initially, just a few particles are pushed out of contact reducing internal friction. However, since the major effect of cohesion is a function of ice-rock bonds and not of the ice itself (Babcock et al., 1978), an increase in ice content by volume causes a direct decrease in the area of ice-rock bonds (provided all other factors remain equal). It follows that frozen soil creep is more likely to occur as the internal friction and ice-rock bonds within a soil are reduced (i.e. as ice content increases). When excess ice volumes are sufficient to cause complete separation of soil particles, as when ice lenses dominate the soil, a critical excess ice content is achieved or exceeded (IV). This critical excess ice content is defined as the volumetric ice content at which the strength of a soil becomes dependent upon the strength of ice rather than rock (Goughnour and Andersland, 1968). Beyond zone IV, the soil is almost totally dependent upon the strength of the ice matrix for its own strength. At very high ice contents, soil particles have a minimal effect upon the creep strength of the frozen material. At this stage, soil creep strength is primarily a function of ice-ice bonds and the internal strength of ice crystals (Vyalov, 1969).

Ice-rich soils are prone to similar processes to those which affect the creep of polycrystalline granular ice (Andersland et al., 1978). Such ice, with random crystal orientation, occurs in most frozen soils (Andersland and Alwalhab, 1983). The Mohr-Coulomb equation is then supplanted by a parabolic equation (Ladanyi, 1972) which, for coarse granular soils, may be represented as follows:

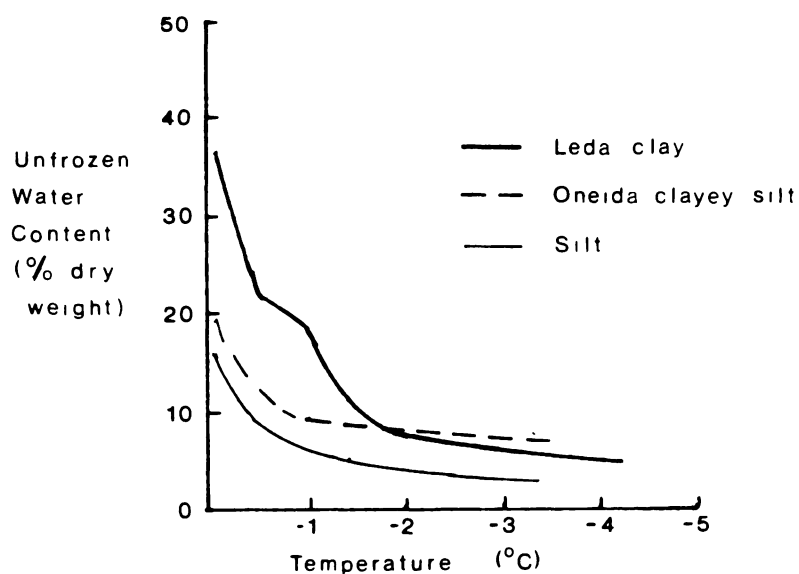
$$\tilde{\zeta} = (c \sigma_n + b)^{0.5} \quad (1.2)$$

where $\sqrt{b}$ is the cohesive force at zero normal stress (σ_n) and c is a function of internal friction (Nickling and Bennett, 1984). Coefficients c and b both decrease exponentially as a function of increasing ice content.

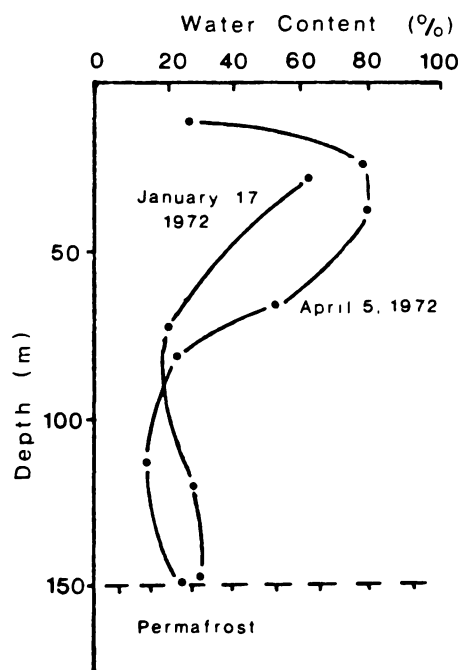
1.3.2 Ground thermal regime

Ambient soil temperature plays a strong role in regulating frozen soil creep. It is well known, from both laboratory and field observations, that not all soil water freezes as ground temperatures drop below 0°C (e.g. Anayam, 1978; Williams, 1982; Anderson et al., 1984; Smith, 1985a). For example, Figure 1.5a illustrates the changes in water content with temperature which occurs when soils of varying grain size are subject to cryotic conditions in a laboratory setting. Such data demonstrate that, in fine-grained soils, adsorbed water may remain in dynamic equilibrium with ice

A



B



C

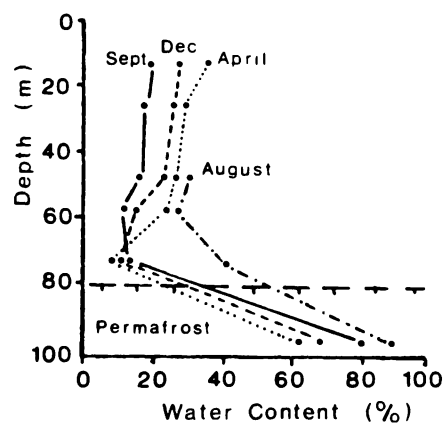


Figure 1.5 A) Changes in unfrozen water content of freezing soils (after Williams, 1982, p. 96); B) Changes in water (ice) content, as measured in China (after Cheng, 1983 [left]) and Siberia (after Parmuzina, 1978 [right]).

crystals at temperatures well below the ice nucleation point. The same is true for that water which is held in capillaries. Field observations from a range of permafrost environments provide convincing evidence that significant changes in ice (water) content occur in freezing and frozen soils (Figure 1.5b). This is discussed in more detail in Section 1.4.

Adsorbed water may film soil particles down to temperatures of at least -190°C (Haynes, 1978). In nature, however, the unfrozen water content probably only affects the creep of fine soils at temperatures above -10°C and becomes progressively more important as temperatures approach 0°C (Vyalov, 1969). The relationship between water content and temperature is a function of several factors. These include 1) specific surface area of soil particles or particle shape, 2) grain size and grain size distribution, 3) mineral type, 4) ice and water pore pressures, 5) solute concentration and 6) void ratio (Konrad and Morgenstern, 1983; Akilı, 1970).

Johnston et al. (1981) suggest that unfrozen water reduces particle contact within soil thereby reducing strength. For example, Tsytoovich (1975) notes that the compressive strength of a medium sand decreases by 15% as the temperature increases from -2.0 to -1.0°C . Under similar conditions, he suggests that the compressive strength of clay decreases by 50%. It follows therefore

that, as the creep strength of material decreases, the creep rate increases, because the unfrozen water weakens ice-cement bonds and changes ice from a brittle to a plastic solid (Hobbs, 1974).

1.3.3 Slope angle

The importance of slope angle as a regulator of frozen soil creep has been the subject of some debate (e.g. Finlayson, 1981, pp. 45, 46), although it is generally agreed that there is some relationship between slope angle and the rate of downslope soil movement. The general equation for shear stresses ($\tilde{\tau}$) in a slope may be approximated as follows:

$$\tilde{\tau} = \gamma z \sin\phi \cos\phi \quad (1.3)$$

where γ is the bulk unit weight of the soil, z is the depth to the potential failure plane, ϕ is the angle of the slope. The equation demonstrates that increasing slope angles cause increasing shear stress. However, as ice will deform under its own weight, the force of gravity should be sufficient to initiate the creep of ice on a slope of any angle. Ice-rich soils differ from pure polycrystalline ice in that they require stress of a particular magnitude (yield stress) to begin non-recoverable deformation. McRoberts (1975) suggests that stress levels on natural slopes vary from

0.003 to 6.0 kgf cm⁻² as a function of slope angle. The greater the slope angle, the more likely the yield stress will be attained.

1.3.4 Cryostructure

In frozen as opposed to unfrozen soils, the soil cryostructure must also be considered. Cryostructure refers to the structural characteristics of frozen, fine-grained earth materials (ACGR, 1988, p. 24). It is defined by the volume and distribution of pore ice and lenses of segregated ice.

The type and arrangement of ice in the frozen material will depend largely upon the initial water content of the material and the extent of moisture migration during subsequent freezing. The Soviet literature recognizes a range of cryostructures, the more important being massive, layered and reticulate (e.g. Kudryavtsev, 1978; see Harry, 1986). Typically, a massive structure is dominated by pore ice and a relatively low total ice content. As such, it is not highly susceptible to frozen soil creep, other things being equal. In soils with a reticulate structure, ice lenses typically form a random net; as such, the cryostructure is an essentially passive component of the creep process. By contrast, in soils with a layered cryostructure, where ice lenses occur in horizontal layers, the cryostructure may accentuate frozen soil creep.

Cryostructure also affects soil creep strength.

The combination of polycrystalline ice and mineral soil produces a unique material possessing characteristics unlike those of its two components. Clearly, if shear stresses are applied in a direction perpendicular to ice lenses in a soil, the soil provides a greater resistance than if the stresses are parallel to the lenses. Thus, Andersland et al. (1978) suggest that cryostructure affects creep strength by providing potential planes of weakness.

Ice type and crystal size alone may affect the nature of frozen soil creep. Creep strength is more than 20 times greater when stresses are applied perpendicular rather than parallel to the basal plane of an ice crystal (Gold, 1973). In polycrystalline ice, random crystal orientation means that only some of the crystals are aligned such that their basal planes are parallel or nearly parallel to the direction of stress. Shear stresses cause these ice crystals to either shear parallel to their basal plane or be destroyed. New crystals then form which are generally smaller, possessing a preferred orientation which allows dissipation of planar creep stresses (e.g. Gold, 1973).

Polycrystalline ice may consist of ice crystals in a wide range of sizes. Ice crystals 1.0 mm in diameter provide the greatest resistance to creep. Baker (1978) suggests that doubling, or halving, average crystal size from that of 1.0 mm increases the secondary creep rate by a

factor of four or five. Although the magnitude of this rate change is not universally accepted (Duval and Le Gac, 1980), it is believed that larger ice crystals creep by means of internal dislocations (glide planes) whereas smaller crystals deform in response to granular adjustment.

1.4 Factors Initiating Permafrost Creep

The primary creep activators are gravity, thermal contraction and expansion and secondary frost heave (Figure 1.2). Gravity stresses occur on all slopes in all environments. Solifluction and frost creep are well known geomorphic processes dependent upon gravity stress (e.g. Harris, 1981). Shear failure on a slope through frozen sediments occurs when gravity stress exceeds the creep strength and shear strength of adjacent sediments.

Thermal contraction and expansion occurs in all soils undergoing temperature change. Where the temperature drop is fast, soil contraction may exceed the tensile strength of frozen soil leading to progressive creep and brittle failure (Lachenbruch, 1962; 1963). This process creates thermal contraction cracks, one of the most distinctive features of the arctic landscape. Where the temperature drop is slower, however, soils may creep to accommodate contraction and remain intact. Both gravity and thermal stresses create "horizontal" creep deformation.

This contrasts with the process of frost heave which creates natural vertical frozen soil creep within sediments.

Secondary frost heave has only been measured over the past ten years. As soils begin to freeze, water migrates in response to thermally-induced gradients (Smith, 1985a, b). The migration is from zones of high potential energy to zones of low potential energy, that is, to the freezing front (Williams, 1982; Cheng and Chamberlain, 1988). The process is particularly effective in fine-grained sediments. If the volume of ice exceeds the volume of voids for the porosity of the soil, frost heave occurs leading to the formation of distinctive bands of ice or ice lenses.

Secondary frost heave occurs not only within seasonally cryotic ground but also within perennially cryotic ground, as demonstrated by Mackay et al. (1979) and Smith (1985b). Telescoping heavemeters are commonly used to monitor the frost heave process. In an early field study near Inuvik, N.W.T. a four-tube system was used (Mackay et al., 1979). The greatest positive heave, measured during the fall, was due to ice lensing at the top and bottom of the active layer. A negative heave, measured during the fall, was due to ice lensing at the top and bottom of the active layer. A negative heave, measured during January and February, was thought to be due to desiccation of the central part of the seasonally cryotic zone. It was also

found that frost heave decreased with depth. Results showed that 10 to 20 mm of surface heave was typical for soils in the region. In a subsequent study involving silty clay hummocks, Smith (1985b) employed a six-tube heavemeter. It was found that frost heave processes were initiated at the surface and then occurred progressively later with increasing depth. Probably the most important result was the confirmation of secondary frost heave; i.e. that significant soil water redistribution and frost heave occurs within frozen soil. Near-surface frost heave tubes showed an active positive heave (approximately 3 - 5 mm/day) several days after the freezing front had passed by them.

1.5 Summary

It follows that there are several key variables involved in any field study of permafrost creep. Since temperature and ice content are the most basic, these parameters are examined separately in Chapters 4 and 5. The field results from the Rea Point vicinity are presented in Chapter 6. The instrumentation and experimental design for the measurement of permafrost creep are outlined in Chapter 3. Background information concerning the Rea Point area, and its representativeness for such a study, are the subject of Chapter 2.

CHAPTER TWO

THE REA POINT AREA AND STUDY SITES

2.1 Introduction

The choice of a study area in remote arctic regions is often limited due to¹ logistical problems. The Panarctic Oils Ltd. base camp facility at Rea Point, Eastern Melville Island was chosen for this study (see Figure 1.1). This locality provided:

- 1) Access to a wide variety of slopes and soils thought typical of the weathered bedrock which commonly occurs in the high arctic;
- 2) access to boring, trenching and other equipment as well as laboratory and workshop space; and
- 3) the option of monitoring the instrumented sites regularly throughout the year.

2.2 Climate

A continuous record of local climatological data was maintained at Rea Point from 1969 to 1985. Representative climatic statistics are given in Table 2.1. The climate is reasonably typical of many Canadian High

Table 2.1
Climatic Statistics for Rea Point, Melville Island

	J	F	M	A	M	J	J	A	S	O	N	D	Annual
Mean Daily Temperature (°C)	-33.4	-34.9	-32.2	-23.6	-11.1	- 0.2	+ 4.1	+ 1.6	- 5.8	-17.0	-26.1	-31.0	-17.5
Mean Daily Maximum (°C)	-29.8	-30.8	-28.8	-19.5	- 7.8	+ 2.0	+ 7.4	+ 4.1	- 3.6	-13.2	-22.3	-26.8	-14.1
Mean Daily Minimum (°C)	-37.2	-38.2	-35.7	-27.7	-14.4	- 2.6	+ 1.5	- 0.5	- 8.1	-20.0	-29.4	-34.3	-20.6
Extreme Maximum (°C)	- 1.1	-12.0	-14.5	- 3.3	+ 3.3	+18.3	+16.7	+15.6	+ 8.0	+ 1.0	- 1.7	-13.3	
Extreme Minimum (°C)	-52.2	-49.0	-48.4	-40.6	-28.9	-16.1	- 5.6	- 7.2	-23.5	-32.2	-43.3	-46.7	
Mean Rainfall (mm)	0	0	0	0	0	3.0	10.6	29.4	1.7	0	0	0	44.7
Mean Snowfall (cm)	1.2	0.3	1.4	1.1	3.4	1.0	1.4	1.7	5.5	7.8	2.0	2.0	28.6
Total Precipitation (mm)	1.4	0.3	1.6	1.1	3.6	4.0	9.1	20.4	6.8	8.4	2.2	2.1	61.0
Mean Windspeed (km/hr)	20.8	20.5	21.5	16.8	19.6	20.2	21.1	21.7	22.4	22.0	20.8	20.9	20.7
Prevailing Direction	NW	NW	NW	NW	N	N	S	N	NW	WNW	NW	NW	NW
Average Frost Free Period	14 days												
Maximum Frost Free Period	41 days												
Minimum Frost Free Period	0 days												
Average Last Spring Frost	July 7												
Average First Fall Frost	July 22												
	Heating Degree Days (<18°C)										12919.0		
	Growing Degree Days (>5°C)										21.4		
	Thawing Degree Days (> 0°C)										222.3		

Source: Atmospheric Environmental Services, Canada

Arctic environments; Figure 2.1 compares data from Rea Point with those from Resolute Bay. It can be seen that there is a broad similarity in conditions between the two sites, in spite of a distance of 300 km.

In general, the climate at Rea Point is characterized by long, dry, cold winters and short, moderately dry, cool summers. The mean annual temperature is -17.5°C . Extreme maximum temperatures range from -12.0 and -14.5°C in February and March to $+18.3$ and $+16.7^{\circ}\text{C}$ in June and July. Extreme minimum temperatures range from -52.2 and -49.0°C in January and February to -5.6 and -7.2°C in July and August. The average frost free period is only 14 days with the last spring frost occurring on July 7 and the first fall frost on July 22.

Although total annual precipitation averages just 61 mm, occasional single day rainfall accumulations of between 10 and 30 mm may occur. Also, up to 17.8 cm of snowfall has been recorded over a 24 hour period. Precipitation is greatest in August, September and October when open water is most abundant. Much of this precipitation falls as snow. The accumulation of snow at a single point is strongly influenced by the relatively constant winds. Mean wind speed ranges between 16.8 km/hr in April and 22.4 km/hr in September. The prevailing northwest winds at Rea Point promote greater snowfall (to accumulate) on east and south-facing slopes.

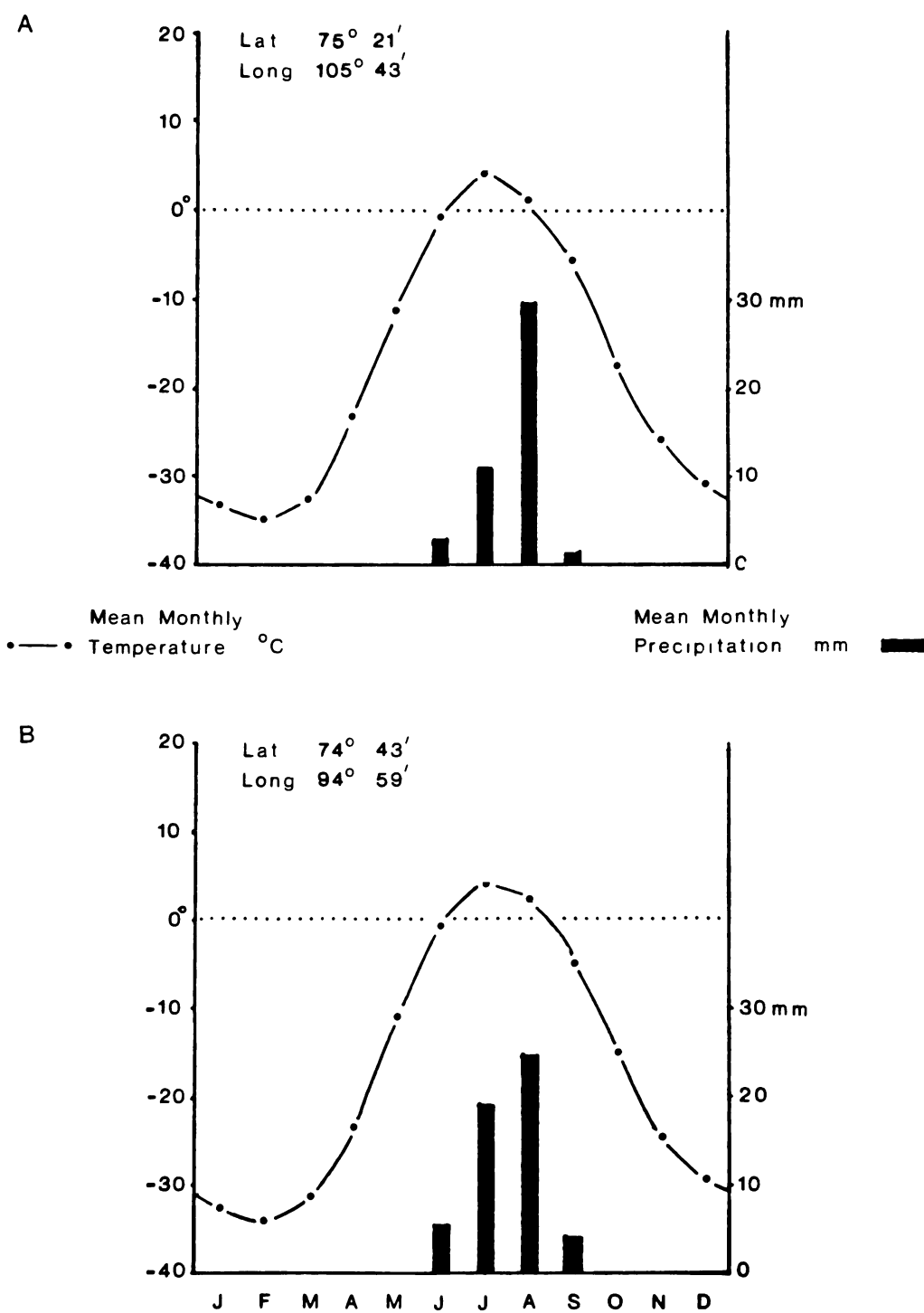


Figure 2.1 Climate data for Rea Point A) and Resolute Bay B).

2.3 Geology

The terrain in the vicinity of Rea Point consists of flat to gently sloping surfaces which rise to form low-altitude ridges or hills. The highest elevations in the area are approximately 100 m a.s.l. The ridges trend in an east-west direction reflecting the bedrock geology which consists of well-folded Paleozoic (Upper Devonian) clastic rocks of the Melville Island group. A geological cross-section extending N - S across Eastern Melville Island is presented in Figure 2.2. The structural control of topography is well illustrated by the accordance of drainage and the formation of major valleys along anticlinal axes, especially where the relatively less resistant Hecla Bay Formation is exposed.

The bedrock geology of Eastern Melville Island can easily be identified upon air photographs. Figure 2.3 shows the area in the immediate vicinity of Rea Point with the major geological boundaries indicated. Since frozen soil creep will be influenced, to varying degrees, by the underlying bedrock and its composition (i.e. weathering characteristics and products, ground ice contents, etc.), the nature of the main bedrock units in the Rea Point area need^c detailed consideration.
L

Three Late Devonian formations outcrop within 2 km of Rea Point: 1) Weatherall, 2) Hecla Bay and, 3) Griper

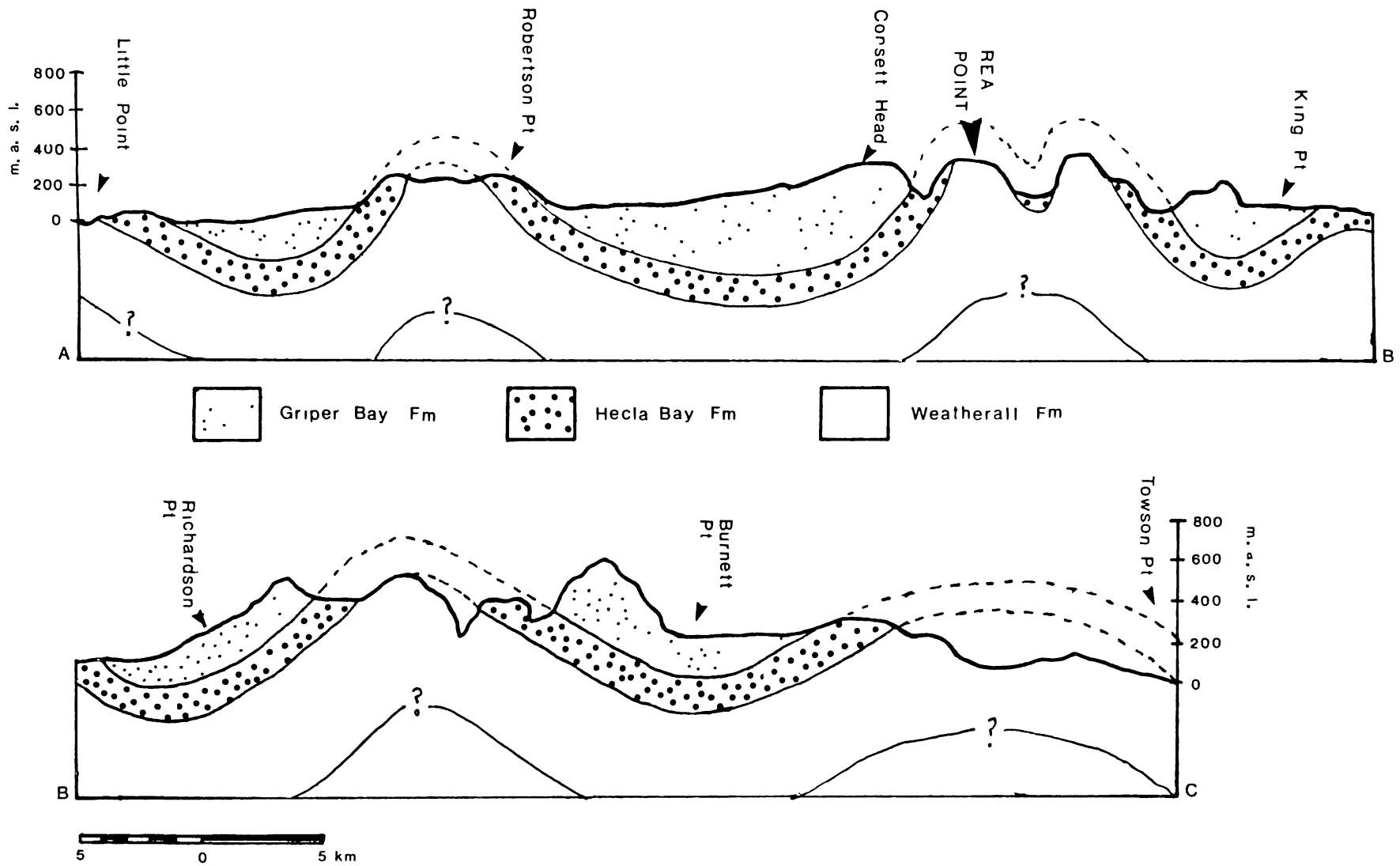


Figure 2.2 Geological cross-section through Eastern Melville Island.
(From: Tozer and Thorsteinsson, 1964)

Bay Formations (see Figure 2.3). Field studies were restricted to sites underlain by the Weatherall and Hecla Bay Formations.

The Weatherall Formation consists of repetitive, thin beds of siltstone and fine-grained sandstone with shale and coal seams. The upper part of the formation, seen just west of Rea Point, is composed of very fine to fine-grained light gray to gray sandstone. Embry and Klován (1976) suggest that the sandstone, with its associated sedimentary structures (ripple cross-stratification, horizontal bedding and low angle cross-bedding), likely reflects deposition as distributary mouth bar and beach deposits. The topography is gently rolling with fluvial dissection creating low relief (Figures 2.4a, b).

The Hecla Bay Formation is composed primarily of white to orange fine to medium-grained well-sorted quartz sandstone. Siltstone, shale and very fine-grained argillaceous sandstone units are occasionally found within this formation (Embry and Klován, 1976). The Hecla Bay sandstones are very resistant where authigenic cements (kaolinite) average or exceed about 10% of the composition. Ridges and fields of felsenmeer (with blocks averaging 1 m in diameter) are typical products of the weathering of resistant units. In the Rea Point area, the Hecla Bay sandstones are unconsolidated resulting in the development of valleys, 'badland' topography (Tozer and Thorsteinsson,

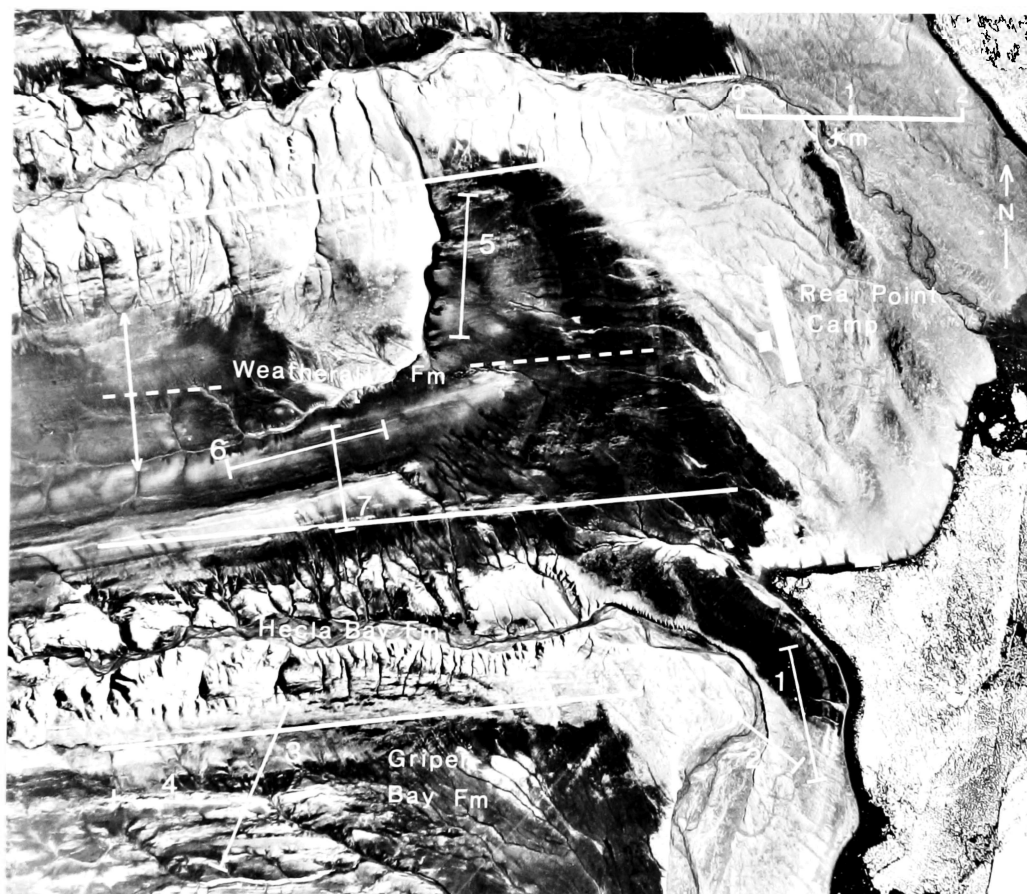


Figure 2.3 Air photograph of the Rea Point area (NAPL #A16763-76).

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A



B



Figure 2.4 Typical terrain-geology relationships developed upon Weatherall Formation siltstone and sandstone: A) Gently sloping uplands, approximately 1.5 km west of Rea Point, showing sparse tundra vegetation and instrumentation (Site W-1). Photo: H. M. French. B) Non-sorted stripes are typical of sloping terrain underlain by Weatherall Formation.

1964) and valley-side hoodoos (Figure 2.5a, b). These sandstones typically contain less than 2% kaolinite.

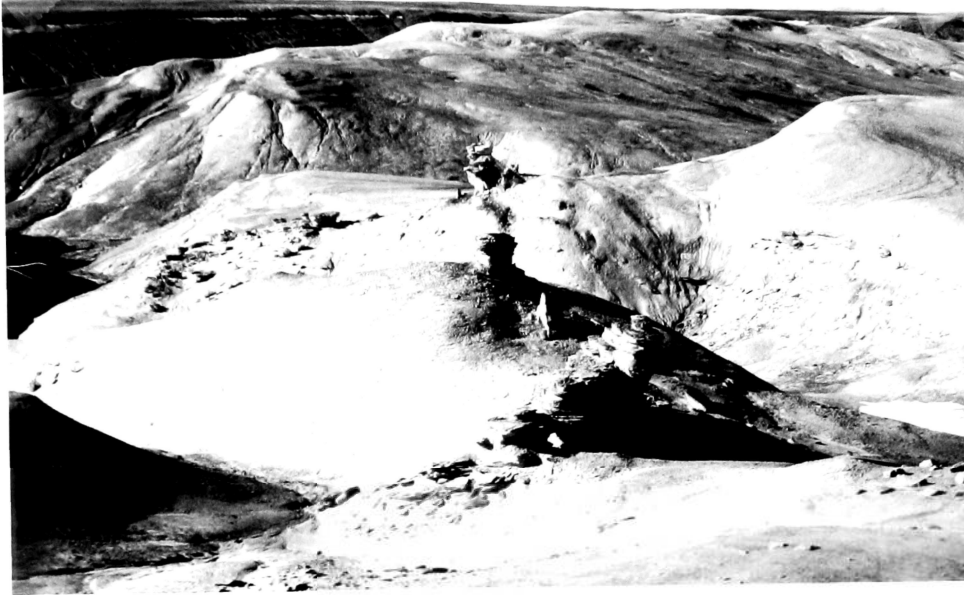
Brief mention should also be made of the youngest Devonian Formation within the Rea Point area, the Griper Bay Formation, and the Cenozoic sediments. The Griper Bay Formation consists of resistant, repetitive fine to medium-grained sandstone and recessive units of green fine-grained sandstone, siltstone and shale. While commonly forming lowland terrain elsewhere on Melville Island (Tozer and Thorsteinsson, 1964), the Griper Bay Formation forms uplands in the Rea Point area.

Cenozoic-age sediments consist of Quaternary deltaic and alluvial fine sand. They are mainly derived from Hecla Bay Formation sandstone following Holocene sea level regression. These sediments now form extensive flat and often poorly vegetated surfaces. The Rea Point facility and airstrip are located at one such surface (Figure 2.6a, b).

2.4 Glacial History

Many geomorphologists with experience in the high arctic agree that distinctive glacial landforms are rare north of Victoria Island. On Eastern Melville Island, there is only limited evidence of glaciation. Extensive and thick till sheets are generally restricted to the south coast of

A



B



Figure 2.5 Typical terrain-geology relationships developed upon Hecla Bay Formation sandstone: A) Both resistant and unresistant beds are present within the Hecla Bay Formation. The resistant beds commonly form tors while streams excavate valleys along weaker beds. Photo: H. M. French. B) Striking mushroom-shaped tors can occur in the valley side or valley bottom locations, as here, approximately 4.0 km west-southwest of Rea Point. Photo: H. M. French.

A



B



Figure 2.6 Typical terrain-geology relationships developed upon Quaternary sediments: A) Quaternary sediments are primarily derived from the weathering and erosion of Hecla Bay sandstones and occur in the form of alluvial and deltaic sands (shown here) near sea level and at the exits of major valleys. Photo: H. M. French. B) Rea Point facility is located on extensive area of Quaternary age sands. This slide is at south end of airstrip.

the island (e.g. Table Hills) and adjacent to the ice caps on Western Melville Island. Otherwise, glacial drift is thin and generally sparse (Craig and Fyles, 1960; Tozer and Thorsteinsson, 1964; Fyles, 1967; Hodgson et al., 1984). In a few localities, there are problematic surficial deposits (Figure 2.7a). It is the relatively common erratics that provide direct evidence of island-wide glaciation (Figure 2.7b). Craig and Fyles (1960) postulate that the widespread erratics and lack of fresh glacial landforms are evidence of a probable pre-Wisconsin glaciation. The erratics, often composed of granite and gneiss, are believed to have a southerly provenance (e.g. Victoria Island and the continental mainland).

A more recent glaciation, however, is indicated by records of Holocene emergence (McLaren and Barnett, 1978). Radiocarbon dating of marine shells collected from emerged beaches on northeastern Melville Island (Craig and Fyles, 1960; Location 14) are dated at 8,300 B.P. According to this line of evidence, coastal emergence increases from southwest to northeast across the island. Marine limits are 61, 72 and 100 m at Nelson Griffiths Point, Consett Head and Towson Point respectively. These data indicate that either the thickest ice mass was located in the northeast, thinning out to the south or the ice melted last in the northeast. Blake (1970) believes that an arctic ice cap (The 'Innuitian

a



b



Figure 2.7 A) Probable Quaternary-age sandy gravels (ice-contact?) occasionally overlie till-like deposits (K-22A; after Barnett et al., 1977) on points of Eastern Melville Island plateau and may indicate previous glaciation. Photo: H. M. French. B) Large striated erratic of Weatherall Formation, located approximately 3 km west of Rea Point at elevation of 95 m a.s.l. (i.e. above marine limit).

Ice Sheet') may have glaciated all of the island, ultimately retreating to the northeast.

England (1976a, b) concludes that the lack of fresh glacial landforms and limited till deposits in the Queen Elizabeth Islands reflect cold-based ice which is deprived of the aggradational and degradational power of more southerly ice sheets that have basal temperatures at or near the pressure melting point (see Shreve, 1984). Although glacial landforms are not obvious at Rea Point, erratics and raised deltaic sands attest to relatively recent glaciation and marine regression in the area.

2.5 Geocryology

2.5.1 Thermal conditions

Eastern Melville Island is underlain by perennially cryotic ground which is over 500 m thick (Taylor et al., 1982). Near-shore thermistor cables show permafrost depths ranging between about 140 m (Hole 200, Hecla I-69) and 290 m (Hole 259, Drake D-73) (Taylor and Judge, 1977). The mean annual ground temperature (MAGT) is approximately -17°C and the depth of zero annual amplitude ranges between 15 to 18 m (Stangl et al., 1982).

A 13 m thermistor string was installed in May 1985 to determine the ground temperature profile in the vicinity of Rea Point. The cable was installed on the crest of a

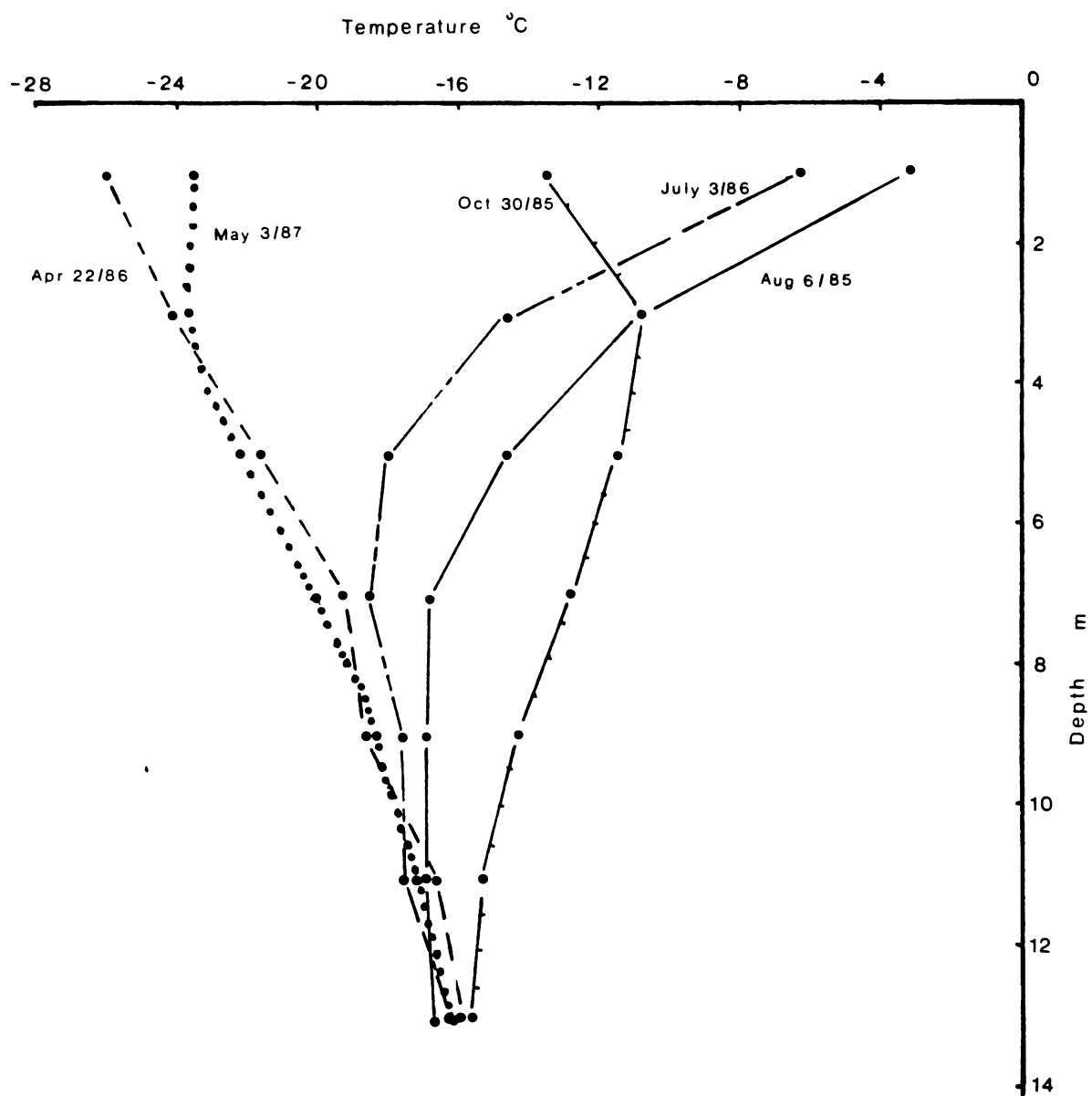


Figure 2.8 Selected ground temperature profiles, Rea Point vicinity.

hill approximately 100 m upslope of Site W (see Figure 2.15) and approximately 1.0 km west of Rea Point.

The trumpet curve, shown in Figure 2.8, indicates that the MAGT is -16.5°C . Ground temperatures above -5.0°C occur only in the top 1.0 to 1.5 m of the soil and then, only for a maximum of 100 days annually. Temperatures warmer than -10.0°C are found no deeper than 2.5 to 3.0 m. The amplitude of temperature with depth ranges from 22.7°C at 1.0 m to 2.7° at 11 m. The lowest near-surface soil gradient is measured in August and April with high gradients in June and September.

Two 4 m five-thermistor cables were also installed approximately 10 m from the 13 m cable. One was located in a trench which had been excavated (2.5 m depth) and then refilled a year before. The second was placed in a standard borehole. These cables enabled an assessment to be made of the speed with which thermal equilibrium, following installation, was attained (see Chapter 5, thermal equilibrium).

A plot of ground temperature versus time for different depths is relevant to any study of permafrost creep since creep will probably be most active when ground temperatures are warmest (see Chapter 1). Mean maximum temperatures exceed 0°C in early June and drop below 0°C during the first week of September at Rea Point. While minimum temperatures occasionally drop below 0°C during the

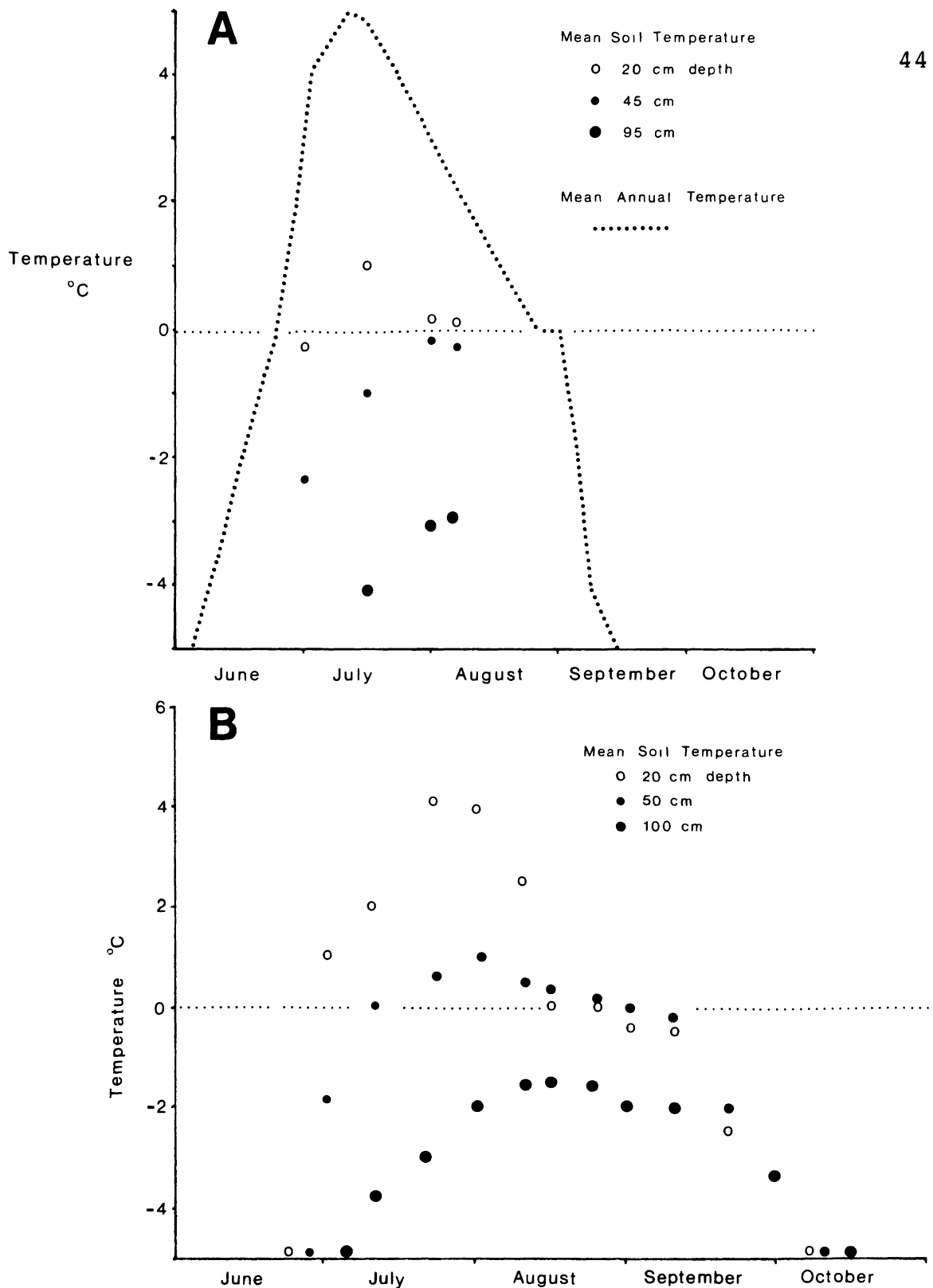


Figure 2.9 Temperature changes with date and time: **A)** Rea Point, in the active layer (20 cm depth) and in near-surface permafrost (45 & 95 cm depth) during summer of 1986; **B)** Resolute Bay, at depths of 20, 50 and 100 cm; average of 30 years' data.

summer, the mean minimum temperature is above freezing usually between the last week of June and the first week of August. Figure 2.9a presents data upon ground temperature and depth obtained at various times during the summer of 1986; each point plotted represents the mean of ten sites in the area (see Chapter 3, experimental design). The data indicate that there is variability, both in time and depth, for soils to experience relatively warm permafrost temperatures. As a consequence, the period of time for permafrost creep to occur will vary from locality to locality, and with depth. The data appear typical of the High Arctic since information provided by Atmospheric Environmental Services (AES) demonstrates that similar ground temperatures and trends occur at Resolute Bay, N.W.T. (Figure 2.9b).

A better indicator of the potential for permafrost creep is provided by Figure 2.10 which shows the approximate number of days with soil temperatures $>-3^{\circ}\text{C}$. At 65 cm depth, soil temperatures are warmer than -3°C for 50 and 75 days. At 100 cm depth, soil temperatures are greater than -3.0°C for 40 and 60 days. It is useful to compare these data with those of Savigny (1980) who measured frozen soil creep in the Mackenzie Valley. At his study site, the mean annual ground temperature was -3.0°C . Thus, the potential for frozen soil creep is provided year round in these 'warm' permafrost soils.

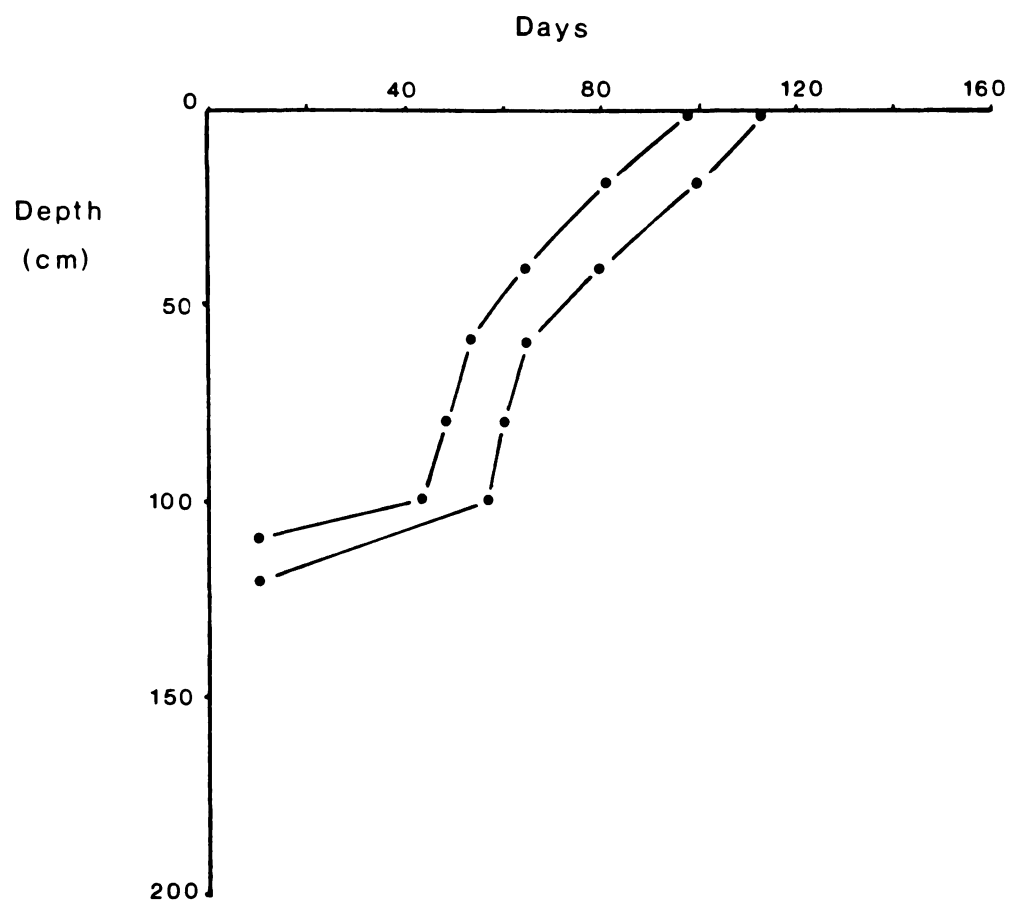


Figure 2.10 Approximate number of days with soil temperatures greater than -3.0°C (all study sites).

Figure 2.10 also indicates that the potential for frozen soil creep decreases with depth at Rea Point. Temperatures do not exceed -3.0°C at depths greater than 1.2 m. This contrasts with the Mackenzie Valley where temperatures are $>-3.0^{\circ}\text{C}$ throughout the permafrost profile (61.3 m). Savigny and Morgenstern (1986a) report frozen soil creep at depths exceeding 30 m.

2.5.2 Thermal contraction cracks

Thermal contraction cracks are ubiquitous in the study area (Figure 2.11a). They produce polygons which range from 10 to 20 m in diameter. The polygons are essentially hexagonal although on slopes, they become rectangular with the long axis perpendicular to the slope. The downslope cracks are difficult to detect and require careful observation. In contrast, cross-slope cracks appear as a series of low ridges with an average spacing of 7 m. Seventeen cross-slope cracks, for example, exist on the east-facing slope approximately 1 km due west of Rea Point (Figure 2.11b). In winter months, cracks averaging 8 mm across have been observed. In spring, as the cracks begin to close, it is possible to find miniature compressive ridges (up to 5 mm in height) where the surface is covered by a thin layer of densely packed snow or ice.

Near the coast and to the south of Rea Point is a small area of low centred polygons, the centres of which are

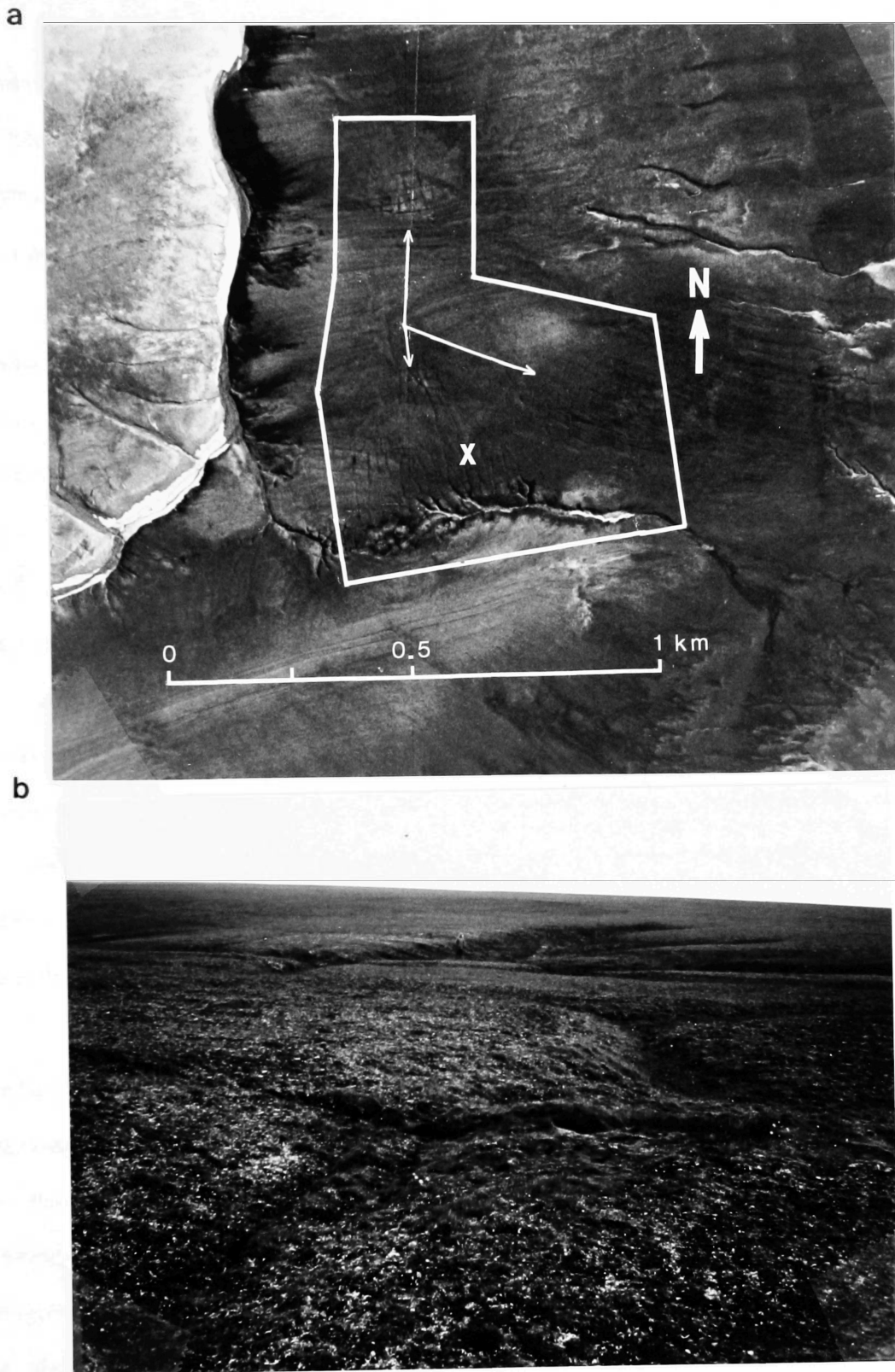


Figure 2.11 Thermal contraction cracks occur widely throughout the Rea Point vicinity: A) General views of thermal contraction crack polygons on upland terrain west of Rea Point (NAPL #A16763-76); B) Surface view of polygons at locality marked "x" on photo in A) above.

generally filled with water throughout the summer. However, to the west of Rea Point, high centred polygons are more common. On ridge crests, the surface expression of most polygons is subdued.

In many of the lowland areas of the Western Canadian Arctic (e.g. Mackenzie Delta, Central Banks Island, etc.), thermal contraction cracks commonly form ice wedges several metres thick and up to 1.0 - 2.0 m in width at the top. In many of the more arid high arctic islands, however, evidence of large ice wedges is not obvious. In a few places, ephemeral streams appear to have eroded small gullies along the lines of thermal contraction cracks, suggesting the presence of ice wedges, and slumping on the sides of the dump pit at Rea Point may also indicate small ice wedges or veins. During three years of fieldwork at Rea Point, however, no natural exposures of ice wedges were observed.

On the other hand, a number of ice wedges up to 1 m wide at the top were exposed in trenches excavated by EBA Engineering Ltd. in early 1981 (Figure 2.12). The ice wedge ice contained vertical foliations and sand infillings while adjacent colluvium displayed typical upward bending. In spring, thermal contraction cracks were observed within most ice wedges ranging from 5 mm width at the permafrost table to less than 1 mm at the base of the trench. Several secondary cracks or fractures at random angles were commonly



Figure 2.12 Photo from EPA trench data showing ice wedges and/or thermal contraction cracks. Photo: EBA Engineering Ltd.

associated with the main crack. In fall, no cracks were observed in the ice wedges.

Although the thermal contraction cracks form nets ranging from 10 to 20 m in diameter, ice wedges are less numerous, occurring at intervals of between 20 to 60 m in some trenches, while totally absent in others. Ice wedges were most common within the near-surface bedrock and residuum of the Weatherall Formation. Ice wedges were also observed within sediments of the Griper Bay Formation and within a local ice-rich vegetated terrace of Quaternary sands. The relative absence of ice wedges on Eastern Melville Island has already been commented upon (French et al., 1986). It is not clear why the distribution is irregular where the surficial geology is essentially the same (e.g. Weatherall Formation sediments). However, it may reflect the limited supply of free water in early summer in exposed areas where snow cover is slight.

As the propensity of a soil to crack following thermal contraction is a function of ice content (Williams, 1982, p. 170), no ice wedges were observed within ice-poor sediments of either the Hecla Bay Formation or the Quaternary sands.

2.5.3 Soils and vegetation

The cold ice-rich permafrost soils of the area are derived primarily from the cryogenic weathering of

underlying weakly lithified sandstone, siltstone, and shale (e.g. Hodgson et al., 1984; Stangl et al., 1982). This leads to the development of considerable quantities of fine sand and silt, materials which are particularly frost susceptible (Anderson et al., 1984) and conducive to the development of excess ice. According to Stangl et al. (1982), soil profiles may be divided into three zones, therefore, to reflect their derivation from cryogenic weathering of bedrock (Figure 2.13a, b). Normally, intact bedrock is overlain by fractured bedrock, which then grades into a residuum or surface regolith of cobbles, gravels and fines. On slopes, and at the base of slopes, the residual soils are overlain by silty, sandy, and gravelly colluvium, the product of widespread mass-wasting (solifluction) processes.

Barnett et al. (1977) observe that the highest proportions of fines occur in soils below marine limit. Such soils permit higher moisture contents than those above marine limit. The variation in moisture content is reflected by increased vegetation below marine limit.

The variable soil cover leads to a sparse and variable plant cover. Most areas have less than 20% vegetation cover (Barnett et al., 1977; Edlund, 1984). Soil moisture and the availability of nutrients control vegetation growth. Uplands are generally dryer and typified by sparse vegetation, except for isolated hummocks or

a



b

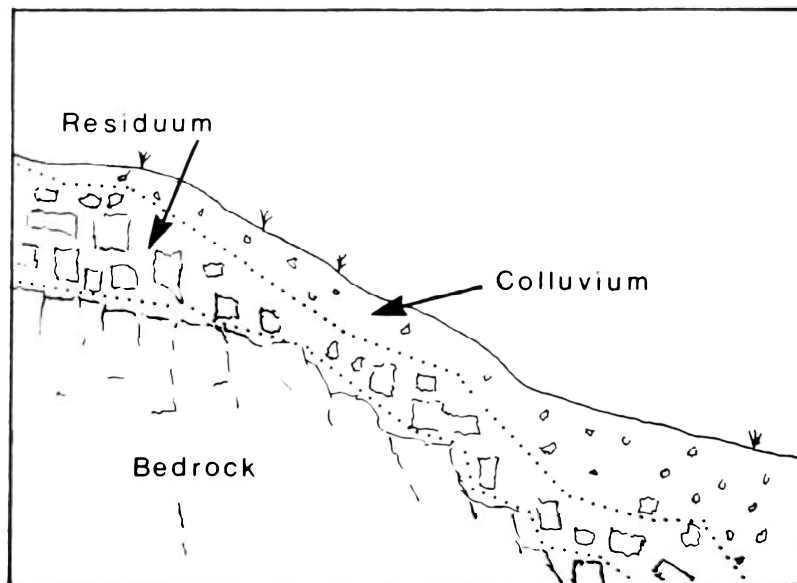


Figure 2.13 Soil and regolith conditions developed on Weatherall Formation shales and sandstones: A) Typical field observation. Photo: EBA Engineering Ltd. B) Sketch of soil and regolith zones, as typified in A) above.

'biomounds', and adjacent to fox dens and lemming burrows.

2.5.4 Patterned ground

Several types of patterned ground occur in the vicinity of Rea Point. These include: 1) non-sorted circles (earth hummocks and mudboils) and 2) non-sorted stripes. Typically, the patterned ground varies according to slope angle; as slope angle increases, circles progressively change to stripes. Then, near the base of the slope, non-sorted circles re-appear, first as elongate forms and then, on slopes $<3^\circ$, as circular forms.

The non-sorted stripes are often delimited by lines of vegetation (Figure 2.14a). The stripes are between 50 to 110 cm wide. Excavations indicate the unvegetated central part of the stripe is composed of fine silty sediment while the vegetated lines are underlain by coarse sediment. These observations suggest that the stripes are the product primarily of meltwater eluviating fines and providing soil aeration suitable for plants while the central part is disturbed due to frost heave and creep.

Non-sorted circles, or earth hummocks, are the most common patterned ground feature on ridge crests and on slopes less than about 2° (Figure 2.14b). They are commonly 1 to 2 m in diameter and 30 cm high. Composed of a mixture of silt and gravel, they are thought to be formed and maintained primarily by cryoturbation processes, as

a



b



Figure 2.14 Patterned ground in the vicinity of Rea Point: A) Non-sorted stripes outlined by vegetation patterning; B) Non-sorted frost-churned feature formed in coarse regolith derived from sandy facies of Weatherall Formation.

c



d



Figure 2.14 C) Mudboils (non-sorted circles) on gently sloping terrain east of Rea Point base camp and situated below marine limit; D) Cross-section through mudboil showing diapiric structures.

described elsewhere by Schunke and Zoltai (1988) and Mackay (1980). In a few localities, mudboils, rather than earth hummocks, occur (Figure 2.14c, d). Ranging from 0.4 to 1.0 m in diameter, they are commonly located at or near the base of slopes. They are thought to be similar to the extrusion features described by Shilts (1978) and Egginton and Dyke (1982) from Keewatin. Located within the deltaic Quaternary sands, the mudboils are predominantly composed of clayey silt to silt (mud) with gravel and capped by a semi-rigid (desiccated) clay. Excavations indicate that the silty mud is extruded from beneath the semi-rigid carapace, probably in later summer at times of soil saturation/high pore pressure. Live willow branches and leaves buried by extruded mud observed in August 1986 attest to the activity of some of the mudboils.

2.6 Study Locations

Five slopes of different aspect and slope angle were selected to measure permafrost creep (Figure 2.15). These were 1) a north and a south-facing slope (N and S) approximately 5 km west of Rea Point, 2) an east and a west-facing slope (E2, W) in a small asymmetric valley northwest of the Rea Point camp, and, 3) an east-facing slope (E1) overlooking the Rea Point camp (herein called the campslope). With the exception of Slope S, all the slopes

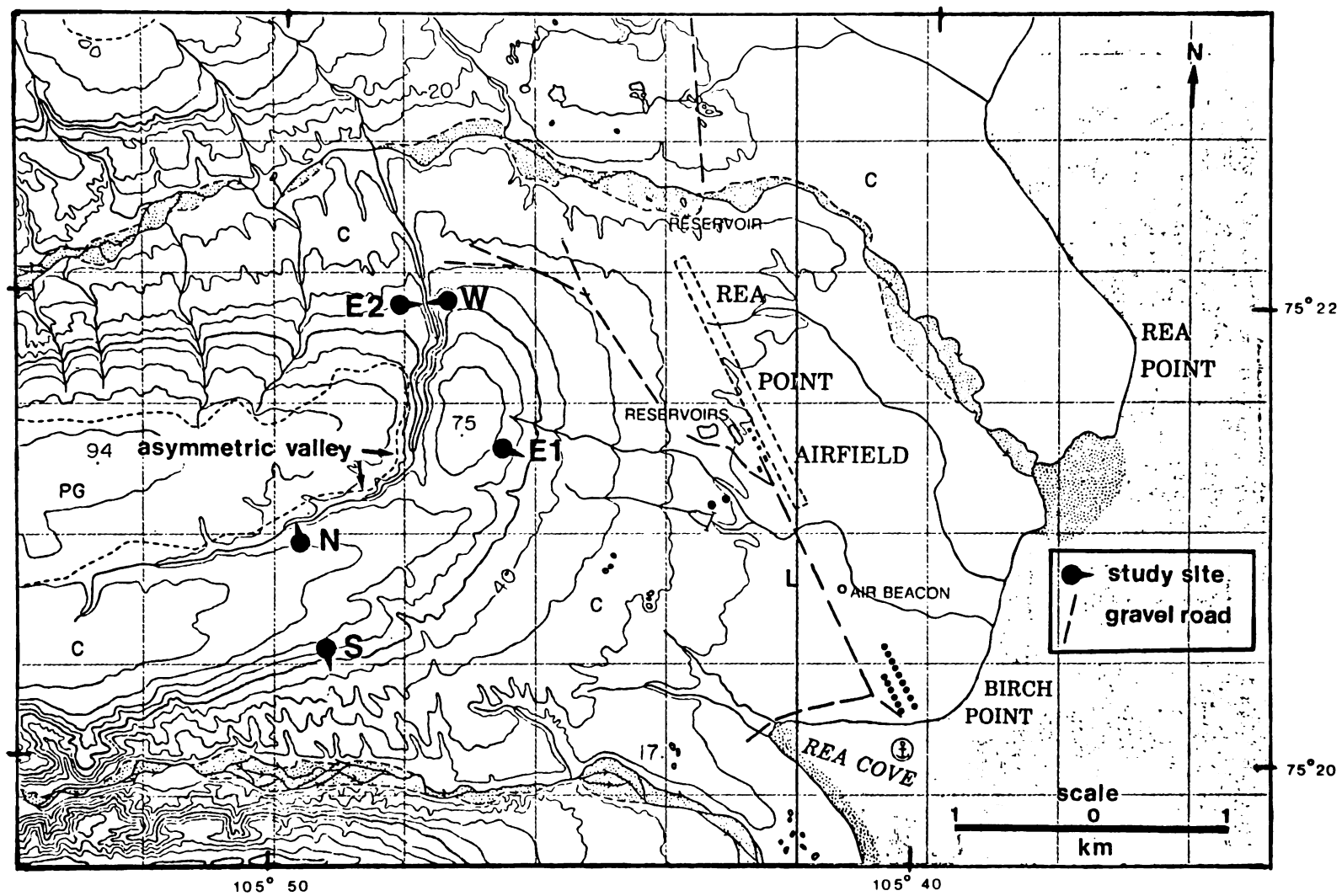


Figure 2.15 Location of the five study slopes.

were developed upon Weatherall Formation sandstones and shales.

Slopes W, E2 and N are located along the slopes of a small asymmetric valley exhibiting gentle south and east-facing slopes and steep north and west-facing slopes. Slope S is located on a gentle south-facing slope which extends to a major east-west valley south of Rea Point. Slope E, located on the gentle east-facing campslope, extends east to the Quaternary deltaic sands upon which the Rea Point camp and airstrip are located. Details of each of the five slopes are provided in Table 2.2 and Figure 2.16.

Slope W is located approximately 50 m south of the gravel pit road, left fork terminus. The dominantly convex slope is adjacent to the small asymmetric valley and possesses vegetated hummocky stripes and unvegetated gravelly stripes. Vegetation on the stripes is restricted primarily to small patches of moss, lichen and patina. Saxifraga sp. is the most common flora on the vegetated stripes. Exposed, steeply dipping bedrock running perpendicular to the valley is exposed 10 to 20 m both north and south of the site.

Slope E2 is located on the opposite side of the asymmetric valley, facing Slope W. Vegetation is sparse. Long-lasting snowbanks, which dominate this side of the valley, collect aeolian and fluvial sediments during the

Table 2.2
Details of the Five Study Slopes

Study Slope	Total Slope Length (m)	Distance from Base of Slope (m)	Height Above Base of Slope (m)	Maximum Slope Angle	Aspect
W	250	100	45-50	24°	West
E1	1000	750	60	6°	East
F2	400	175	50	14°	East
N	250	100	13	21°	North
S	500	325	80	8°	South

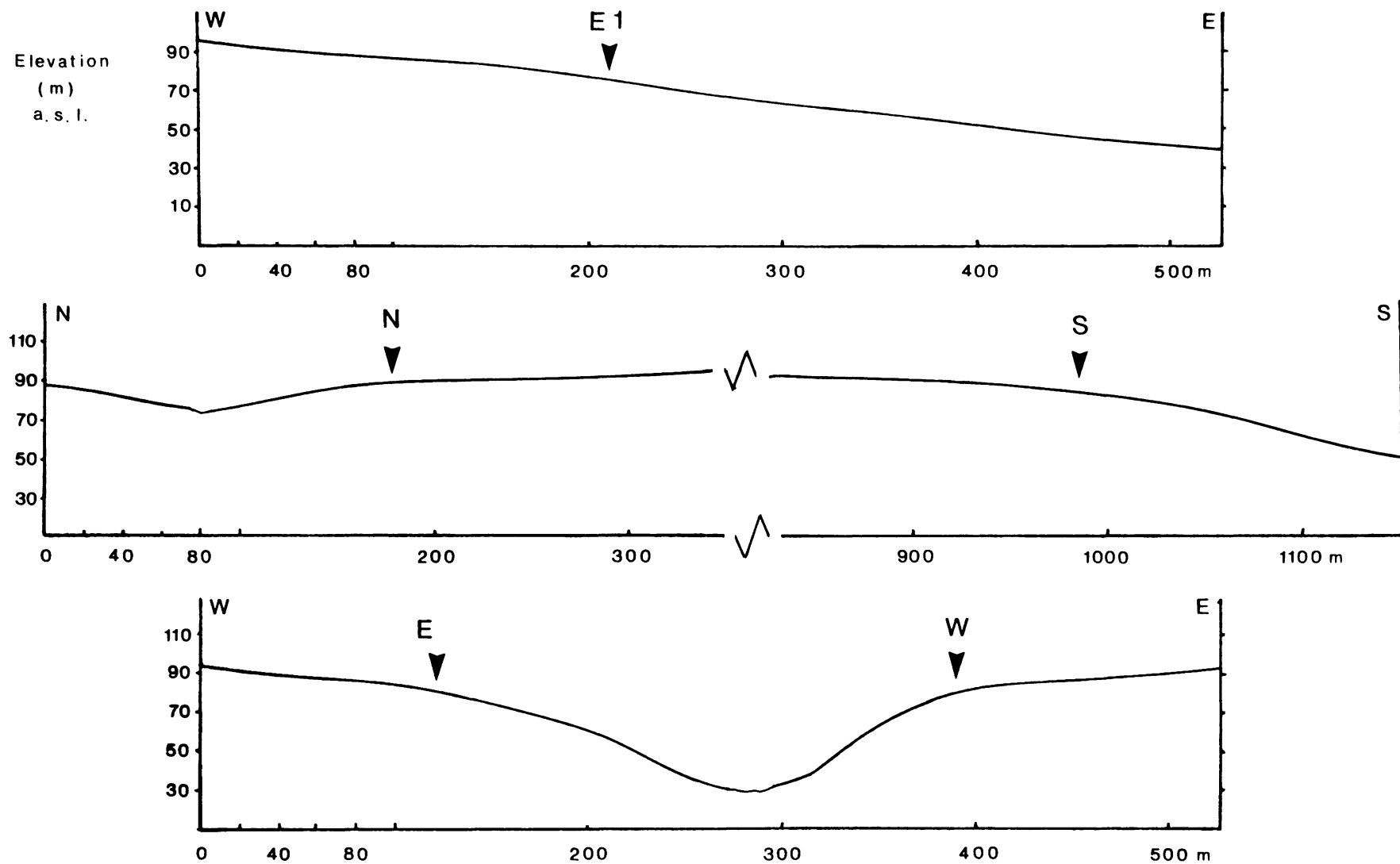


Figure 2.16 Morphology of the 5 study slopes. The locations of the instrumented sites are indicated by the black arrows.

winter and spring, and these, in turn, are deposited during the spring melt.

Slope E1 is located on the east-facing camp slope. Gravelly stripes dominate the slope surface. A polygonal network of frost cracks are located both on the slope and the ridge crest.

Slope S is the only site located on Hecla Bay Formation colluvium. The surface materials are composed primarily of sands and sandy gravel rather than silt. Stripes are somewhat subdued at this site. Slope S is on the shallow south-facing slope of the major east-west valley located just south of Rea Point.

Slope N is separated from Slope S by a wide (1 km) east-west trending ridge, the latter marking the approximate position of the geological boundary between Hecla Bay and Weatherall Formations. The slope is dominated by gravelly stripes.

All of the instrumented sites were created on slope segments of between 3 - 8° in angle. This was deliberate since much of the lowlands of the Western and High Arctic consists of gently undulating topography with slopes of this magnitude. The morphology of the five slopes was also thought to be typical of extensive areas of Eastern Melville Island and the Western Arctic. Slope E1 was a near-rectilinear slope with a very gentle lower concavity. Slopes N and S constituted the two sides of a broad shallow

convex interfluve. Slope E2 was a near typical convexo-concavo slope, while Slope W was a dominantly convex slope. Maximum slope angles ranged from 24° (Slope W) to 6° (Slope E1).

CHAPTER THREE

INSTRUMENTATION AND EXPERIMENTAL DESIGN

3.1 Introduction

The cold temperature inclinometer is generally considered to be one of the most sophisticated instruments for use in the measurement of downslope creep (Bromhead, 1984). It is rapidly becoming a research standard in the field study of soil and ice rheology (e.g. Hooke and Hanson, 1986; Hooke et al., 1987; Savigny and Morgenstern, 1986a). In this study, thermistor strings and heavemeters were used in conjunction with the inclinometer since permafrost creep is significantly affected by temperature and changing ground ice conditions. To complement this approach, more traditional movement data (i.e. solifluction or gelifluction) have been obtained by the measurement of objects inserted within the active layer (e.g. see French, 1974; Egginton and French, 1985) and by marked pebbles on the ground surface (e.g. see Caine, 1981; Perez, 1988).

A weakness of the study was the absence of any measurement of the unfrozen water content, as determined by TDR or neutron probes. In this study, heavemeters were used to monitor changes in water content and provided a general indication of its importance within permafrost.

3.2 Inclinometer System

Since many mass-wasting processes are slow and subtle, they require sensitive instruments that are dependable over the long term and capable of operating through a wide range of temperatures. Although many types of inclinometers exist (e.g. Franklin, 1984), a servo-accelerometer type downhole or probe inclinometer was used in this study. The inclinometer, manufactured by Terra Technology of Redmond, Washington, provides both accurate and precise data at the extreme sub-zero temperatures experienced in the north. It is designed to generate reliable data in cases of very limited deformation with negligible non-linearity and hysteresis.

The inclinometer system consists of 1) PCV casing, 2) inclinometer probe or torpedo, 3) cable, and 4) readout unit. The system is designed to detect changes in the downslope configuration of slope soils to 0.0003 cm over a 30 cm vertical distance. Accuracy of the system, however, as measured in field tests is demonstrated by maximum variability of 0.0061 cm for repeated readings at a single point. Results are well inside the resolution of the system.

The PCV casing (SINCA, Richmond, B.C.) came with a standard outside diameter (OD) of 7.09 cm. The casing contains 4 longitudinal pre-cut grooves which are located



Figure 3.1 FN60 Nodwell with seismic drill in operation, Slope W, May 1985.

equidistant around the interior. Though highly flexible lengthwise, the casing is virtually incompressible. Prior to installation, the casing was capped and sealed at the base with silicon and a rubber (butyl) tape. The top was capped and sealed with cold temperature (-18°C) electric tape.

Drilling was undertaken in May 1985 using a seismic drill (10.2 cm diameter) mounted on a FN60 Nodwell (Figure 3.1). Boreholes, approximately 2.8 m deep, penetrated the regional bedrock (see Section 2.4.6) and allowed 20 cm of casing to be exposed above the soil line. The PVC casing was surrounded by a slurry of grout mixed in situ and damped in place by a long thin metal rod. The grout was developed by Haliburton Services Ltd. of Edmonton and intended for use in filling all void space between casing and soils in a permafrost environment. Sold under the trade name 'Permafrost Cement', it is designed to set rapidly at sub-zero temperatures with minimum expansion. This Permafrost Cement was used successfully by Savigny (1980) in similar circumstances. The grout was left to set and temperatures allowed to stabilize prior to data collection during August, 1985.

The inclinometer is a Terra Technology, Mini Probe MP20. It is connected to a digital readout unit (Terra Technology, Model TPC-101) by 15 m of flexible, epoxy-encased cable. The grooves within the borehole casing allow

a precise directional positioning of the inclinometer each time a study site is monitored. Markings at 1 foot intervals (30.48 cm) allow for a precise depth positioning of the probe.

While the inclinometer functions properly in soils to temperatures as low as -40.0°C , LCD's on the readout unit become inoperable at temperatures less than -10.0°C . Thus, the readout unit was protected from the cold when measurements were taken between October and April. A portable, collapsible tent and naphtha gas camp stove were used. After readings were taken, the readout box was wrapped in bubble plastic packaging and stored within a large wooden box for transit. The connection between the probe and readout was maintained throughout the day to prevent condensation on the contacts while moving the tent from site to site.

3.3 Thermistors

The thermistor strings were constructed of 16-strand wire cable protected by a flexible PVC cover. Standard YSI-44033 thermistors with a factory-rated accuracy of $\pm 0.1^{\circ}\text{C}$ were utilized. The resistance of each thermistor, first tested in the lab, was also measured in the field prior to installation with a Beckman Tech 310 multimeter connected to the desired thermistor and common wire.

Thirteen multi-thermistor cables were installed in boreholes to a maximum depth of 13 m in May 1985. The 10 cm diameter boreholes were carefully backfilled with soil while the thermistor strings were held in position. A description of the thermistor cables constructed, and their employment in the field, is given in Table 3.1a and Figure 3.2.

In summary, thermistor cables were of three types. A single 13 m cable of type A (Figure 3.2) was installed to provide a ground temperature profile for the Rea Point area and aid in the estimation of mean annual ground temperature (MAGT). Seven thermistors placed at 2.0 m intervals along the cable provided the data.

Two 4 m cables of type B were installed to measure the rate at which thermal equilibrium was re-established after disturbance (by drilling). Four thermistors placed at 1 m intervals along the cable provided the data. One cable was set in a 2 m long, 2 m deep trench; the other in a standard inclinometer casing borehole.

Ten cables of type C were installed in order to determine the ground thermal regime at each instrumented site (see Figure 3.4). Seven thermistors placed at 12.5 cm intervals provided thermal data in the near-surface permafrost (upper 95 cm of soil). Four more thermistors, placed at 25 cm intervals, measured ground temperatures to 1.95 m depth. A twelfth thermistor was placed on the cable at most sites to give temperature readings at 2.40 m depth.

Table 3.1
List of Design Specifications
(A) Termistor cables; (B) Heavemeters

(A)

Category	Number of Cables	Number of Thermistors per Cable	Common Wire	Designated Length (m)	Location
A	1	7	Black	14.0	Undisturbed bed ridge-top location
B	2	4	Black	5.0	Trench site undisturbed soils
C	11	12	Black	2.5	E1, E2 N, S, W

(B)

Tube Number	Length	Outside Diameter	Tube Wall Thickness
1	120 cm	1.27 cm (0.5")	0.124 cm (0.049")
2	100 cm	1.59 cm (0.625")	0.124 cm (0.049")
3	80 cm	1.91 cm (0.75")	0.124 cm (0.049")
4	40 cm	2.22 cm (0.875")	0.124 cm (0.049")

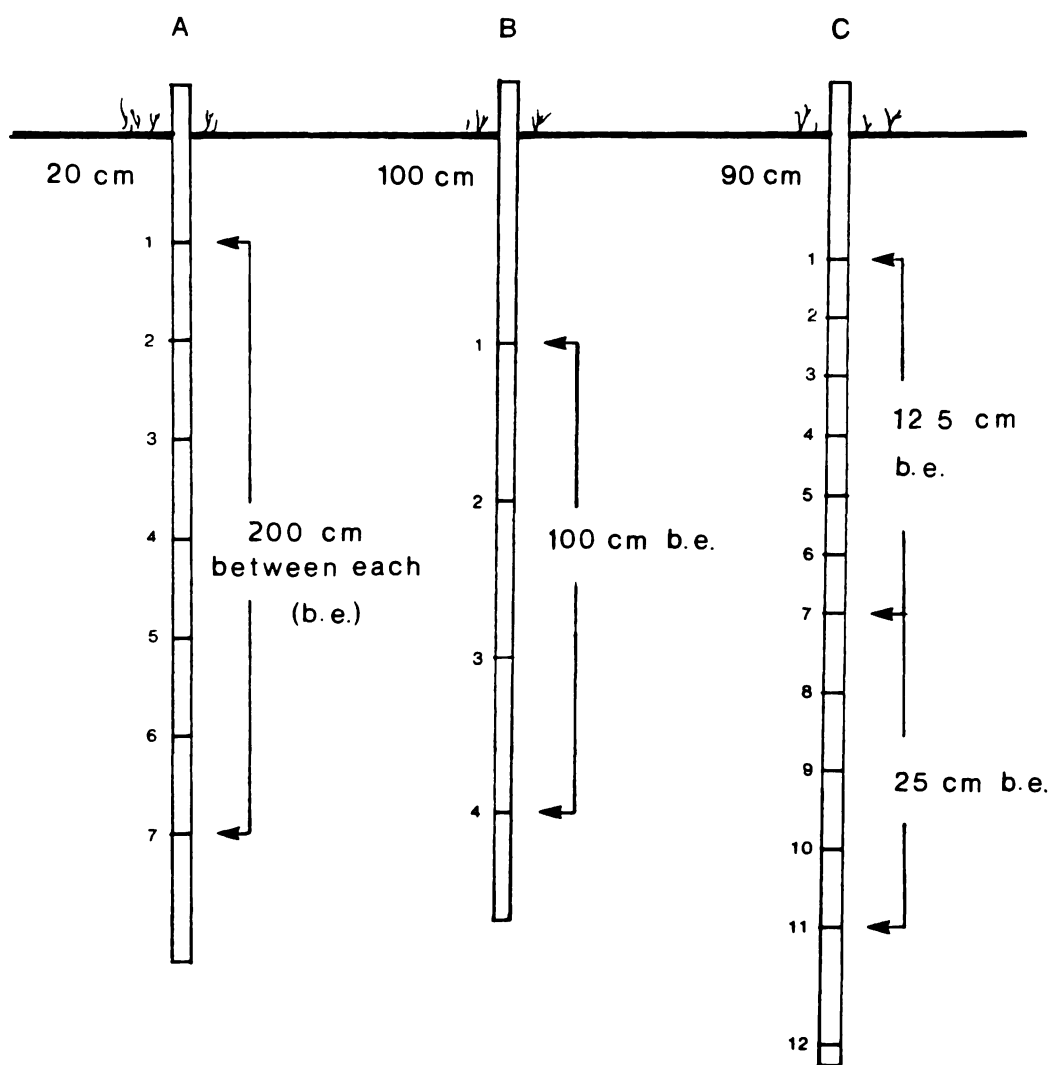


Figure 3.2 Diagram illustrating thermistor string design and emplacement.

3.4 Heavemeters

Four-tube telescoping heavemeters were constructed using drawn aluminum tubing of quality grade D6061-T6 (see Table 3.1b). A 3.8 cm hand auger was used to drill holes for the heavemeters in conjunction with a 3.0 cm pointed steel rod and sledge hammer (the 'permafrost punch'). These were located within 1 m of the thermistor cable and inclinometer casing. After the heavemeter was installed in the vertical hole, careful backfilling assured contact between the meter and the wall.

Heavemeters reflect the volumetric expansion of freezing water and the degree of ice segregation (ice lense formation); i.e. vertical deformation stresses within the soil. The accuracy of this information depends upon correct placement of the individual heavemeter tubes. At Rea Point, where the average depth of the active layer ranges between 20 and 60 cm, depths of 15, 45, and 55 cm were chosen for heave measurements. Figure 3.3 illustrates the heavemeter design and its emplacement. In calculating heave, measurements from one set of readings were compared with those from the previous set, thereby indicating whether heave or settlement occurred over the short term.

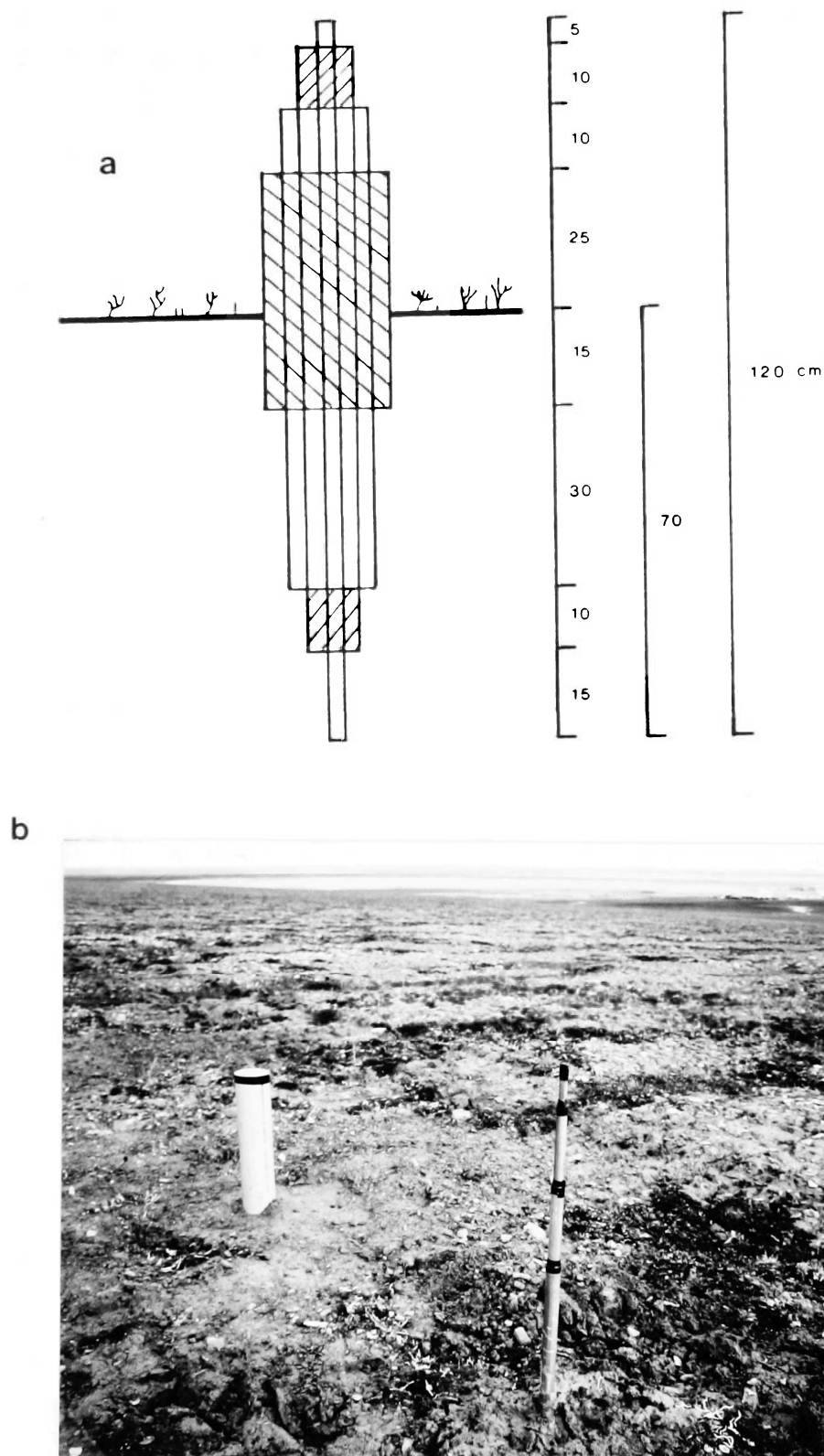


Figure 3.3 A) Heavemeter design; B) Photograph of heavemeter and inclinometer casing in place at Site E1 late July 1986.

3.5 Traditional Solifluction Measurements

Marked pebbles and objects placed within the active layer were employed to establish rates of movement in the seasonally cryotic zone, and to identify any variability of movement across the slope.

At each measurement site, lines of painted pebbles (3 to 4 cm maximum diameter) were placed across the slope at 10 to 20 cm distance and tin foil strips placed in the soil. Subsequent measurements were made against a datum line established by extending a taut polypropylene rope between two steel stakes hammered at least 50 cm into the permafrost. Measurements made annually recorded the difference between the downslope side of the rope and the upslope edge of the pebble or tin foil strips.

3.6 Experimental Design

3.6.1 Permafrost creep

A number of sites were selected on each slope for instrumentation. At one site, the inclinometer casing emerged from the ground within 24 hours of installation, either due to bouyancy effects of the freezing and expansion of excess water in the borehole. Slope W had four instrumented sites, the most of any of the slopes under study. This concentration of instrumented sites reflected two factors:

- 1) ground ice

data from an adjacent EBA trench provide excellent control on this variable, and 2) the slope exhibits well developed non-sorted stripes indicating active movement of surface layers, while at the base of the slope, several small rotational slumps indicate instability of the slope. Slopes E2 and E1 each had two instrumented sites, while Slopes N and S each had one instrumented site. In total, 10 sites were instrumented, each consisting of a heavemeter, a 12-string thermistor cable and an inclinometer casing. Figure 3.4 illustrates the location of the sites and Figure 3.5a, b illustrates the typical instrumented site in the field.

Inclinometer data were gathered four times a year. Measurements were taken on August 6 and October 31, 1985; April 22, July 3 and November 22, 1986; May 3 and July 22, 1987. Following the procedure outlined by Savigny (1980), inclinometer measurements were taken in the same order and all on the same day.

Inclinometer readings were recorded at every 30 cm commencing at the 35 cm depth. As the inclinometer torpedo is 60 cm long, creep measurements at the 35 cm depth actually incorporate all deformation between 35 cm and 95 cm depth (Figure 3.6). Similarly, creep measurements at 65 cm incorporate all movement between 65 cm and 125 cm depth. Readings were taken to 190 cm depth.

Critical to the accuracy and reliability of the inclinometer used in this study is the immobility of the

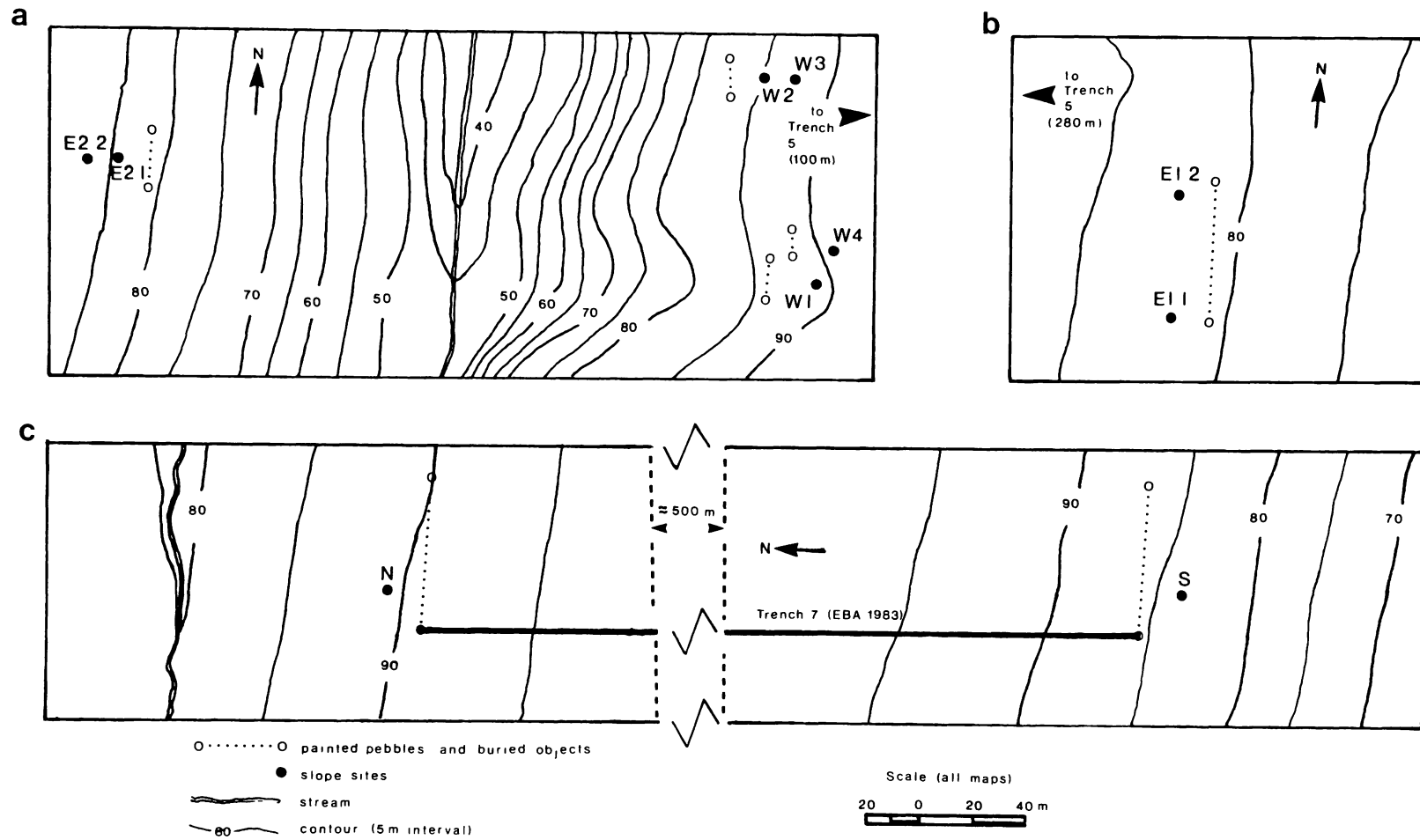
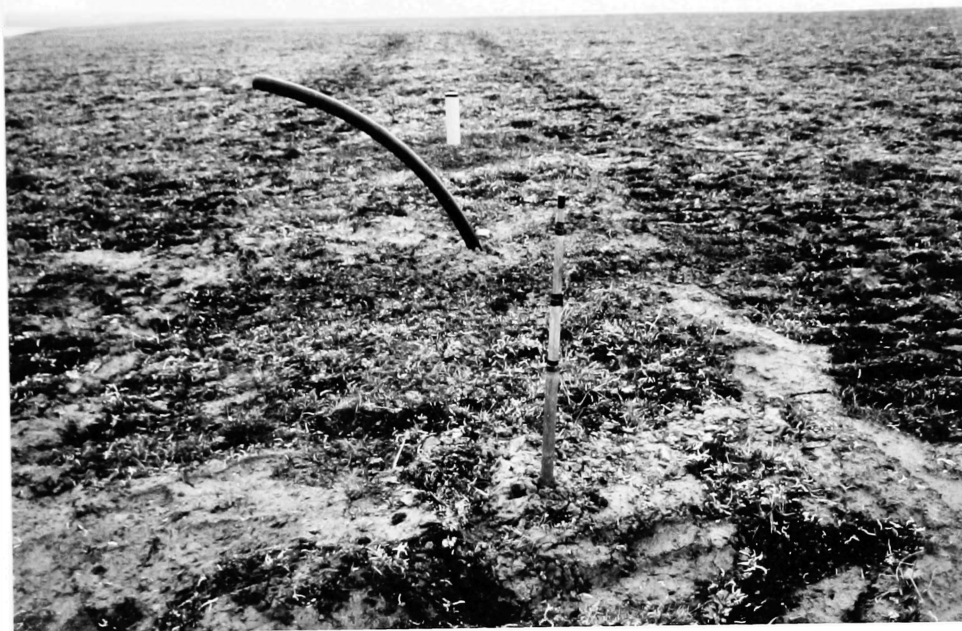


Figure 3.4 Detailed maps of instrumented sites: A) E2 and W; B) E1; C) N and S. For locations of A), B) and C), refer to Figure 2.15.

a



b



Figure 3.5 Instrumented Sites N (August 1985) and E2 (April 1986). Similar instrumentation to this was installed also at Sites W, E1 and S.

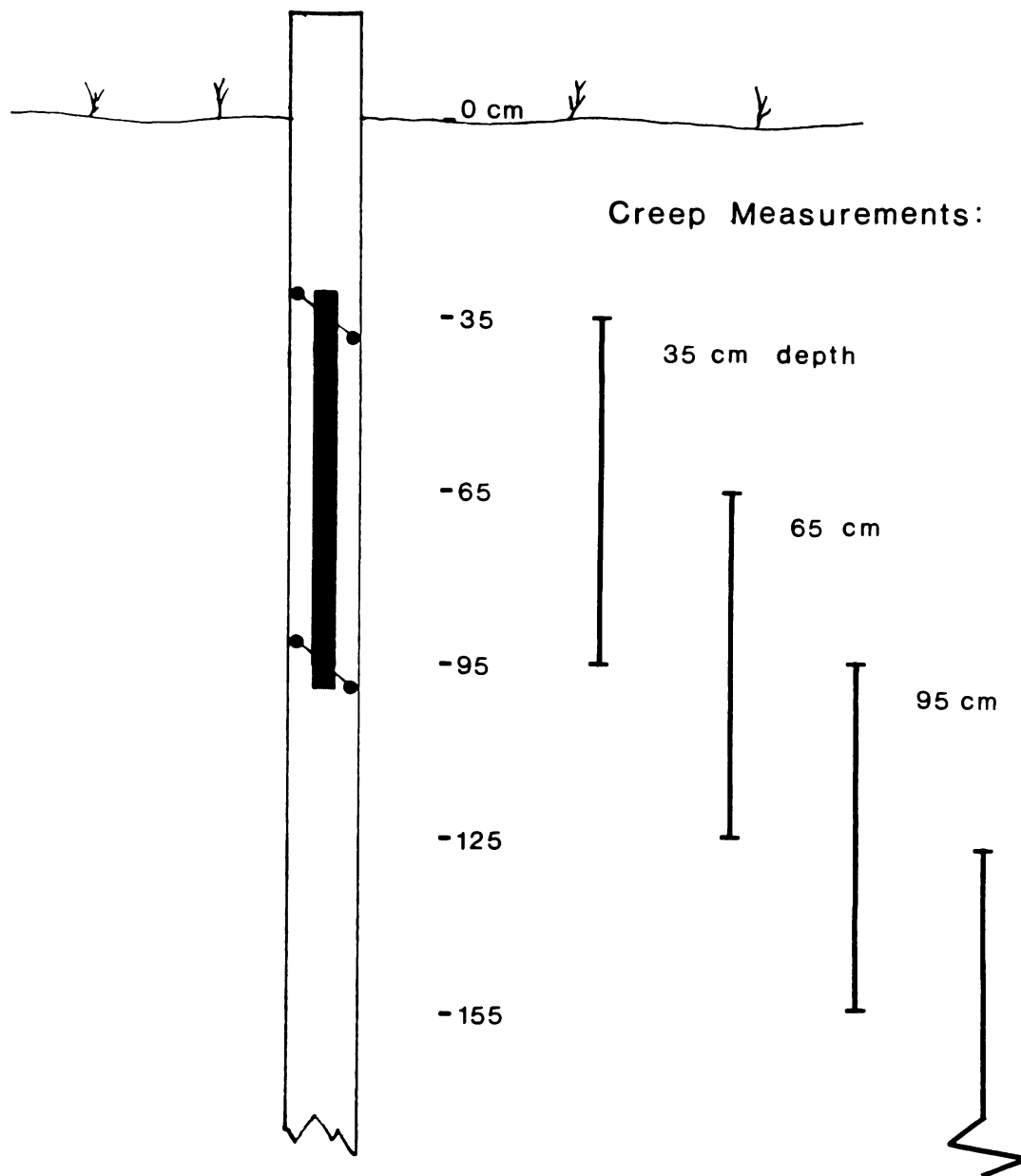


Figure 3.6 Relationship between inclinometer torpedo, casing and frozen soil creep results.

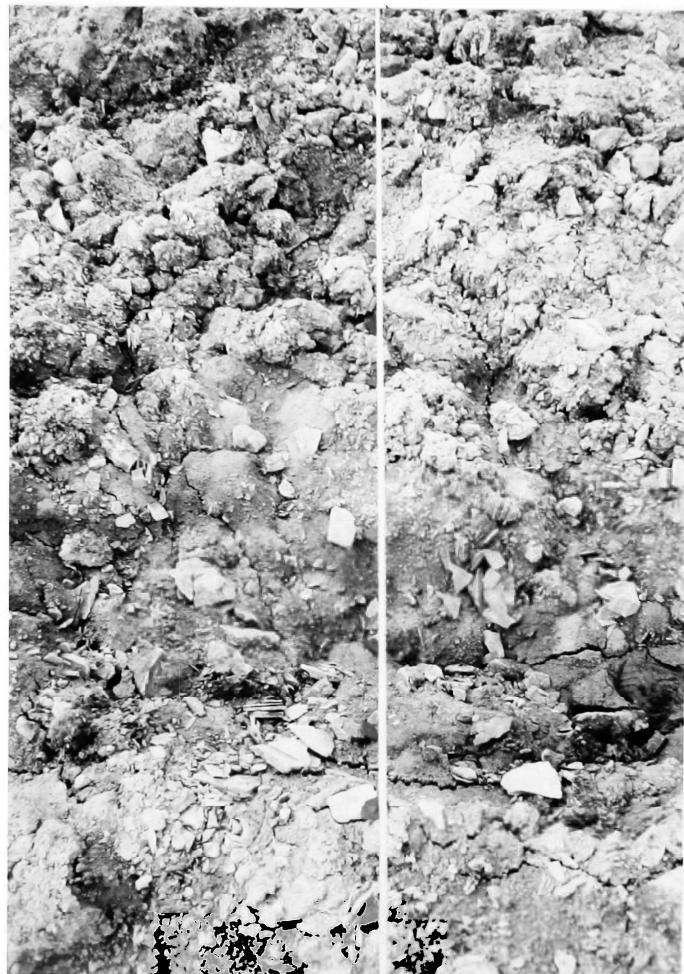
base of the casing. In all cases, the inclinometer casing (which extends 2.8 m below the surface) is installed in bedrock. Thus, permafrost creep is assumed to be $0.0000 \text{ cm yr}^{-1}$ at the base of the casing.

3.6.2 Solifluction

Totals of between 200 and 500 painted pebbles were placed on the surface of each of the 5 slopes along a number of surveyed lines (see Figure 3.7a). In addition, 37 shallow pits were excavated at various positions along each of the lines of marked pebbles. Four strips of aluminum foil, 3 cm by 10 cm in dimension, were placed on the downslope wall of each pit (e.g. see French, 1974; Figure 3.7b) at depths of 1, 10, 15 and 30 cm.

The marked pebbles were measured during the summer of each year. The buried objects were measured only after two years of deformation and several pits were left intact for future studies.

a



b



Figure 3.7 Solifluction measurements: A) Painted pebbles, Slope 5, July 1984; B) Tin foil markers being inserted, Slope N, July 1984.

CHAPTER FOUR

GROUND ICE CONDITIONS

4.1 Introduction

This chapter describes ground ice conditions associated with different geologic formations on Eastern Melville Island. A summary paper is already available (French, Bennett and Hailey, 1985). These relationships are of great relevance to the study of permafrost creep at Rea Point since all five slopes are underlain by Paleozoic age siltstones, shales and/or sandstones (i.e. Weatherall and Hecla Bay Formations). The nature of the surface regolith and the associated ground ice amount and distribution will be intimately related to the underlying bedrock. Moreover, given the potential for hydrocarbon development on Eastern Melville Island, it is necessary to establish whether or not the results from Rea Point can be extrapolated to other areas. In situ permafrost creep, if of a critical magnitude, together with thaw of ice-rich permafrost, may pose a threat to the long term stability of a pipeline or other engineering structure.

With respect to permafrost creep, ground ice is a primary creep regulator since the creep strength of a frozen soil varies directly with the ground ice content. Ground ice is also a primary creep activator since frozen soil

creep can be initiated by ice segregation processes or secondary frost heave.

As part of a geotechnical evaluation undertaken for the proposed Melville Island pipeline, field ditching trials were performed at Rea Point during the winter of 1981 - 1982 (EBA Engineering Consultants Ltd., 1983; Hayley et al., 1984). Following an air photo interpretation study of the entire route, seven sites judged to be representative of the terrain conditions along the majority of the pipeline route and all located within a 10 km radius of Rea Point, were selected for trials. The wall logs compiled from the trenches were kindly made available to the author by EBA Engineering Ltd. These data, together with other permafrost observations obtained from exploratory wellsites and boreholes on Eastern Melville Island, enable the calculation of total ice content in the upper 2 to 4 m of permafrost at localities similar to those instrumented in this study. This chapter attempts to relate these ice contents to bedrock, surficial geology and in situ permafrost creep.

With certain qualifications, the data contained in this chapter are thought typical of large areas of Eastern Melville Island. It is recognized, however, that ground ice conditions will vary according to local slope and drainage conditions. For example, data are from sites of generally low relief and, therefore, may not be typical of more rugged terrain known to exist on the island. EBA Engineering

(1983, p. 18) has found, by means of a series of test boreholes across the island, that ice contents generally increase as one moves from ridge crest to colluvial slope. The difference in ice content by volume averages about 20%. Furthermore, surficial geology is an important influence upon ground ice condition. As stressed by Barnett et al. (1977, pp. 14, 16), localities below marine limit are usually characterized by higher silt and ground ice contents than areas above marine limit, and the relationship between ground ice and vegetation cover is not straightforward.

4.2 Determination of Volumetric Ice Content

Ground ice observations were made by EBA personnel in seven trenches cut through Paleozoic (Devonian) clastic sediments and Quaternary silts and sands near Rea Point in the winter of 1980 - 1981 (EBA Engineering Consultants Ltd., 1983). The trenches ranged between 61 and 782 m in length and averaged about 2 m in depth (see Table 4.1). In this thesis, data are extracted from trenches 3, 5, 6 and 7. Frozen soil samples had been taken at various depths at 10 - 20 m intervals along each trench. These samples provided moisture content and frozen bulk density information. Visible ice contents were added to the detailed trench wall logs.

Table 4.1
Summary of Trench Locations Near Rea Point*

Site No.	Ditch Length (m)	Terrain Conditions	Comparable terrain on Eastern Melville Island
1	467	Vegetated sand terrace with polygons and un-vegetated deltaic deposits	Vegetated terrace and un-vegetated deltaic plains of Sabine Lowland
2	212.5	Deltaic and alluvial deposits: fine sand derived from Hecla Bay Formation, small stream crossing	Unvegetated deltaic plains of Sabine Lowland, numerous stream crossings in lowlands
3	378.6	Residual and colluvial fine-grained soils overlying both Hecla and Griper Bay bedrock	Typical of Edmund Lyon Hills
4	61	Residual and colluvial fine-grained soils overlying Griper Bay bedrock	Typical of Edmund Lyon Hills
5	529.6	Residual soil along hill crest underlain by Weatherall Formation	Typical of Bridport Inlet region, Edmund Lyon Hills, and Sabine Bay region
6	300.9	Residual and colluvial soils derived from Weatherall Formation	Typical of Bridport Inlet region, Edmund Lyon Hills, and Sabine Bay region
7	782.1	Residual and colluvial derived from Weatherall Formation Ridge crest with gentle slopes to north and south	Typical of Bridport Inlet region, Edmund Lyon Hills, and Sabine Bay region

*Source: EBA Engineering Consultants Ltd. (1983, Table 1).

Volumetric ice contents (I/C) were calculated using a method developed by Pollard and French (1980). When given the laboratory results of moisture content (w) tests (EBA Engineering Consultants Ltd., 1983), volumetric ice content (percent) is

$$I/C = 100(w + 0.09w) \quad (4.1)$$

This equation simply reflects the 9% increase in the volume of water as it freezes in an unsaturated soil. However, this equation is not valid if the calculated ice content is greater than the natural porosity (η) of the soil (i.e. excess ice is present). When soils are saturated, the freezing water cannot simply expand into voids or air pockets within it; rather, the soil must expand to accommodate the ice. This effectively changes the moisture content of a given volume of soil. Clearly, the volume fraction of the solids (V_s) is reduced.

In saturated soils, ice content may be determined as follows:

$$I/C = V_1 \times 100 / (V_1 + V_s) \quad (4.2)$$

where

$$V_1 = Mw + 0.09Mw \quad (4.3)$$

and

$$V_s = M_s / P_s \quad (4.4)$$

where V_i is the volume fraction of ice, M_w is the mass of water, M_s is the mass of soil solids, and P_s is the soil particle density.

4.3 Ground Ice and Bedrock

Volumetric ice contents associated with Paleozoic age Weatherall, Hecla Bay and Griper Bay Formation regoliths, and of Cenozoic sediments, are described below.

4.3.1 Weatherall Formation

All slopes except S are located on Weatherall Formation colluvium (see Figure 2.3). Three trenches together with smaller trenches in adjacent terrain give a reasonably complete record of volumetric ice contents and its distribution for this formation. In Weatherall-derived sediments, ground ice appears to be influenced by the distribution of colluvium, residuum, and the depth to unweathered bedrock.

For permafrost creep to occur, high ice contents and/or supersaturated (i.e. ice) sediments are critical. According to Mackay and Black (1973) sediments can become ice-saturated with a volumetric ice content which may be as low as 10 - 15% while supersaturated frozen sediments commonly have ice contents between 30 - 40% by volume. In this context, the data from the EBA trenches (Nos. 5, 6 and

7) are especially useful vis-à-vis Weatherall Formation. Analysis of the EBA data, using the procedures outlined above, indicate that the upper 2.0 m of material contains >40% ice by volume. Minimum values of 10 - 15% are obtained from bedrock while maximum values of >70% occur in the colluvium. The distribution of ice content with depth for Weatherall Formation is illustrated in Figure 4.1a.

Whereas Figure 4.1a shows consistently high ice contents in the colluvium, considerable variability of ice content over the length of each ditch occurs (Figure 4.1b); ice contents, for example, vary between 50 and 100%. The highest ice contents are found where trenches intersect ice-wedge ice. In the absence of wedge ice, pore and segregation ice average 50% by volume in the soil. An ice wedge observed by EBA personnel in the Weatherall Formation colluvium is illustrated in Figure 4.2a. This 'inactive' wedge did not display any surface expression and in profile provided no evidence of recent cracking. The upward bending of ice-rich colluvium at the edge of the wedge is clearly evident.

The siltstone strata of the Weatherall Formation is vertically oriented at the north end of trench 5. Alternating bands of grey and olive-coloured bedrock at the ditch bottom can be traced upwards through the residuum to the residuum-colluvium transition. At one point the beds are observed clearly to bend downslope (EBA Engineering

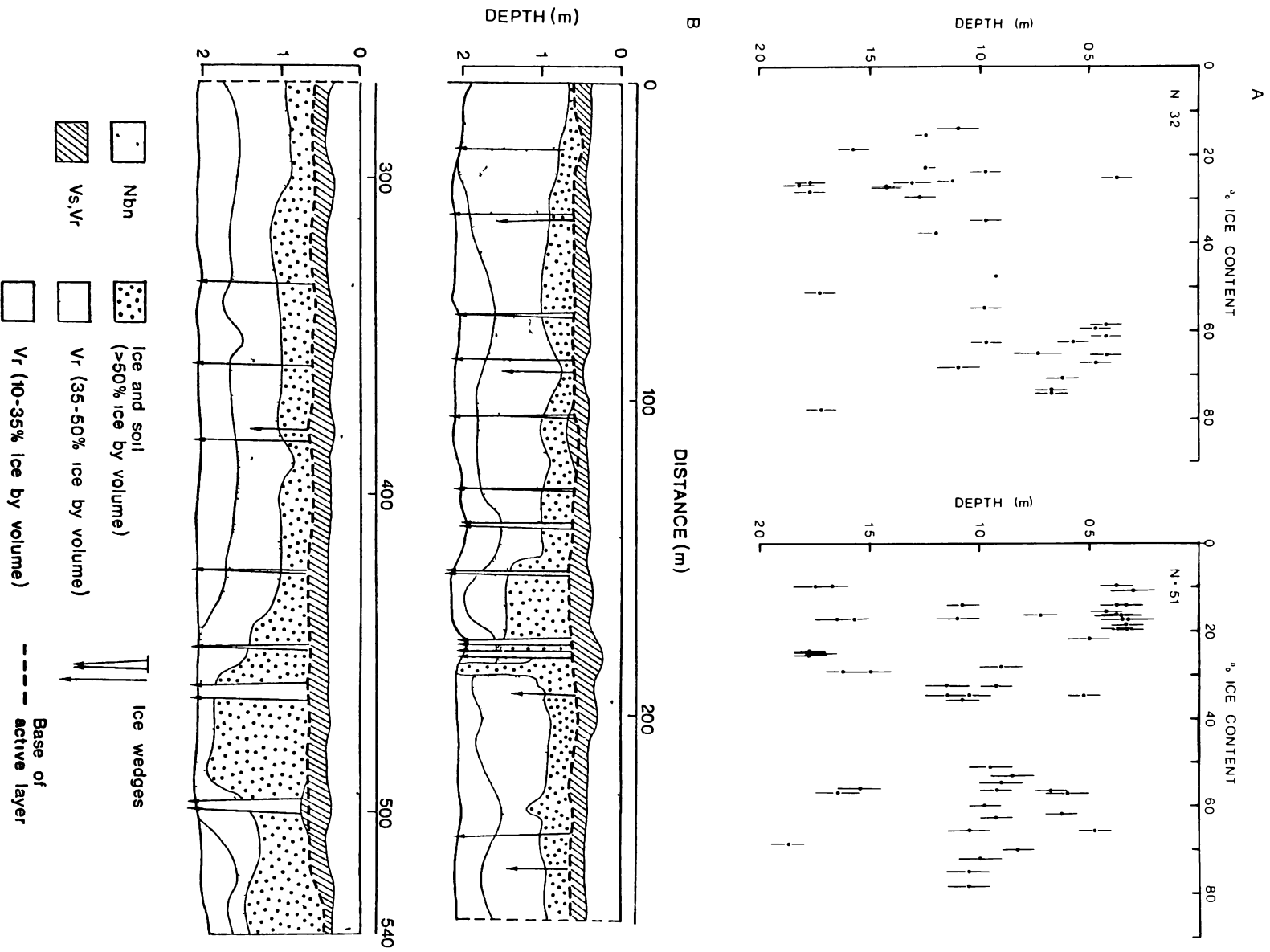


Figure 4.1 Ground ice conditions in Weatherall Formation silt and shale: A) Variation in the depth (EBA Trench Data 5 and 7); B) Variations along trench (EBA Trench 5).

a



b



Figure 4.2 Observations in EBA trenches excavated in Weatherall Formation sandstone, winter 1981-82: A) Large ice wedge in Trench 5. Note bending of colluvium at side of wedge. B) Bending of vertically-oriented Weatherall siltstone strata at colluvium/residuum contact in trench 5. (Photos courtesy of EBA Engineering.)

Consultants Ltd., 1983, p. 39; Figure 4.2b). Angular clasts within the ice-rich colluvium can be traced upslope to their respective declined ('bent') source beds. It is believed that the downslope bending of strata and trains of clasts in the colluvium observed by EBA provide direct evidence of the creep within permafrost which has now been measured (see Chapter 6).

A dominance of silty materials, which induce high negative pore pressures, probably leads to the concentration of ground ice within the colluvium (Figure 4.3). Water may be drawn from both the overlying thawed active layer and the underlying soil by primary and secondary heave processes (Miller, 1972; Smith, 1985a; Williams, 1982).

One to three random block samples were cut from the trench walls every 20 metres. In all, 96 samples were analysed for grain size, moisture content, frozen bulk density and unconfined compressive strength (EBA, 1983). Grain size analysis confirmed that the prime constituent of samples (e.g. silt, sand or gravel by weight) increased in size with depth although sand-sized sediments were not common. Silt is the predominant material in the upper 1.0 m of sediments with gravel becoming increasingly important with depth (Figure 4.4a). Figure 4.4b illustrates the relationship between the prime soil constituent of each sample and volumetric ice content. The average ice content of all samples, whether primarily silt, sand or gravel, is

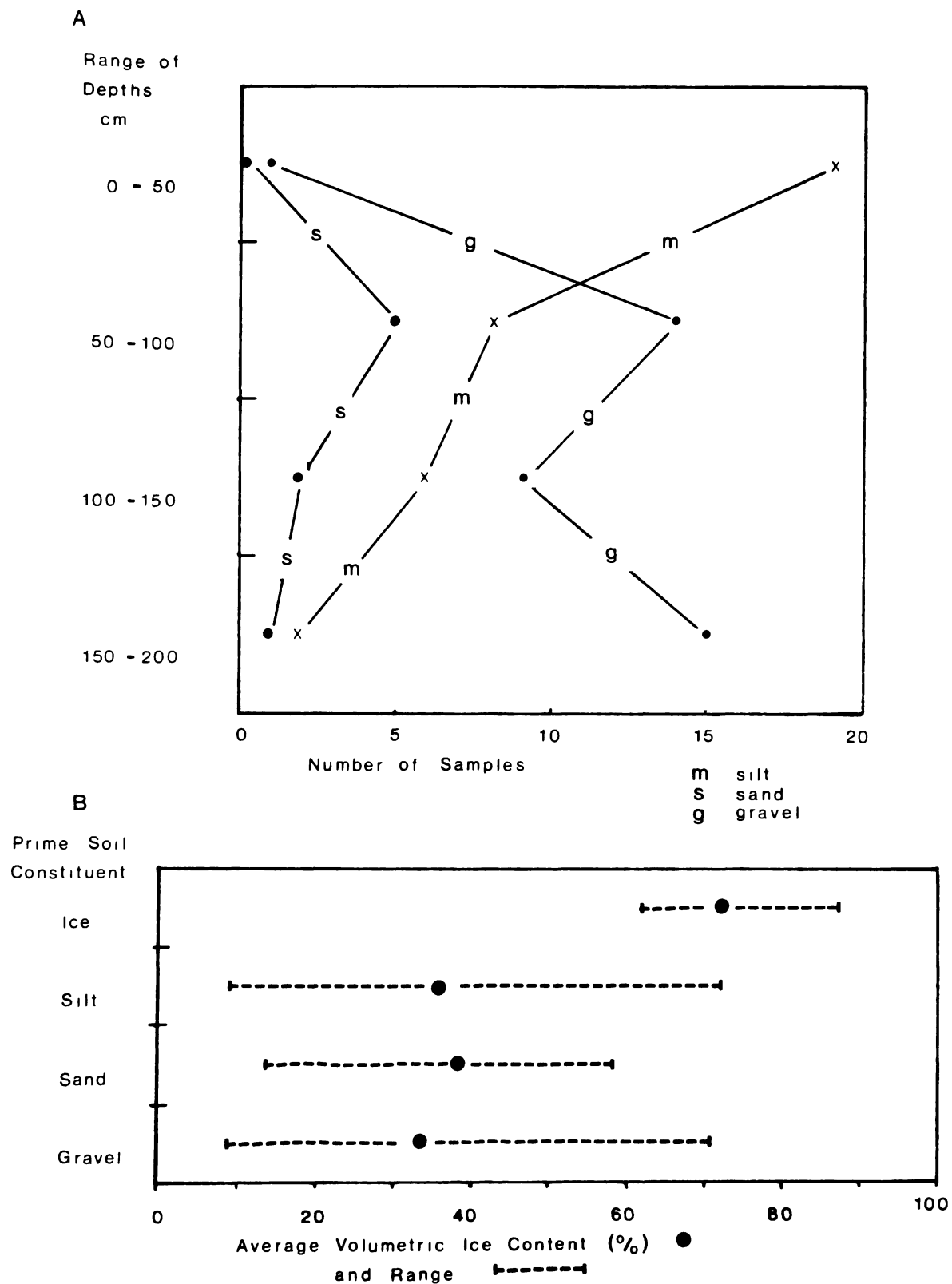


Figure 4.4 A) Changes in grain size with depth (EBA Trench 5, 6, 7); B) Ice content and grain size relationships (EBA Trench 5, 6, 7).

similar ranging between 34 and 37%. However, the average ice content values for silt are unusually low because 13 samples were collected from the ice-poor zone of the seasonally cryotic layer. Figure 4.5 shows that silt samples collected from the base of the seasonally cryotic layer and near-surface permafrost generally contain 55 to 65% ice by volume. The average ice content of silt samples is 46% if the desiccated samples from the seasonally cryotic zone are excluded. Boundaries between the colluvium-residuum and residuum-bedrock cannot be drawn on Figure 4.5 due to their considerable variability.

The residuum-colluvium boundary, ranging from 0.6 to 1.8 m in depth, is characterized by a concentration of excess ice. The cause of this is not obvious but it may reflect a change in soil type (grain size), wherein changing porosity or permeability may act as a barrier to free flow of water under negative pore pressures. In the residuum the concentration of ground ice decreases with depth from 50% by volume at the surface contact to 10% by volume at the bedrock contact. The residuum is typified by pore ice which fills joints and bedding planes in the weathered bedrock. Higher ice contents are found where the residuum contains segregated ice or ice-wedge ice.

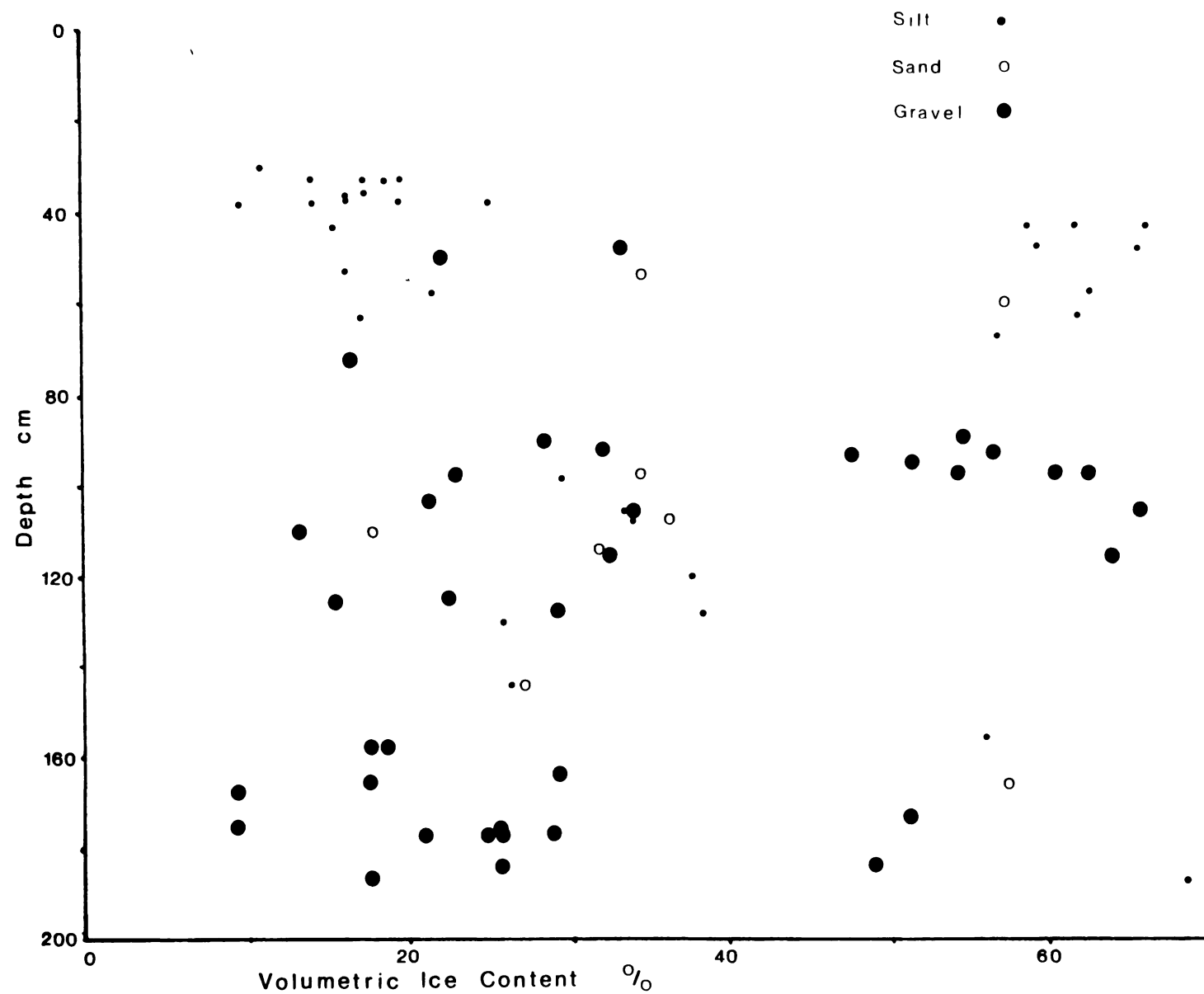


Figure 4.5 Following grain size analysis, samples were classed according to their primary soil constituent (silt, sand, gravel) and then plotted against ice content.

In summary, the Weatherall Formation is characterized in the near-surface layers by a silty, ice-rich colluvium. At greater depths, ice-rich soils are usually found in association with either ice wedges or soil type boundaries. Ground ice contents increase by 20% on average as one moves from ridge crest to colluvial slope (EBA, 1983). The increase may be related to the increasing thickness of the ice-rich colluvium.

4.3.2 Hecla Bay Formation

One short trench and others crossing the Weatherall-Hecla Bay and Griper Bay-Hecla Bay Formation boundaries provide volumetric ice contents and the pattern of vertical ice distribution for this formation. In general, the trench data indicate that the upper 2.0 m of the Hecla Bay Formation contains less than 20% ice by volume. Typically, the more resistant units where unweathered bedrock is encountered at depths as little as 0.4 m below the surface contain no visible ice. Block samples collected from the trench walls revealed a volumetric ice content of just 3.0% (EBA, 1983). The absence of silt and/or shale facies means that this well-sorted unit has limited capacity for the formation and preservation of ground ice in any bedrock-derived colluvium.

It follows, therefore, that slopes underlain by sandstone of the Hecla Bay Formation are not highly

susceptible to in situ permafrost creep in comparison to other lithologies of higher ice content.

4.3.3 Griper Bay Formation

Two trenches (Sites 3 and 4, Table 4.1) were excavated on slopes underlain by Griper Bay bedrock. Trench 3 was aligned perpendicular and Trench 4 aligned parallel to the regional strike on a low hill 8 km west of Rea Point. Sorted stone stripes and circles are typical of this terrain, and depths of seasonal thaw range from 0.6 to 1.0 m. Ice content shows a general increase with depth (Figure 4.6a). Although data are few below 1.0 m depth, the wall log (Figure 4.6b) indicates the presence of ice-rich soils to a depth of at least 2.0 m.

Of the 22 samples analysed from this location, the most ice-rich are those composed of a combination of gravel and silt. However, the wall log (Figure 4.6b) reveals that the highest ground ice contents are invariably associated with silty clay colluvium. These materials constitute the upper 0.5 to 1.0 m of the permafrost. This is in contrast to the Weatherall-derived sediments, where maximum ice contents occur at or in very close proximity to the permafrost table.

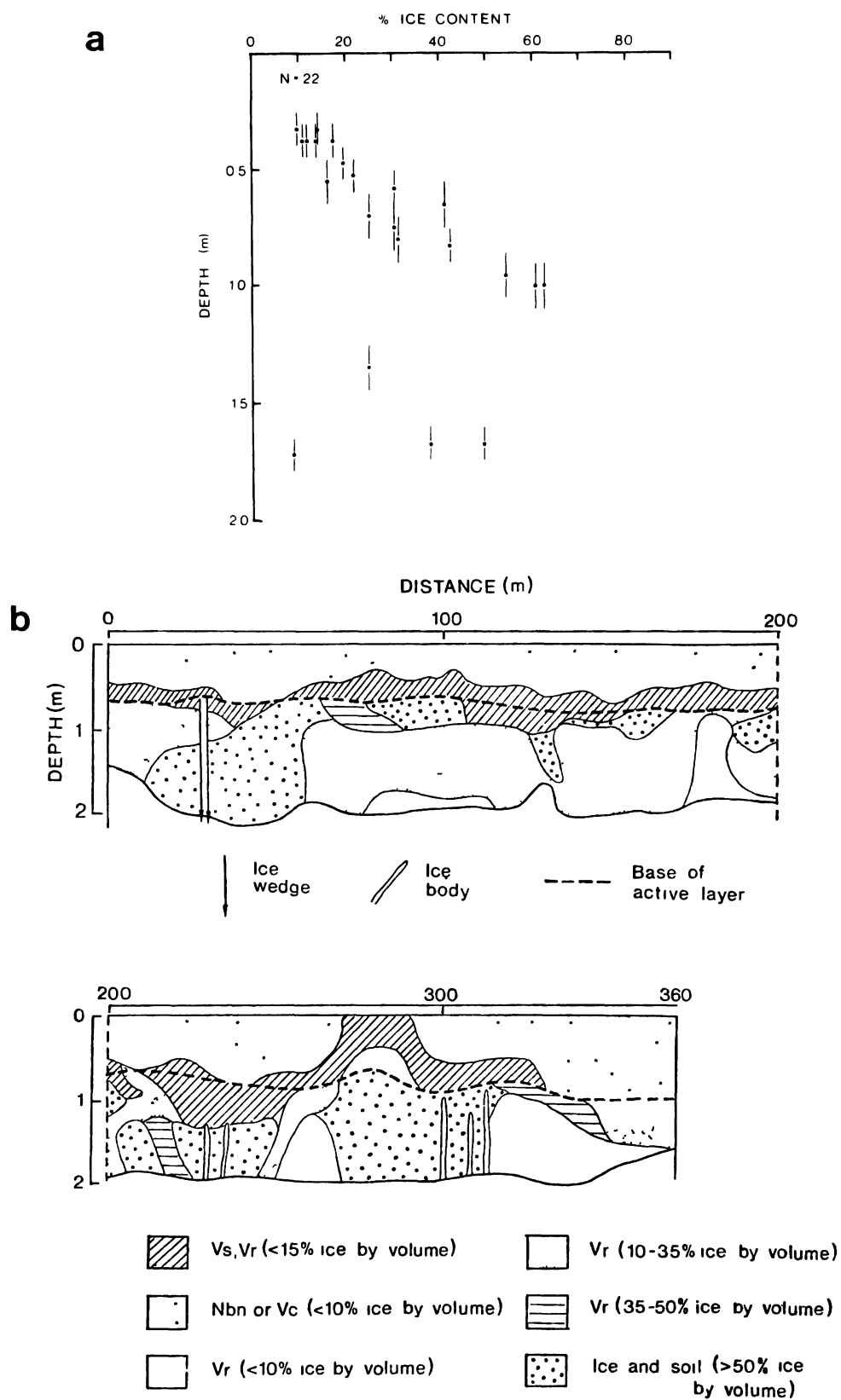


Figure 4.6 Ground ice conditions in Griper Bay Formation sandstone: A) Variations with depth (EBA Trench Data 3, 4); B) Variations along trench (EBA Trench Data 3, 4).

4.3.4 Cenozoic sediments

Two trenches were excavated in the Quaternary alluvium deposits near Rea Point (Sites 1 and 2, Table 4.1). The distribution of ice with depth shows no distinctive trend unless analysed according to surface conditions. Vegetated terrain, possessing ice-wedge polygons, has a higher (>40%) near-surface volumetric ice content than adjacent unvegetated sands. In the absence of ice wedges, however, ice contents in Quaternary alluvial and deltaic sands are extremely low (approximately 22% by volume). It follows that the potential for permafrost creep will be relatively low on these sediments.

4.4 Ground Ice and Frost Heave

One of the primary creep activators is thought to be secondary frost heave. The latter occurs in response to ice segregation following moisture migration through cryotic or partially frozen ground. It was for this reason that heavemeters were installed in each instrumented site.

Results from the four-tube heavemeters are shown in Table 4.2. The minimum surface heaves which were recorded, ranging from 11 to 25 mm, are similar to those measured by Mackay et al. (1979) and Smith (1985b). The greatest positive heave was measured in fall while the greatest negative heave was measured during the spring (see

Table 4.2

The Heavemeater Results 1985-1987 (Data in mm)

Site	Minimum Surface Heave	Maximum Heave vs. Depth			Active Layer Thickness (cm)
		15 cm	45 cm	55 cm	
W2	11	9	3	2	25
W3	25	4	3	2	40
W4	17	2	6	1	53
W5	19	6	10	8	29
E2-1	20	8	5	3	34
E2-2	16	10	3	0	29
S	17	2	1	1	35
N	20	0	3	3	30
E1-1	19	0	0	0	41
E1-2	21	5	6	1	28
<u>Summary</u>					
W	18	5.5	5.5	3.5	37
E/	18	9	4	2	32
S	17	2	1	1	35
N	20	0	3	3	30
F1	20	2.5	3	1	34
<u>Average</u>	18.6	3.8	3.3	2.1	33.6

Figure 4.7). In all probability, the positive frost heave was due to ice lense formation during freeze-up while the negative heave was due to the melt of those lenses.

Table 4.2 clearly indicates that frost heave is greatest near the surface and decreases with depth as observed by Mackay et al. (1979). While surface heave averages 18.6 mm, heave within the active layer (at 15 cm) averages about 4 mm. Frost heave within the permafrost (i.e. secondary heave) averages just over 2.0 mm during the fall and early winter. This is 40 to 67% less than that measured by Smith (1985b, p. 289). Furthermore, the data suggest that the greatest primary frost heave occurs on west and east-facing slopes, especially E2. In addition, limited secondary heave occurs on all slopes, particularly the west and north-facing slopes.

These results confirm that one of the primary permafrost creep activators occurs on all five study slopes. However, in comparison to similar data measured elsewhere, the magnitude of this activator is much reduced at Rea Point. This must be borne in mind when comparing and interpreting rates of permafrost creep from Melville Island with those from other localities such as the Mackenzie Valley.

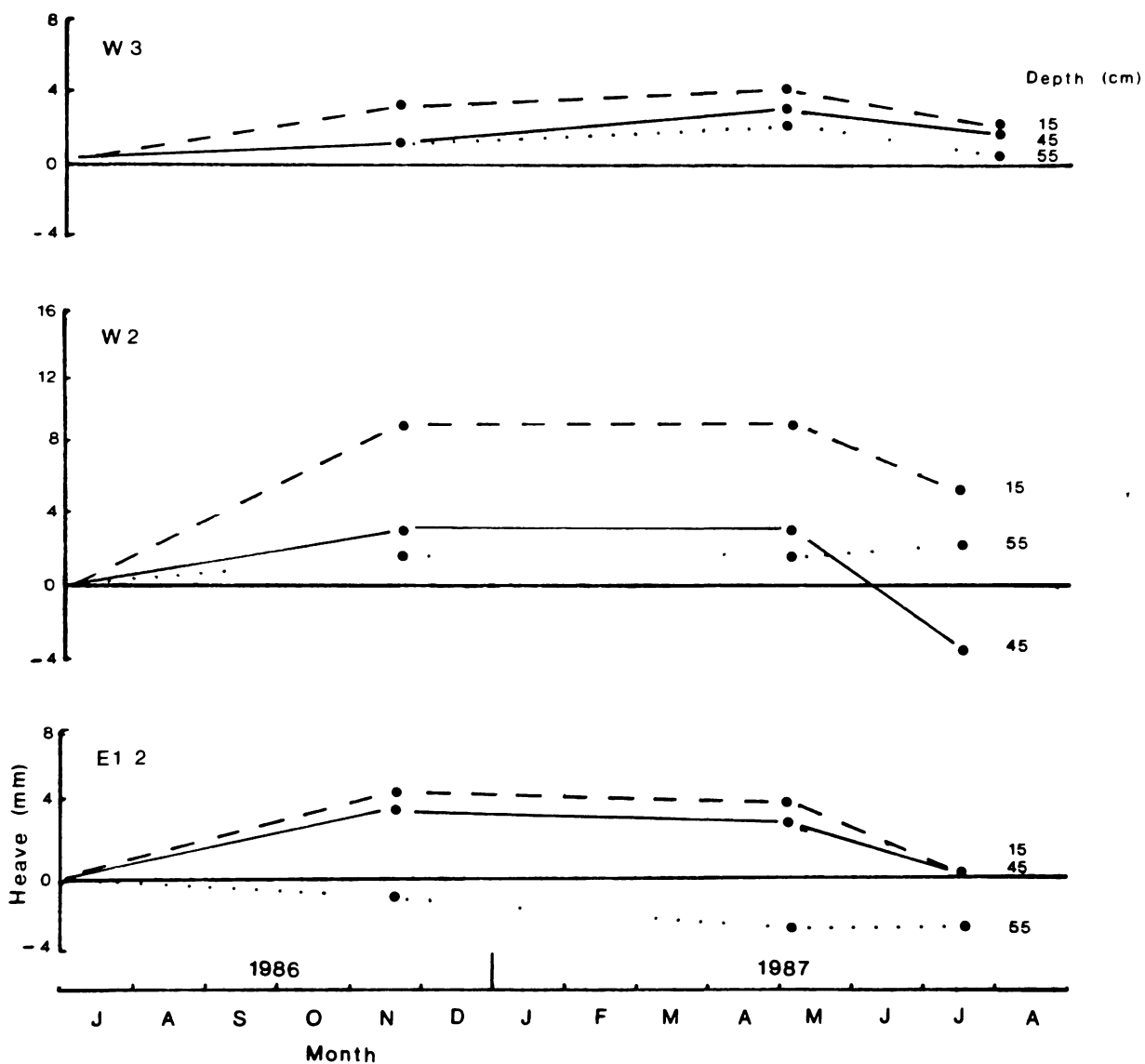


Figure 4.7 Heavemeter results over one year at 3 sites showing fall heave and spring subsidence.

4.5 Summary

In Eastern Melville Island, slopes underlain by the Weatherall and Griper Bay formations probably have the highest ice contents in the upper 1 - 2 m of permafrost, on account of the silty nature of their associated residuum and colluvium. Ice appears to be concentrated at the 0.5 to 1.0 m depth in the Weatherall Formation and at the 1.0 to 1.5 m depth in the Griper Bay Formation. In both units, maximum ice contents at these depths are between 50 and 70% by volume. Sandy and well-drained units, such as the Hecla Bay Formation and Quaternary age alluvial and deltaic deposits, possess extremely low ground ice contents. It should be expected, therefore, that rates of permafrost creep, as determined by ice content, should be greater on slopes underlain by Weatherall and Griper Bay formations than those on slopes underlain by Hecla Bay Formation and Quaternary sands.

The heavemeter data confirms that secondary frost heave occurs on all slopes but at magnitudes considerably less than those reported from the Mackenzie Delta. There is also considerable slope-to-slope variability. It should be further expected, therefore, that the component of in situ permafrost creep at Rea Point which is caused by frost heave will be less than in the Mackenzie Delta.

Collectively, these data may provide a degree of explanation for the relatively low and quite variable rates of deformation observed at Rea Point (see Chapter 6).

CHAPTER FIVE

GROUND THERMAL REGIMES

5.1 Introduction

The process of permafrost creep is inherently linked to ground temperatures (see Section 1.3.2). First, fluctuating ground temperatures can activate creep by initiating thermal expansion and contraction of the ground. Second, ambient temperature regulates the creep process by effectively controlling the amount of unfrozen water in 'frozen soils' (e.g. Smith and Riseborough, 1985, p. 18). It is ground temperature which determines whether ice is 'soft' or hard, ductile or brittle (e.g. Hobbs, 1974). Each of these factors affects the ease at which a frozen soil creeps.

In this chapter, thermal data are presented from each of the five study locations. The aim is to identify any variations in ground temperatures between the slopes since such variations may influence the rate of permafrost creep recorded.

5.2 Thermal Equilibrium

The thermistor cables were installed in May 1985. Since readings commenced in July and August of that year, it

was necessary to establish that thermal equilibrium had been re-established by that time. The data presented in Figure 5.1 indicate that this was indeed the case. Temperatures recorded in the two 4 m cables (see Table 3.1) were essentially the same in August 1985 and August 1987. Given the accuracy of the thermistors, and this evidence that thermal equilibrium was re-established following drilling, it follows that a high level of confidence can be attached to the ground temperature data.

5.3 Depth to Permafrost

Annual maximum and minimum ground temperatures to a depth of 2.0 - 2.5 m for the period 1985 - 1987 are presented in Figure 5.2. They indicate that the depth to permafrost is between 25 - 53 cm. These values compare well with the 20 - 60 cm quoted by Stangl et al. (1982) elsewhere on Eastern Melville Island. As demonstrated by Brown (1973), Linnell and Tedrow (1981) and others, these differences between sites can be related to variations in aspect, surface conditions and near-surface materials.

Determination of the depth to permafrost is especially relevant to the study of in situ permafrost creep, since only movements recorded at depths in excess of the depth of seasonally cryotic ground can unequivocally be interpreted as permafrost creep and not as solifluction.

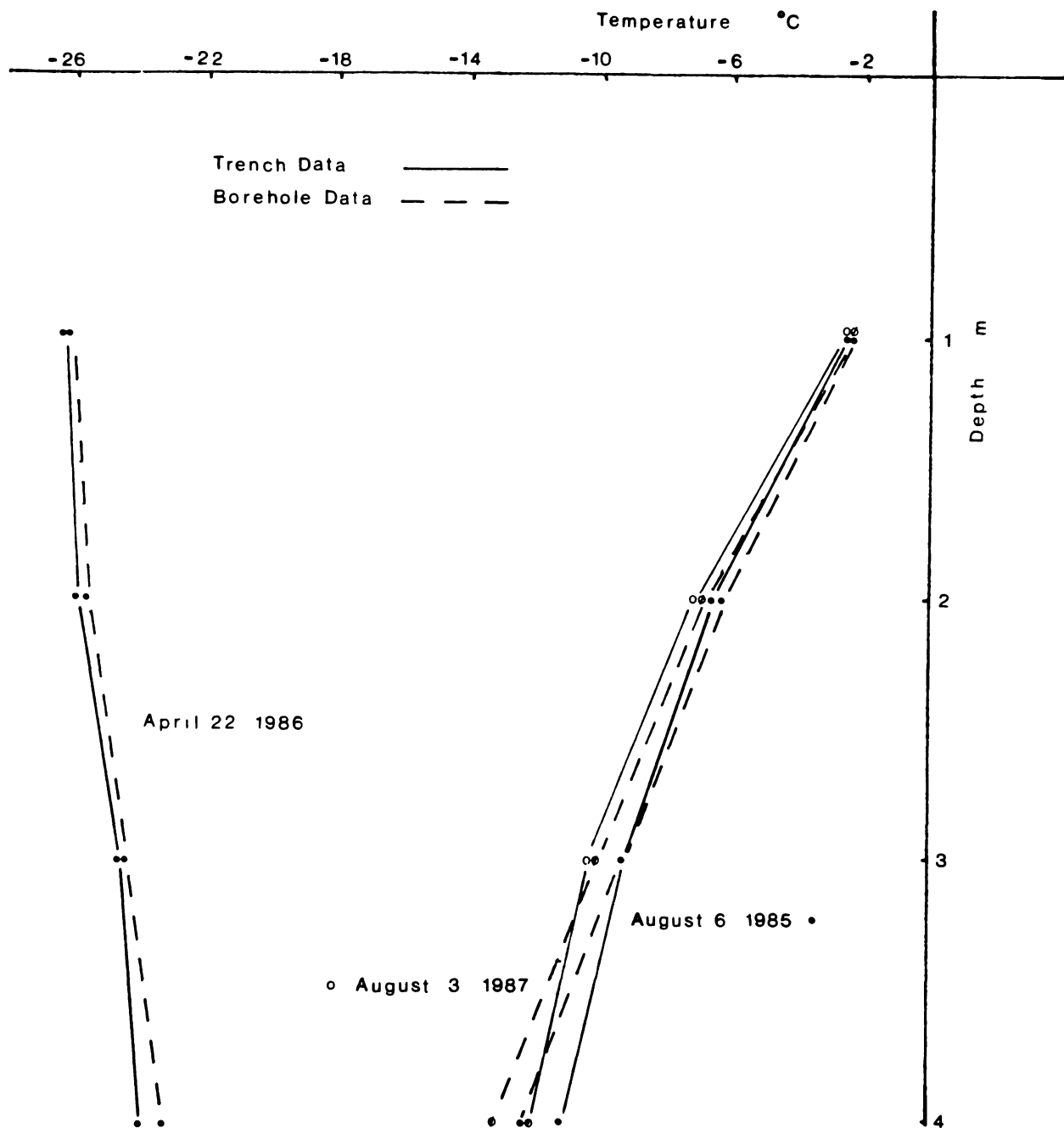


Figure 5.1 Diagram illustrating the re-establishment of thermal equilibrium at Slope W by August 1985, only 3 months subsequent to installation of thermistor cables.

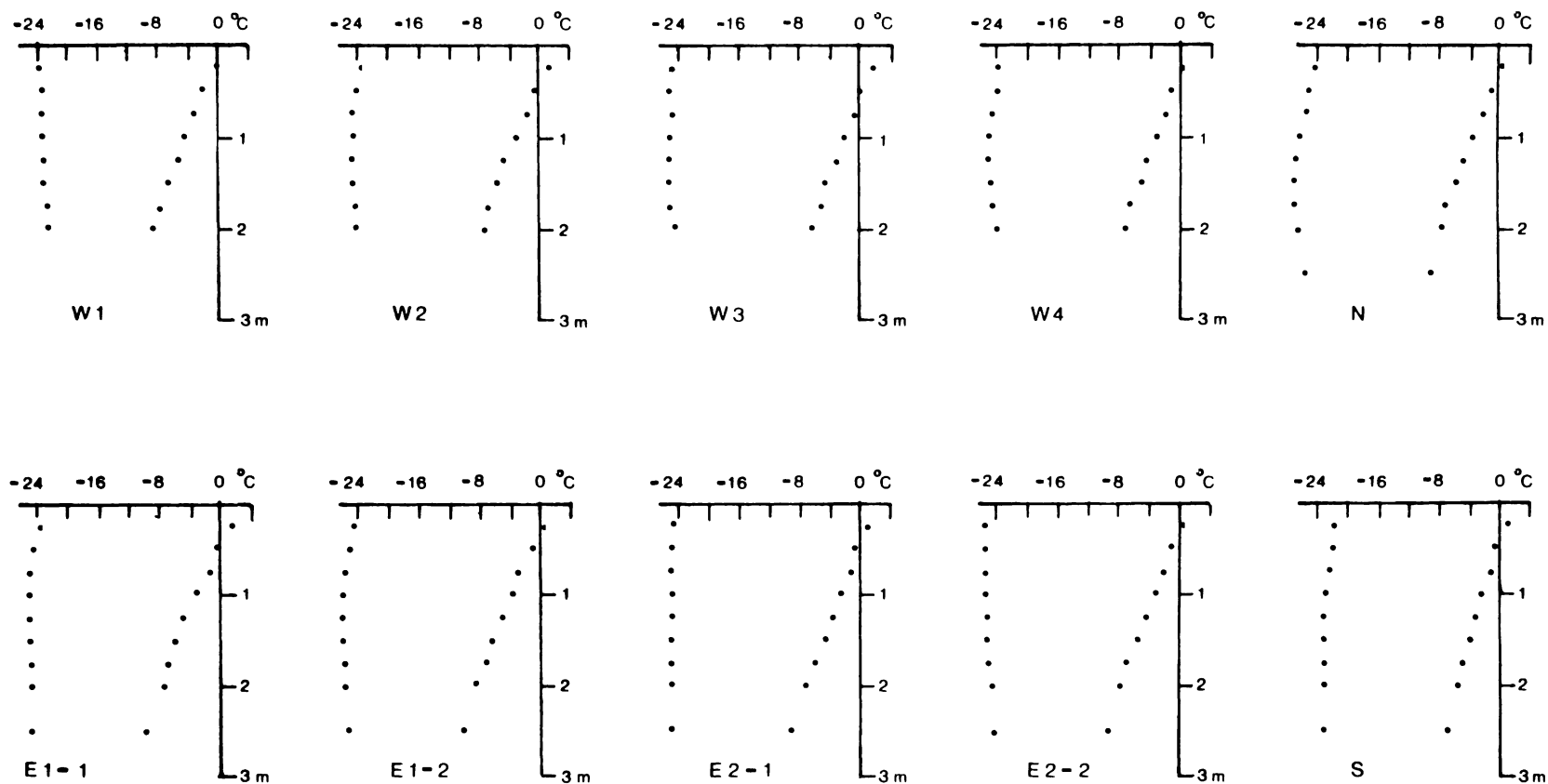


Figure 5.2 Depths to permafrost, as indicated by maximum ground temperatures, at all 10 sites, 1985-1987. Minimum ground temperatures are also indicated.

5.4 Factors Affecting the Ground Thermal Regime

Since the study sites are located in relatively close proximity to each other, factors affecting net radiation, such as latitude and season, cloud cover and atmospheric dust, are considered equal. Of particular concern, however, are differences between sites in the slope aspect, surface conditions and the nature of near-surface materials.

5.4.1 Slope aspect

At Rea Point, the coldest ground temperatures are generally recorded on the north-facing slope (Table 5.1). This is typical of many slopes in the Arctic and can be related to the amount and intensity of net radiation throughout the year (Linnell and Tedrow, 1981; Brown, 1973). In summer, for example, the near-surface permafrost on the north slope is 0.5 to 1.0°C colder than on other slopes. At 195 cm depth, the temperature difference between permafrost on the north and south-facing slopes exceeds 2.0°C. A similar temperature trend may be observed in winter.

A complication to this general trend is that the south-facing slope cools very rapidly in the fall while the west-facing slope remains comparatively warm. Clearly, this is not simply related to net radiation; rather, differences in surface conditions must be considered.

5.4.2 Surface conditions

Since vegetation cover is generally sparse throughout the study area, differences in surface condition are brought about primarily by differences in snow cover. Data presented earlier (see Table 2.1) show that prevailing winds are from the northwest. Thus, snow tends to accumulate on, and at the base of, east and south-facing slopes at Rea Point. Sites E1 and S are near the top of east and south-facing slopes where snow cover is variable. Site N, W2, W3 and W4 are directly exposed to the prevailing winds and are generally swept free of snow. Considerable snow accumulates in the valley between E2 and W1 and this snow covers E2-1, E2-2 and W1.

Near-surface permafrost measured at Site E1-2 was almost as cold as that measured on the north-facing slope. Unlike other east slope sites, E1-2 was consistently thinly snow covered throughout the winters of 1985-86 and 1986-87 (Figure 5.3). The limited amount of snow at this site allows considerable heat flow between the snow/ground interface (Goodrich, 1982). In the Schefferville area Desrochers and Granberg (1988) suggest that more than 10 cm of snow is required before it becomes an effective insulation; in the colder environments of the High Arctic, probably 20 - 30 cm of snow is necessary before significant thermal modification is induced. Although the area surrounding Site E1 is exposed, snow cover is highly

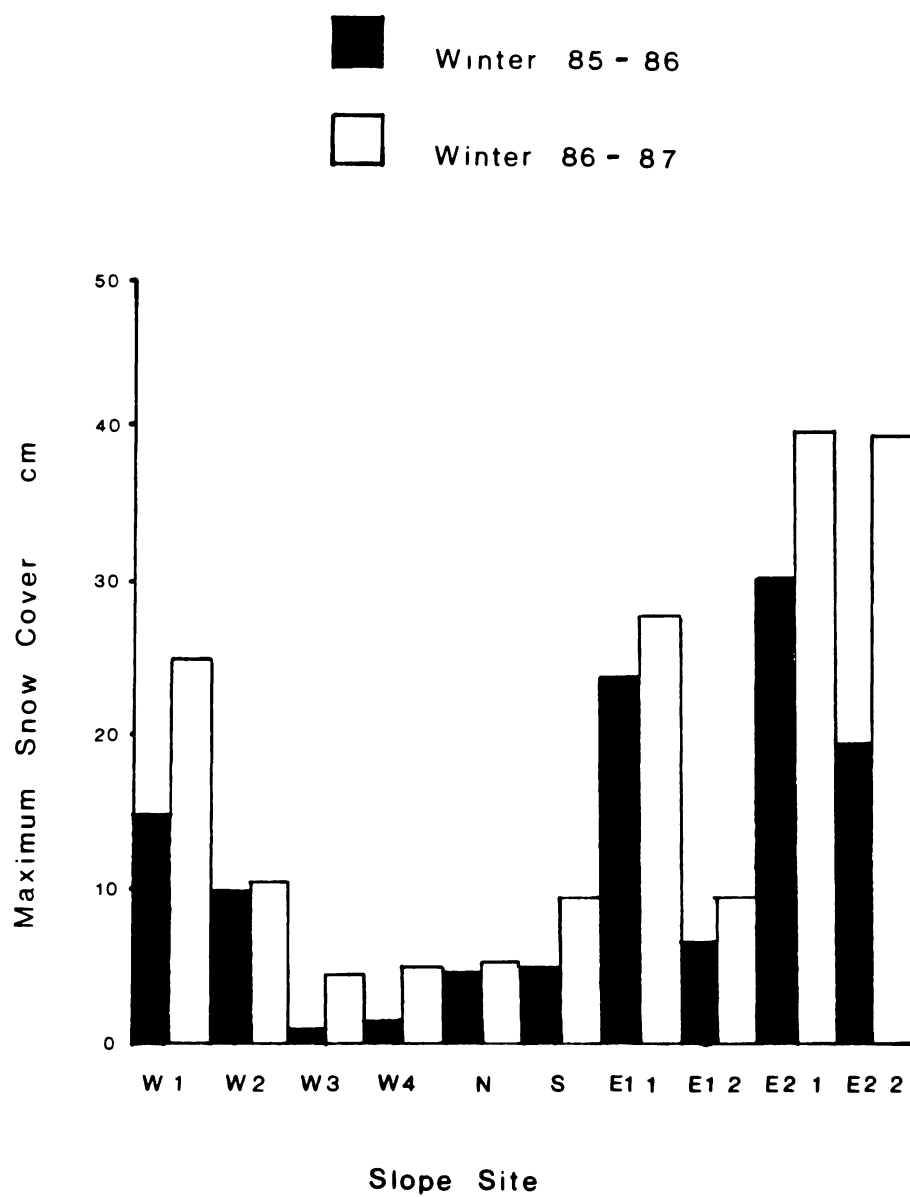


Figure 5.3 Maximum winter snowfall cover measured at all study sites.

variable. By late winter each year drifting snow patches, up to 40 cm thick, are randomly distributed over the upper slope surfaces and the ridge crest. In the spring of 1986, near-surface permafrost at E1-2 was 0.1 to 0.4°C colder than that measured at E1-1. At some depths permafrost temperatures were more than a degree cooler (Figure 5.2). Site E1-1 was covered by about 25 cm of snow. However, although more snow was measured at E1-1 in both 1985-86 and 1986-87, there is no reason to assume that this is always the case.

The time of year at which snow accumulates also affects the ground thermal regime (Goodrich, 1982). Twenty to 40 cm of snow were measured on Slope E2 each winter (see Figure 3.4; Table 2.1). In 1985, tracks made in the snow while working at the sites in late October and early November were still visible the following April when follow-up measurements were taken. It is believed that the fall/early winter snowfall insulates the near-surface permafrost keeping temperatures equal to, or warmer than, those of the south-facing slope which has up to 75% less snow cover. Near-surface permafrost in the south-facing slope cools rapidly in the fall (Table 5.1).

The amount of time required to melt snow in the spring affects the ground thermal regime. While Site W1 experiences the warmest winter temperatures (this site is covered by the deepest snow of all sites on the west-facing

Table 5.1

Average Ground Temperatures At Three Depths Through the Year

	60 cm	Depth 75 cm	195 cm
<u>August (Summer)</u>			
North	- 1.5	- 3.3	- 7.7
East	- 1.2	- 2.8	- 6.9
West	- 1.5	2.9	- 7.0
South	- 1.1	- 2.4	- 5.4
<u>April (Spring)</u>			
North	-25.6	-26.0	-26.1
East	-25.3	-25.3	-25.1
West	-24.2	-24.5	-24.2
South	-23.9	-24.5	-25.2
<u>October (Fall)</u>			
North	-17.0	-15.1	-11.9
East	-14.9	-13.2	-11.5
West	-13.9	-12.7	-11.4
South	-16.9	-15.5	-12.7

slope) it is thought that this site experiences the coolest summer temperatures due to the time and energy required to melt the surface snow. Such a condition has previously been described by Walker and Harris (1976) and Bakkehoi and Bandis (1988). The seasonally cryotic layer at W1 is only 25 cm thick. In contrast, Site W3 experiences the coldest winter temperatures and the warmest summer temperatures. This site is the most exposed on the west slope and is relatively free of snow in the winter. In spring, solar energy is not required for the melting of snow and instead is used to heat the soil. The seasonally cryotic layer at W3 is more than twice the thickness of that measured at W1.

5.4.3 Nature of near-surface materials

Surface snow conditions and aspect play important roles in determining the thickness of seasonal thaw and temperatures in the near-surface permafrost. However, where surface snow conditions and aspect are similar but ground temperatures are different, the nature of the near-surface soil materials must be considered (e.g. Linell and Tedrow, 1981). Of relevance are the thermal conductivity (K) and heat capacity (C) values of materials at and below the ground surface. Although it is beyond the scope of this thesis to determine values of K and C, general relationships may be identified.

K and C are governed by soil type, soil density (porosity), moisture content and/or pore filling gas, water and/or ice (e.g. Linell and Tedrow, 1981). K and C are not constant through time and will change as the contents of the pore space change.

While the surficial soils at sites W2, W3 and W4 are generally dry by mid-July, W1 remains moist. At W1, the thicker snow cover and extended period of melt likely contributes to the higher moisture content. This, in turn, increases the heat capacity of the soil reducing the distribution of heat.

At Site E2, the upper 40 cm of soil consists primarily of fine sands. These sediments tend to dry out relatively rapidly reducing heat capacity and increasing thermal diffusivity. Soil temperatures at E2 are warmer than those at E1, at least in part, due to this effect. E2-1, which is located only 5 m downslope of E2-2, experiences milder temperatures in summer (by as much as 0.7°C in the near-surface permafrost). However, the depth of seasonal thaw is similar at both sites.

In many areas, the Hecla Bay Formation is covered by colluvium composed of cobbles and felsenmeer with a limited sand matrix. The thermal conductivity of dry, coarse-grained porous materials is generally low although heat capacities are low as well. The Hecla Bay Formation is generally reported as having a shallow depth of seasonal

thaw (e.g. often a maximum of only 20 cm [EBA, 1983]). At Site S, the colluvium is composed of fine sand and gravel. The southerly aspect which provides exposure to relatively intense solar incidence accounts for the relatively thick depth of thaw at this site. The south-facing slope provides the warmest permafrost temperatures of all study sites.

5.5 Summary

Measureable permafrost creep is most likely to occur when ambient soil temperatures are relatively warm. At Rea Point, only the upper 1 m of permafrost exceeds -3.0°C , and then only for about 100 days per year. This contrasts with the Mackenzie Valley where ice-rich permafrost has a MAGT of -3.0°C . Thus, it follows that frozen soil creep rates will be much less than those reported from the Mackenzie Valley.

At Rea Point, the ground thermal regime is affected by a number of variables. These include aspect, differential snow cover, surface, soil and vegetation conditions, and the nature of near-surface materials. The warmest permafrost was measured at the south slope site; thus, with all other factors being equal, one should expect the greatest rate of permafrost creep to occur on this slope. However, the greatest amount of snow was measured on east-facing slopes. Temporal variability further

complicates the picture since an early snowfall reduces the permafrost cooling rate, enhancing the potential for permafrost creep. Moreover, a prolonged spring snowmelt delays warming of permafrost and inhibits creep.

Collectively, these factors must provide part of the explanation for the relatively low and variable rates of in situ permafrost creep which were measured at Rea Point. These data are now presented in the following chapter.

CHAPTER SIX

IN SITU FROZEN SOIL CREEP

6.1 Introduction

This chapter presents the results of the inclinometer measurements and relates their magnitude and variability to ground ice content, ground thermal regime and slope angle. Two preliminary papers, analysing data for slopes W and E1, are already available (Bennett, 1988; Bennett and French, 1988) and this chapter incorporates these data with data from other slopes, in order that a more typical or average pattern of permafrost creep can be recognized, as well as any site-to-site or slope-to-slope variability.

Traditional rates of solifluction movement (i.e. movement in the seasonally cryotic zone, or active layer) were also measured on all five slopes over the same period 1985 - 1987. These data are presented in order that a comparison can be made between these two related but quite distinct processes.

6.2 Magnitude of In Situ Permafrost Creep

Since the maximum thickness of the seasonally cryotic zone varies between 20 - 53 cm, any movement

observed at depths greater than 53 cm must unequivocally be related to permafrost creep. Figures 6.1 to 6.3 indicate that significant movement does occur in the upper 0.5 m of perennially cryotic ground (i.e. permafrost) lying immediately below the seasonally cryotic zones. In general, the amount of creep decreases from a maximum immediately beneath the permafrost table to virtually no deformation at 2.0 m within permafrost. The maximum deformation was recorded at Site W1 where movement of 0.22 cm at a depth of 65 cm was measured over the period August 1985 to August 1987. Typically, deformation was greater on the west and north-facing slopes.

At depths greater than approximately 1.0 m below the permafrost table, where ground temperatures rarely rise above -4.0°C and ice contents are low, no measureable net downslope creep occurs.

6.2.1 Average annual rates

On August 6, 1985, the position of the inclinometer casing within each slope was determined. All movements taken after this date were referenced to the casing when in its initial position. Annual rates of in situ permafrost creep were calculated by determining the deformation of the casing between August 6, 1985 and August 1, 1987 and dividing by two. The results are illustrated in Table 6.1.

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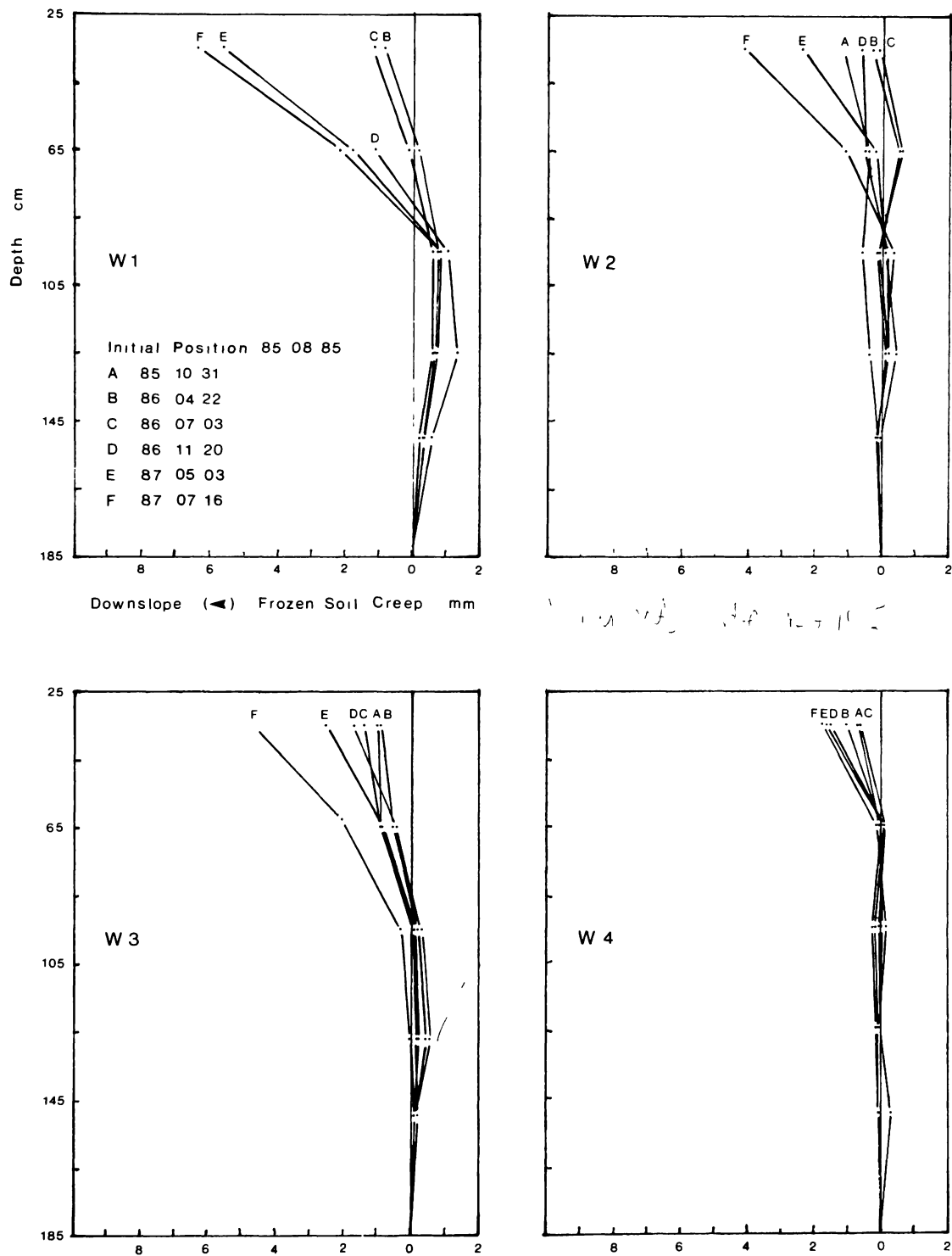


Figure 6.1 Diagram showing A-direction (parallel-to-slope) inclinometer data for Slope W.

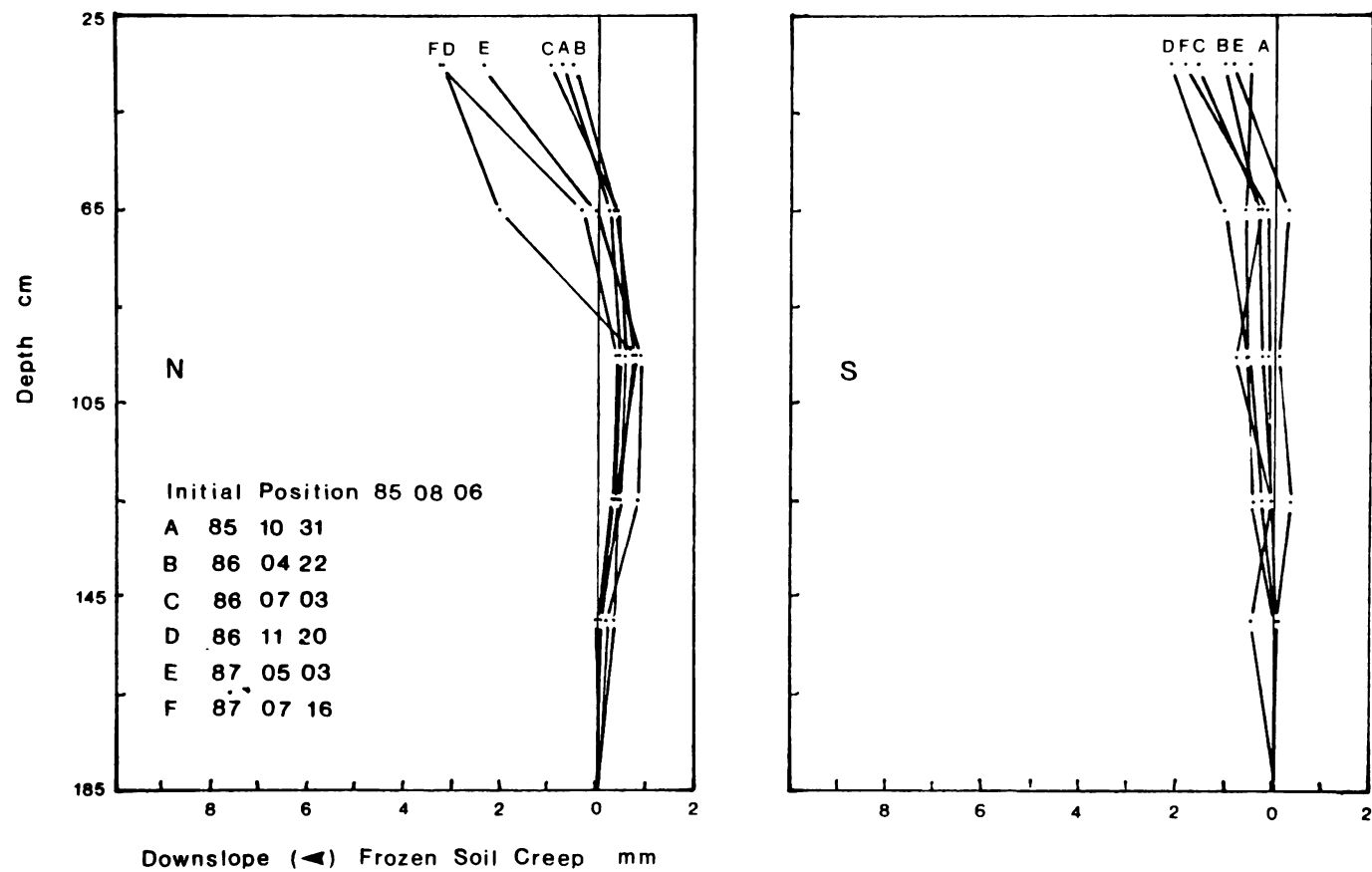


Figure 6.2 Diagram showing A-direction (parallel-to-slope) inclinometer data for south and north slopes.

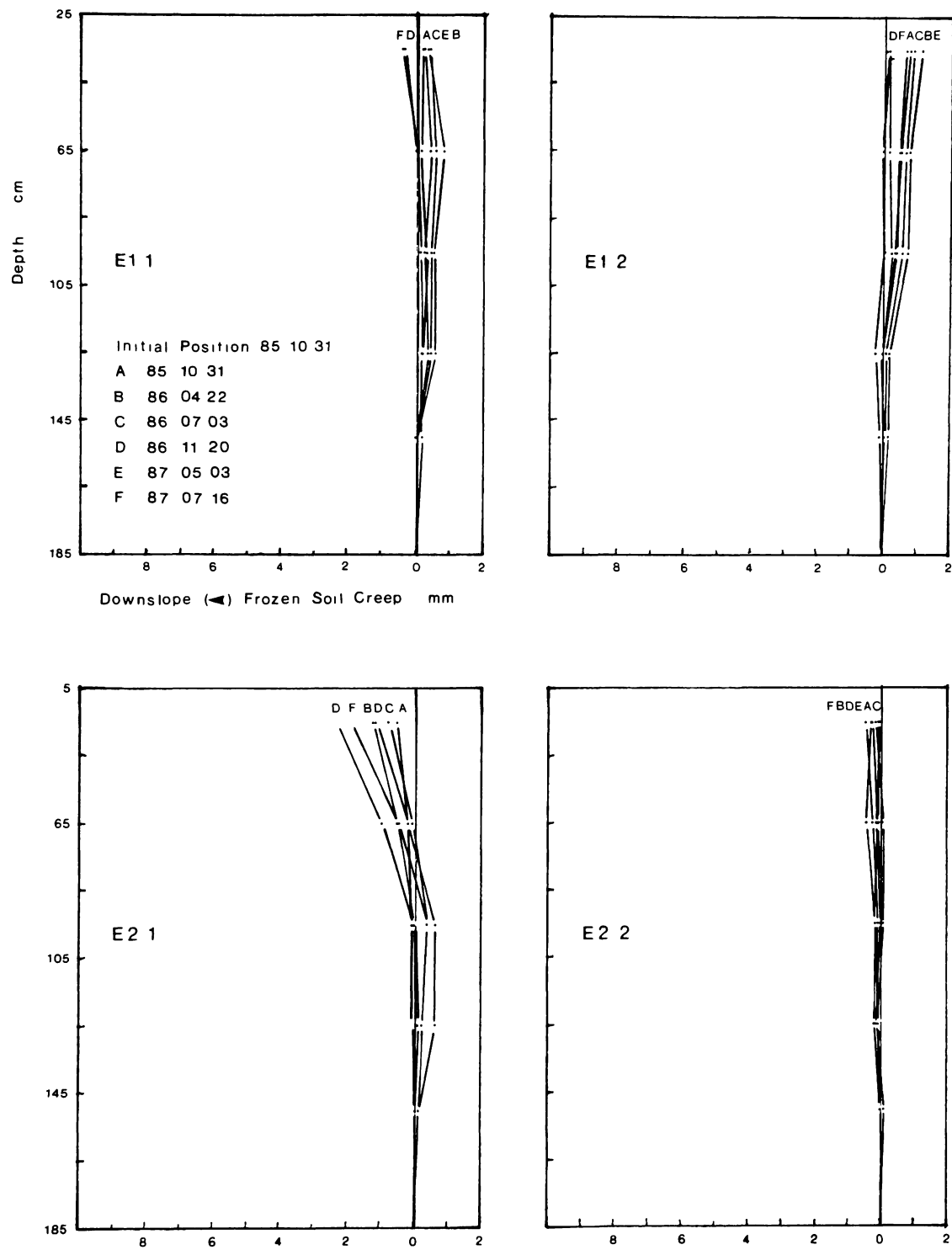


Figure 6.3 Diagram showing A-direction (parallel-to-slope) inclinometer data for east slopes (E1, E2).

Table 6.1

Average Annual Rates of In situ Frozen Soil Creep for the Period August 1985 - August 1987
Together With Information on Frozen Soil Creep Activators and Regulators

Study Site	Net Downslope Frozen Soil Creep (cm/yr)			Primary Frozen Soil Creep Regulators			Primary Frozen Soil Creep Activators			Secondary Frozen Soil Creep Regulators				Active Layer Depth	Depth to -5°C Iso-therm
				Ground Ice Content	Ambient Soil Temperature	Slope Angle	Secondary Frost Heave	Thermal Expansion & Contraction	Gravity Stress	Maximum Snow Cover	Vegetation	Aspect	Stratigraphy		
	35 cm	65 cm	95 cm	(%)	(65 cm)	(°)				(cm)			(USC)		
W1	0.30	0.10	0.00	40	-2.7	6.05	Yes	Yes	Yes	15-25	Mosses, Grasses	West	GM	25	121
W2	N/A	0.06	0.00	42	-1.3	6.14	Yes	Yes	Yes	10	Sparse	West	GM	40	157
W3	N/A	0.10	0.00	33	-0.8	5.08	Yes	Yes	Yes	0-5	Sparse	West	GM	53	171
W4	0.09	0.00	0.00	34	-1.6	5.95	Yes	Yes	Yes	0-5	Sparse	West	GM	29	144
N	0.16	0.10	0.00	53	-1.9	2.54	Yes	Yes	Yes	5	Sparse	North	GM	30	130
S	0.09	0.00	0.00	40	-1.4	4.37	Yes	Yes	Yes	5-10	Sparse	South	GP	35	183
E1-1	N/A	0.00	0.00	25	-1.0	2.08	Yes	Yes	Yes	25-30	Sparse	East	GM	41	145
E1-2	0.00	0.00	0.00	41	-1.9	2.75	Yes	Yes	Yes	5-10	Sparse	East	GM	28	135
E2-1	0.10	0.03	0.00	22	-0.9	5.65	Yes	Yes	Yes	30-40	Sparse	East	SP-SC	30	162
E2-2	0.03	0.00	0.00	27	-1.6	3.55	Yes	Yes	Yes	20-40	Sparse	East	SP-SM-SC	29	150

While specific depths are indicated in Table 6.1, the results, in fact, are an integration of all movement over the length of the inclinometer torpedo. The torpedo is 60 cm in length. Thus, data given for creep at 35 cm in fact reflect all movement between 35 and 95 cm depth. Similarly, data given for creep at 65 cm reflect movement between 65 and 125 cm depth. No data are given for sites W2, W3 and E1 at 35 cm because the depth of seasonal thaw exceeds that depth.

Annual maximum rates of in situ permafrost creep ranged from 0.3 cm yr^{-1} at 35 cm depth to 0.1 cm yr^{-1} at 65 cm depth. Average annual creep rates range from 0.11 cm yr^{-1} at 35 cm depth to 0.04 cm yr^{-1} at 65 cm depth. In general, permafrost creep rates at Rea Point are approximately one-third less than those reported by Savigny and Morgenstern (1986a) in the relatively warm permafrost of the Mackenzie Valley.

6.2.2 Seasonal variability

The typical pattern of in situ permafrost creep consists of a recognizable downslope component during summer and early autumn, followed by a weaker upslope component during winter and early spring. Results from three sites are plotted in Figure 6.4a. Data from slope Site E2-1 best exemplifies the pattern of movement. Such movement is also apparent at Site W2 and W3 although the former experienced a

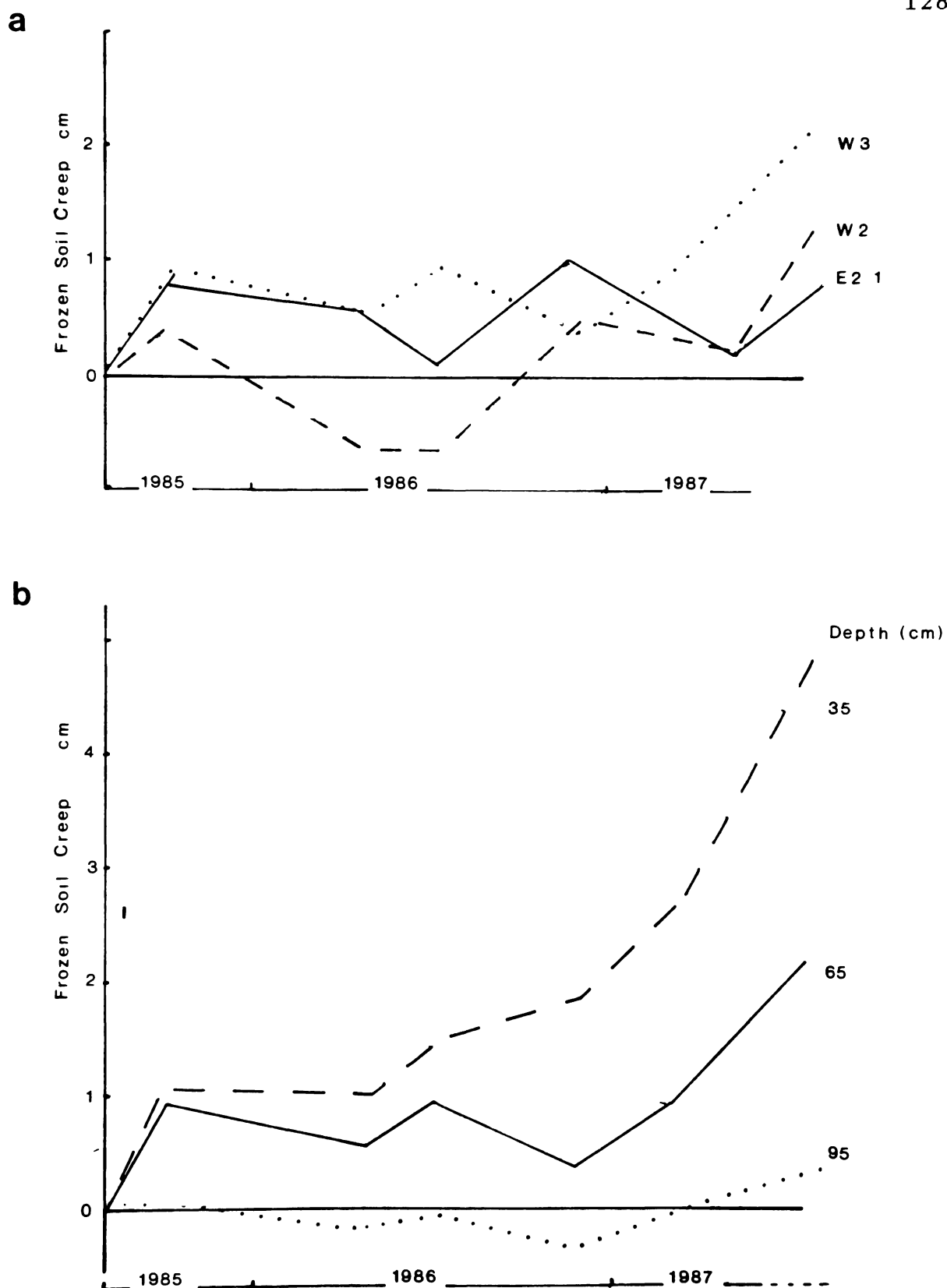


Figure 6.4 Seasonal variability of in situ frozen soil creep at: A) 65 cm depth for the Sites W3, W2 and E2-1; B) At site W3 for depths 35, 65 and 95 cm.

strong upslope component during the first winter and the latter underwent strong downslope movement during the spring of 1987.

At certain sites (e.g. E1-1, E1-2) and at depths of 65 cm or greater (e.g. E2-2, S, W4) downslope creep is equal to upslope creep. At Sites W1, W3 and N1, at 65 cm depth where greatest downslope creep was measured, upslope creep averages 33% of downslope creep (total upslope creep divided by total downslope creep x 100). This value varies a few percentage points either way as a function of the time at which creep measurements were made.

Seasonal variability of in situ permafrost creep with depth is illustrated in Figure 6.4b. The results illustrate that, in general, the trend (or direction of movement) remains the same although the magnitude of the process decreases with depth. However, this is not always the case. It was not unusual to measure upslope creep at one depth while measuring downslope creep immediately below.

6.2.3 Site variability

The magnitude of in situ permafrost creep was found to vary considerably from slope to slope, and even between sites on the same slope. This makes it essential that any interpretation of average annual rates be made with caution. Figure 6.5 shows the variability of creep at depths of 65 and 35 cm for all sites.

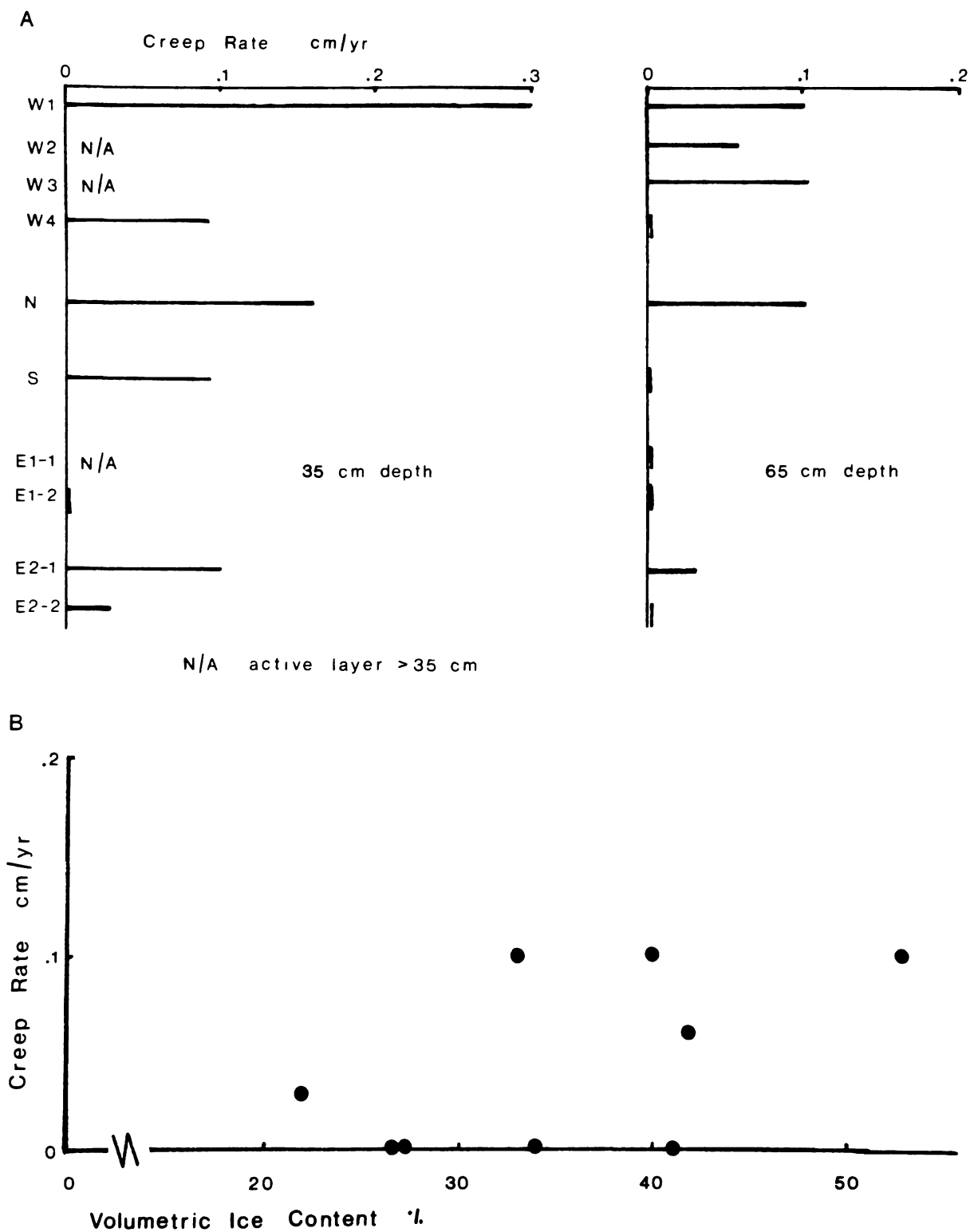


Figure 6.5 Average creep rates for all sites plotted against: A) slope aspect; B) ice content.

The greatest movement was measured on Slopes W and N. At 65 cm depth, that movement was three times more than the creep recorded on slope E2. No movement was measured on Slopes S and E1 (65 cm depth).

On Slope W, only three of four sites experienced movement yet all were located within 150 m of each other and all slopes were approximately 6° in angle. Likewise, on Slope E2, creep rates were three times greater at E2-1 than at E2-2, yet both sites were located within 10 m of each other.

This variability appears difficult to interpret. Since all but one of the slopes were located upon Weatherall Formation siltstones, the regolith does not appear to be the controlling factor. Instead, variations in ground ice conditions, ground thermal regime and slope angle appear more relevant. These are discussed below.

6.3 Factors Affecting In Situ Frozen Soil Creep

6.3.1 Ground ice content

As demonstrated in Chapter 4, the permafrost at Rea Point varies in composition from frozen rock (e.g. competent Hecla Bay Formation) to pure ice (e.g. lenses within the ice-rich colluvium). The latter are found within the near-surface permafrost of all slopes. Abundant horizontal lenses up to 5 mm or more in thickness were found

within the colluvium of Slopes N, W and E1. Thin horizontal lenses were observed within the finer soils on Slope E2. On Slope S (Hecla Bay Formation), ice lenses are randomly distributed within the gravel although close to the base of the seasonally thawed zone horizontal lenses become more common.

It has been shown (see page 17) that the cohesion and internal friction within a frozen soil decrease exponentially with increasing ice content. Thus, with all other variables remaining the same, a higher creep rate should be expected for soils with higher ice contents. Figure 6.5b illustrates the relationship between ice content and creep rate for all sites located upon Weatherall Formation sediments. A positive correlation exists between ice content and frozen soil creep rate (i.e. as ice content increases, creep rate increases). However, this relationship is not a strong one since the linear correlation coefficient (r) is only +.5442 (r for an exponential relationship is +.5146).

6.3.2 Ground thermal regime

At Rea Point, measurable permafrost creep was experienced only in the upper metre of permafrost. At these depths, ambient soil temperatures fluctuate by over 20°C during the year (see Figure 5.1). The following discussion focusses, therefore, upon permafrost creep as it is affected

by time of year, snow cover and aspect.

The greatest downslope movement occurs in summer and autumn (Figure 6.6a). One may speculate that this is because permafrost temperatures are warmest at this time. It is well known that as ice approaches pressure melting point, it becomes 'soft' or ductile (e.g. Hobbs, 1974, p. 236). As a consequence, the creep strength of the material is reduced. For example, Huang et al. (1986), while measuring 'continuous' creep within the CRREL permafrost tunnel at Fox, Alaska, found that an exponential relationship exists between increasing temperature and increasing creep, probably reflecting the increase in unfrozen water content.

The east slope best reflects the exponential relationship between soil temperature and permafrost creep (though data are limited and must be viewed with caution) (Figure 6.6b). At 65 cm depth, no creep is discernible until soil temperatures exceed -1.0°C . A similar exponential relationship may be shown with the west slope though the curve for data at both 35 and 65 cm depth is shifted to the right. This indicates that creep will occur on Slope W at cooler temperatures suggesting that other factors (e.g. higher average ice content and/or increased slope angle) may have a stronger influence. This suggestion is supported by the fact that Site W1 appears inconsistent with the other data portrayed in Figure 6.6b.

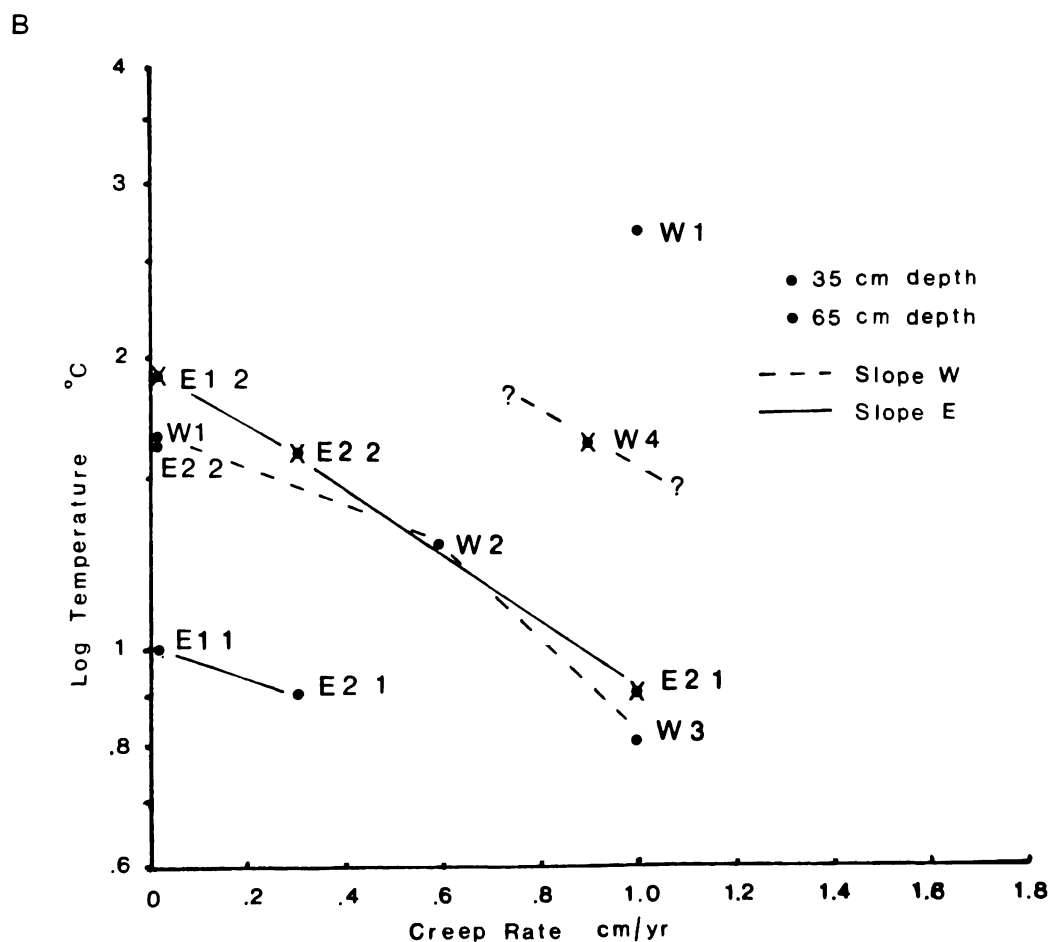
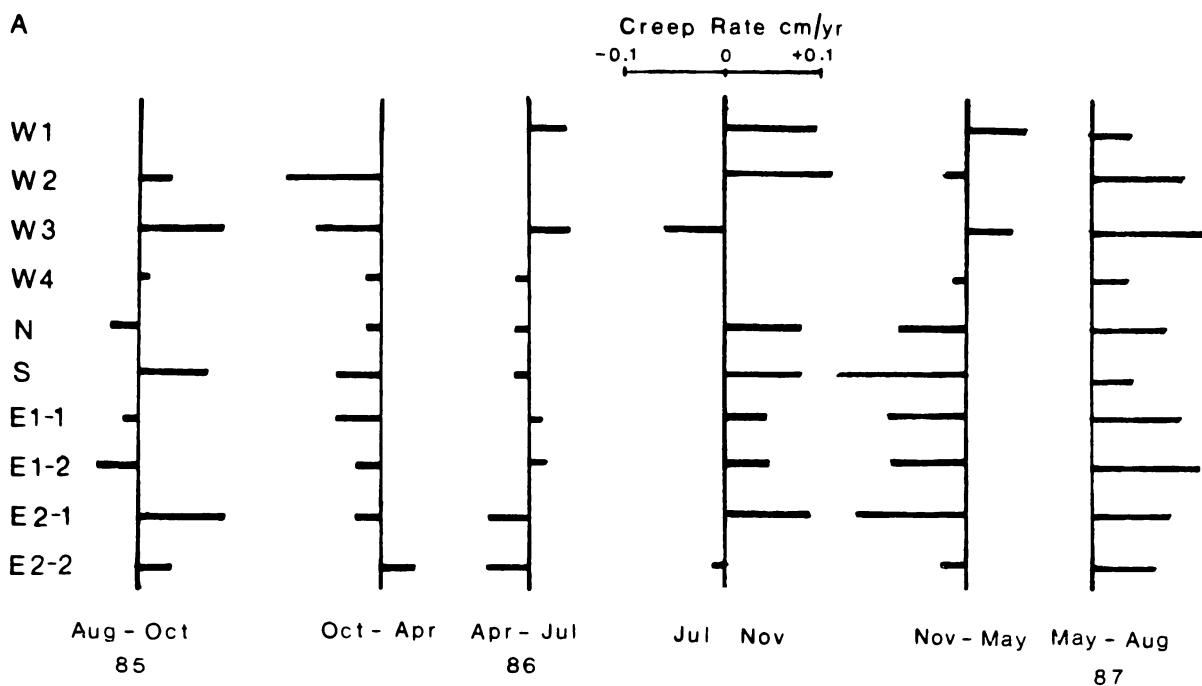


Figure 6.6 Variability of creep rates: A) According to time of year; B) Against temperature.

Site W1, although experiencing the coolest summer temperatures, also exhibits the highest permafrost creep rate. It is believed that Site W1 is influenced by deeper snow cover during winter which acts to prolong the time at which the permafrost is at a temperature suitable for creep.

The roles of snow and vegetation cover in modifying ground temperatures are well known (Goodrich, 1982; Mackay and Mackay, 1974). Site W1 is consistently snow covered during the winter and saturated during snowmelt and into early summer (Figure 6.7a). As a consequence, when compared with the other sites, ground temperatures are warmer by approximately 3 - 4°C in the autumn and early winter, but due to latent heat of evaporation, or high heat capacity, cooler (approximately 1 - 2°C) in early summer. The comparatively warm ice-rich permafrost soils at this site may explain, therefore, why this site experiences the highest amounts of permafrost creep. By contrast, Site W4, located only 8 m away, is covered by relatively little snow during the winter months (Figure 6.7b) and becomes dry and well drained shortly after spring melt. Very little downslope permafrost creep was recorded at this site.

At all sites, creep decreased exponentially with depth. Permafrost creep was not recorded at depths exceeding 100 cm. The data presented in this thesis suggest that the control over this pattern, however, is not internal friction (as cited in creep studies in non-permafrost

a**b**

Figure 6.7 Variability of snow cover from site to site at W1 and W4, April 1986: A) Site W1; B) Site W4.

regions: Carson and Kirkby, 1972), but rather, temperature. Measureable permafrost creep was limited to ice-rich soils at temperatures generally exceeding -3.0°C . This condition is met only by soils in the top 50 to 75 cm of permafrost. Moreover, it has already been demonstrated that the sediments within this zone attain temperatures $\geq -3.0^{\circ}\text{C}$ for progressively shorter periods of time with increasing depth (see Figure 2.10).

The third factor to be considered is slope aspect. It is generally accepted (e.g. Linell and Tedrow, 1981) that south-facing slopes in the Northern Hemisphere receive a greater incidence of relatively intense radiation and, thus, are usually warmer. This is even the case in high latitudes although other factors are increasingly important. Intuitively, one might anticipate greater rates of permafrost creep on the warmer slopes. However, even though the south-facing slopes are the warmest (see Chapter 5, Section 5.3.1), the greatest creep at Rea Point occurs on west and north-facing slopes.

6.3.3 Slope angle

There is usually a positive relationship between soil creep and slope angle, with gravity being the connecting factor. This should be true for both permafrost and non-permafrost situations.

Figure 6.8 shows the relationship between slope angle and frozen soil creep for all sites except N. The

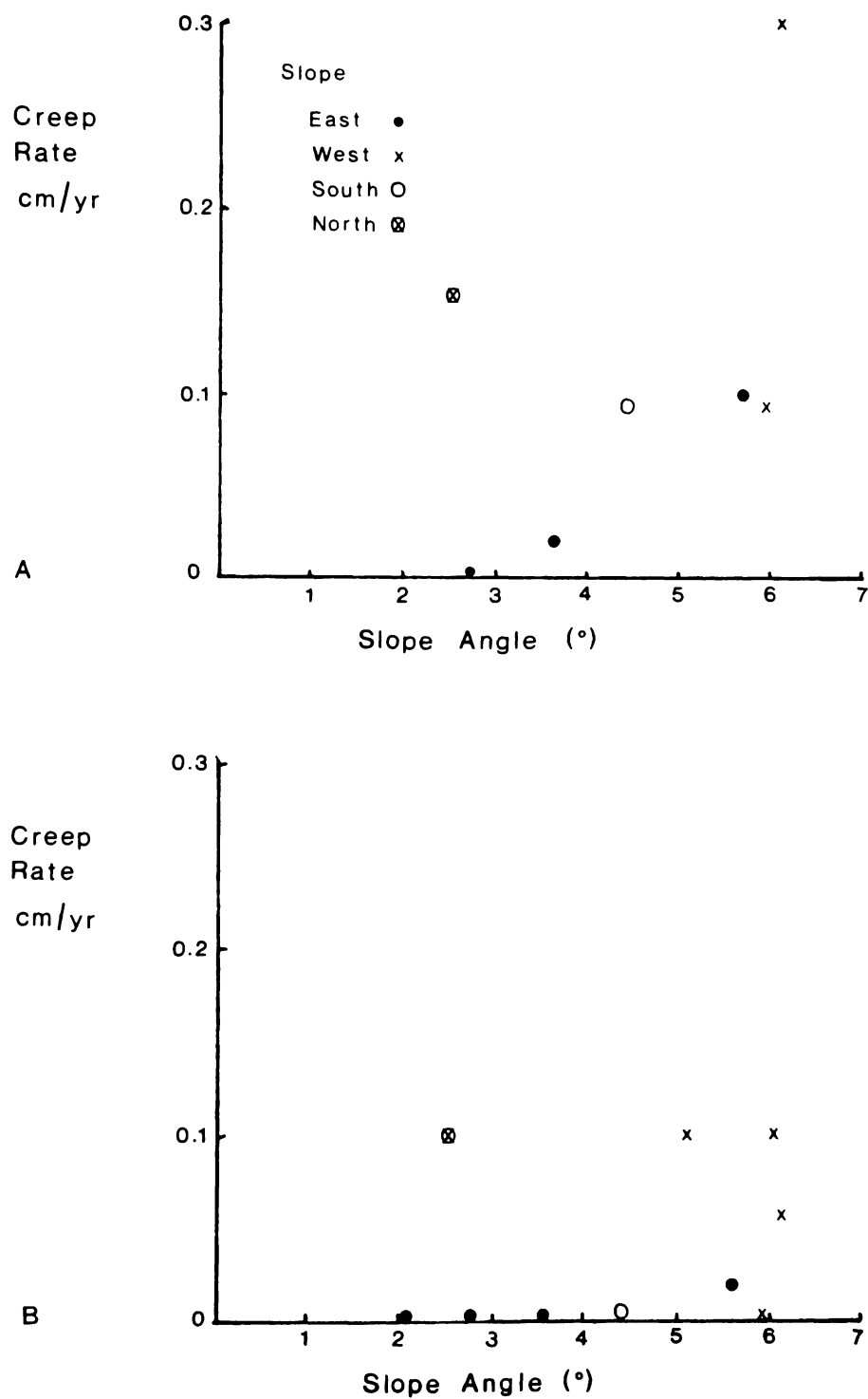


Figure 6.8 A plot of creep rate versus slope angle for all study sites:
A) 35 cm depth; B) 65 cm depth.

positive correlation indicates that frozen soil creep increases with increasing slope angle. The relationship holds at both 35 and 65 cm depth. This may explain the high rates of creep measured on the west-facing slope. However, the reason why Site N experiences comparable creep while located on a moderate slope is unclear. One possibility is the high ice content of this slope (see Table 6.1).

The north and west-facing slopes are both located on the convex or steep side of an asymmetric valley (see Figure 2.16). Slope N is located 13 m above the ephemeral stream channel while Slope W is 45 to 50 m above the same stream. It may be that the oversteepened slopes in question increase creep or shear stress. It can be argued that this is especially relevant for the west slope which is in closer proximity to the valley bottom (see Figure 2.16).

6.4 Movements in Seasonally Cryotic Ground

Marked pebbles and buried objects were used to establish rates of movement in the seasonally thawed zone. Such movements constitute gelifluction and/or frost creep and are typically referred to in the literature as 'solifluction'. Because the pebbles and buried objects were placed across the slopes, it was possible to identify variability of movement across the slope.

Pebbles were measured in both 1986 and 1987 on Slopes S and N while measured only in 1987 at all other slopes. The results are illustrated in Table 6.2. A few pebbles from most lines were lost and a few were found several centimetres downslope. These spurious results, likely due to meltwater flow or to the intentional or accidental actions of local wildlife, were not used in the calculations shown in Table 6.2.

At Rea Point, greatest surface movement was measured on the east-facing slopes. This contrasts with the low values for frozen soil creep at the same locality. While some permafrost creep was measured at E2 (0.03 cm yr^{-1}), no creep was measured at E1. Also, while significant permafrost creep was measured on the west-facing slope, only limited surface movement was recorded. These data show clearly that the surficial movement (i.e. gelifluction and frost creep) is unrelated to deformation in perennially cryotic ground.

The objects placed within the seasonally cryotic layer more closely reflect creep within the underlying permafrost (Table 6.2). For example, two pits excavated at E1 showed no measureable movement since installation two years earlier. Also, three pits within the north and west-facing slopes revealed comparatively high movement. However, the most active movement was located on Slope E2:

Table 6.2

Movement of (A) Surface Pebbles and (B) Buried Objects Within the Seasonally Cryotic Layer (i.e., Active Layer) at the 5 Slopes Where Permafrost Creep Was Measured

(A)

Site	Date Installed	Measured	Line Length	Number of Pebbles Installed	Measured	Pebble Spacing	Movement Total (All) Distance	Movement /Pebble Average Distance	Movement /Year Average Distance	Permafrost Table
			(m)			(cm)	(cm)	(cm)	(cm)	(cm)
E2	05/08/85	23/07/87	25	110	108	20-25	512.6	4.7	2.4	32
E1	05/08/85	23/07/87	30	141	133	20-25	570.8	4.1	2.1	35
S	02/07/84	02/07/87	50	395	385	10-15	1981.1	5.1	1.7	35
N	01/07/84	23/07/87	50	381	360	10-15	1368.3	3.8	1.3	30
W3	04/08/85	23/07/87	20	84	76	20-25	172.1	2.3	1.2	40
W1	04/08/85	23/07/87	20	87	82	20-25	140.8	1.7	0.9	25
W5	04/08/85	23/07/87	15	56	56	15-20	103.4	1.8	0.9	30

(B)

Site	Date Installed	Measured	Number of Pits	Number Opened	Pit Number	Foil Depth	Movement (cm)			Permafrost Table
						Near Surface	10 cm Depth	15 cm Depth	30 cm Depth	
						(cm)	(cm)	(cm)	(cm)	(cm)
E2	05/08/85	23/07/87	6	2	1	1.9	1.8	1.8	1.4	32
					2	0.9	1.0	0.8	0.5	
W2	04/08/85	23/07/87	2	1	1	1.1	1.1	1.0	0.9	25
W3	04/08/85	23/07/87	3	1	1	1.1	1.0	0.8	0.8	40
N	01/07/84	23/07/87	10	2	1	1.4	1.4	1.2	0.9	30
					2	0.5	0.2	0	0	
S	02/07/84	23/07/87	8	2	1	0.9	0.8	0.7	0.6	35
					2	0.8	0.6	0.5	0.3	
E1	05/08/85	23/07/87	6	2	1	no measureable movement				35
					2	no measureable movement				
W5	04/08/85	23/07/87	2	0						30

Total number of pits 37; pits remaining to be opened 27.

this slope did not experience the greatest rate of permafrost creep.

The objects in seasonally cryotic ground also show considerable variation in deformation across the slope, as already highlighted by Egginton and French (1985). One pit at Site E2 showed twice as much deformation as the other over the same time period. While one pit at Site N showed near-surface movements of 0.9 to 1.4 cm yr⁻¹, the second pit was essentially inactive. Such results emphasize the basic heterogeneity of frozen soils and, perhaps, point to the limited value of test results based upon single boreholes or point (i.e. spot) observations.

6.5 Discussion

The data presented above enables one to speculate, in qualitative terms, about the dynamics of permafrost creep. With knowledge that supersaturated (ice-rich) frozen sediments creep in a manner similar to ice, Keusen and Haeberli (1983) were able to predict the downslope deformation of frozen unconsolidated sediments in the Swiss Alps to be approximately 1.0 to 10.0 cm yr⁻¹. Subsequent measurement revealed creep rates of 5.0 to 8.0 cm yr⁻¹. Their predictive results were calculated using Glen's Flow Law combined with data derived from photogrammetric measurement of rock glaciers. It was necessary to combine

these quite different data sets since Glen's Flow Law (e.g. Glen, 1955) alone tends to overstate the rate of deformation of frozen soils.

Glen's Flow Law predicts the creep or deformation of ice (Glen, 1955; Nye, 1956). It assumes laminar flow of an essentially plastic body. The flow law predicts the relationship between an applied stress (σ) and the resulting strain rate ($\dot{\epsilon}$):

$$\dot{\epsilon} = \sigma^{\beta} \quad (6.1)$$

where β is a constant which is a function of stress and time, and n is a dimensionless exponent. According to Savigny and Morgenstern (1986b, p. 516) the latter approaches 3 for a polycrystalline ice. Although the equation gives reasonable estimates of the flow of glacier ice, Hooke and Hanson (1986) report that it may not always apply in certain low stress situations where a single dominant stress is lacking. Morgenstern (1985) suggests that the creep modulus for frozen soils must be reduced by a factor of over 6.

It follows, therefore, that frozen ice-rich soils should react to gravitational stresses in a manner similar to that of glacier ice. However, not only does the rate of flow appear to be an order of magnitude lower, but movement of any particular point on the slope is more difficult to

predict. The heterogeneous mixture of ice and soil combined with the increasing influence of internal friction assures that the dynamics of frozen soil creep differ from those of a glacier (e.g. Parameswaran, 1985).

The creep of frozen soils combines the properties of 'laminar' (unidirectional) flow within the ice lenses and 'turbulent' flow between soil or ice and soil particles. Thus, near-surface sediments are subject to a net discontinuous downslope creep (discontinuous in space and time). However, both upslope and upstream creep were measured in the field at Rea Point. Creep against gravity stresses may occur regularly (e.g. annually) or randomly (Anderson and Cox, 1986, pp. 147, 148). Culling (1983a, b) suggests that the best way to envisage the process is to consider individual soil particles, rather than the soil mass. Permafrost creep occurs as soil particles receive kinetic energy which is relieved in relation to constraints imposed by surrounding particles. Thus, while gravity is the most constant stress, it is not always the most dominant stress. This may explain the apparently anomalous upslope movements.

On a macroscale, two major stresses dominate frozen soils, tension and compression. It is these stresses to which the inclinometer casing is most sensitive. Gravitational stresses produce tension (extension) where slope angles increase and compression where slope angles

decrease. Thermal stresses induce compression (expansion as temperatures increase) and tension (contraction as temperatures decrease). Ice segregation of secondary frost heave produces tension (uplift) in soils while thaw of excess ice leads to soil compression (or subsidence). Both tension and compression will displace soil, though the degree of displacement depends on the volume and content (air, water, ice, brine) of soil voids.

These considerations may help explain the high rates of movement experienced at Slopes N and W (see Table 6.2). It can be shown that shear stresses (τ) in a slope may be approximated by the following general equation:

$$\tau = \gamma z \sin\theta \cos\theta \quad (6.2)$$

where γ is the bulk unit weight of the soil, z is the depth to the potential failure plane, and θ is the angle of the slope. It can be argued, following Terzaghi and Peck (1967), that convex slopes undergo tensile stress and that the strength of frozen soils varies as a function of the stress to which they are subject. For example, frozen sediments undergoing tension are 12 to 15 times weaker than those undergoing compression (Vyalov, 1986) and are more likely to approach steady state creep and/or progressive creep than others. This may explain the high creep rates measured on Slopes W and N, and why limited creep was

measured on Slopes E1 and S, located on relatively long, gentle surfaces.

Vyalov (1986, p. 50) states that the deformation of frozen soil subject to unidirectional stress will lead to particle reorientation resulting in reduced resistance to stress. In the EBA trenches excavated at Rea Point, a preferred orientation of soil particle in the upper 50 cm of permafrost was not observed. This suggests that the motion is random and flow is turbulent. Multi-directional stresses occur and simple laminar flow is restricted. In fact, gravitational and thermal stresses tend to operate horizontally (or parallel to the ground surface) while ice segregation stresses operate vertically. Thermal stresses, however, may induce upslope, downslope or cross-slope deformation. It is believed that thermal stresses are the primary cause of the cross slope deformations (b axis) recorded by the inclinometer (Figure 6.9). No trends are discernible on the graphs. The casing appears to be simply shifting back and forth, exactly the type of motion one might expect as a response to thermal stresses.

At Rea Point, it is suggested that damped or attenuating permafrost creep processes dominate. In summer, moisture contents are generally highest at the base of the seasonally cryotic layer. The water which migrates in the direction of coldest temperatures will flow through and freeze within the permafrost. Secondary frost heave

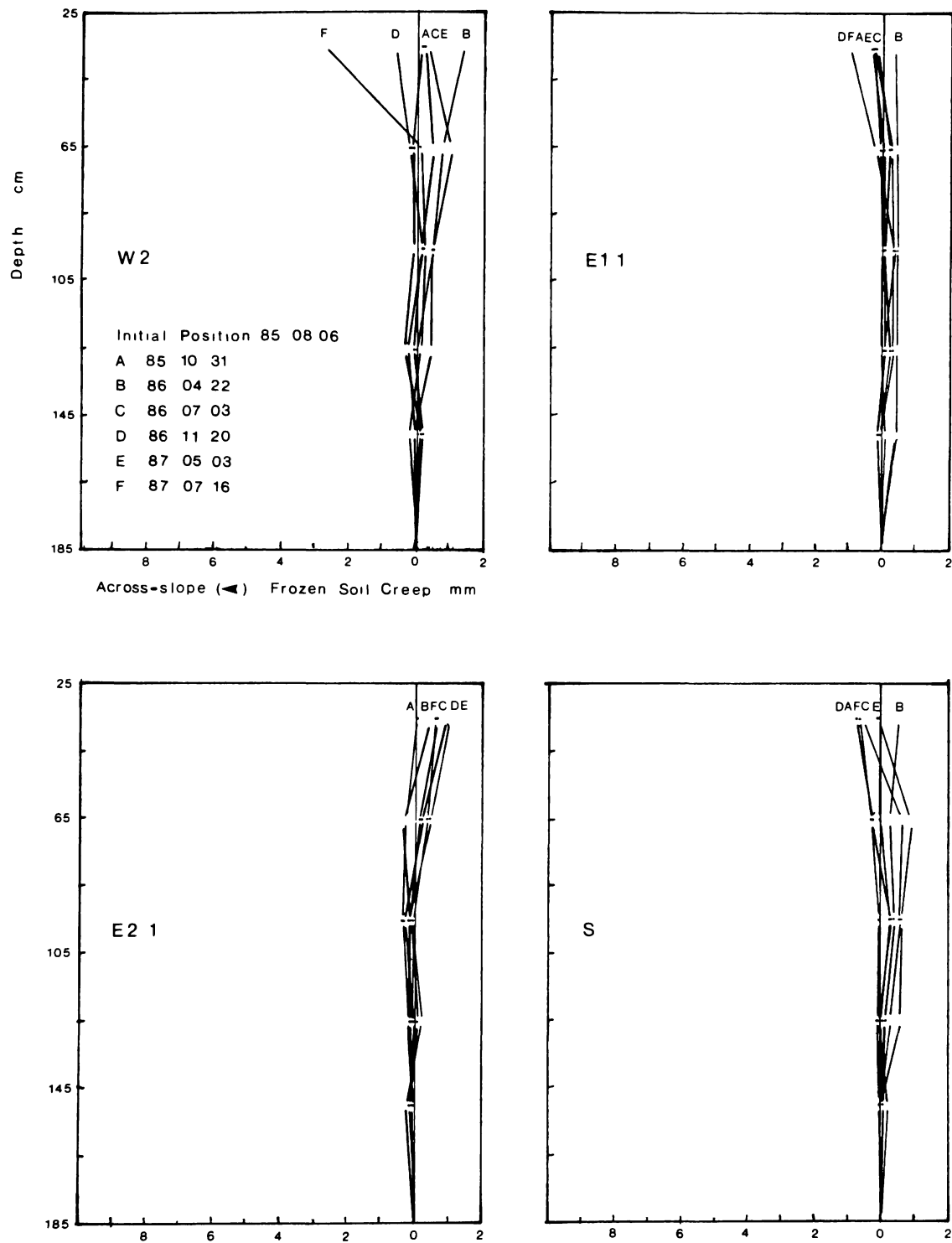


Figure 6.9 Diagrams showing B-direction (i.e. perpendicular to slope) inclinometer data for 4 sites.

produces ice lenses which become unstable at a critical thickness and temperature. Attenuating creep is initiated which leads to a thinning of the ice lense. Eventually, as more and more soil particles come into contact, internal friction will create sufficient strength to resist creep. Deformation will cease. The process will recur elsewhere as water from the active layer (as well as water created within the permafrost by regelation and heat of fusion) produces new ice lenses which reach critical thickness. The distribution of ice within permafrost may change simply due to changes in temperature (e.g. Washburn, 1985, pp. 167, 168).

Clearly such creep would lead to the reorientation of some of the soil particles. It is believed, however, that these may be disturbed by subsequent secondary heave processes. The continuous creep of a soil mass brought about by attenuated creep processes distributed randomly throughout the soil profile gives the effect of secondary or steady-state creep. Thus, for many engineering purposes, ice-rich soils may be represented by a secondary creep model (Sayles, 1985).

The creep data presented earlier in this chapter also prompt speculation as to the magnitude of deformation which might occur when ice-rich permafrost strata thaws (Morgenstern, 1981; p. 322). This has relevance to the Pleistocene periglacial and stratigraphic literature since

thawing permafrost conditions are sometimes invoked to partially explain phenomena such as valley bulges, cambers and gulls (e.g. Kellaway and Taylor, 1953; Horswill and Horton, 1976). Valley bulges are anticlinal structures developed in valley floors while the cambering process results in a progressive valleyward thinning of strata and gulls consist of crevasses, and cavities in otherwise coherent bedrock.

In southern England, such structures are widespread and occur in both glaciated areas and areas that have experienced periglacial conditions (e.g. Hollingsworth et al., 1944; Akermann and Cave, 1967; Chandler et al., 1976; Hawkins and Privett, 1981). They are best developed in Jurassic and Cretaceous sediments which contain thick, often overconsolidated, argillaceous sequences. Vertical displacements as great as 40 - 45 m are known (Kellaway, 1972, p. 143). Camber structures also occur in both glaciated and 'periglacial' areas, usually in terrain of undulating to low relief.

Possible mechanisms for the development of these non-diastrophic structures include lateral movement due to stress relief, vertical loading due to overlying ice, and the downslope sliding of frozen ground. In the case of the latter, Vaughan (1976, p. 458) suggests that horizontal stress equal to vertical stress would develop if there were viscous creep in ice-rich frozen soil. High pore pressures

might also develop temporarily if the melting of icy beds occurred at depth. Equally, shearing at the base of permafrost might occur. According to Vaughan (1976, p. 460):

" . . . ground freezing would result in expansion of the material in the valley slopes, with creep within and shear below the frozen ground, resulting in valleyward displacement decreasing with depth and increasing towards the valley centre."

In all probability, the slow degradation of warm permafrost at the end of each cold period during the Pleistocene might have been characterised by widespread permafrost creep, with resulting ground deformation.

6.6 Summary

Permafrost creep was measured in the ice-rich soils which are typical of the colluvium near Rea Point, Eastern Melville Island. Creep rates ranged from 0.0 to 0.3 cm/yr at 35 cm depth and 0.0 to 0.1 cm/yr at 65 cm depth. The typical pattern of in situ permafrost creep consists of a recognizable downslope component during summer and early autumn, followed by a weaker upslope component during the winter and early spring.

At Rea Point, permafrost creep is initiated by gravity, thermal contraction and expansion, and secondary

frost heave stresses. The rate of creep is affected by ground ice content, ambient temperature, slope angle and associated agents. While surface features such as thermal contraction cracks, earth hummocks and mud boils indicate frost action and suggest the possibility of permafrost creep, downslope deformation of materials in the seasonally thawed layer (by frost creep and gelifluction) generally appear unrelated to the process of in situ permafrost creep.

There is speculation that the creep of frozen soil can partially explain phenomena such as valley bulges, cambers and gulls in certain mid-latitudes which experienced permafrost conditions during the Pleistocene. It is believed that the data provided here, and those of Savigny (1980), lend additional circumstantial support to these hypotheses.

CHAPTER SEVEN

CONCLUSIONS

7.1 Threshold Conditions

Gravity, temperature change and secondary frost heave are the major factors inducing permafrost creep. Although ground ice content, ambient temperature, slope angle and associated components (e.g. cryostructure, aspect, snow cover and depth) are the major factors affecting creep, it is difficult to ascertain the specific contribution of each agent to the creep of frozen sediments. However, certain generalizations may be made.

Gravity stresses leading to net downslope creep are most effective when permafrost is relatively warm and thus operate best within the upper 0.5 m of permafrost. Permafrost creep (upslope, downslope or cross-slope) resulting from temperature change may be measured in the upper 1.0 to 1.5 m of perennially cryotic ground. Thermal contraction and expansion is not limited to a particular range of temperatures and its effects may be measured over the winter months. Vertical creep resulting from secondary frost heave is most notable during fall freeze-up and spring thaw. However, vertical creep was measured during summer and also between November and April when maximum soil temperatures were less than -10.0°C . Secondary frost heave

was measured in the upper 0.3 m of permafrost.

In general, it was determined that as ice content increases, the role of in situ permafrost creep increases. As it is impossible to separate the various agents of creep, a plot of ice content versus creep rate shows considerable scatter. At Rea Point, the majority of the Weatherall Formation sites display creep ranging from 0.06 to 0.10 cm yr⁻¹ (65 cm depth) when volumetric ice content is greater than 32%. It is possible that Hecla Bay Formation colluvium requires a higher ice content to promote the same rate of creep though data are limited.

An exponential relationship between the rate of permafrost creep and soil temperature was determined. On east-facing slopes, a soil temperature of -1.0°C was required to initiate creep (65 cm depth). On the west slope, where ice content and slope angle are greater, creep was initiated when soil temperatures exceeded -1.6°C. One site on the west slope experienced significant creep in spite of experiencing cooler than average summer temperatures (e.g. -2.7°C). It is believed that soil temperatures greater than -4.0°C were maintained at this site for an extended period of time in the fall due to the insulating effects of deeper snow cover.

Permafrost creep was measured primarily on slopes exceeding 5° in angle. One exception was that of Site N where significant creep occurred on a slope of 2.5°. This

site is located on the upper part of a convex slope. On such a slope shear stresses are comparatively high while creep strength (a function of the tensile strength of the frozen soil) is comparatively low. The potential for a higher creep rate on this slope in combination with a relatively high ice content may explain the relatively high creep rate measured.

7.2 Future Research

Further field measurements of in situ permafrost creep are required to confirm the earlier results of laboratory creep tests. They should be located in a variety of periglacial environments, particularly where soils are ice-rich, and should occur on slopes of varying angle, aspect and snow cover.

Laboratory tests are primarily a study of induced creep rather than natural creep. These tests are designed to reveal the bearing capacity of a frozen soil: the vertical or lateral stress imposed by a structure on surrounding soil is measured, rather than the natural downslope stresses imposed by soil on a structure.

The relationship between creep and natural (i.e. in situ) creep has yet to be determined. Natural permafrost creep is a slow process operating in heterogeneous materials at temperatures near 0°C; thus, laboratory study of natural

creep requires long tests using undisturbed samples at constant temperatures.

In situ permafrost creep is a subtle process that creates few distinctive surface features (except for rock glaciers). This explains perhaps the relative lack of field data. At Rea Point, Melville Island, the nature and magnitude of annual permafrost creep was limited in both time and place. It is limited to about 100 days/year during which time the permafrost is warm enough to promote measureable creep. Also, it is restricted to the upper metre of permafrost. In the Mackenzie Valley, however, permafrost creep may occur during much of the year and at greater depths. Much remains to be learned about this process.

This thesis incorporates data collected between August 1985 and August 1987. Time constraints (a consideration in all fields of study) limit the data base and thus, results require careful interpretation. It is hoped that the Rea Point study sites can be measured again in five and ten years time.

Monetary constraints limited the number of instrumented sites. In total, there were ten sites on four slopes. Although eight sites were concentrated on west and east-facing slopes, only one instrumented site was upon each of the north and south-facing slopes. Thus, the discussion of permafrost creep upon the north and south-facing slopes

is based upon the results of only two inclinometer casings. Moreover, since only one casing was situated on bedrock other than Weatherall Formation or regolith derived from that formation, the results should only be extrapolated to other lithologies with extreme caution.

Time and monetary constraints of this project in the High Arctic islands made it impossible to contemplate manipulation of any of the field variables. For example, the ground thermal regime could have been varied by using snow fences to alter snow cover. The effects of slope angle could have been determined by placing a series of instrumented locations perpendicular to slope. Further studies along these lines will help clarify the contribution of in situ permafrost creep to mass wasting and landscape development in northern regions.

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