A GEOMORPHIC INVESTIGATION OF RETROGRESSIVE THAW SLUMPS AND ACTIVE LAYER SLIDES ON HERSCHEL ISLAND, YUKON TERRITORY

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Abstract

This thesis investigates the geomorphology of retrogressive thaw slumps and active layer slides on Herschel Island, northern Yukon Territory. In particular, it examines the formation and morphology of both landforms, and the ground ice characteristics of retrogressive thaw slumps. During 1988-1989 a number of retrogressive thaw slumps and active layer slides were surveyed and monitored. Field and laboratory investigations involved (1) documentation of landform distribution, setting and morphology, (2) examination of processes of landform formation, and (3) the examination of cryostratigraphy, ground ice characteristics and material properties.

Retrogressive thaw slumps developed in areas of low to moderate slopes underlain by a variety of sediments with ice contents up to 4500% (on a dry weight basis). Slump headwall retreat rates of up to 19.5 m/yr were recorded. By comparison, active layer slides developed on steeper slopes underlain mainly by marine silts and clays. The sediments exposed in the slide floors and headwalls displayed no visible ground ice, but moisture contents were between 15-35%. Retrogressive thaw slumps and active layer slides form by entirely different processes. However, they do occur in close association and are influenced by many of the same parameters.

Resume

Le présent mémoire de maîtrise porte sur la géomorphologie des éboulements de fonte (retrogressive thaw slumps) et des glissements de mollisol sur l'île Hershel, dans le nord du Yukon, et plus particulièrement sur leur formation et leurs aspects morphologiques ainsi que sur les caractéristiques de la glace dans les éboulements de fonte. Un certain nombre de sites ont été observés et arpentés en 1988 et en 1989. Les travaux sur le terrain et en laboratoire ont couvert trois aspects, à savoir 1) l'évaluation de la distribution, de l'environnement immédiat et de la morphologie des formes de terrain, 2) l'étude des processsus de formation et 3) l'étude de la cryostratigraphie, des caractéristiques de la glace et des propriétés des matériaux.

Les éboulements de fonte se manifestent au droit de pentes faibles à moyennes, dans des sédiments variés ayant jusqu'à 4 500 % de contenu en glace (par poids sec). On y a mesuré un taux de recul des falaises maximum de 19,5 m/an. Par contre, les glissements de mollisol s'inscrivent dans des terrains plus pentus riches en limons et en argiles marines. Bien que le contenu en eau de ces sédiments soit de 15 à 35 %, aucune concentration de glace n'a été observée soit en coupe ou encore le long de la surface de glissement. Quoique issus de processus géomorphologiques fort différents, les éboulements de fonte et les glissements de mollisol n'en sont pas moins des formes associées régies en grande partie par les mêmes paramètres.

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Table of Contents

List of Figures

List of Tables

The Formation of Ground-Ice Depressions - A Record of an Alternative Explanation.

Isbrants Ides, writing in 1706 of an expedition to Siberia, (Sloane, 1728) reported that:

"Amongst the hills, which are situated to the north-east of Makofskoi, not far from thence, the Mammut's tongues and legs are found; as they are also particularly on the shores of the rivers Jenize, Trugan, Mongamesa and Lena... Concerning this Animal there are very different reports. The heathens of Jakuti, Tugusi, and Ostiacki say, that they continually, or at least by reason of the very hard frosts, mostly live under ground, where they go backwards and forwards; to confirm which, they tell us, that they have often seen the earth heaved up, when one of these beasts was on the march, and after he was past the place, sink in, and thereby make a deep pit. They further believe, that if this animal comes so near the surface of the frozen earth, as to smell, or discern the air, he immediately dies, which they say is the reason that several of them are found dead on the high banks of the river, where they unawares come out of the ground."

Chapter 1

Introduction

1.1 Introduction

The Yukon Coastal Plain and Mackenzie Delta are possibly the most ice-rich permafrost areas of the Canadian Arctic (Pollard and French, 1980). These areas are also characterized by a large number of natural thermokarst landforms, including: retrogressive thaw slumps, thaw lakes, beaded drainage, and collapsed pingos. However, with the exception of studies by Kerfoot (1969) and Mackay (1966) there has been little research on thermokarst processes in this part of the Arctic. Furthermore, apart from a number of site specific thermokarst studies (e.g., Are, 1988; Black, 1969; French and Egginton, 1973) this aspect of the geomorphic environment has received limited consideration. A similar problem also does exist concerning active layer slides. As a result, confusion has arisen from conflicting and often geomorphologically incorrect use of terms in the description of these features.

Surficial deposits of the Yukon Coastal Plain, including Herschel Island, contain considerable volumes of massive ground ice and ice-rich permafrost. It is not surprising then that retrogressive thaw slumps and active layer slides are some of the most active landscape-forming processes. The general aim of this study is to investigate retrogressive thaw slumps and active layer slides in the Herschel Island area of the Yukon Coastal Plain. A study of this nature will contribute to a better understanding of the role of thaw-induced processes in landscape evolution. Furthermore, information leading to an improved comprehension of geomorphic processes may help to resolve problems of interpretation brought about by contradictory terminology. Finally, there is still a need for detailed site specific permafrost studies to help develop a more complete data base of permafrost and ground ice conditions in the Canadian Arctic. In view of the planned development of oil and gas resources of the Beaufort region and the anticipated stress upon the environment associated with predicted global climate warming, there is an immediate need for detailed investigations of both thermokarst and thaw related slope processes. Retrogressive thaw slumps are of particular significance because they are extremely dynamic modifiers of permafrost landscapes. A better understanding of the geomorphology of retrogressive thaw slumps and active layers slides, together with an improved data base on ground ice and permafrost conditions for Herschel Island, will assist in the planning and management of the Herschel Island Heritage Park and the North Yukon National Park, as well as with any further development of Beaufort Sea oil and gas.

The landscape of the Yukon Coastal Plain is characterized by evidence of previous thermokarst activity, such as thaw lakes, cemetary mounds and stabilized retrogressive thaw slumps. The cyclic nature of thermokarst, together with the high degree of thermokarst activity, forms the basis for a number of geomorphic questions. Is the present period of thermokarst activity triggered by global change? What causes the polycylic nature of retrogressive thaw slumps? What is the triggering mechanism for both retrogressive thaw slumps and active layer slides? It is not possible for a site-specific short-term study such as this to fully answer these questions; however, it is hoped that by investigating some of the processes and landforms involved we may be better equipped to focus on these fundamental questions.

1.2 Study Aims

The specific aims of this research are:

(1) to document the distribution and setting of retrogressive thaw slumps and active layer slides on Herschel Island,

(2) to investigate retrogressive thaw slump and active layer slide processes, including (a) factors influencing their initiation, (b) mechanisms and rates of retreat and (c) the polycyclic nature of retrogressive thaw slump occurrence,

(3) to investigate and compare the morphology of retrogressive thaw slumps and active layer slides, through the measurement of a number of standard morphometric parameters (e.g., total length, total width,

scar length, and scar depth),

(4) to examine material and ground ice characteristics (grain size, Atterberg Limits, moisture contents) of retrogressive thaw slump and active layer slide materials, in order that the role of moisture in slump and slide morphology and process may be assessed, and

(5) to assess the use of traditional mass wasting and thermokarst terminology to describe these features.

Chapter 2

Background Literature

2.1 Introduction

The study of earth materials having temperatures less than or equal to $0^{\circ}C$ is called geocryology. The recognition of geocryology as a science (Shumski, 1964; Washburn, 1980) has given rise to a specialized terminology (van Everdingen, 1976) in which the prefix "cryo" and term "cryotic" refer specifically to subzero ($^{\circ}C$) thermal conditions, and the term "frozen" refers to the solid phase of moisture found in cryotic materials.

Permafrost geomorphology is concerned with landforms and processes associated with the occurrence of permafrost (Brown, 1974; Mackay, 1971a; Rampton, 1974). This includes both aggradational and degradational phenomena. Geocryology and permafrost geomorphology are relatively new branches of the earth sciences. Although they have their origins in climatic and periglacial geomorphology, their relevance today is directly related to problems associated with climate change and northern development.

Over the past four decades, an extensive body of permafrost literature has been produced, yet several important topics remain shrouded in confusing and contradictory terminology, or are poorly understood. This is the case for some aspects of thermokarst processes and landforms. The following will briefly summarize important concepts and background information on permafrost, ground ice and thermokarst.

2.2 Permafrost

Permafrost is defined as ground (soil or rock) that remains at or below 0°C for at least two years (Permafrost Subcommittee, 1988). Although the thermal nature of permafrost is implicit in this definition, its applied significance is directly related to the behaviour of water/ice at or below 0°C. Permafrost underlies approximately 20% of the earth's land surface (Figure 2.1), including as much as 50% of Canada and the Soviet Union and 80% of Alaska (Brown, 1978; Péwé, 1983). The occurrence of permafrost reflects the complex interaction between climate (i.e., mean annual and winter air temperature), geothermal gradient and material characteristics. In North America permafrost distribution is divided into continuous and discontinuous permafrost. Alpine and subsea permafrost are also recognized.

The continuous and discontinuous zones of permafrost in Canada are defined on the basis of spatial extent, areal coverage and permafrost depth. Within the continuous zone, permafrost exists almost everywhere beneath the land's surface and may extend to depths of 500-1000 m (Brown, 1972). In theory, the colder air temperatures and corresponding shallower active layers experienced in areas of continuous

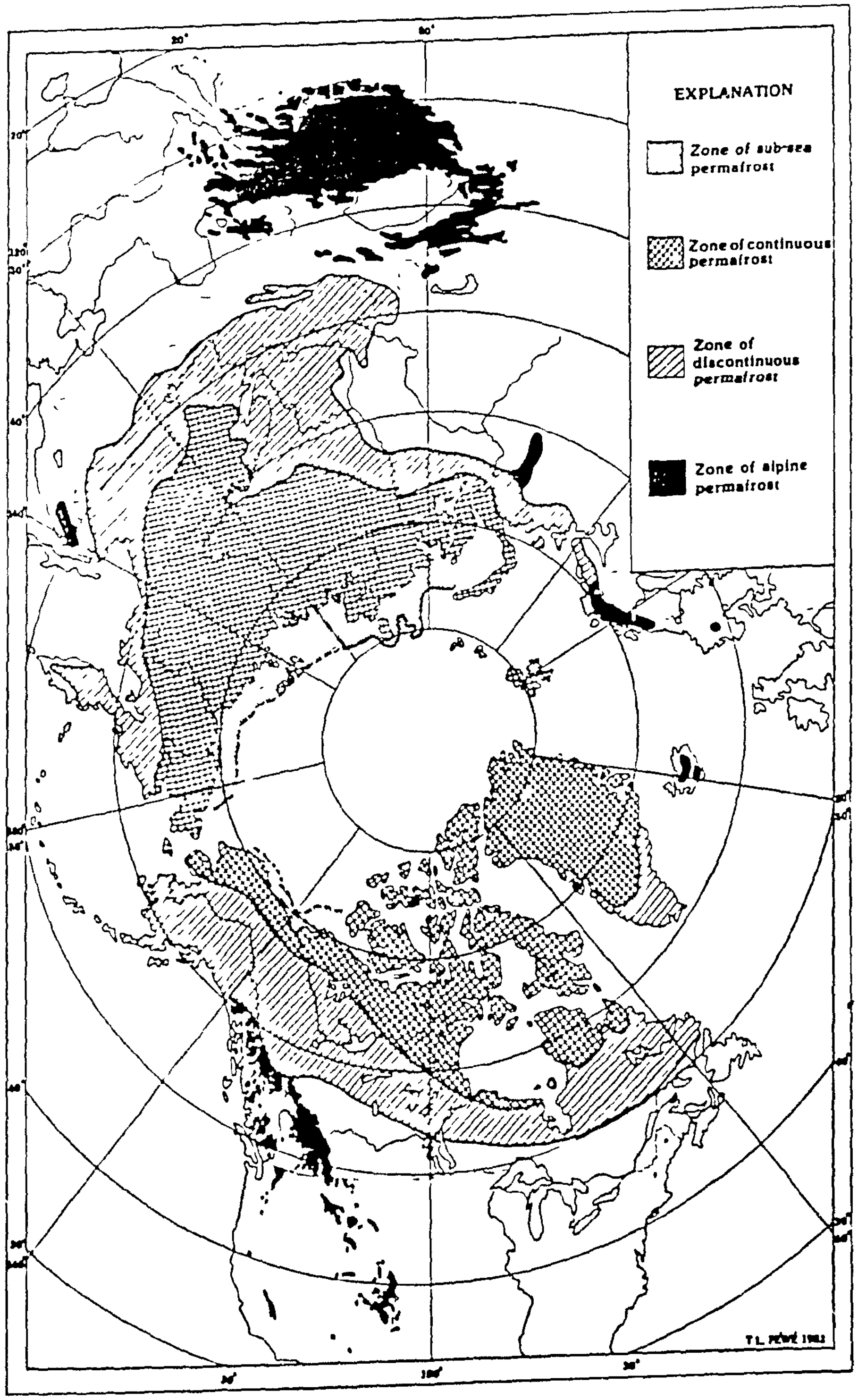

Figure 2.1. Map of global permafrost distribution (Péwé, 1983).

permafrost will limit the nature and intensity of thaw-induced processes. At lower latitudes permafrost distribution becomes discontinuous in nature and is characterized by the presence of permafrost and nonpermafrost areas in close association. As discontinuous permafrost grades from widespread to sporadic, site specific factors, such as topography, vegetation, snow cover become increasingly important variables influencing the distribution of cryotic and noncryotic materials (Brown, 1963a, 1963b, 1973; Nicholson and Granberg, 1973). The discontinuous zone is further subdivided into widespread permafrost in the north to sporadic permafrost at its southern limit (Figure 2.2). Discontinuous permafrost ranges in depth from 100 m to only a few centimetres near its southern limit, where its presence may be quite transient (Brown, 1970). Altitudinally (azonal) induced permafrost is termed alpine permafrost. Subsea permafrost exists beneath coastal shelf areas of the Arctic Ocean, and may either be in equilibrium with negative seabottom temperatures or a relic of cold climates during former emergent conditions (Are, 1988; Lewellen, 1973; Mackay, 1972a).

The temperature of permafrost in the continuous zone, measured at the depth of zero annual amplitude, ranges from about -15°C in the High Arctic to about -5°C at lower latitudes. In the discontinuous zone, the temperature ranges from approximately -5°C to 0°C (Brown, 1978; Judge, 1973). Alpine and subsea permafrost temperatures are comparable with

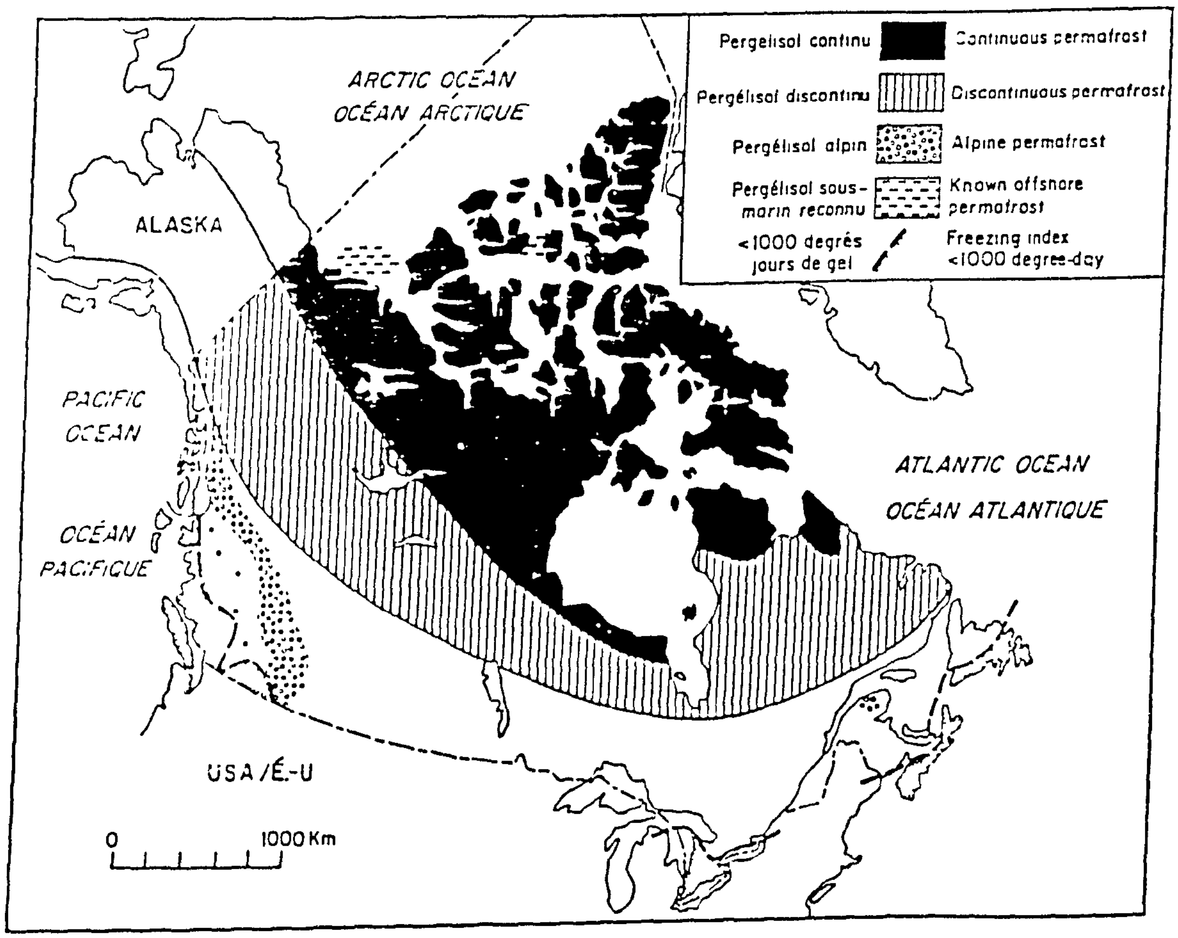

Figure 2.2. Permafrost Distribution Map of Canada (after Brown, 1978).

the discontinuous zone. In most permafrost regions a surficial ground layer, termed the active layer, lies above the permafrost table and experiences seasonal freezing and thawing. In the continuous zone, active layer thickness ranges from 10-20 cm in the coldest areas, but often reaches one metre near the southern limit. In the discontinuous zone active layers tend to increase in thickness from north to south, up to a maximum of two to three metres (Gold, 1967; Gold and Lachenbruch, 1973; Goodrich, 1982; Lachenbruch, 1959).

In North America, permafrost or geocryological science is a relatively recent scientific field. For example, the term 'permafrost' was introduced in 1947, by S.W. Muller as a short form for "perennially frozen ground". Some of the earliest scientific observations of permafrost and ground ice conditions in North America were for the Yukon and Alaska coasts as part of a geological study undertaken by Leffingwell (1919). The Soviet Union has a longer history of permafrost research which began in the early 1900's; such research is now a well established discipline (e.g., Tsytovich, 1975). This is in part due to their long history of northern settlement and resource development. The rapid development of permafrost science in North America, the Soviet Union and, recently China is well illustrated by the proceedings of five International Conferences on Permafrost (1963, 1973, 1978, 1982, 1988).

2.3 Ground Ice

Ground ice is a major component of permafrost. Its aggradation and degradation contribute significantly to the evolution and modification of periglacial landscapes (Brown, 1974; Harry, 1988; Rampton, 1974). For example, ground ice aggradation associated with the formation of ice-wedge polygons and ice-cored mounds (pingos) is responsible for the unique appearance of many permafrost areas. Equally, ground ice degradation may dramatically modify a landscape and contribute to high rates of erosion. From an applied perspective, the presence of ground ice within permafrost has long been recognized as a key factor influencing the design and the high cost of construction and maintenance of northern engineering projects (e.g., Brown, 1970; Ferrians *et al.*, 1969; French, 1979; Muller, 1947).

Ground ice is broadly defined as all ice occurring beneath the surface of the ground, irrespective of age and origin (Brown and Kupsch, 1974; Mackay, 1966). In theory, this definition includes ice occurring within both seasonally and perennially cryotic ground; however, in North America the term is generally limited to ice occurring within permafrost. Ground ice may be either syngenetic or epigenetic in origin. Syngenetic ice forms simultaneously with sediment deposition and, thus, is part of the original depositional sequence. For example, ice-wedge growth in a river flood plain. Epigenetic ice is younger than its enclosing sediments (e.g., ice aggrading into sediments). The volume of ground ice varies both horizontally and vertically, and is dependent upon the type of enclosing sediments, the availability and source of moisture, the rate of freezing, and the thermal and geologic history at a given location. Ice content may be expressed as a percentage by weight or volume of the enclosing sediment. Massive ground ice is a term used to describe large masses of pure ice. It is characterized by a gravimetric ice content equal to or greater than 250% (Mackay, 1971b, 1972b) or approximately 85-90% by volume (Pollard, 1990).

During the past three decades considerable research has focused on the nature and occurrence of ground ice, dealing specifically with its origin, distribution and geotechnical significance (Brown, 1978; Johnston, 1981; Mackay, 1989; Mackay *et al.*, 1978; Pollard, 1989, 1990; Williams, 1979). Areas of Alaska, Canada and Siberia have undergone extensive drilling programs in support of engineering investigations for pipeline construction, water resources, research purposes and hydrocarbon exploration. This has provided the opportunity to record the depth and character of frozen material and subsurface conditions. Examples include Mackay's (1971b) use of borehole logs to interpolate the geometry of massive ice on the Tuktoyaktuk Peninsula. Bouchard (1974) used a similar method to map the distribution of ground ice on Herschel

Island. Using the Mackenzie Valley Geotechnical Data Bank, Pollard and French (1980) calculated an approximation of ground ice volume for Richard's Island in the Mackenzie Delta.

In Canada, two general ground ice classifications have been adopted, the genetic classification by Mackay (1972b) and a field classification by Philainen and Johnston (1963). Mackay's classification (Figure 2.3) is based upon the origin of water prior to freezing, the mode of water transfer, and the types of ground ice formed. Philainen and Johnston's classification (Figure 2.4) is based on visual examination of frozen samples. It is qualitative in nature and was designed for engineering purposes as part of a larger permafrost analysis procedure.

Ground ice types frequently observed in permafrost include: ice-wedge ice and vein ice, segregated ice (massive tabular ice, ice layers, ice lenses), intrusive ice and pore ice (Mackay, 1972b; Pollard and French, 1980) (Figure 2.5). In addition, buried surface ice (buried snowbank ice, glacier ice) is being documented in many areas (Pollard, 1989; Pollard and Dallimore, 1988).

Ice wedges occur where thermal contraction cracking during winter months is common. They are vertically foliated and characteristically wedge shaped bodies of massive ice, which may appear "milky" in colour (due to high concentrations of gas inclusions) compared to near pure ice. They grow by the incremental accumulation of annual ice veins, which form

Origin of Water Prior to Freezing	Principal Transfer Process	Ground Ice Forms
Atmospheric water	Vapor diffusion	Open cavity ice
	Thermal contraction	Single vein ice
Surface water	Gravity transfer	Ice wedge ice
	Tension rupture	Tension crack ice
	Vapor diffusion	Closed cavity ice
		Epigenetic ice
	Thermal and pressure potential Segregated ice	Aggradational ice
		Sill ice
Subsurface water	Pressure potential Intrusive ice	Pingo ice
Expelled	In place freezing	Pore ice

Source: Mackay (1972)

Figure 2.3. Ground ice classification by Mackay.

Main Group	Symbol	Subgroup Description	Symbol	Field Identification
Ice not visible	N	Poorly bonded or friable	Nf	Hand examination
		Well bonded	Nb	Thaw sample to determine excess ice (supernatant water)
		No excess ice	Nbn	
		Excess ice	Nbe	
Visible ice less than 2.5 cm thick	V	Individual ice crystals or inclusions	Vx	Visual Examination Observations on: location orientation thickness length spacing size shape pattern of arrangement hardness structure colour
		Ice coatings on particles	Vc	
		Random or irregularly oriented ice formations	Vr	
		Stratified or distinctly oriented ice formations	Vs	
Visible ice greater than 2.5 cm thick	ICE	Ice with soil inclusions	Ice + soil type	Visual examination Observations on: hardness - hard, soft structure - clear, cloudy, porous, candled, granular, stratified, colour admixtures
		Ice without soil inclusions	ICE	

Source: Pihlainen and Johnston (1963)

Figure 2.4. Ground ice field classification proposed by Philainen and Johnston.

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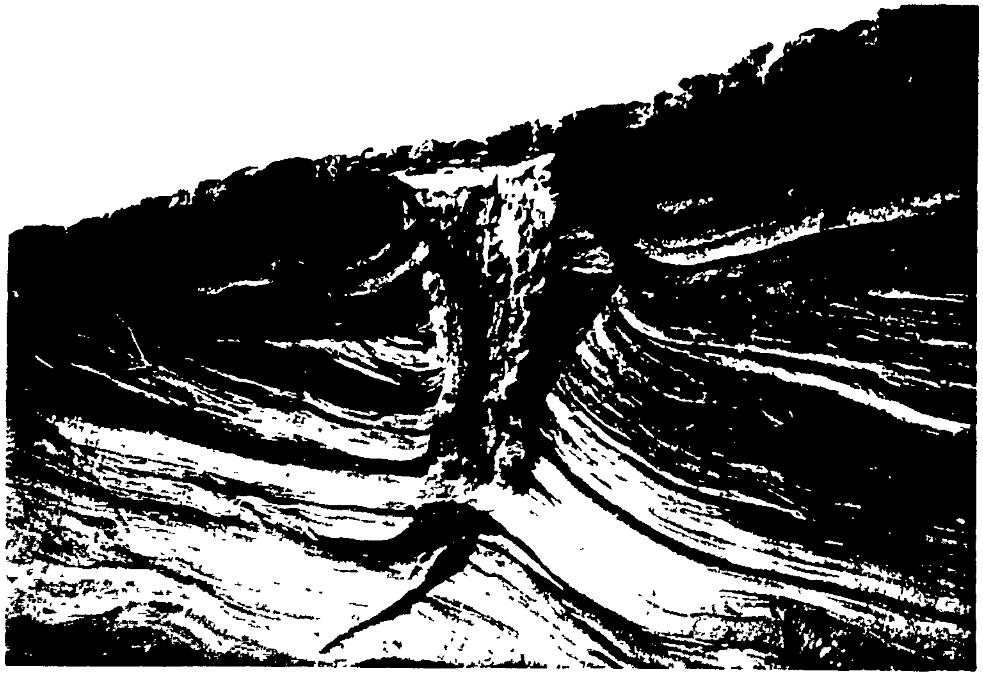

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Figure 2.5. Photographs illustrating different ground ice types: (a) ice-wedge ice penetrating massive segregated ice body, and (b) reticulated ice lenses and pore ice.

by the freezing of water that enters thermal contraction cracks. An excellent theoretical discussion on thermal contraction cracking is presented by Lachenbruch (1962). The aggradation and degradation of ice wedges is responsible for the distinct polygonal appearance of many arctic landscapes. Our current understanding of ice wedge growth has been greatly advanced by studies by J.R. Mackay in the Mackenzie Delta (1974a, 1974b, 1975b, 1976a, 1980, 1984).

Segregated ice forms by the migration of groundwater to the freezing front as permafrost aggrades into unfrozen material (e.g., Mackay, 1966; Penner, 1972). Segregated ice may occur as discrete lenses, layers or massive bodies of ice and range from thin discontinuous hairline lenses to extensive tabular ice bodies. The complex thermodynamic processes responsible for ice segregation may also produce alternating layers of ice and soil. Some of the thick bodies of massive ice described in the Mackenzie Delta and Yukon Territory may have been formed by a combination of ice segregation and injection. Although the depth to which segregated ice can occur in permafrost remains uncertain, massive segregated ice has been documented from an appreciable fraction of drill holes which exceed a depth of 40 m (Mackay, 1966, 1976b). The permafrost literature of the U.S.S.R. also includes numerous references to massive ice that appears to be segregated in nature. Vtyurin (1973) reports that test borings have revealed the presence of segregated ice in unconsolidated deposits to depths of 130 metres or more. The terms "massive ground ice" and "massive icy bodies" were proposed by Mackay (1971b) to describe the thick horizontally extensive masses of nearly pure ice and icy sediments which are common in the Mackenzie Delta. Such bodies of massive ice are frequently responsible for prominent topographic rises (Rampton and Mackay, 1971).

Pore ice forms as a freezing front moves through the soil column, freezing interstitial water *in situ*. It acts as a bonding or cementing agent by holding soil grains together and occurs as ice crystals in pore spaces or as a film of ice surrounding individual particles. The main distinction between pore and segregated ice, besides their genetic processes, is that pore ice yields no supernatant water. The thawing of pore ice will not cause ground subsidence, although there may be a loss of bearing strength (Brown and Kupsch, 1974; Johnston, 1981). By comparison, segregated ice often contains excess ice, defined as the volume of ice that exceeds the total pore volume that the ground would have under nonfrozen conditions. As a result, a soil containing excess ice will settle under its own weight during thawing until it attains a consolidated state. This is of considerable importance when considering slope stability (Rampton, 1974).

Apart from the varied character of the mineral and organic fractions within permafrost beds, it is necessary to describe the multiplicity of other conditions which may exist. Ice lenses and veins may be interspersed with frozen material thus resulting in complex configurations. A considerable body of research now exists examining ground ice cryotextures (Mackay, 1989; Pollard, 1989, 1990; Pollard and Dallimore, 1988). Ground ice shows great variations in soil both qualitatively and quantitatively. For example, in some areas it may consist of almost pure clear ice as exemplified by ice wedges. On the other hand, drier permafrost will generally have only small quantities of ice present, usually as pore or cement ice (Linell and Tedrow, 1981).

2.4 Thermokarst

Thermokarst is defined as the process by which characteristic landforms result from the thawing of ice-rich permafrost (Permafrost Subcommittee, 1988). The thawing of permafrost containing excess ice results in surface subsidence and the development of thermokarst terrain. Recent research has addressed both the processes and landforms associated with this phenomenon. Both are of considerable importance, since naturally occurring thermokarst represents one of the most dynamic processes modifying arctic and subarctic landscapes (French, 1974). In addition, the avoidance of human-induced thermokarst represents a major challenge to northern development (e.g., construction of the town of Inuvik). A process-response model which defines the factors related to the disturbance of tundra and forest-tundra environments has been presented by Mackay (1970). Surface disturbance results from the disruption of the thermal equilibrium of the underlying permafrost due to changes in geomorphic, vegetational and/or climatic conditions.

Mackay (1970) suggests that thermokarst subsidence be differentiated from thermal erosion. For example, melt-out of a pingo ice core is a thermokarst process which results in surface subsidence and the formation of a circular depression with a "doughnut shaped" rampart. By comparison, processes such as flowing water thermally erode an ice body (e.g., ice wedge or thaw face). This may be an important process in tundra environments, particularly during spring snowmelt. For example, thermal erosion along ice wedge troughs can lead to degradation of polygon terrain (Mackay, 1974a), the development of thermokarst gullies and the formation of isolated polygon remnant cemetary mounds (French, 1974). In coastal and fluvial environments, thermal erosion of ice-rich sediments contributes to rapid rates of shoreline recession (e.g., Harry *et al.*, 1983). Rates of coastal erosion by wave action may also be influenced by ground ice distribution. For example, Are (1988) has documented catastrophic block failure in areas of ice-wedge polygon terrain along the Beaufort shoreline. Similar processes were observed in 1988 by the author on the north and west sides of Herschel Island.

In Siberia, long-term thermokarst processes has produced a distinctive relief, the end stage being the development of large "alas" depressions (Czudek and Demek, 1970). Alas topography is rare in North America, indicative of differences in climate and the geomorphic history between the two regions. A close equivalent in North America may be the extensive lacustrine plains of north Alaska and the Canadian western Arctic, which appear to have developed in response to initiation, growth, and drainage of thaw lakes (Sellman *et al.*, 1975). Several cycles of this process may have been repeated during Holocene times (Billings and Peterson, 1980). Black (1969) identified a number of problems regarding the initiation and subsequent development of thaw lakes, of which, unfortunately, many remain unresolved. Some thaw lakes appear to originate as small tundra ponds formed by the thawing of ice wedges (French and Egginton, 1973), while others may grow by coalescence of ponds located within low-centered polygons (Harry and French, 1983). However, the processes associated with thermokarst lake development are still poorly understood.

Thermokarst may develop in response to local disturbance and regional environmental change. For example, a forest fire near Inuvik, N.W.T., resulted in a significant increase in active layer thickness, as well as melt-out and redistribution of ground ice (Heginbottom, 1973, 1974). Similarly, many of the active thermokarst processes on eastern Banks Island are believed to occur in response to changes in local geomorphic conditions (French, 1974; French and Egginton, 1973). However, there is also evidence in the western Arctic of regional thermokarst activity induced by Hypsithermal climatic warming approximately 9,000-10,000 yrs. B.P. (Rampton and Bouchard, 1975). It is clear when flying over the Yukon Coastal Plain that many areas owe much of their present landscape characteristics to local and regional thermokarst processes.

Retrogressive Thaw Slumps

One distinctive form of backwearing thermokarst (Figure 2.6) produces bowl-shaped depressions that resemble scars resulting from slumps or flows in nonpermafrost environments. Several terms have been used to describe this landform, including ground ice slump, thaw slump, thawflow slide, thermocirque and retrogressive thaw slump. In this case, the term "slump" refers to the resemblance of the landform shape to slumps occurring in nonpermafrost areas, and not to its genetic processes. The term retrogressive thaw slump is currently recommended by the Permafrost Subcommittee of the National Research Council of Canada.

Studies have documented the morphology and dynamics of retrogressive thaw slumps observed in northern Canada (French, 1974; Kerfoot, 1969; Lewkowicz, 1988; Mackay, 1966), but few studies in the North American literature relate the

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Figure 2.6. Photographs of retrogressive thaw slumps: (a) Herschel Island (June 1987), and (b) Yukon Coastal Plain (July 1987).

morphology of thermokarst landforms to the mechanics of slump development. Furthermore, most studies have been concerned with thermokarst initiated by human-related disturbances (French, 1975). Retrogressive thaw slumps occur where bodies of massive ice or ice-rich sediments are exposed by mass wasting or erosion and they are extremely active agents in landscape modification. French and Egginton (1973) recorded scarp retreat rates on eastern Banks Island of up to 10 m per year and Edlund *et al* (1989) have documented 25 m of retreat over a period of 38 days on Ellesmere Island. Radiational energy input into the ice face of retrogressive thaw slumps may be high during the brief summer period (e.g., Pufall and Morgenstern, 1980a, 1980b) particularly on south - facing exposures. Radiometer measurement of a retrogressive thaw slump near Tuktoyaktuk, N.W.T., revealed temperatures of up to 10°C at the ice-water interface (Mackay, 1978) Many thaw slumps develop in a cyclical fashion, with periods of activity separated by periods of stability (Mackay, 1966). This process appears to be a function of changes in climate, slope, rate and pattern of erosion, and the distribution and quantity of ground ice (Lewkowicz, 1988).

Retrogressive thaw slump formation involves a combination of thermal and mass movement processes, namely: (1) ablation of the ice-rich face, (2) falling and slumping of sediment and organics from the headwall and overlying active layer, (3) sliding and flowing of material off the ice face, and (4) flow in the slump floor of supersaturated mud derived from the headwall and sediment contained within the ice (Lewkowicz, 1988; Mackay, 1966). These processes are self perpetuating until the slump is stabilized by one or more of the following mechanisms: (1) the accumulated sediment either covers the thaw face or assumes a gradient insufficient to remove material by mudflow, (2) the body of ground ice completely melts, resulting in a standstill of headwall retreat, (3) the ice content of the frozen ground decreases, or (4) the depth of the overburden changes. Any of these processes results in a decrease in the available meltwater and causes burial of the face (Lewkowicz, 1988). The resulting landform is often shaped like an elongated horseshoe, although, dimensions are highly variable. Lewkowicz (1988) suggests that slump size decreases with increasing latitude. However, it remains uncertain whether differences in size are the result of variations in ground ice distribution, slope characteristics or the local energy balance.

2.5 Mass Wasting

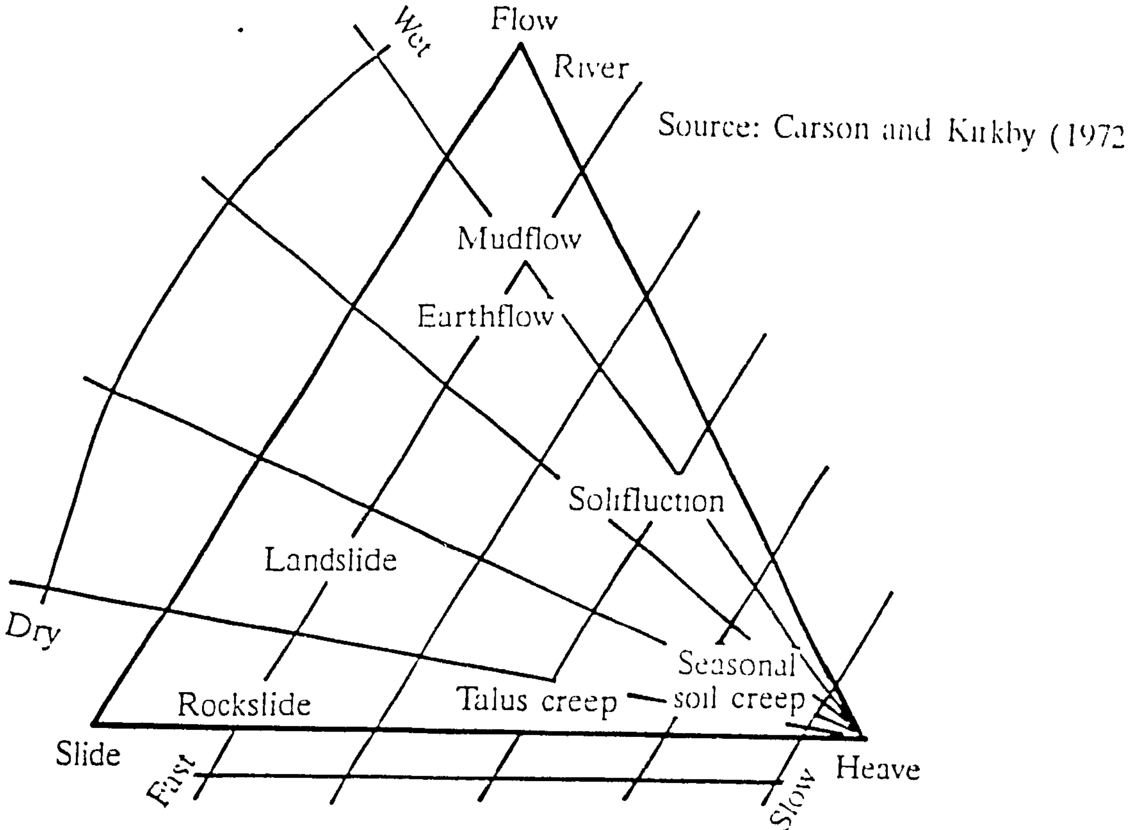
Mass wasting is the collective term for all gravity induced downslope movements of weathered debris (Permafrost Subcommittee, 1988). As the definition implies, gravity is the sole driving force, although other agents of erosion may contribute by inducing unstable slope conditions. For example, groundwater may reduce internal strength through increased pore water pressure, or erosion by running water may cause oversteepened slopes; in both cases unstable conditions may result (Rampton, 1974). Forms of mass wasting most frequently described in arctic environments are solifluction, blockfailure and skinflows. There are few references in the general mass wasting literature to other forms of thaw-induced mass movements. However, considerable work has been undertaken addressing slope problems in permafrost environments (Algus, 1986; Chandler, 1972; Harris, 1973; McRoberts, 1973; McRoberts and Morgenstern, 1974a, 1974b).

Analysis of mass wasting of any one particular slope failure is a difficult problem, since most mass wasting occurrences display more than one mechanism of motion, and since most field studies attempt to reconstruct slope failures *a posteriori* based upon resultant landforms. Thus, evidence concerning the processes in operation may not be conclusive. Nevertheless, two broad classes of mass wasting are generally described: (1) surface mass movements, which are characterized by the transfer of debris downslope as a thin layer often overlying bedrock (creep, solifluction), and (2) deep seated mass movements, which affect and alter the bedrock surface (rockfalls and landslides). Each type of movement is influenced by a number of variables, including: climatic conditions, slope angle, nature of debris and vegetation cover.

There are numerous classifications which attempt to categorize the various forms of mass movement by their mechanism of transport. A widely accepted classification developed by Carson and Kirkby (1972) separates the mechanism of mass movement into three categories: (1) slides, which include falling, slumping and sliding movements, (2) flows, and (3) heaves. By placing these particular processes on the corners of a triangle diagram, and using moisture content and rate of movement to differentiate the mechanisms (Figure 2.7.a), they have demonstrated that most mass movement events are a combination of two or three mechanisms.

Another classification by Varnes (1978) is based primarily on the type of material and the type of movement. The categories defined in Varnes' study include falls, slides and flows, with no consideration of heave mechanisms. This classification does take into account moisture content, amount of deformation and rates of movement (Figure 2.7.b). Falls are a type of mass movement which involve the descent of rock or soil material of any size from steep slopes or cliffs. Common types are stone, boulder, rock and soil falls. Observations of fall mass movements in the arctic include those occurring within thaw slumps and along coastal bluffs as a result of melting-out of ground ice (Are, 1988; Lewkowicz, 1988; Mackay, 1966). Thermal erosion of an exposed ice face may cause the formation of overhangs well in excess

A



B

Type of Movement	Type of Material		
	Bedrock	Coarse	Fine
Falls	Rock	Debris	Earth
Topples	Rock	Debris	Earth
Slides: Rotational / Translational	Rock slump Block slide Rock slide	Debris slump Block slide Debris slide	Earth slump Block slide Earth slide
Lateral spreads	Rock	Debris	Earth
Flows	Rock	Debris	Earth
Complex	Combination of two or more types of movement		

Source: after Varnes (1978).

Figure 2.7. Classification of mass movement (a) Carson and Kirkby (1972), and (b) Varnes (1978).

of one metre which may collapse and fall from a height of several metres.

Active Layer Slides

The most common sliding phenomena are those which occur on planar surfaces, termed translational slides. Initiation of these slides occurs when the material's shear strength along a predetermined failure plane is exceeded. Active layer slides are a form of rapid mass wasting found in permafrost environments (Figure 2.8). The term active layer slide is one of several adopted in the literature; other terms include active layer glides, active layer detachments, skinflows and thaw slides (Lewkowicz, 1988). They all refer to the downslope movement of seasonally thawed material as a discrete mass over a translational surface (McRoberts and Morgenstern, 1974a; 1974b). Active layer slides are long and shallow, and usually involve movement over an inclined planar surface (the top of permafrost). The resultant landform is a small ribbon-like scar oriented parallel to the slope (Lewkowicz, 1988). Although these features have been widely reported in the Arctic (e.g., Lamonthe and St. Onge, 1961; Lewkowicz, 1988; McRoberts and Morgenstern, 1974a, 1974b), few studies have been concerned with their genesis, morphology, material and ground ice characteristics, and possible association with thaw slumps. The most useful research to date is a series of studies concerned with the relationship between active layer

A

B

Figure 2.8. Photographs of active layer slides: (a) Thetis Bay, Herschel Island (July 1988), and (b) Eureka, Ellesmere Island (August 1988).

slides and material properties in the Mackenzie Valley (McRoberts and Morgenstern, 1974a, 1974b; Morgenstern and Nixon, 1971).

Flows are mass movements in which the displaced body of material deforms as a viscous fluid. They are characterized by internal shear, acceleration downslope, and a velocity profile which decreases with depth into the flowing mass. Flow dominated processes are commonly described as mudflows (Lewkowicz, 1988). They are produced by the downslope movement of saturated soil and are frequently associated with retrogressive thaw slumps (Mackay, 1966). Retrogressive thaw slump flows occur on the floor of the slump, where supersaturated fine-grained material moves downslope away from the headwall as the supernatant water drains. Flows have traditionally been defined as rapid mass movements; however, in many cases their rate is extremely variable and dependent upon water content.

Heave mechanisms are the main process in slow mass movements which are frequently termed creep. Frost heave is defined as the upward or outward movement of the ground (or objects in the ground) caused by the formation of ice in the soil, and is an important component of pattern ground (Permafrost Subcommittee, 1988). A common heave process in the arctic is solifluction, defined as the downslope flow of unfrozen earth material on a frozen substrate (Permafrost Subcommittee, 1988). Geomorphologically, solifluction is a highly significant form of slow mass movement, more effective than those forms generally occurring in temperate regions. Our present understanding of heave type processes has been greatly advanced as a result of engineering projects such as pipeline and road construction (Johnston, 1981; Konrad and Morgenstern, 1984; Williams, 1979).

Chapter 3

Study Area and Methodology

3.1 Introduction

Herschel Island is located in the northern Yukon Territory at 69°41'N 139°01'W, approximately 60 km east of the international boundary between Yukon and Alaska (Figure 3.1). It is situated 3 km north of the Yukon coast, and forms part of the Yukon Coastal Plain physiographic region which is bordered by the British and Richardson Mountains to the south and south-east, the Mackenzie Delta to the east and the Beaufort Basin to the north.

Herschel Island was selected as the study area for this research for five reasons:

(1) the ice-rich nature of the permafrost, together with steep slopes and rapid coastal erosion, is ideal for the initiation of retrogressive thaw slumps and active layer slides;

(2) there is a paucity of detailed permafrost process studies for the Yukon Coastal Plain, yet generally good background geological information (Bouchard, 1974; Rampton, 1982; Mackay, 1959);

(3) there is relatively easy access to the island from either Tuktoyaktuk or Inuvik by helicopter or fixed wing aircraft;

(4) the study is a natural compliment to the research studies

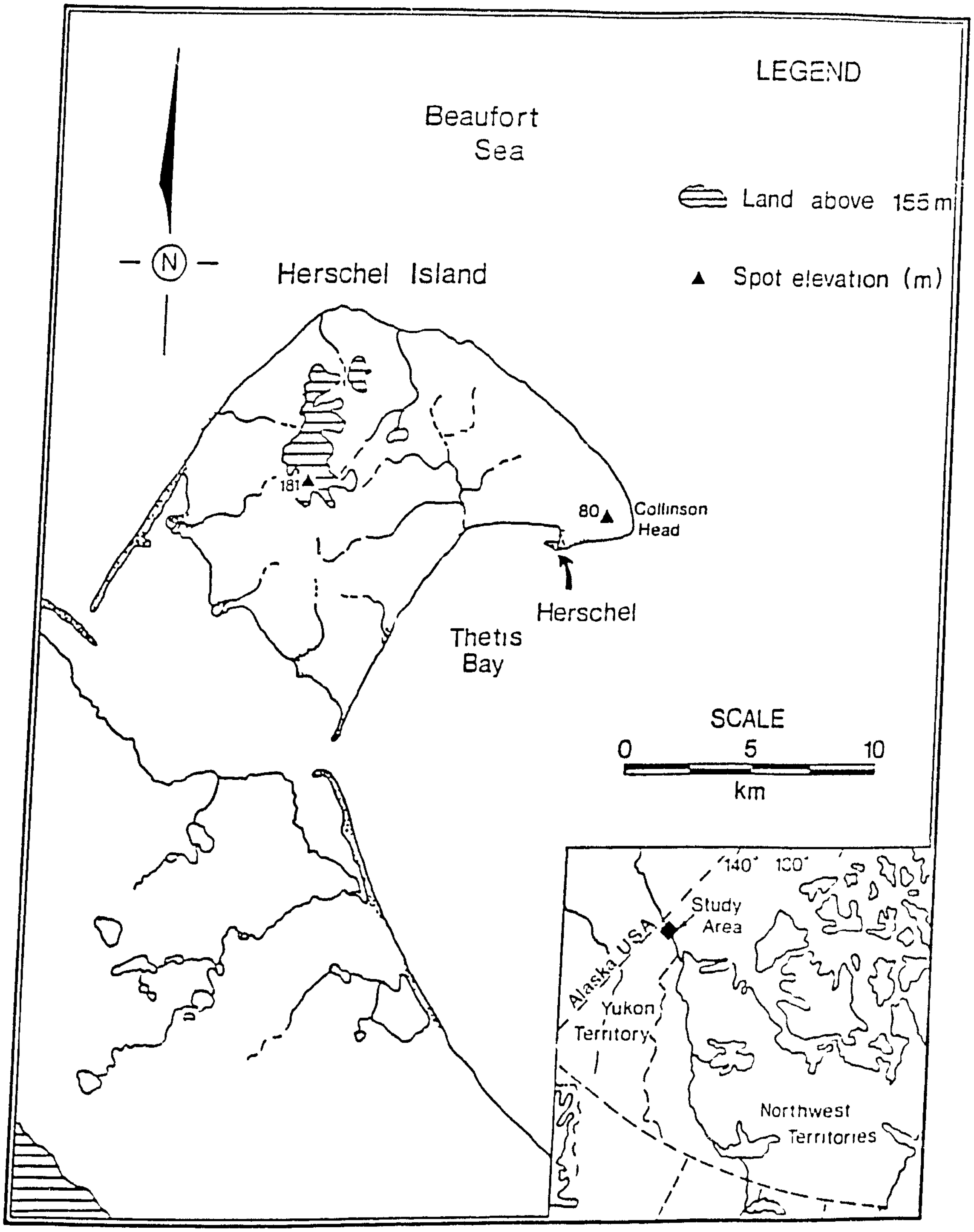


Figure 3.1. Study area map, Herschel Island, Yukon Territory.

on massive ground ice nature and genesis currently being carried out by Dr. W.H. Pollard;

(5) in recent years, Herschel Island has become a centre of interest for three reasons: (1) it has a colourful and relatively long cultural history; in 1987 it was designated as the first Territorial Heritage Park, (2) it lies within the Herschel Basin which is a well known hydrocarbon area, and (3) its high elevation and offshore position make it an ideal location for a remote early warning station (Department of National Defence).

3.2 Historical Perspective

Archaeological investigations indicate that Herschel Island was inhabited by western Thule peoples during the period from about 1000-1400 A.D. These inhabitants exploited seal, fish, caribou, whale and other species during their occupation of the area (Yorga, 1980). The abundance of various wildlife and game species like caribou (Porcupine Caribou Herd), musk-ox, Arctic Fox, wolf, grizzly and polar bear, wolverine, moose and various marine mammals (seal, beluga and bowhead whale) have attracted, and continue to attract, seasonal hunting activities of the Inuit and Dené people.

The earliest recorded visit by Europeans to Herschel Island was by John Franklin's first expedition to map the arctic coast in 1823 (Francis, 1984). Franklin named the island after the British astronomer, William Herschel, changing it from its local name Kigirktayuk.

During the late 1800's and early 1900's Herschel Island took on an entirely new significance. Since Pauline Cove on the east side of the island provides the only safe deep water harbour between Point Barrow, Alaska, and the Mackenzie Delta, it became the centre for Beaufort whaling activities by American whaling companies. By wintering up to 15 ships at Herschel, Americans were able to obtain a head start on the season in the Beaufort Sea. Between 1892 and 1905 the presence of the whalers had a strong impact on wildlife and the Inuit. Epidemics such as the measles (1902) nearly eliminated the Inuit population and hunting depleted the caribou and bowhead whale numbers. The net result was a tremendous erosion of the native culture. In 1903, an Anglican mission, a trading post (Hudson's Bay Company) and a Royal North West Mounted Police (RNWMP) post were established at the same time as buildings for the whaling companies. This resulted in the settlement of Herschel at Pauline Cove. In 1914 whaling was abandoned, and by 1931 the Hudson's Bay Company decided to return to supplying arctic posts by river; thus, the importance of Herschel Island diminished. The RNWMP maintained a detachment there until 1932. The RNWMP did reopen the post following the Second World War; however, it only lasted for a short period.

The earliest scientific studies of the Herschel Island area were by O'Neil (1924) and Bostock (1948). As part of the 1913-1918 Canadian Arctic Expedition O'Neil made a number of interesting geological observations. For example, he noted a similarity between the Pleistocene marine deposits found on Herschel Island and those covering much of the Yukon Coastal Plain between the Firth River and the Mackenzie Delta. He also proposed the theory that Herschel Island may be a remnant of a former or paleo Mackenzie River Delta. O'Neil (1924) documented the occurrence of widespread erratic boulders on the island's surface and large bodies of ice in its sediments. At approximately the same time, Leffingwell (1919) produced his detailed observations of materials, permafrost and ground ice for part of northern Alaska. Roughly two decades later Bostock (1948) lead a Geological Survey of Canada expedition to the northern Yukon. However, it was the work of J.R. Mackay in the 1950's, as a member of the Geographical Branch (University of British Columbia), that shed light on the origin of Herschel Island. In 1959 Mackay suggested that Herschel Island was the result of glacial tectonic processes. He observed a number of structures in support of this theory, including folds, inclined beds, overlap faults and shear planes. Mackay also noted the presence of erratic boulders at an elevation of approximately 165 m and the occurrence of massive ground ice bodies. Bouchard (1974) conducted a detailed study of the surficial geology of Herschel Island and commented on ground ice conditions and topography. Rampton

(1982) summarized the Quaternary geology of Herschel Island and Yukon Coastal Plain based on his extensive field work as well as that of co-workers. More recently, work on Herschel Island has been concerned with examining soil and vegetation (Smith *et al.*, 1990) and the investigation of ground ice characteristics (Pollard, 1989, 1990).

3.3 Regional Setting

Compared to the rest of the Yukon Coastal Plain, Herschel Island displays a number of distinct characteristics resulting from its glacial origin, exposed position and the ice rich nature of its permafrost. It is interesting to note that Herschel Island is the only sizeable island found along the Yukon and Alaska Coastal Plains west of the Mackenzie Delta. The following will highlight some of its distinctive characteristics.

3.3.1 Physiography

Herschel Island has a crude rhombohedral plan shape that has been streamlined by more than 40,000 years of wave action by the Arctic Ocean. It covers an area of 108 km^2 and has dimensions of approximately 15 km along a north - south axis and 8 km along an east - west axis. It has a maximum elevation of 180 m above sea level, which is curiously higher than other coastal areas of the Yukon Coastal Plain. Approximately 30% of its area is 90 m above sea level (Bouchard, 1974).

The topography of the island displays two marked trends which roughly divide the island in half. The north and north eastern portion is characterized by the highest elevations and the steepest relief. The south and south western portion is lower in elevation and the coast is marked by a gently sloping plain. The morphology of high relief areas is dominated by rolling and hummocky moraines which form a series of high parallel ridges and hills with narrow asymmetrical valleys in between. The ridges are breached by a series of deep gullies (Figure 3.2) which, in places, form steep valleys up to 45 m deep. The parallel ridge and asymmetrical valley sequence is interpreted as glacially ice-thrust ridges (Mackay, 1959). The coastline of the north-northeastern side of the island is dominated by steep bluffs up to 50 m high. Undercutting of frozen sediments by wave action results in coastal retreat by block failure (Are, 1988). Where coastal sediments contain high ice content numerous retrogressive thaw slumps and active layer slides contribute to rapid bluff retreat.

The leeward side of the island (the southwest, south and east sides) have large aggrading spits composed of coarse gravel, sand and sandy silt. The largest, Avadlek Spit, is roughly 6 km long while Herschel Spit and Osborne Point are each between 2 - 3 km in length. Beaches are best developed

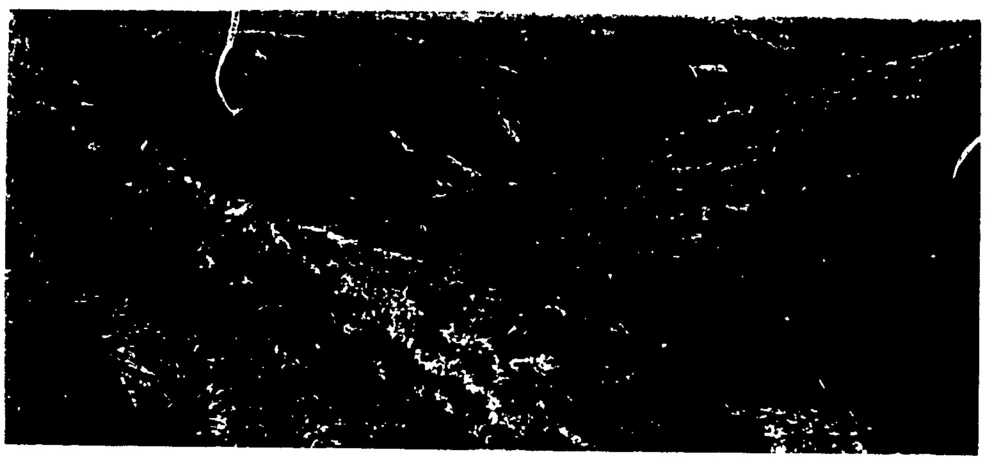


Figure 3.2. Photograph showing gully formation in the vicinity of Collinson Head (July 1987).

along the southwest side of the island, but are still only a few metres wide, contain gravel, and are locally strewn with boulders (Bouchard, 1974). On the southeast side, the shoreline is mantled by a thick accumulation of supersaturated clays, clayey silts and organics. This is a residue from thaw-induced mass wasting and thermokarst activity. Elsewhere, the beaches are generally narrow to non-existent. The island's interior contains lower concentrations of retrogressive thaw slumps and active layer slides, but other forms of thermokarst are present, including melt-out of ice wedges, beaded drainage and thaw lakes. There is also evidence of human-induced thermokarst in the form of gullying along 1960's seismic lines and wheel ruts from a Twin Otter landing in 1987 (Figure 3.3).

3.3.2 Geologic History

Rampton (1982) mapped Herschel Island as part of the Yukon Coastal Plain geologic unit. The Yukon Coastal Plain is considered a pediment, formed by the parallel retreat of escarpments in areas of arid or semi-arid non-periglacial climatic conditions. The pediment surface of the coastal plain slopes gently from the Barn and British Mountains towards the coast, where the surface is covered by a thick mantle of unconsolidated glacial, fluvial-glacial and lacustrine deposits. The age of the erosional surface of the Yukon coast is uncertain and difficult to determine. Hughes

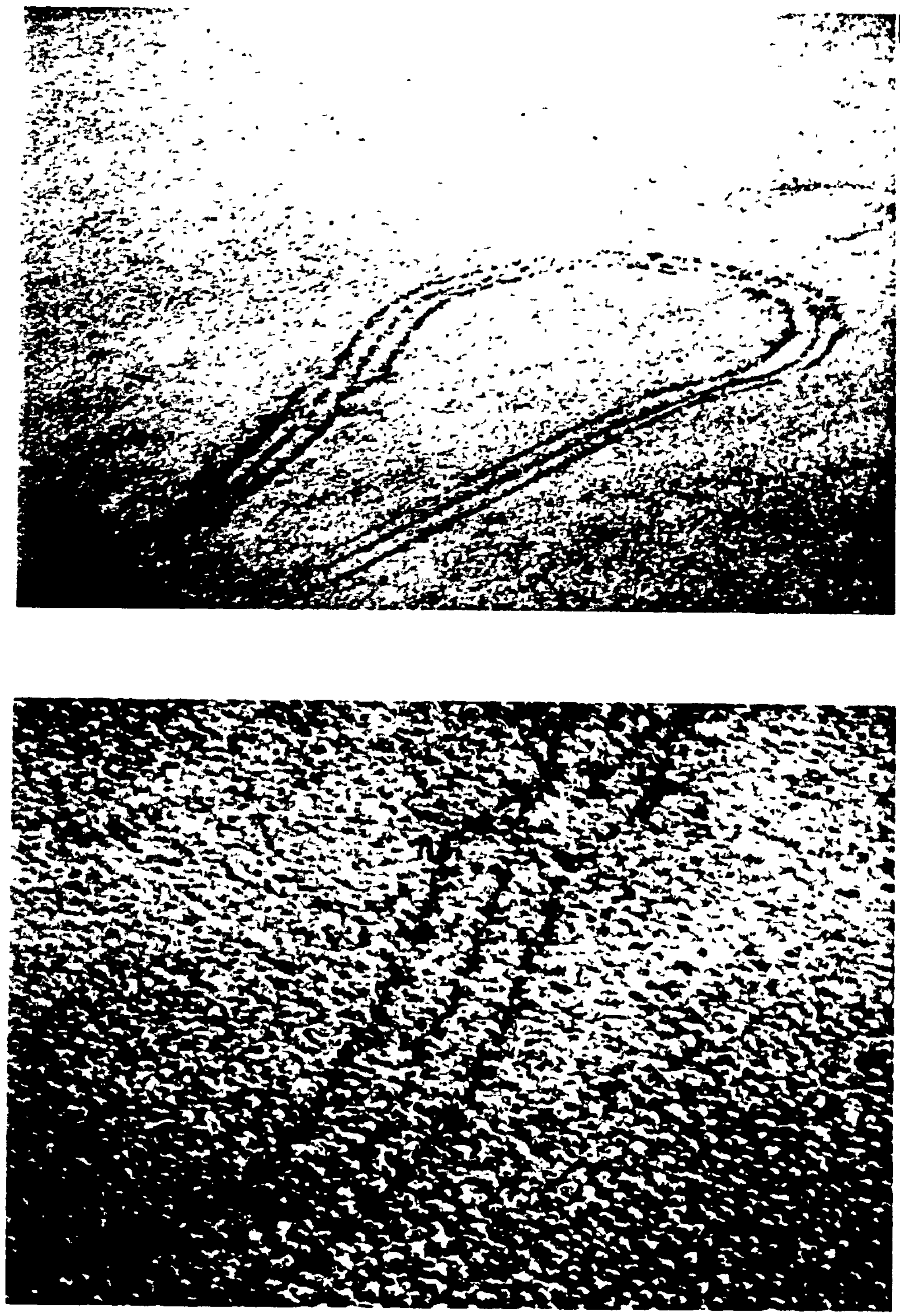


Figure 3.3. Photographs of thermokarst as a result of wheel ruts caused by the landing of a Twin Otter in the late summer of 1987.

(1972) noted similar surfaces throughout much of the Yukon (e.g., Klondike Plateau and Eagle Plains), but has been unable to determine their age or the climatic conditions which produced them. However, it is apparent that they formed under conditions very different from those which prevail today (Hughes, 1972). It is believed that the age of the Yukon Coastal Plain postdates regional tilting and faulting associated with the Laramide Orogeny of the late Cretaceous. This is evident where stream channels and a fault escarpment incise the surface of its northwest slopes, near the Mackenzie Delta (Rampton, 1982). The maximum age of this surface is defined by its youngest rock, the Aklak Member of the Reindeer Formation which is late Cretaceous or early Tertiary (Young, 1975). Rampton (1982) believes that all evidence points to a late to middle Tertiary age for the formation of this surface because climatic conditions are known to have been warmer than those of the present.

Generally, Herschel Island owes its origin to Pleistocene glaciotectonic activity. Topographic and structural characteristics of the island and the Mackenzie Basin support a glacier ice-thrust hypothesis first proposed by Mackay (1959). In theory, a lobe of the Laurentide ice sheet (Buckland Glaciation) moved northwest ward from the Mackenzie Valley and excavated material from the current position of Herschel Basin more than 40,000 years B.P. (Rampton, 1982). As the lobe impinged upon the Yukon coast it produced an ice-thrust moraine complex which currently forms the main body of Herschel Island (Bouchard, 1974). Evidence in support of this theory includes (1) the presence of a submarine ridge of probable glacial origin 40 km west of Herschel Island (Rampton, 1982), (2) the volume of sediments missing from Herschel Basin located southeast of the island is proportional to the size of the island (Mackay, 1959), and (3) the deformed nature of the sediments is believed to have been caused by the overriding of Pleistocene glaciers (Mackay, 1959; Mackay and Mathews, 1964; Mathews and Mackay, 1960). Deformations are preserved in frozen sediments as folds, inclined beds, thrust fault planes and simple shear planes. These structures are readily exposed along stream cuts and coastal bluffs, and in the surface of ridges along the northwestern side of the island (Bouchard, 1974). Some deformation of sediments may have also occurred as permafrost aggraded into earth materials exposed subaerially following deglaciation. Other evidence supporting a glacial origin is the presence of erratic boulders, meltwater drainage channels, glacial fluvial deposits and glacial till (Bostock, 1948; Mackay, 1959, 1963; O'Neil, 1924; Rampton, 1970, 1973).

Deposits predating glaciation are found at the base of coastal exposures on the north and west sides of Herschel Island and along the coastal plain toward the west. Bouchard (1974) classifies the pre-glacial deposits into three categories: (1) marine sediments, containing shells or fragments of shells, (2) mixed sediments, containing occasional shells and detrital terrestrial organics, and (3) non-marine, containing only organic materials. The marine unit is a brown and grey silt with sand laminae, which ranges from 1.2 m to more than 10 m in thickness. The mixed sediments consist of interbedded brown to grey silty clays and fine-grained sands with pebbles, organic detritus, wood and shell fragments. The mixed sediment unit is up to 20 m thick. Bouchard's non-marine pre-Buckland unit consists of 60 cm of peat underlain by sand and silt. Paleoecological studies from the northern edge of Thetis Bay indicate that deposition of marine sediments took place under relatively shallow nearshore conditions of an arctic region (Rampton, 1982). Naylor *et al* (1972) analyses of molluscs from the same section suggests that marine conditions were similar to those of the present, and that the mixed sediments were deposited in brackish shallow marine waters during a period of marine transgression. Quaternary sea levels for this region are perhaps the least documented and most poorly understood of all arctic coastal regions in Canada (Forbes, 1980; Hill, 1985).

3.3.3 Surficial Deposits

The surficial geology of Herschel Island, as well as materials from the Mackenzie Basin support the glacier ice-thrust origin proposed by Mackay (1959). The general absence of till from the Herschel Island geology is somewhat problematic; however, the widespread bouldery clay diamicton and erratics may be interpreted as reworked till. Since the mid-Wisconsin, the topography and surficial deposits have been reworked by permafrost aggradation, thermokarst and slope processes. The majority of research undertaken on the geology and geomorphology of the Yukon Coastal Plain (e.g., Bouchard, 1974; Forbes, 1980; Mackay, 1959; Pollard, 1989, 1990; Rampton, 1982) have focused primarily on processes and events that post-date glaciation.

Post-glacial materials form the uppermost geologic materials on Herschel Island, and include a combination of littoral, alluvial and colluvial deposits, as well as relatively thick accumulations of peat (Bouchard, 1974). Littoral deposits form a number of prominent coastal landforms such as Avadlek Spit, Osborne Spit, Herschel Spit and Lopez Point. With the exception of these areas of deposition the shoreline of Herschel Island is dominated by steep bluffs and narrow ephemeral beaches. Sands and clayey silts with varying amounts of organic matter are found on the upper parts of the ephemeral beaches along the southwest coast of the island. Alluvial sediments containing clayey silt, silt, sand and gravel are found along the southwest side of the island. Bouchard (1974) documented the presence of a large alluvial fan near the base of Herschel Spit. This fan consists predominantly of interbedded silt and fine sand, and may also cont n ice lenses (Bouchard, 1974). Colluvial materials consist of a widespread surface diamicton (Bouchard, 1974). This layer is up to 1 m in thickness and consists primarily of pebbles and silty-clay sediment, varying in colour from light brown to dark grey. Other colluvial deposits are the result of mass wasting and thermokarst activity. Organic sediments occur as peat layers and wood fragments, most of which are associated with ice wedge polygons. Bouchard (1974) described peat bogs in the eastern section of the island 180 m in length with ice-wedge polygon dimensions of 25 m X 35 m.

3.3.4 Regional Climate

Climate is the single most important variable influencing permafrost and thermokarst processes (Brown, 1963a, 1973; Gold and Lachenbruch, 1973). The climate of Herschel Island clearly reflects the combined influence of the Arctic Ocean, the close proximity of the island to the continental landmass and the regional physiography. Climate data are collected at two locations on the Yukon Coastal Plain, namely, Shingle Point (Figure 3.4) and Komakuk Beach. These Distant Early Warning meteorological stations have been in operation since 1957 and 1961, respectively. There are also climate records from Herschel Island for the periods 1896-1918, 1974-1975 and 1986-1988; however, these records are very fragmented and provide little useful information. The harsh, cold arctic climate which characterizes the Yukon Coastal Plain is

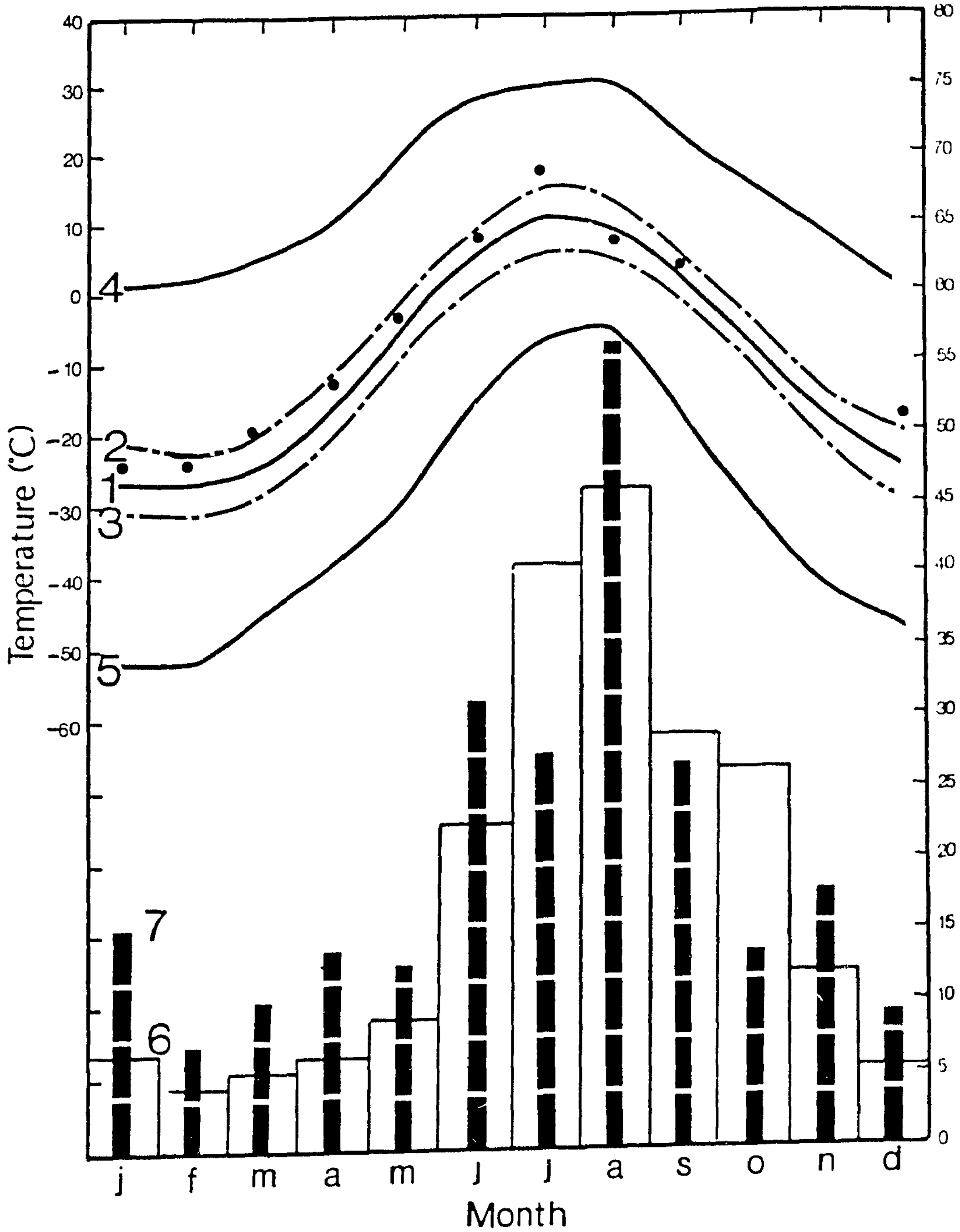


Figure 3.4 Temperature and precipitation data from Shingle Point, Yukon Coastal Plain (1957-1981, and 1988). (1) mean daily temperature, (2) mean daily maximum, (3) mean daily minimum, (4) extreme maximum, (5) extreme minimum, (6) mean total precipitation, and (7) greatest precipitation in 24 hours. The symbol (●) represents mean monthly temperature for 1988. No data is available for the months of October and November (Source: Wahl *et al*, 1987).

generally dominated by continental arctic air in winter and maritime arctic air in summer. The mean annual air temperature is -10°C at Shingle Point and -11° C at Komakuk Beach. Mean daily temperatures for Komakuk Beach average 8°C and -28°C, and at Shingle Point 11°C and -27°C, for the months of July and January, respectively. Maximum recorded temperatures rarely exceed 25-26°C on the Yukon Coastal Plain, while the extreme minimum temperature generally is about -50°C (Wahl *et al.*, 1987). No ground (subsurface) temperatures are available for Herschel Island. However, in the Richard's Island and Tuktoyaktuk region (approximately 200 km east of Herschel Island) the average annual temperature of the ground 15 m below the surface is approximately -7 to -10°C (Mackay and Mackay, 1972). Ground temperatures along the Yukon Coastal Plain, including Herschel Island, are probably quite similar. The total annual precipitation is less than 20 cm, most of which falls as rain or drizzle during the short summer when the Beaufort Sea is ice-free. Snow cover averages 50 cm on the coastal plain (Rampton, 1982), but on Herschel Island it varies from a few centimetres on exposed ridges to 2-3 m in narrow valleys and in the lee of major obstructions. Winds are an important element of the regional climate, with a dominant direction roughly parallel to the Yukon coast - from the west to the northwest. Its exposed position along the Yukon Coast means it is highly susceptible to wind activity.

3.3.5 Permafrost and Ground Ice

The Yukon Coastal Plain is underlain by continuous permafrost (Brown, 1978, Figure 2.2) which is believed to have existed prior to glaciation. Permafrost thickness is in excess of 300 m (Judge, 1973), although it is thinner under active or recently drained lake basins and may be absent under aggrading spits such as Herschel Spit. The age of permafrost and ground ice of Herschel Island are believed to be quite similar to the rest of the Yukon Coastal Plain; however, insufficient research has been done to confirm this.

On Herschel Island ground ice is an important constituent of permafrost materials. It occurs mainly in the form of ice-wedge ice, epigenetic and syngenetic segregated ice layers, ice lenses and massive tabular ice bodies, pore ice, buried snowbank and glacier ice (Pollard, 1989, 1990). The island also exhibits periglacial features such as tundra tussocks, patterned ground, frost mound phenomena (palsas), solifluction and frost creep.

Air photo analyses and field reconnaissance revealed that ice wedges are widespread on Herschel Island and can be found in various stages of development. In North America, few ice wedges have been reported extending more than 7-8 m below the ground surface or exceeding 3 m in width. In the coastal lowlands of Siberia syngenetic ice wedges frequently exceed 5 m in width and 40-50 m in depth (Harry *et al.*, 1985;

Leffingwell, 1919). On Herschel Island, ice-wedge ice exposed in the headwall of a retrogressive thaw slump measured 3 m wide and 10 m deep (Figure 3.5). Segregated ice exposed in slump and coastal sections range from ice lenses a few centimetres thick to horizontal massive tabular ice bodies 4-5 m thick. Pore ice is found in most exposures, but its content varies considerably. Buried snowbank ice occurs in areas of previous thermokarst activity and is limited to near-surface deposits. Buried glacier ice was tentatively described in areas of glacially deformed and ice-thrust sediments (Pollard, 1990).

Bouchard (1974) mapped the depth and distribution of ground ice based upon field evidence and drill logs (Figure 3.6). Sixty-one drill logs were compiled for Herschel Island - the deepest being 35 m in length - with an average depth of 25 m. Bouchard found that most massive ground ice encountered was at depths of 5 m and at 18 m. For 15 cores, the drill did not completely pierce the massive ice body or icy sediments. For the 12 cores that did cut through massive ice and icy sediments the materials immediately below were either dry or ice-poor. Drill logs have revealed that the average maximum depth of the massive ice and ice-rich sediment is approximately 13 m, whereas on the adjacent Yukon Coastal Plain it is approximately 5-9 m (Rampton, 1971). Cores which encountered deep massive ice where commonly overlain by a clay or a clayey gravel unit, probably the main unit forming the

Figure 3.5. Photograph of ice-wedge ice, Thetis Bay, Herschel Island (July 1987).

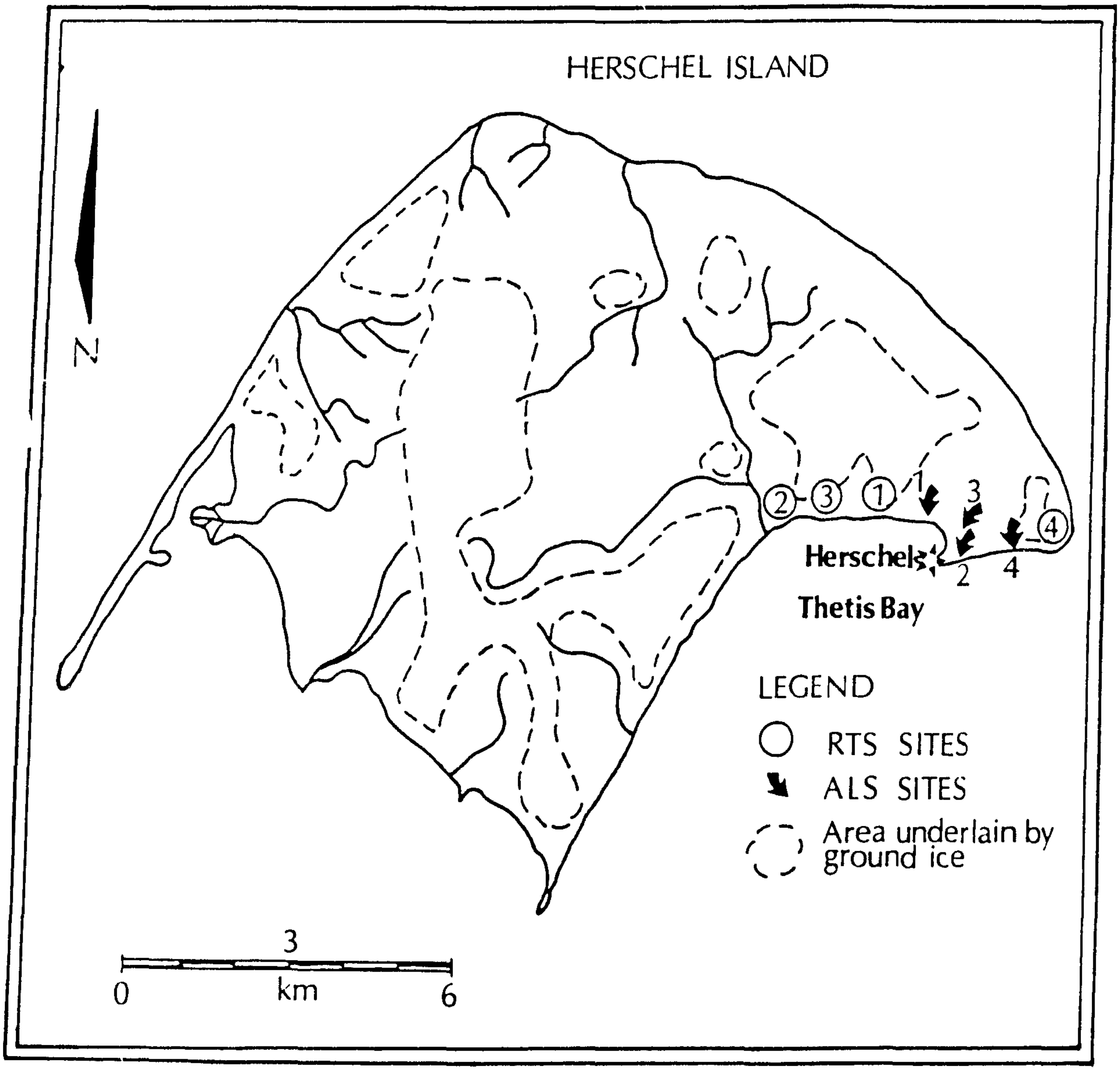


Figure 3.6. Study site location map. The map shows the location of RTS 1-4 and ALS 1-4. It also illustrates the distribution of ground ice on Herschel Island, after Bouchard (1974), revised by Pollard (1989).

preglacial materials.

3.4 Study Sites

Thetis Bay was selected as the primary study area because both retrogressive thaw slumps and active layer slides occur in close proximity. Also, based on the presence of numerous stabilized and revegetated scars, there appears to be a long history of these processes. Another reason for this decision was the accessibility of this area from the base camp on Herschel Island. Within this thesis retrogressive thaw slump sites have been abbreviated to RTS (RTS 1-4). Although the geomorphic accuracy of the term "slump" is questionable, it is retained throughout this thesis since it is recommended by the Permafrost Subcommittee (1988), and also for the purpose of internal consistency.

Three retrogressive thaw slumps up to 3.5 km west of the camp were selected for detailed investigation. These slumps occur on south - facing slopes adjacent to Thetis Bay. Two thaw slumps were chosen for detailed investigation (RTS 1 and 2) and a third (RTS 3) for less intensive study. Another retrogressive thaw slump located near Collinson Head (RTS 4) is included in the discussion of permafrost stratigraphy, ice content and geochemistry. RTS 4 is located 2.5 km east of the Herschel base camp in an area of ice-push moraine, steep coastal bluffs and deep drainage channels (Figure 3.6).

The four areas of active layer slide activity examined are in the vicinity of Thetis Bay and Collinson Head. Generally, active layer slides are found along seepage lines in areas experiencing various types of mass wasting (e.g., creep, solifluction) (Figure 3.6). Within this thesis active layer slide sites have been abbreviated to ALS (ALS 1-4).

3.5 Methodology

Research methodologies used in this thesis are divided into two components: first, the collection of field data on the physical characteristics of retrogressive thaw slumps and active layer slides, and factors which influence their occurrence, and, secondly, a laboratory component concerned with the analysis of samples obtained in the field. Fieldwork for this study was carried out during June-July 1988, April and June 1989.

3.5.1 Field Work

Field work focused on three aspects of retrogressive thaw slump and active layer slide occurrence: morphology, material properties and ground ice characteristics, and process. Examination of morphology involved the mapping of location, measurement of dimensional parameters and examination of topographic conditions. Detailed surveys and slope profiling of slump and slide form and the surrounding terrain was undertaken using a TO-184244 Wild theodolite, 50 m survey tape, range pole and Abney level. Examination of slump processes involved installation of a grid of survey stakes around the slump headwall with regular measurement of headwall position in order to record rates of retreat. Survey stakes installed within the slump floor allowed determination of the rates of movement within the slump floor. Examination of processes involved in slide occurrence was attempted by reconstructing processes based upon the resultant landform, "an eye witness" account of a slide occurrence and laboratory analyses. Although technical and logistical support was not available to undertake a complete micrometeorological study, daily air temperature was recorded at base camp, as well as inside and outside of the slump bowl at RTS 1. Several types of ground ice were documented on Herschel Island; however, this thesis only focuses on three ice-rich slumps containing massive ground ice. Ground ice investigations involved (1) the description of ground ice types and stratigraphic relationships, and (2) sampling at 25 cm intervals for calculation of ice content and geochemical analyses. Sediment samples were collected from both thaw slumps and active layer slides for geotechnical analyses.

3.5.2 Laboratory Work

Laboratory analyses focused on (1) the measurement of soil properties, including Atterberg Limits and grain size distribution, and (2) the analysis of ground ice, including the calculation of moisture content and geochemistry.

Atterberg Limits provide a measure of the rheological or flow properties of a soil, and were determined for both thaw slumps and active layer slides. The limits are defined by the soil water content required to produce a specified resistance to deformation.

The liquid limit of a soil is the water content (expressed as a percentage of the weight of the oven dried soil) at the boundary between the liquid and plastic states. Using a Casagrande ASTMD 423-66 instrument, the soil sample is molded wet, placed in a bowl, and levelled parallel to the base. A grooving tool is used to separate the soil mass into two parts. A cam allows the bowl to fall from a predetermined height at a rate of 2 drops/sec on to a standard hard rubber base. After several turns, the two parts of the soil mass flow together and join for at least 25 mm. This procedure was repeated three to four times per sample to obtain readings between 15 and 35 drops. Water content was calculated for each test as a percent dry weight of soil. Results were graphed with water content on the y-axis (arithmetic scale) and number of drops on the x-axis (semi-logarithmic scale). A best fit line was then plotted through the points. The liquid limit is read as the water content corresponding to the point of intersection of the twenty-five drop ordinate with the best fit line (Wray, 1986).

The plastic limit of a soil sample is the water content at the limit between plastic and semi-solid states. The plastic limit is defined as the lowest water content at which soil can be rolled into threads approximately 3 mm in diameter without the threads breaking into pieces (Wray, 1986). Three tests were made per sample and the average was taken.

Knowledge of the size of soil particles forming a certain soil type, and their relative proportion in the soil mass, is used to determine how a soil will behave with respect to shear strength, settlement, permeability and susceptibility to frost action and ice segregation processes. Grain size analyses were undertaken to assess the importance of soil type and soil size on the occurrence of active layer slides, retrogressive thaw slumps and ground ice. Samples were taken from the headwall and floor of both landforms, for sieve and hydrometer. Sieve analysis (U.S. numbers 4, 10, 30, 50, 100 and 200) determined the range of particle sizes and their relative distribution for particles greater than and equal to the No. 200 sieve. Hydrometer analysis, using a standard ASTM 152 H hydrometer, determined particle size and distribution of particles which passed through the No. 200 sieve. Procedures discussed in Lambe (1951) and Wray (1986) were followed for this study. Results are plotted in the form of a grain size distribution curve. The analyses of material properties was undertaken at the Centre for Geotechnical Studies laboratory, McGill University.

Ice content is an important variable influencing the behaviour of permafrost sediments upon thawing and are expressed as a percentage of the dry weight of soil. It is obtained by measuring the weight of the frozen sample, subtracting the weight of the oven dried sample (i.e., remaining sediment), and dividing by the weight of the dry soil. Samples were thawed, filtered and oven dried in the soils laboratory at the Inuvik Research Laboratory, Inuvik, N.W.T.

Ice geochemistry provides an indication of the origin of water which forms the ground ice. Samples for each of the massive ground ice types were analysed for major ion concentrations. Water from ice samples was filtered through Whatman GF/C paper and stored between 2 - 4°C. Electrical conductance and pH measurements were made on a Yellow Springs model 32 conductance meter and a Fisher Accumet pH model 210 meter. Cations (Ca, K, Na, Mg) were determined using atomic absorption spectrophotometry, anions (Cl, SO_4) through ion chromatography. Geochemical analyses were undertaken at the Department of Renewable Resources, MacDonald Campus, McGill University.

3.6 Organization of Results

In the investigation of any aspect of the physical environment there is a distinction between form and process (Young, 1972). This notion is fundamental to geomorphology, as is the understanding that "form" refers to the morphology, or shape and structure, of a portion of the land surface at a given moment in time, and that "process" refers to the agents active in producing a change in form. Traditionally, geomorphologists often studied form and process independently. This segregation is no more apparent than in early slope and mass wasting studies, in which the focus on form gave rise to a distinct body of morphometric literature (Crozier, 1973; Strahler, 1950), and the study of process concentrated more on slope stability analysis and the investigation of slope failure as a function of material properties (e.g., Carson, 1975). However, the current emphasis on process geomorphology has forced researchers to resolve geomorphic problems within a dynamic framework in which form and process are considered together. The remainder of this thesis presents the results and discussion of a geomorphic investigation of retrogressive thaw slumps and active layer slides by first presenting characteristics of form and setting (Chapter 4). Since environmental conditions such as geology, climate and vegetation influence form (Young, 1972) then form, *sensu lato*, is not restricted to the ground surface. Because permafrost and ground ice are the two most important controls influencing thaw slump and active layer slide occurrence, then the next body of observations is concerned with cryostratigraphic controls and material properties (Chapter 5). In many cases, process is inferred from a combination of morphologic and material property analyses, particularly in the study of rapid forms of mass movement such as slides and slumps, for which direct process measurements are few. However, in the present study the predictable nature of, at least, thaw slumps has allowed for some direct observation of process (Chapter 6). Several conclusion can be drawn from this study, they are the focus of Chapter 7.

Chapter 4

Setting and Morphology

4.1 Introduction

Despite the geomorphic differences between retrogressive thaw slumps and active layer slides, or slump and slide phenomena discussed in the literature, the traditional morphometric approach to landform analysis retains considerable merit and is used in this study for several reasons, including:

(1) it provides a basis for the application and assessment of standard morphometric indices,

(2) it allows for direct comparison of measurements with other studies using one or the other approach,

(3) it provides the framework for an analysis of form and process relationships.

In this study, morphologic investigations focus primarily on the dimensional characteristics of the landforms and slope characteristics. Process studies focus on changes in form over time as well as the dynamics of this change. The latter involves analysis of several parameters (e.g., air temperature, moisture content) that directly influence the frequency and magnitude of process. Since the landforms and processes under investigation are thaw-induced or thaw-related, then the analysis of material characteristics must focus upon permafrost and ground ice conditions.

This chapter presents results on the morphology of retrogressive thaw slumps (RTS 1 and 2) and active layer slides (ALS 1-4). Retrogressive thaw slumps 1 and 2 are similar in morphologic character and are therefore addressed collectively.

4.2 Retrogressive Thaw Slumps

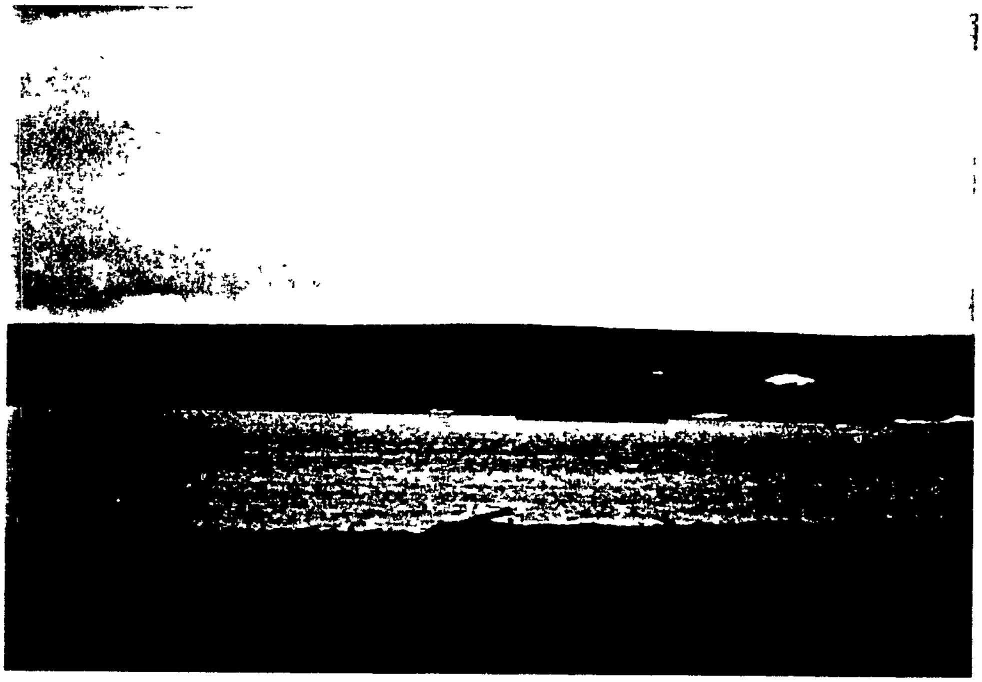
Setting

Three of the retrogressive thaw slumps studied (RTS 1, RTS 2, RTS 3) are situated on south to southeast - facing slopes at 22-30 m above sea level, and up to 400 m inland from Thetis Bay. The terrain surrounding these sites is characterized by rolling hills dissected by deep drainage channels. Retrogressive thaw slumps RTS 1-3 all occur in an area mapped by Bouchard (1974) as underlain by massive ground ice (Figure 4.1.a). Site RTS 4 is situated in an area mapped as an ice-push moraine and is characterized by steep coastal bluffs and a deep drainage channel which separates two ice push ridges (Rampton, 1982) (Figure 4.1.b).

Morphology

The morphology of retrogressive thaw slumps bear a resemblance to that of landslides found in more temperate environments (Ritter, 1978; Varnes, 1958, 1978). However, apart from a resemblance the two landforms have little in common. Since retrogressive thaw slumps form by a slowly

A



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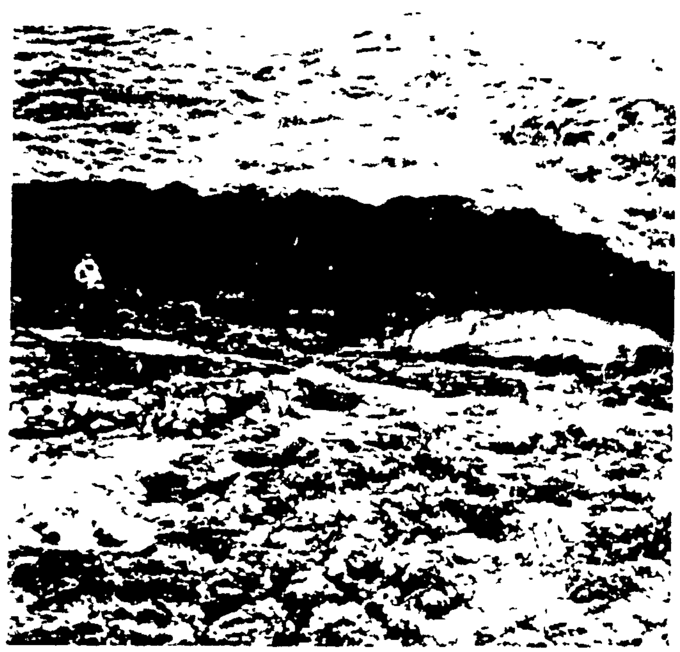


Figure 4.1. (a) A series of four retrogressive thaw slumps, Thetis Bay (July 1987), and (b) retrogressive thaw slump, Collinson Head (July 1987).

retreating headwall, they lack many of the structures indicative of rapid slope failure. For example, the rupture plane, the rotational failure plane, tension cracks in the parent slope above the crown and the displaced mass are all absent (Figure 4.2). However, retrogressive thaw slumps can be divided into three parts (1) the active layer, which is a component of all permafrost areas, (2) a headscarp consisting of a thermal erosional belt which retreats retrogressively by melting of exposed ground ice and ice-rich permafrost, and (3) the slump floor which consists of residual thaw material and flow deposits. The headwall is the most dynamic part of the retrogressive thaw slump. The processes in thaw slump development will be discussed in Chapter 6.

Both RTS 1 and RTS 2 display the classic amphitheatre or bowl shape; however, plan dimensions of other slumps around Thetis Bay are variable, as are those found along the Yukon Coastal Plain. For example, coastal erosion at King Point exposed a body of massive ground ice more than 1.3 km in lateral extent (Figure 4.3).

Retrogressive thaw slump 1 was the larger of the two slumps surveyed in June 1988. It was 194 m long, 67 m wide, and had a headwall ranging between 5-7 m in height. The total area covered by RTS 1 was 13,000 m^2. RTS 2 measured 83 m in length, 76 m in width and 5-7 m in height. The total area covered by this slump was 6,300 m^2. In 1989, RTS 1 measured 204 m in length and 70 m in width (14,280 m^2), and RTS 2 was

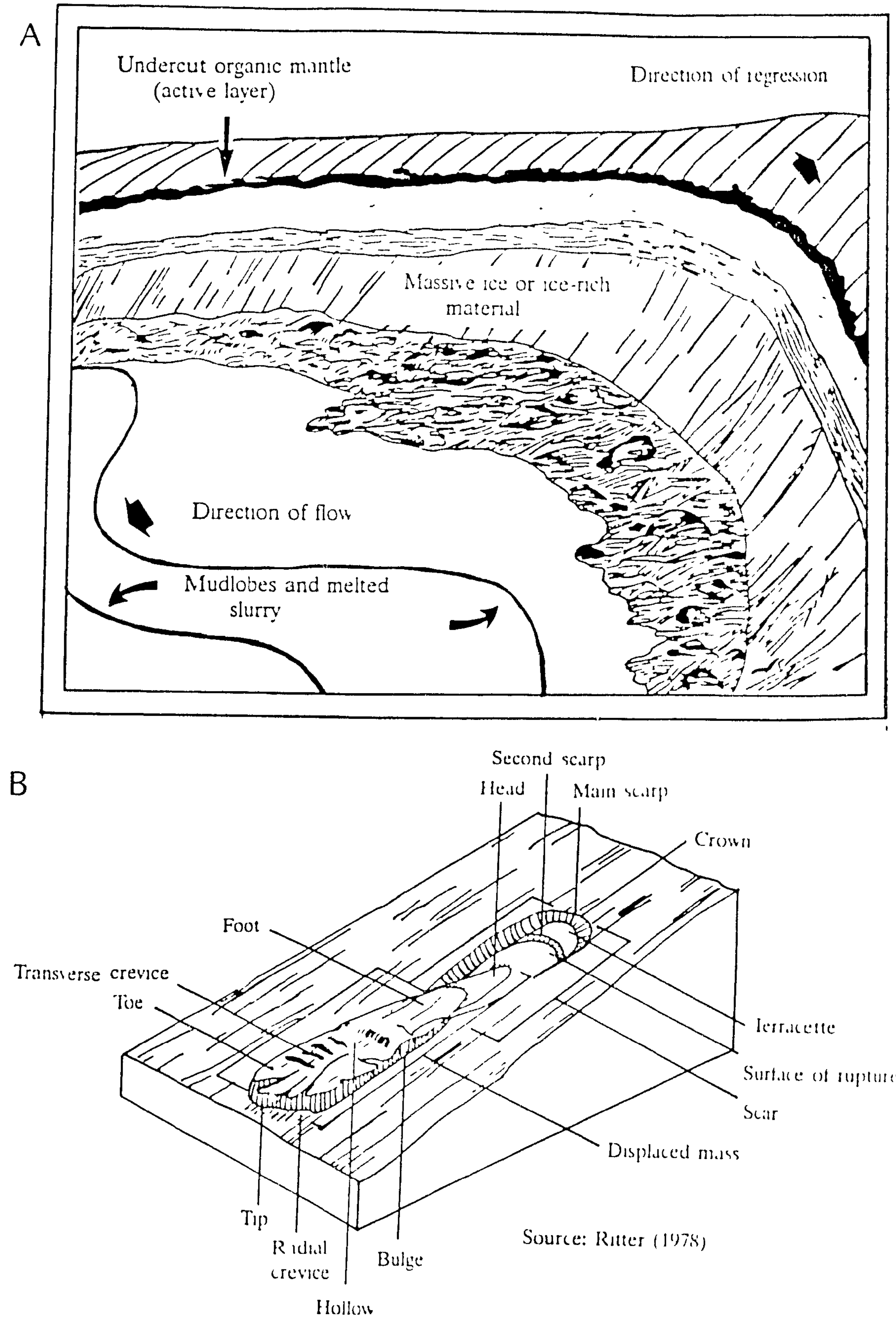
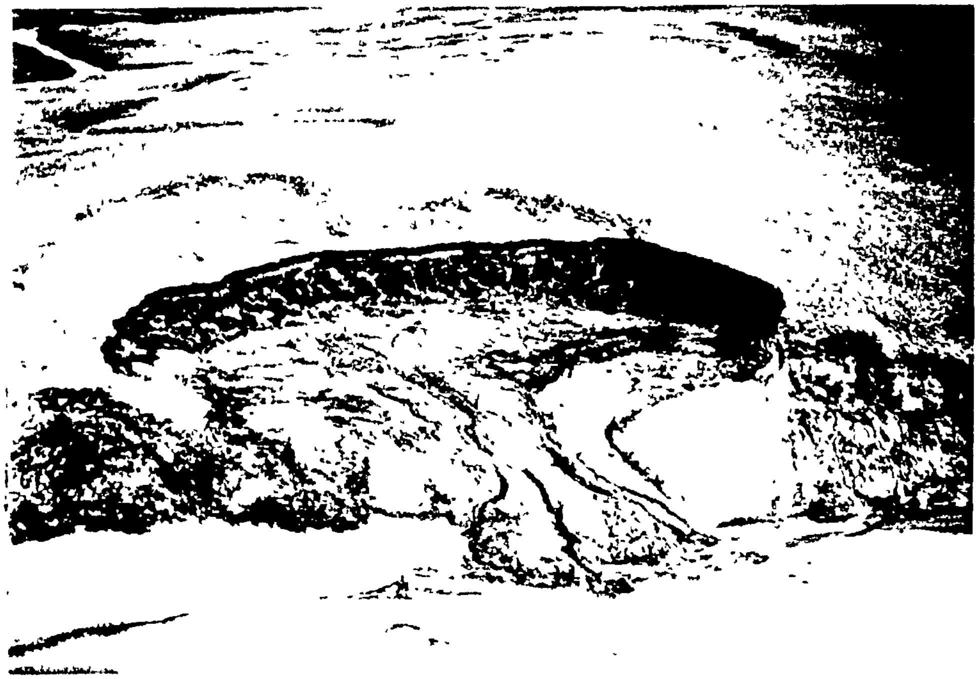


Figure 4.2. Schematic representations of (a) retrogressive thaw slumps, and (b) landslip (Ritter, 1978).

A



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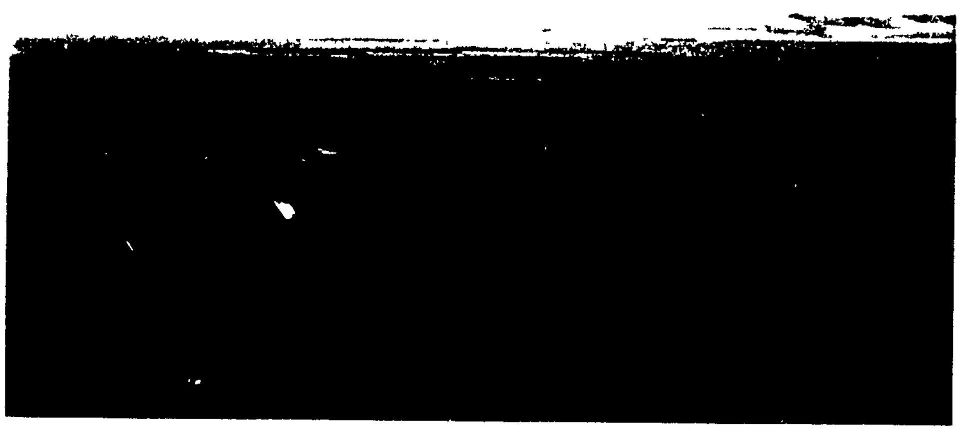


Figure 4.3 Photographs illustrating (a) classic bowl shape of RTS 1 (July 1987), and (b) the lateral headwall at King Point (July 1987).

92 m in length and 86 m in width (7,900 m^2). In the vicinity of RTS 1 and RTS 2 there are several slumps of variable size, with lengths up to 300 m. Based on 1970 air photos (A21921 102-163) taken at an elevation of 7000 m ASL with a lens of 6", the mean size of 33 slumps on Herschel Island was 10,120 m^2, averaging 110 m X 92 m (Table 4.1). This range in size is comparable to that for slumps documented by MacKay (1966) in the Mackenzie Delta and French and Egginton (1973) for Banks Island. However, a number of the slumps on Herschel Island exceed the maximum dimensions reported for those areas (200 m X 300 m). For example, based on the 1970 air photos, RTS 3 was 245 m long and 230 m wide. Helicopter reconnaissance along the Yukon coast revealed only one slump with greater dimensions. At King Point, a retrogressive thaw slump had a headwall 1.3 km long and was roughly 60 m deep (Harry *et al*., 1985; Rampton, 1982) (Figure 4.3.b).

The floor at the base of slump headwalls is characterized by supersaturated soils and pools of standing water. The microtopography downslope from the headwall is very irregular. It contains turf blocks, slumped debris, mud lobes and runoff channels which frequently form a dendritic pattern (Figure 4.4).

Morphometric Measurements of Retrogressive Thaw Slumps
on Herschel Island, Yukon.

Slump Number	Aspect°	Length(m)	Width(m)	Area(m^2)
1	180	193.75	66.50	12,998
2	180	82.60	76.0	6,277
3	170	244.80	231.20	56,597
4	250	190.40	122.40	23,304
5	250	95.20	149.60	14,241
6	250	81.60	102.00	8,323
7	260	91.80	68.00	6,242
8	250	98.60	68.00	6,704
9	240	71.40	68.00	4,855
10	120	84.32	54.40	4,587
11	350	74.80	54.40	4,069
12	120	74.80	47.60	3,560
13	150	149.60	102.00	15,259
14	0	122.40	68.00	8,323
15	10	115.60	47.60	5,502
16	270	69.36	40.80	2,829
17	280	149.60	108.80	16,276
18	330	129.20	95.20	12,299
19	300	136.00	122.40	16,646
20	325	115.60	183.60	21,224
21	300	98.60	149.60	14,750
22	265	224.40	136.00	30,518
23	270	40.80	54.40	2,219
24	345	95.20	78.20	7,444
25	80	95.20	57.12	5,437
26	35	149.60	88.40	13,224
27	355	95.20	40.80	3,884
28	360	68.00	108.80	7,398
29	10	27.20	95.20	2,594
30	180	54.40	54.40	2,959
31	180	61.20	61.20	3,745
32	90	108.80	149.60	37,818
33	90	126.40	94.80	11,982

Table 4.1. Morphometric parameters for 33 slumps found on Herschel Island.

A

B

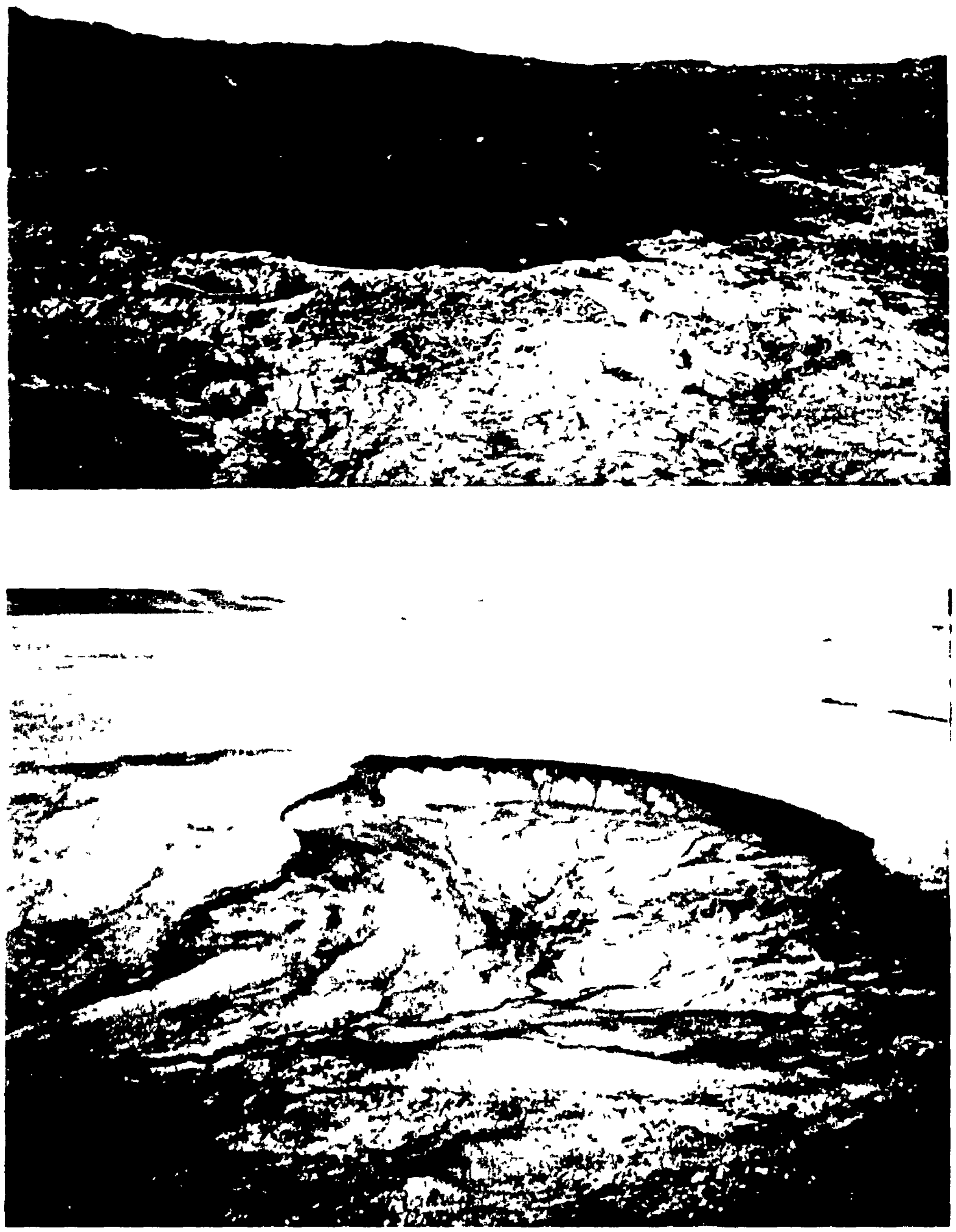
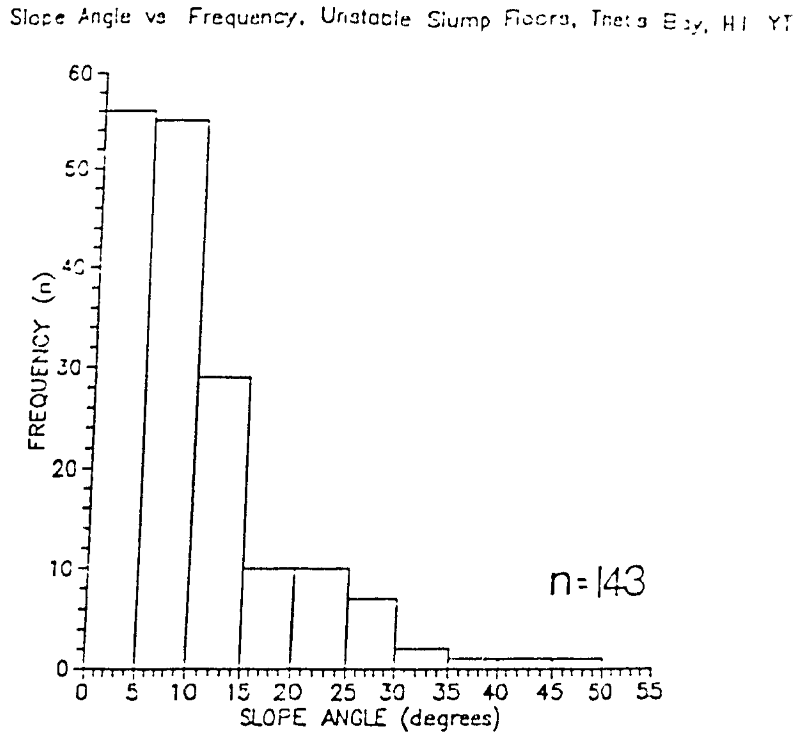


Figure 4.4. Photographs illustrating characteristics of slump floors: (a) mudlobes near the headscarp, and (b) the dendritic pattern of the lower slump floor.

Slope Conditions

Slope conditions constitute an important component of form analysis. Information on slope form is often treated as a source of evidence for process and landform evolution. A series of slope profiles and a large number of slope angles were obtained for undisturbed surfaces above and beside the slumps and for disturbed surfaces within slump bowls (Figure 4.5). Slope angles within slump bowls are highly variable. In the case of RTS 1, values range between 0-40°, with an average of 9° and a standard deviation of 9.3°. For RTS 2, values are between 0-35°, with an average of 7° and a standard deviation of 7.5°. Steepest slopes (up to 40°) are found at the base of the slumps where a coastal bluff is actively being eroded by wave action, thermokarst, thermal erosion, and mass wasting. The slump headwalls have low to moderate slope angles, with a maximum value recorded of 35°. Lower slope angles (5-10°) were recorded for stable sidewalls. The scarp face at these sites are concave, except for areas at RTS 2 where there is oversteepening of sidewalls. Slope angles of terrain adjacent to slump sites and above the active headwalls are between 2° and 8°, with an average of 4° and a standard deviation of 1.3° (Figure 4.6 and Table 4.2).

A



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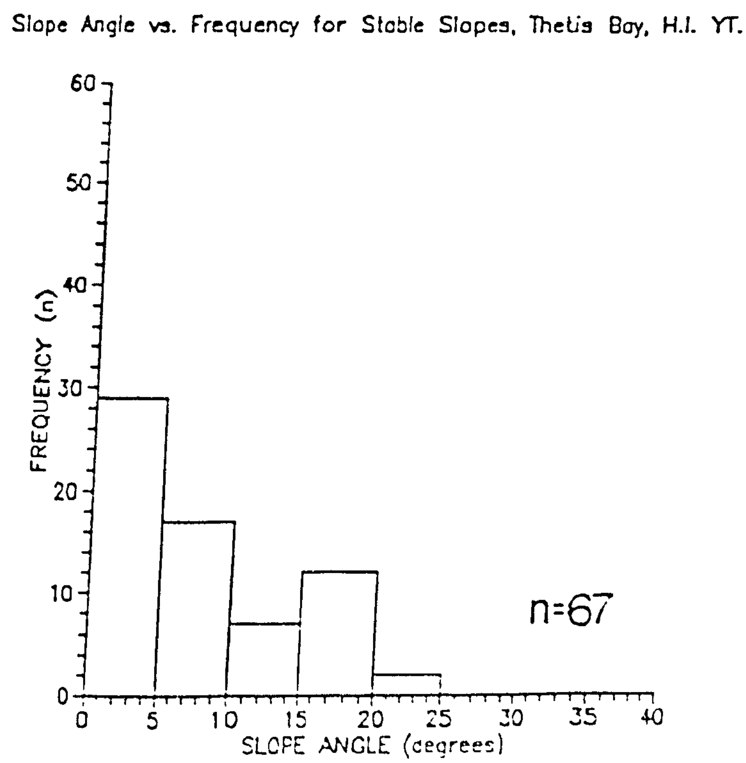


Figure 4.5. Frequency histograms for (a) disturbed slopes within thaw slump floors, and (b) undisturbed slopes surrounding thaw slumps.

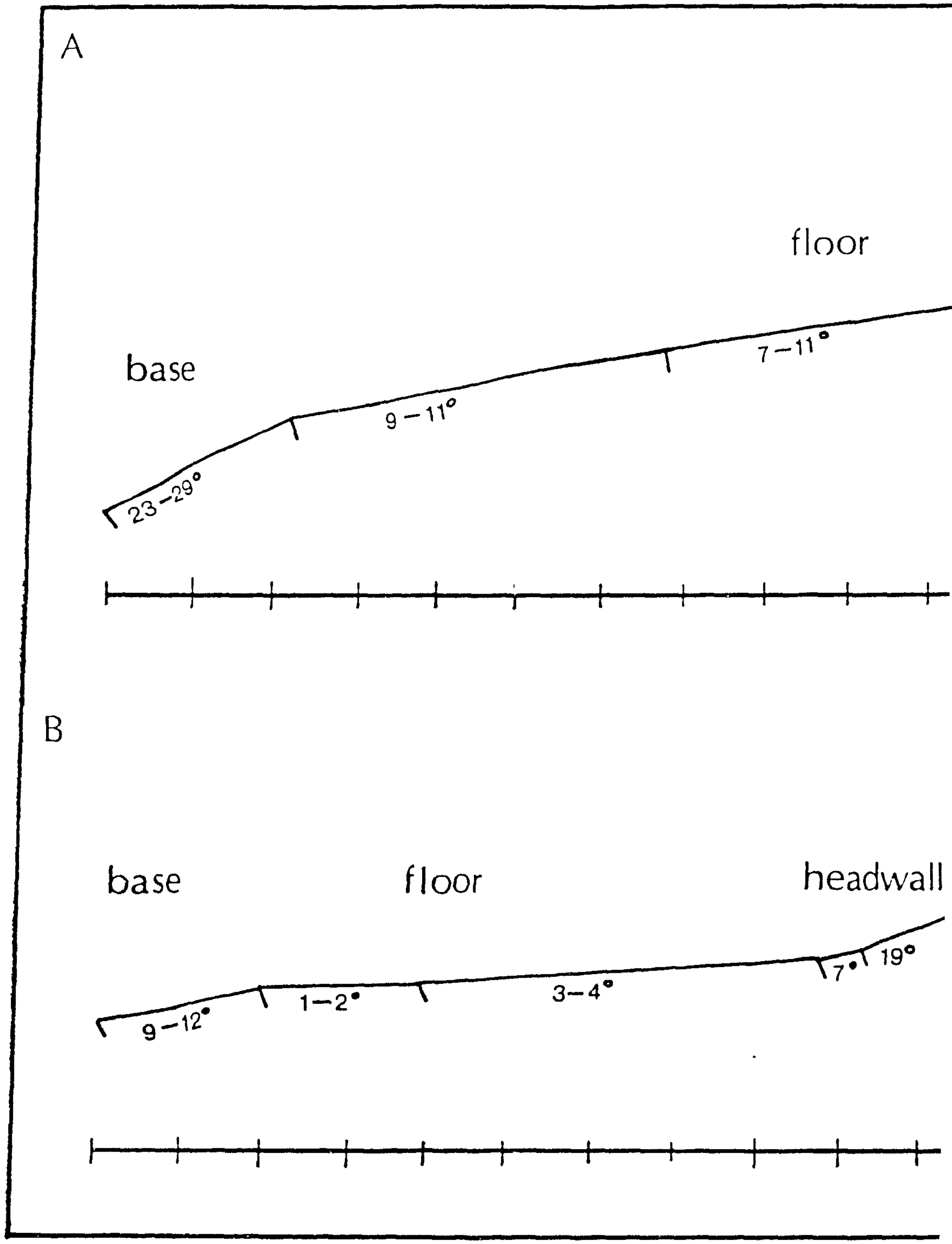
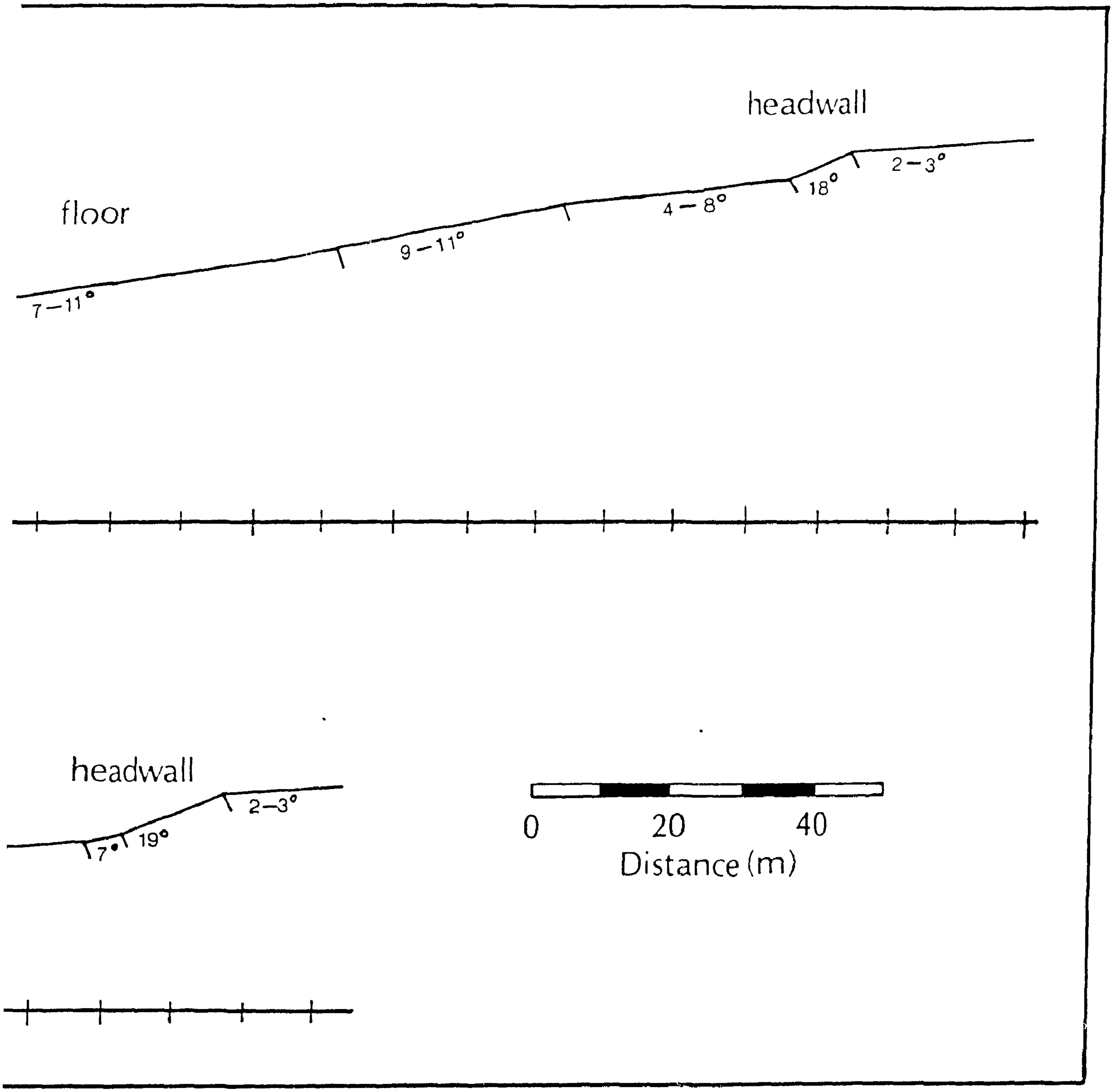


Figure 4.6. Slope profiles for (a) RTS 1, and (b) RTS 2.


headwall
floor
2–3°
18°
4–8°
9–11°
7–11°
headwall
2–3°
7°
19°
0
20
40
Distance (m)

l (b) RTS 2.

Site	Total Number of Readings	Min	Max	Average	STD	Variance
RTS 1	70	1	40	9.3	9.3	86
RTS 2	49	1	31	6.8	7.5	56.4
ALS 1	66	1	37	10.7	8.6	73.6
ALS 2	52	1	36	7.4	7.1	50.7
ALS 3	33	1	33	11.0	10.0	99.9
*1	17	2	8	3.9	1.3	1.6
*2	20	1	11	4.8	3.0	9.3

(*1) for stable terrain near RTS 1, and (*2) for stable terrain near ALS 1.

Table 4.2. Tabulated results of slope angle measurements for retrogressive thaw slumps, active layer slides and stable slopes adjacent to the sites.

4.3 Active Layer Slides

Setting

The active layer slides examined are situated on a variety of slopes displaying a range in aspect, slope angles, material properties and vegetation characteristics. Five slide scars are present at site ALS 1 and one each at ALS 2, 3, and 4.

ALS 1 is located 1.5 km from retrogressive thaw slumps RTS 1 and RTS 2, and contains five closely spaced active layer slide scars. All of the slide scars at this site begin between 35-40 m above sea level, approximately 200 m inland, and are situated on south - facing slopes. Terrain conditions surrounding this site are characterized by steep, thinly vegetated slopes dissected by deep gullies. An interesting aspect of these slide scars is their initiation at approximately the same elevation.

The slide scar at ALS 2 is situated on a south to southeast - facing slope directly upslope from a large alluvial fan and spit that forms Herschel Spit. Terrain in the vicinity of this site can be described as gently rolling, although steep coastal bluffs, dissected by deep gullies, are prevalent towards the east.

The slide scar at ALS 3 is situated on a south to southwest - facing slope, also upslope from the same alluvial fan as ALS 2. The slide is confined to an area between ridges and has abrupt, steep sidewalls. The terrain at this site is characterized by steep, poorly vegetated slopes.

The slide scar at ALS 4 is located on a south to southeast - facing slope near Collinson Head and is the largest of the slides examined. The topography of this area is gently rolling and well - vegetated, with steep coastal bluffs characterizing most of the shoreline.

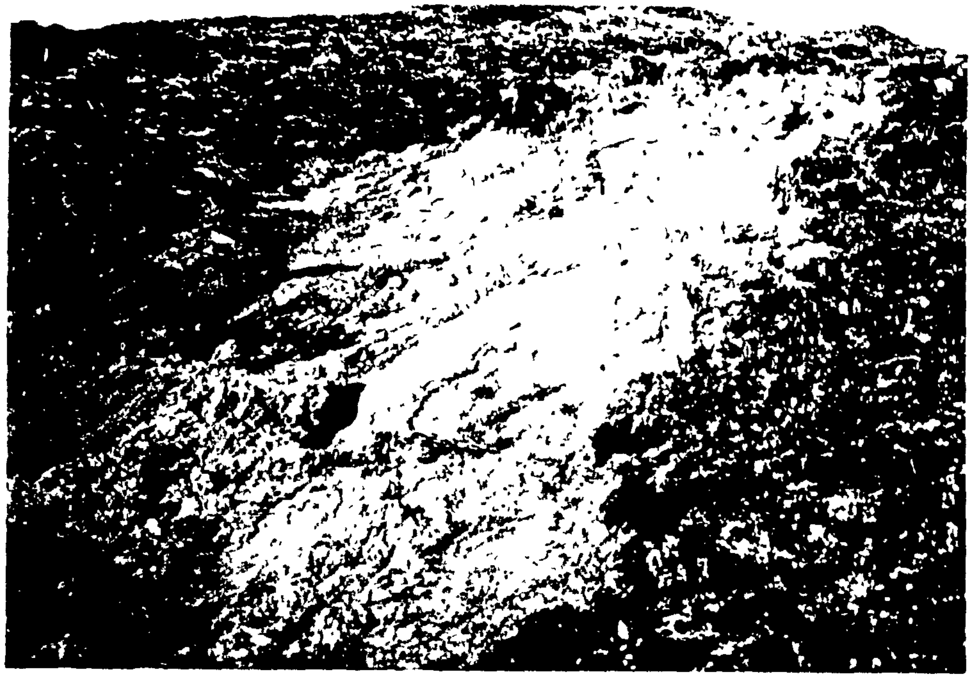
Morphology

Active layer slide scars (ALS 1-4) are ribbon-like features resembling mudslides found in more temperate environments, except that they tend to be shallower in nature (Figure 2.8).

The slide scars at ALS 1 measure up to 200 m in length and 20 m in width, and slide headwalls measure only 70 cm in depth. Average slide area was 3,240 m^2 with a maximum of 4,080 m^2. The most extensive slide disrupted a large section of the hillside; the displaced mass forming the slide base was up to 2 m in height and had slid approximately 25 m into Thetis Bay. The slide floor and lower slide sidewalls are abrupt and are characterized by slickenslides. The slide floors also contain residual blocks of tundra and material which had broken away from the headwalls probably some time after the initial failure (Figure 4.7.a).

The slide scar at ALS 2 has a headwall-to-toe length of 177 m, width of 52 m, headwall height of 1 m, and covers an area of 9,200 m^2. The slide floor at this site contains

A



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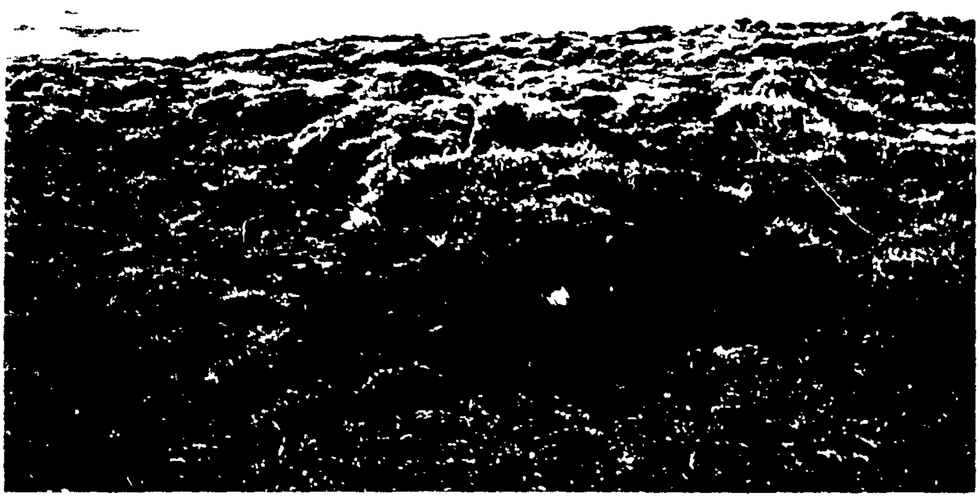


Figure 4.7. Photographs illustrating microtopography of (a) slide floor (July 1988) and (b) the base of slide (July 1988).

numerous residual blocks of tundra. Displaced material that formed the base of the slope had moved as an intact layer. At the base of the slide the mass took on a ridged or corrugated appearance. These ridges were up to 1.5 m in height and had a compressional index of 0.89 (undeformed length/deformed length) (Figure 4.7.b).

The active layer slide scar at ALS 3 measures 60 m in length, 12 m in width, and has a headwall height of 1.5 m. It covers an area of 700 m^2. Unlike many of the active layer slides on Herschel Island, this slide was deeply incised. At its mid-point it measured 2.5 m in depth. This slide can be described as a textbook example of an active layer slide - it is wide at its uppermost point of slope detachment, narrow near its mid-section and widens at its base. The slide floor is characterized by residual blocks of tundra and the sidewalls display numerous slickenslides. The base at this site also has a corrugated appearance with a compressional index of 0.85.

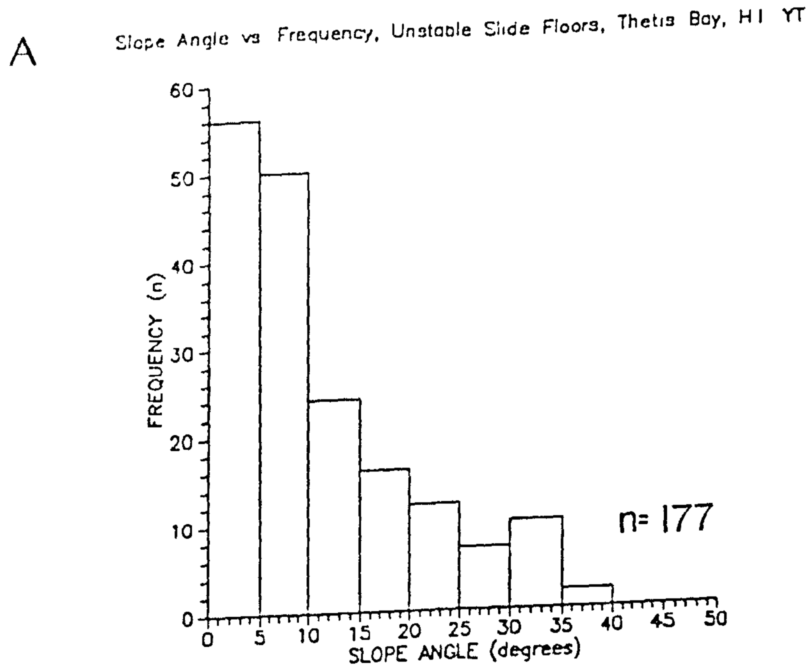
The large failure at ALS 4 consists of three small slide scars which coalesce 45-50 m downslope from their point of initiation. This resulted in the disturbance of an area over 9,860 m^2. It appears that the slides must have been initiated simultaneously, otherwise there would be evidence of overriding of material. Total slide length is 290 m, with a width of 34 m at the point where the three small slides coalesce. The height of the headwall averaged 90 cm. The slide floor is characterized by slickenslides and numerous partially to fully-detached blocks of material. The failure followed the topography of the rolling terrain, taking on a "roller coaster" appearance.

Slope Conditions

Slope profiles and a large number of slope angles were obtained both for disturbed surfaces within slide floors and for undisturbed surfaces above and adjacent to active layer slide scars (Figure 4.8 and 4.9). Slope angles of the slide floor at ALS 1 range between 1-37°, average 10°, and have a standard deviation of 8.6°. At ALS 2 angles within the slide floor range between 1-36°, average 7° and have a standard deviation of 7.1°. Slope angles at ALS 3 range between 1-33°, average 11°, and have a standard deviation of 10° (Table 4.2). Greatest angles (up to 37°) were recorded at the slide headwall and slide base. These were areas of slide initiation, gully formation or coastal bluff development.

4.4 Discussion

Although RTS 1 and 2 display the classic amphitheatre or bowl shape, very large slumps containing a variety of ground ice types may develop very irregular shapes (Figure 4.10). The shape of RTS 3, situated between RTS 1 and RTS 2, is influenced by several large ice wedges exposed in the slump headwall. In 1988 it was observed that ice wedges exposed in



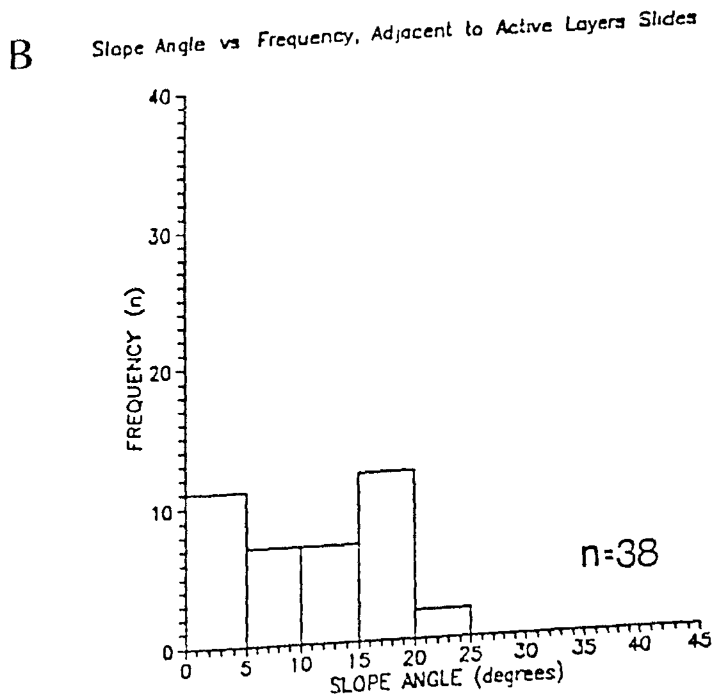


Figure 4.8. Frequency histograms for (a) disturbed slopes within active layer slide floors, and (b) undisturbed terrain surrounding active layer slides.

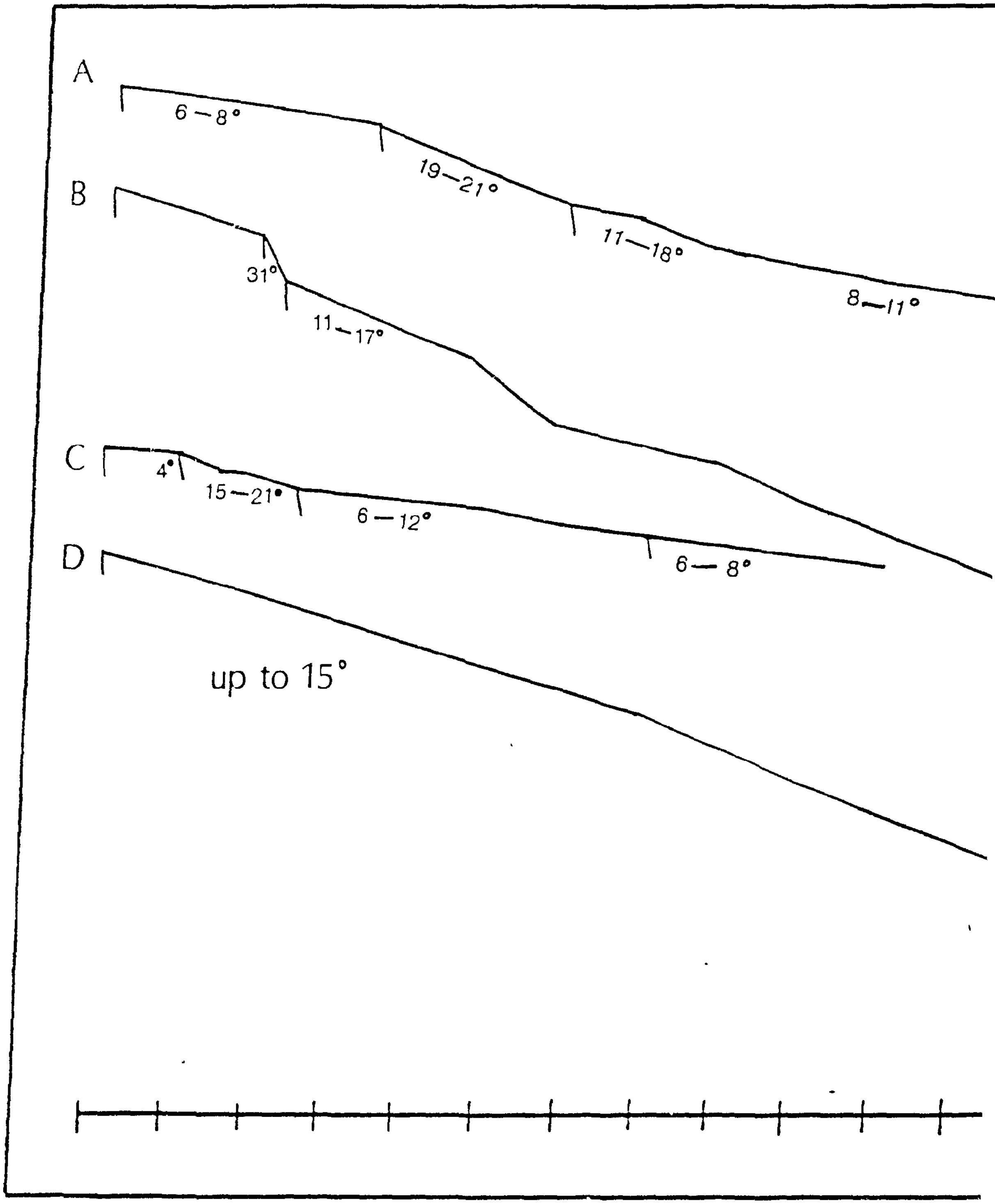
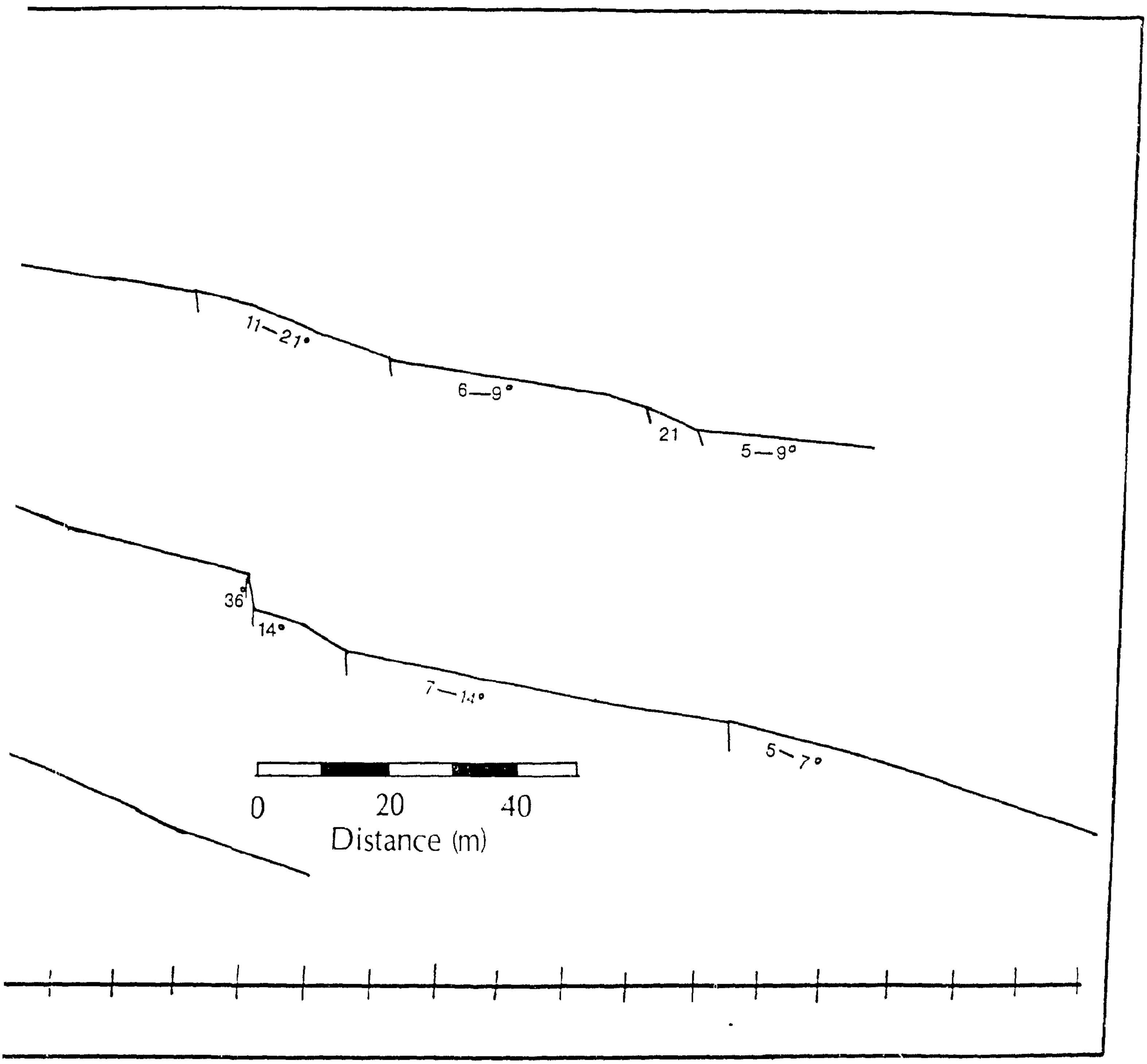


Figure 4.9. Slope profiles for active layer slides (A-C) and undisturbed terrain surrounding them (D).



11—21°
6—9°
21
5—9°
36°
14°
7—14°
5—7°
0
20
40
Distance (m)

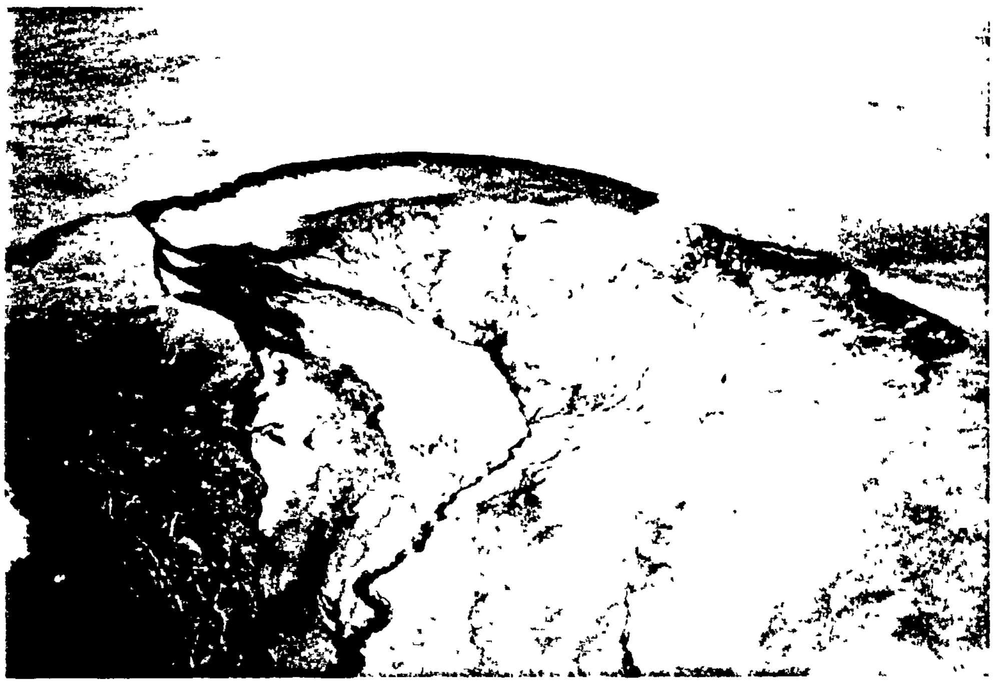


Figure 4.10. Photograph illustrating the irregular shape of RTS 3 (July, 1989). The photo also shows the cyclic nature of slump occurrence. Note the presence of a slump scar to the right of the active slump.

the slump headwall tended to melt slower than the enclosed massive ground ice and ice-rich sediment. The irregular ablation rates across the slump caused a jagged or serrated appearance. By comparison, RTS 1 and RTS 2 which exhibit a symmetrical amphitheatre shape contain no ice wedges and two small wedges, respectively. Thus, morphologic differences between slumps may be the result of subtle differences in ground ice conditions and character. The shape of RTS 1 and RTS 2 reflect relatively homogeneous ground ice conditions.

The age of the two slumps is unknown. RTS 1 and 2 are not visible on air photos from 1951 (A13470 138-140; A14361 90-95), and are thus less than 39 years old. In 1951 the current locations of RTS 1 and 2 were areas of stable terrain (Figure 4.11.a). Based on the air photos and field work, it is evident that both sites had experienced former thermokarst activity in the form of retrogressive thaw slumps. This is clearly indicated by the presence of stabilized revegetated slump scars above the active headwall at both sites. The air photos from 1970 (A21921 135-136) show a headscarp present at RTS 1 and a small active slump (Figure 4.11.b). In 1970 the headscarp of RTS 1 was approximately 110-120 m from the coast and unstable (freshly disturbed) hummocky terrain occurred within the slump floor. In addition, melt-out along ice-wedge troughs and a small thaw lake are evident above the retreating headwall. In the case of RTS 2, the 1970 air photos show a polygonal pattern of ice wedges, as well as slumping and block

A

B

Figure 4.11. Photographs of airphotos of the Thetis Bay shoreline in (a) 1951 (A13470 140), and (b) 1970 (A21921 136). RTS 1 and 2 are not visible on 1951 photos. However, by 1970 RTS 1 is visible and RTS 2 displays slumping and block failure along the shoreline.

failure along the shoreline.

Unlike retrogressive thaw slumps. active layer slides are not a thermokarst landform, but are the result of unstable slope conditions. They are usually elongated, shallow, and variable in size.

Active layer slides are influenced by topography. Sites ALS 1 and 3 are classic textbook examples of slope failure in permafrost environments; they are wide at the point of detachment, narrow at the midsection and wide at the base. The morphology (dimensions, in particular) of active layer slides cease to change to any great extent subsequent to the sliding event. However, Chapter 6 does describe the development of a retrogressive thaw slump within the floor of an active layer slide scar.

Chapter 5

Cryostratigraphic and Material Controls

5.1 Introduction

Since retrogressive thaw slumps are a form of thermokarst, then their cryostratigraphy, including the type and volume of ground ice is the dominant variable controlling slump location and growth. By comparison, since active layer slides are a thaw-induced mass wasting process, then slope conditions, material properties, depth of thaw and soil moisture are the main factors influencing their location, magnitude and frequency. In both cases, a complexity of other related factors such as air temperature, radiation, snowmelt, runoff, precipitation, and vegetation also determine site specific processes and morphologic characteristics. This chapter presents observations and analyses of rheological properties of materials and ground ice characteristics associated with retrogressive thaw slumps and active layer slides on Herschel Island.

5.2 Cryostratigraphy and Ground Ice

The deformed fine-grained lacustrine sediments and glacial sand and silt of Herschel Island are known to contain considerable volumes of ground ice (Bouchard, 1974; Mackay, 1959; Pollard, 1989, 1990). Many of the primary depositional sedimentary structures and stratigraphic sequences have been deformed and modified by the epigenetic aggradation and, in some cases, melting of ground ice. Figure 5.1 illustrates the complex nature of ground ice and deformed sediments found on Herschel Island. The reworking of ice-rich sediments by thermokarst activity has produced a deep thaw unconformity observed at several locations, and an upper diamicton unit which mantles much of the landscape. This stratigraphic relationship has also been described for a number of sites along the Yukon coast (Harry *et al.*, 1985; Pollard, 1989; Pollard and Dallimore, 1988). Ground ice observations were initially made at ten locations between 1987 and 1989. However, this thesis will present stratigraphic observations only from three retrogressive thaw slumps located along Thetis Bay and in the vicinity of Collinson Head.

Since active layer slides are a thaw-related mass wasting landform, their occurrence is not dependent upon the melting of exposed massive ground ice or ice-rich sediments. Therefore, unlike thaw slumps, which are often characterized by headwalls more than 10 metres in height, active layer slide headwalls reach a maximum height of only 1.0-1.5 m. Thus, detailed stratigraphic analyses for comparison with retrogressive thaw slumps could not be undertaken, and consequently, field and laboratory analyses of active layer slides focused upon material properties.

Figure 5.1. Deformed sediments and ground ice exposed in a slump headwall, Thetis Bay (July 1988).

5.2.1 Retrogressive Thaw Slumps

Stratigraphy

The headwall of RTS 1 ranged between 5 and 7 m in height and contained 3 distinct stratigraphic units (Figure 5.2.a). The lowest unit (Unit 1A) was a 2.5 m to 3.0 m thick massive ice body. It consisted of dirty ice interbedded with numerous layers (20-30 cm thick) of pure ice which contained thin (2-5 cm) discontinuous silty clay bands. The latter were inclined 15-20° towards the northeast. The foliated appearance of the ice was enhanced by high concentrations of gas inclusions associated with sediment bands. Unit 1A was unconformably overlain by an ice-rich, dark brown to grey-brown, silty clay layer (Unit 1B). This unit contained clear ice veins and lenses 2-3 cm thick which, in places, formed a reticulated cryotexture. Isolated horizontal ice layers and lenses 5-10 cm thick occurred randomly throughout this unit. Where the headwall lay beneath a former thaw slump scar a thin diamicton unit (Unit 1C) unconformably overlay Unit 1B. This unit consisted of an icy brown clay which graded upwards into 10-40 cm of organic silt and surface root mat. The base of Unit 1C, interpreted as a thaw unconformity, was marked by a dark organic-rich clay layer and a distinct change in ice lens pattern. A steeply dipping band of pale brown ice 20-30 cm wide cut discordantly across the entire face, terminating at the thaw unconformity (Figure 5.2.b). This structure is interpreted as an ice dyke, which has recently been described

A

B

Figure 5.2. (a) Exposed headwall of RTS 1 (July 1987), and (b) ice dyke exposed in the headwall of RTS 1 (July 1987). Unit 1A, the lower unit, was a 2.5-3.0 m thick massive ice body which overlay Unit 1B, an ice-rich, dark brown to grey brown silty-clay layer. Unit 1C, the upper unit, consisted of an icy brown clay which graded upwards into 10-40 cm of organic, silt and root mat.

in detail by Mackay (1989). Ice dykes consist of intrusive ice injected from depth into overlying frozen sediment, and occur primarily along fractures and faults. Since this ice cuts across existing structures it must be younger than other ice bodies in the sequence.

Ground ice contents for RTS 1 ranged between 150% in the thawed organics which make-up the active layer, and 4500% for massive ice bodies. Moisture contents within the slump floor ranged between 10-30%. Ground ice types observed at this site are segregated ice, intrusive ice (ice dyke) and pore ice.

The headwall of RTS 2 ranged between 5 to 7 m in height and contained two stratigraphic units (Figure 5.3). Unit 2A, exposed at the base of the section, was a foliated, massive ice layer greater than 3 m in thickness. This unit contained layers of dirty ice, interbedded with layers of grey-brown silty clay and pure segregated ice. Cobbles 5-10 cm in diameter and fine sand also occurred as sediment bands. Gas inclusions and thin discontinuous bands of sediment occurred within dirty ice and pure ice layers. The unit had a foliated appearance which resulted from the discontinuous horizontal sediment layers. These were gently inclined at 10-12° towards the north-northeast. The lower contact of this unit was below the level of the slump floor. A dark brown silty clay 1.5-2.5 m thick unconformably overlay Unit 2A. Unit 2B had a blocky cryotexture and contained lenses and layers of clear and sediment - rich ice. A reticulated pattern of thin ice

Figure 5.3. Exposed headwall at RTS 2 (July, 1987). Unit 2A, the lower unit, was a foliated, massive ice layer greater than 3 m in thickness, and was overlain by Unit B, a dark brown silty clay layer 1.5-2.5 m thick. Unit B was overlain by 45 cm of thawed organics.

veins occurred immediately above the contact between units 2A and 2B. The top of Unit 2B graded into a 45 cm layer of thawed organic silt which contained pods of peat and root fibres (active layer). Two small ice wedges 30 cm wide and 1.0-1.5 m deep are exposed at the base of the active layer and penetrated the upper part of Unit 2B.

Ground ice contents at RTS 2 varied from 35%, for partially dessicated clays and organics, to greater than 4500% for massive ice. Moisture contents of thawed material within the slump floor were between 10-30%. The dominant ground ice types observed in the slump headwall include segregated ice, ice-wedge ice and pore ice. Sediments and stratigraphic characteristics described for RTS 1 and 2 are consistent with the "mixed sediment" unit described by Bouchard (1974). According to Bouchard, this unit represents preglacial deposition in a nearshore marine and shoreline environment. The occurrence of both shell fragments and detrital organics in these sediments is also consistent with Bouchard's description.

A small retrogressive thaw slump occurred at Collinson Head (RTS 4) in 1987-1988. RTS 4 was situated at the head of a deep gully between ice-push ridges, and contained roughly 3.0-3.5 m of massive ice and ice-rich diamicton (Figure 5.4). The massive ice and ice-rich sediments were divided into three units. The lowest unit (Unit 4A) consisted of a 2.0-2.5 m arched body of layered "whitish" ice that pinched out at both

Figure 5.4. Exposed headwall of RTS 4 (July, 1987). Unit 4A, the lower unit, is a body of "whitish ice" which overlay Unit 4B, which consisted of an irregular, ice-rich, grey-brown clay diamicton. Unit 4C, the upper unit, consists of horizontal ice layers and lenses.

ends of the exposure. This ice contained oval to elongated bubbles and fine suspended sediment which formed distinct horizontal bands. The massive ice of Unit 4A was composed of two different ice types - a lower body of white layered ice was overlain by a massive body of brown colored ice. These ice bodies, particularly the lower, more extensive arched ice, are very different from any other type of ice documented for Herschel Island. Stratigraphic studies and the detailed analyses of ice crystallography and analyses of ice chemistry lead to the speculation that this ice unit was buried glacier ice and segregated ice preferrentially incorporated during the ice-thrusting process that formed the nearby ridges (Pollard, 1990). Unit 4B consisted of 1.0-1.5 m of irregular and undulating ice-rich, grey-brown clay diamicton. The contact between Units 4A and 4B was abrupt and unconformable, representing a thaw unconformity produced during either a previous thermal disturbance or during syngenetic ice formation which is consistent with a buried glacier origin. The upper unit (Unit 4C) contained thick horizontal ice lenses and layers and irregular bodies of massive pale brown ice - possibly a buried snowbank. Reticulated ice veins occurred throughout the diamicton. The active layer, 40-50 cm in thickness, contained a thawed portion of the diamicton, as well as organics and other detrital materials.

Ground ice contents at RTS 4 varied from 35% in surface organics to a maximum of 8000% in massive ice layers. Ground ice types observed at RTS 4 include segregated ice, pore ice and possibly buried glacier and snowbank ice.

By 1989, the Collinson Head site no longer displayed the clear massive ice or ice-rich permafrost that it had in 1987-1988. The slump had stabilized, although the area as a whole is experiencing rapid fluvial erosion and gully formation. These processes are perpetuated by the melting of snow and exposed ground ice.

Sediments and stratigraphic characteristics at RTS 4 are typical of the "marine" and "mixed" sediments described by Bouchard (1974). These sediments are the result of deposition in a nearshore marine environment which were subsequently repositioned by glacial processes.

Material Properties

Grain size characteristics, moisture contents and Atterberg Limits were analysed to assess the role of material properties on ground ice occurrence and to evaluate the relationship between ice content and material behaviour upon thawing.

Silt and clay sized sediments are the dominant particle sizes in materials where thaw slumps and massive segregated ground ice bodies occur; however, sand and gravel sized (cobbles 5-10 cm diameter) materials occurred as bands within massive ice and formed a residue on the slump floor. These observations are comparable with data from the Mackenzie Delta where fine materials, such as silt and clay, are associated with massive segregated ice (Mackay, 1966) and with Bouchard (1974) for Herschel Island. Sand sized sediments are found as layers within and beneath some of the massive ground ice bodies at Thetis Bay (Figure 5.5). This association supports the segregated ice interpretation for massive ground ice at sites RTS 1 and 2.

The liquid limits for sediments taken from retrogressive thaw slump headwalls range between 30-40% (Figure 5.6), while plastic limits were between 15 and 25%. The ice content of material in the slump headwall normally exceeds the liquid limit of the sediment. As a result, thawing produces a slurry of supersaturated sediment that initially pools near the base of the headwall and eventually flows onto the slump floor. Sediment samples taken from distances up to 10 m from the headwall had the highest moisture content. This is believed to be the result of gradual settling of the coarser sediment fraction. Fresh material at the base of the headwall often displayed much lower moisture content because it frequently consists of blocks of drier sediment from the active layer; it eventually, however, becomes part of the mud slurry. At a distance of 15 m from the headwall, moisture contents of sediments drop below their plastic limit (Figure 5.7).

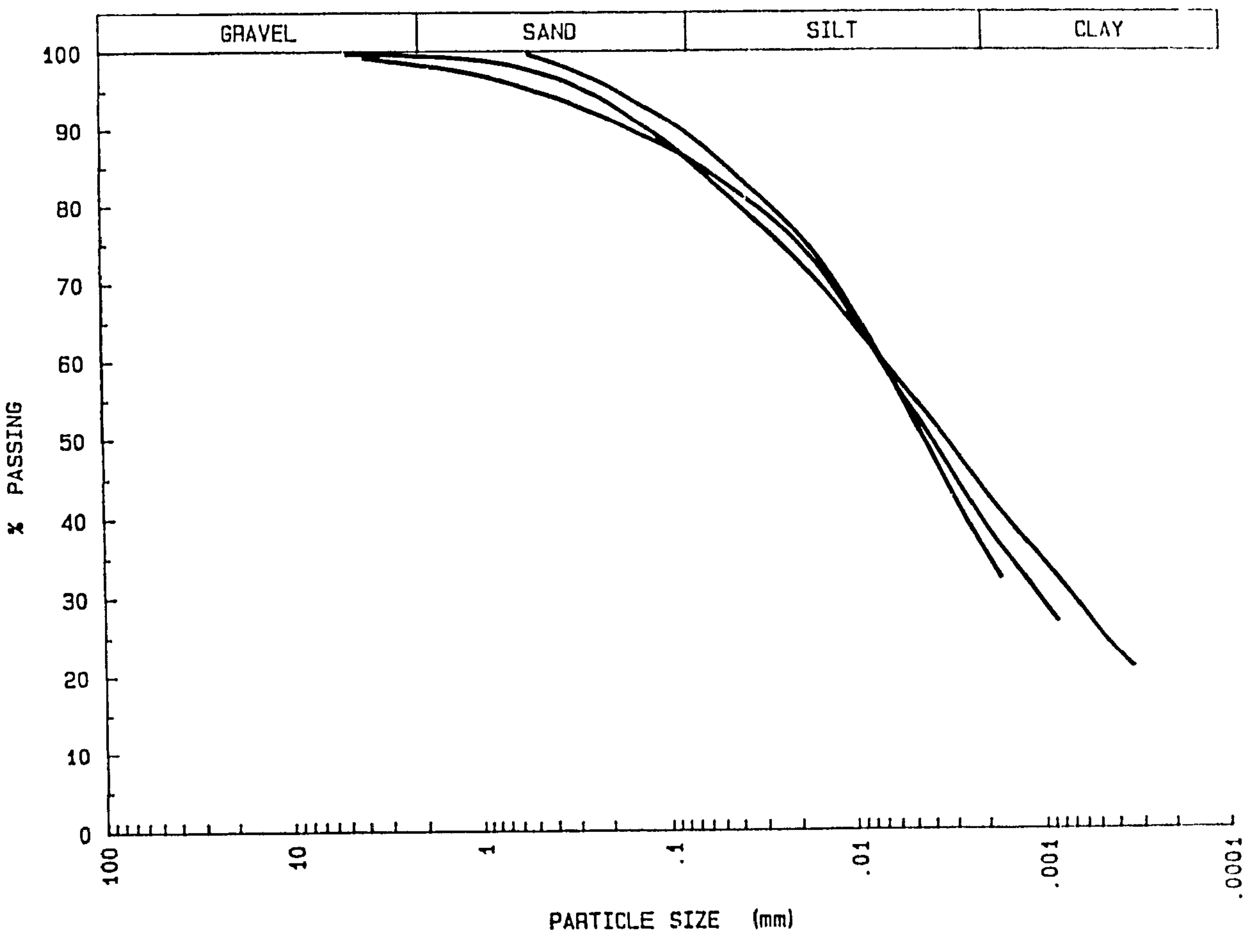


Figure 5.5. Representative grain size distribution curves for retrogressive thaw slump sediments.

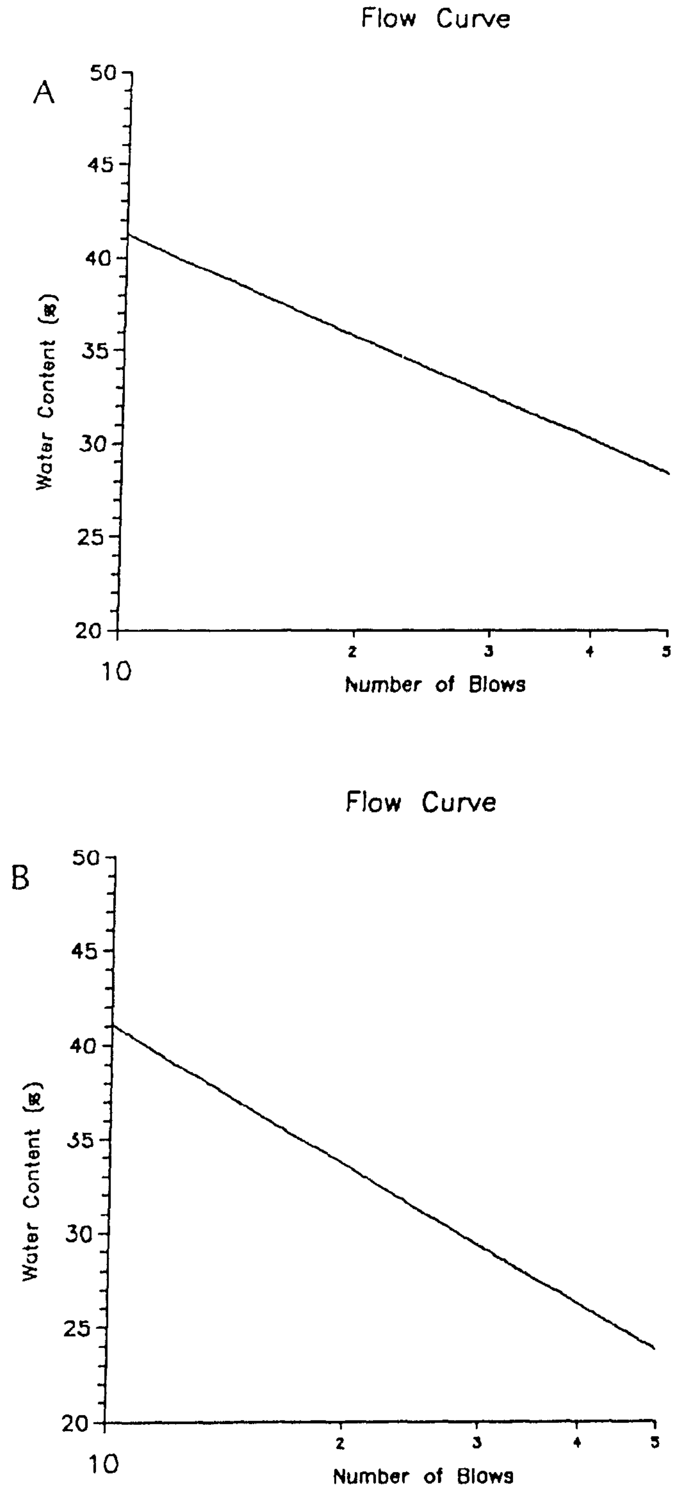


Figure 5.6. Representative flow curves (liquid limits), for sediment samples from retrogressive thaw slumps (a) RTS 1 and (b) RTS 2.

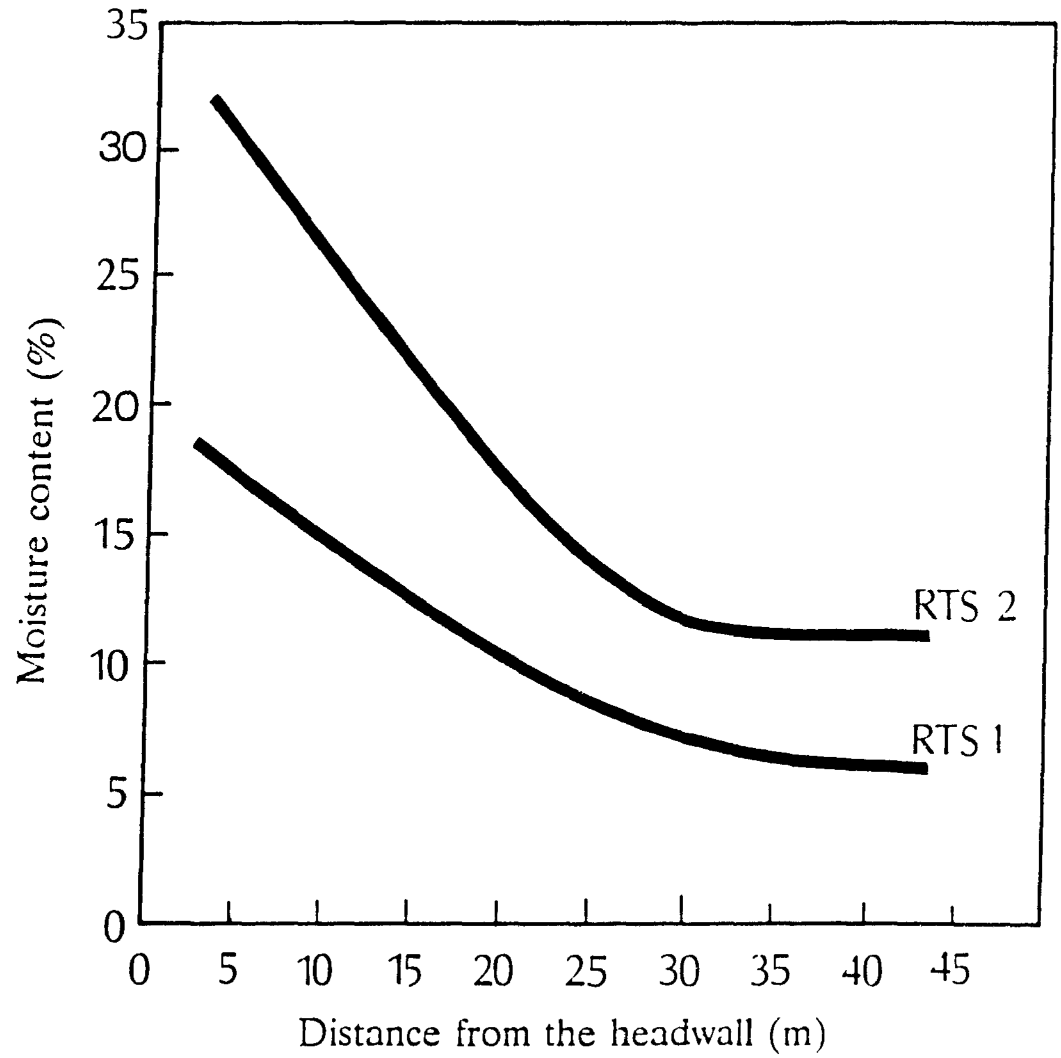


Figure 5.7. Graph of moisture content vs. distance from the headwall at RTS 1 and RTS 2. There is a rise in moisture content associated with pooling of water near the base of the headwall, and a gradual decrease at distances greater than 10 m, a result of dewatering of sediments.

5.2.2 Active Layer Slides

Observations from four sites (ALS 1-4) containing one or more active layer slide scars are described. In this part of Herschel Island, the majority of slides occur on steep slopes forming the Thetis Bay shoreline; the most extensive slide occurred in the vicinity of Collinson Head where relief is the greatest. The following summarizes material properties of sediment from four slides and shallow stratigraphic observations documented during headwall excavation of ALS 2.

These sites are all located in Bouchard's (1974) pre-glacial deposits. According to Mackay (1959), emplacement of these sediments is related to glacier ice-thrusting. Bouchard's map (1974) designates the area around ALS 1 as non-differentiated pre-glacial materials, and the area where ALS 2, 3 and 4 occur as marine sediments. Marine sediments are the most abundant of the preglacial deposits, and consist primarily of clays and clayey silts or clayey silts with sand laminae.

Due to their shallow nature, slide headwall exposures all contain two stratigraphic units consisting of (1) an overlying vegetation and root mat corresponding to the Ahy cryosol horizon of Smith *et al* (1990), and (2) a marine clayey silt and clay unit. The latter contains scattered marine shells, shell fragments and sand laminae. Since these materials form part of the modern active layer, the absence of massive ice was expected; however, discontinuous horizontally layered ice lenses up to 2 cm thick and thin continuous ice layers less than 0.3 cm are common in frozen exposures.

The massive sediments contain 10% sand, 60% silt and 30% clay (Figure 5.8). This unit ranged from massive silt and clay to faintly bedded silt and clay with sand laminae. Occasional gravel or cobble sized particles also occur, but are most noticeable as lag deposits with sand on the slide floors. The liquid limits of these materials vary between 25-30% (Figure 5.9), with corresponding plastic limits of 10% or less.

Moisture contents of soils having small ice lenses from the slide headwall range from 15-35%. By mid-August an active layer thickness of up to 70 cm is not unusual. Around the same time, soil moisture content at the base of the active layer frequently is near saturation (30-40%). This is particularly the case near seepage lines. Excavation of a slide headwall revealed a relatively continuous and uniform zone containing ice lenses from 2-5 cm thick at a depth of 70-80 cm. This ice-rich zone is common and is generally interpreted as aggradational ice. Ice contents of this ice-rich zone ranged between 100-250%.

5.3 Ground Ice Geochemistry

The chemical regime of ground ice is a potential tool for elucidating the genesis of water for ground ice formation and

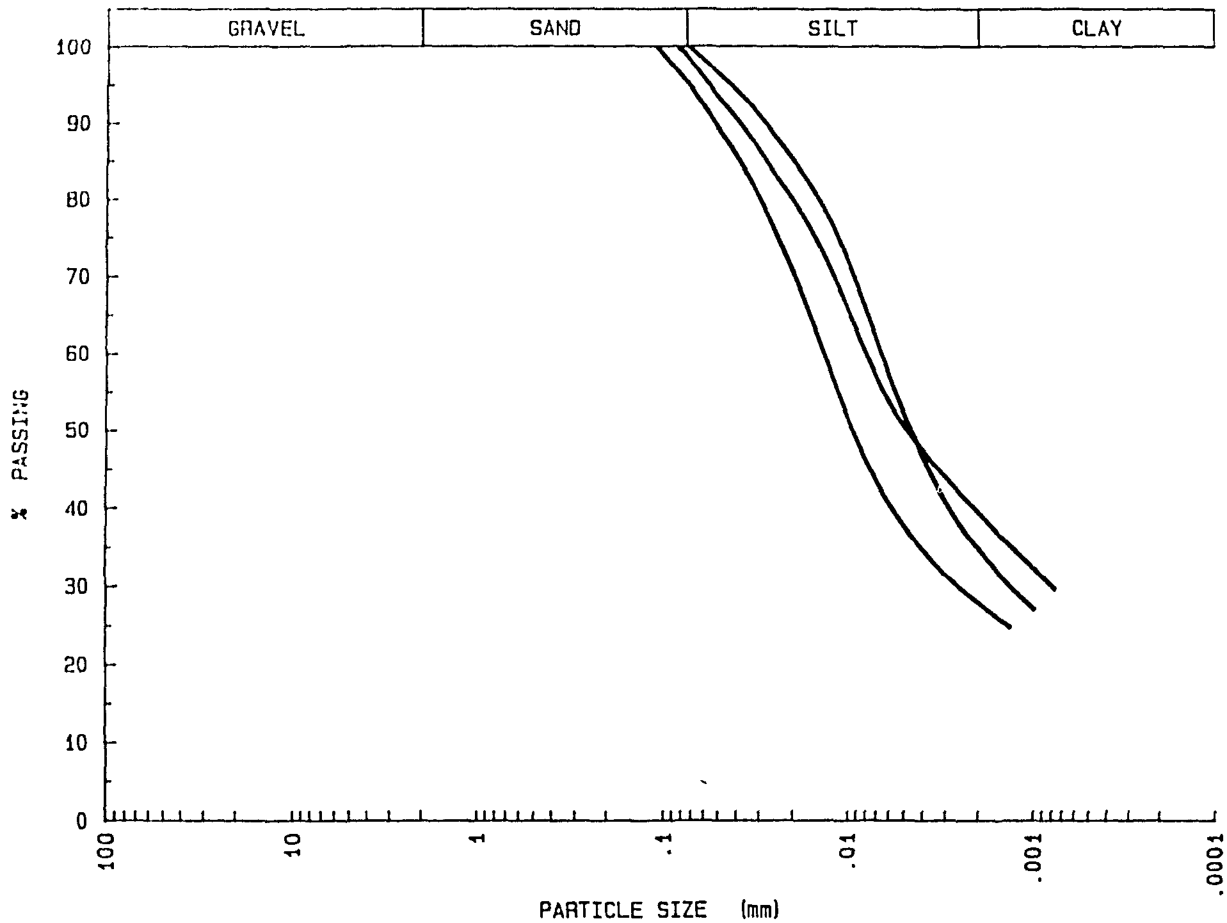
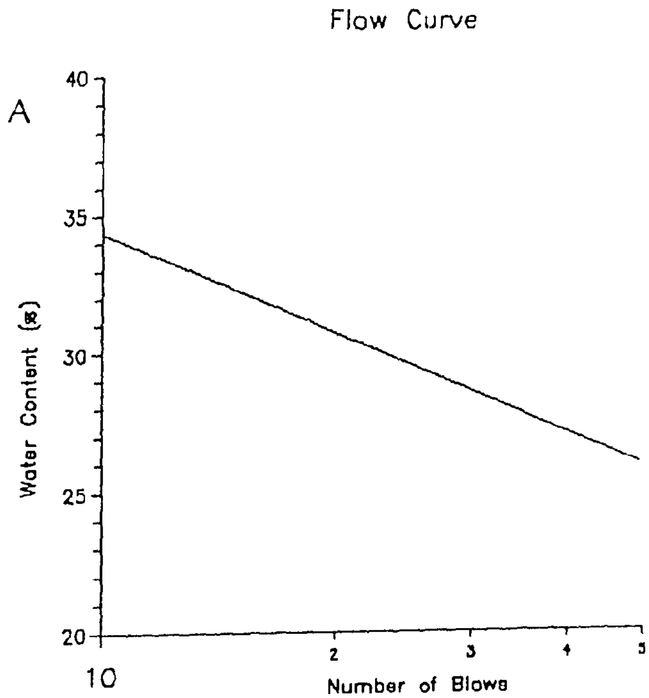


Figure 5.8. Representative grain size distribution curves for sediments from active layer slides.



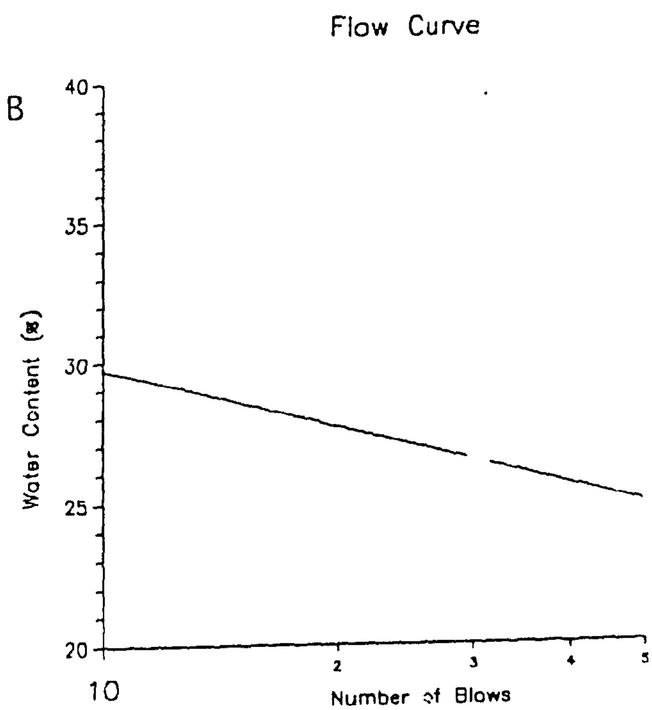


Figure 5.9. Representative flow curves (liquid limits) for sediment samples from active layer slides (a) ALS 1 and (b) ALS 2.

the complex mechanisms involved. For example, if massive ice is the product of ice segregation then one would expect it to display chemical characteristics consistent with those of soil and groundwater from its point of origin. By comparison, ground ice formed from recent atmospheric water (ice wedges, buried snowbank ice, ice lenses and pore ice in the active layer) will reflect the chemical characteristics of recent precipitation and generally have low total dissolved solids (TDS). In theory, buried glacier ice may show distinctly different chemical characteristics from ice in enclosing sediment or other massive ice types.

Samples were collected in polyethylene containers, thawed and the supernatant waters were analysed. Due to limitations arising from cost and time, only selected samples were analysed (Table 5.1). The following will discuss the nature and significance of ground ice hydrochemical characteristics and thus the influence of ice type on slump development. Ice types have been categorized into two broad types: Type A, formed from meteoric water frozen *in situ*, and Type B, ground ice formed from ground and soil water by segregation and/or injection processes.

5.3.1 Type A Ground Ice

According to Mackay's (1972b) classification, a variety of ground ice types may have meteoric water as their immediate

	No.	Ph	Ec (mS/cm)	Cl	SO_4	F	Na	K	Ca	Mg
				--------- units are in (mM) ------- ----						
Site 1	1	6.88	.11	10.14	9.60	.09	5.47	1.56	8.70	2.80
	2	7.83	.38	11.06	51.74	.19	25.45	7.62	28.80	9.20
	3	8.10	2.01	60.20	917.66	.20	173.34	26.59	163.93	76.10
Site 2	4	7.58	.21	3.94	31.87	.26	9.40	3.91	21.80	1.10
	5	7.37	.12	6.70	.77	.11	3.91	.86	13.30	2.30
	6	7.48	.75	132.60	66.82	.45	114.58	7.82	15.80	9.30
	7	8.15	2.49	572.60	174.53	.46	386.11	2.39	30.50	40.00
	8	7.18	.93	170.25	115.01	.36	134.15	8.41	21.60	13.00
	9	7.73	3.23	761.93	347.52	.38	513.82	23.46	27.30	63.00
	10	7.96	1.60	342.74	74.69	.35	235.81	12.20	22.90	11.30
	11	7.22	3.76	834.26	501.12	.28	606.93	26.39	33.90	69.00
	12	8.10	.32	33.15	34.56	.13	33.38	4.50	18.00	6.60
Site 3	13	6.21	.01	.80	.29	.00	10.67	.66	1.50	.70
	14	7.91	.35	33.50	45.12	.12	18.97	3.01	32.80	9.60
	15	7.55	.23	14.64	7.87	.09	8.30	2.62	28.20	1.60
	16	7.64	.76	24.46	205.92	.13	25.70	4.69	46.20	22.10
	17	8.15	2.27	169.58	859.20	.14	145.16	11.18	199.00	129.00
	18	8.44	4.12	936.01	453.12	.42	671.30	28.15	70.00	73.00
Site 4	19	7.53	.09	12.66	.77	.02	8.51	.08	6.20	1.04
	20	7.46	.05	4.89	2.02	.01	7.77	1.76	4.33	.96
Others	21	7.55	.89	74.53	215.04	.08	74.85	.39	91.60	11.60
	22	7.75	.16	15.60	14.69	.09	12.67	.39	11.50	1.20
	23	7.84	.31	82.15	20.35	.07	68.74	5.08	13.50	4.34

Table 5.1. Ground ice chemistry of various ice types from Herschel Island.

source (e.g., pore ice, ice-wedge ice, some forms of aggradational ice, and shallow segregation ice). Ground ice fed by atmospheric moisture and surface water should have low TDS values (Brown, 1963c; Hallet, 1978; Levinson, 1974).

Ice chemistry analyses (Table 5.1) indicates that samples 13, 14, 15, 16 and 20 formed by meteoric water have low concentrations of TDS. Samples 13 and 14 (snow), form an ideal standard because recent snow is known to have low TDS (Levinson, 1974). Although ice types fed by meteoric water have low concentrations of TDS, anomalies may result from freeze/thaw activity, fractionation and residue inclusion (Brown, 1963c; O'Sullivan, 1963). Freezing not only reduces the diluting effect of recharge by precipitation, but also increases the concentration of the dissolved mineral content at the freezing front. This includes ice formed near the base of the active layer or at the contact of two stratigraphic units (Levinson, 1974).

Significant differences in ice chemistry occur between relict ice-wedge ice (16), which has a high sediment inclusion content, and modern ice-wedge ice (15, 20) which have low sediment inclusion content. Brown (1963c) has explained differences in chemical composition of ice-wedges as a function of water supply (meteoric) and residue inclusion. Mineral and organic materials either drop into thermal contraction cracks, or are washed in during spring melt. Another possible explanation for higher TDS of ice wedges (e.g., sample 9) is that sediments and some ions may be incorporated from the sides of the wedge, especially at depth (Brown, 1963c. Sample 9 (modern ice-wedge ice) taken from a depth of 1.0-1.5 m has a high sediment inclusion content which may possibly explain the increase in TDS compared to that of sample 20, which was taken from the surface of the ice wedge and had no visible sediment inclusion. It is suggested that differences in chemistry of ice-wedge ice (formed by meteoric water) at Herschel Island are possibly the result of the incorporation of minerals and organics during ice wedge formation.

5.3.2 Type B Ground Ice

The second ice type, massive segregated ice, is believed to have a ground or soil water origin. Massive segregated ice samples 3, 7, 8, 10, 11, 17, 18 have higher levels of TDS compared to that of ice Type A. This may be a function of (1) the inclusion of minerals and organics, and/or (2) solute rejection (fractionation) during the freezing process. Ice samples 6, 8, 17, and 18 have a high mineral and organic content and a corresponding higher concentrations of TDS. Massive ground ice with low TDS include samples 1, 2, 4, 5, 12, 19, 22, and 23. These low concentrations are best explained as a combined result of low inclusion content of minerals and organics and fractionation, a process of solute rejection as water changes to ice. Fractionation may cause surrounding materials to have high TDS. Similarly, fractionation may occur during the formation of segregated ice layers and lenses, resulting in an enrichment of pore water in the surrounding sediment. This may be the case for sample 8, which has a low TDS, and the enclosing ice-rich material sample 11 which has higher concentrations.

The results of ice chemistry analysis for samples from Herschel Island are supported by findings presented in the permafrost literature. Hallet (1978) documented solute rejection from soil water as soil freezes from the surface down. Solutes tend to accumulate at the freezing front, but may also diffuse downwards (samples 10, 9, 11 show increasing TDS with increasing depth). With continued preferential rejection of solutes by the growing ice lens or layer through moisture migration (samples 1, 2, 4, 5 and 12), the solute concentration in the unfrozen soil progressively increases. The end result is a zone of high concentration of TDS at the permafrost table (samples 3 and 10). High TDS may also be influenced by the residence time and the migration route of water during the ice segregation process (Brown, 1963c). In both cases, there is likely to be a greater incorporation of unfrozen dissolved minerals and leached organics. There may well be buried glacier ice on Herschel Island particularly along the north side of the island. However, stratigraphic analyses, ice petrofabrics documented by Pollard (1989, 1990) and ice chemistry (samples 22 and 23) do not conclusively support a glacial origin. Overall, most massive ice bodies sampled have low TDS; the incorporation of organics and minerals may be held responsible for higher TDS in samples described as dirty massive ice.

An interesting component of massive ice samples 7, 9, 11 and 18 is the higher level of sodium and chloride ions compared to other massive ice samples. Sodium and chloride ions generally occur to a limited extent under normal conditions, but these samples contain at least 2-3 times that of other segregated ice samples, and several hundred times that of ice-wedge ice and snowbank ice. Higher levels of these ions would predominate in a marine environment, and particularly in ground ice formed within sediments deposited in either marine or coastal marine conditions. These ice samples come from an area within Bouchard's "mixed sediments" classification. According to Bouchard (1974) this unit represents preglacial deposition in nearshore marine and shoreline environments. Thus the presence of higher concentrations of these salts provide strong support for the interpretation of these ice bodies as segregated ice. However, the higher concentrations leave a number of unanswered questions concerning the permafrost history of Herschel Island. Since both sodium and chloride ions are highly mobile they would be readily removed by groundwater movement and surface water infiltration. Assuming an abundance of freshwater following deglaciation sodium and chloride ions would have readily been removed during the early - mid Wisconsin. The presence of sodium and chloride ions suggests that permafrost developed shortly after emergence and possibly even before being glacially deformed during the Buckland Glaciation.

5.4 Summary

The nature and distribution of ground ice is the important factor influencing retrogressive thaw slump development. This is not strictly the case for active layer slides, however, the marine silty clays which occur in the Thetis Bay and Collinson Head areas may have a low yield strength and a decrease in cohesion upon saturation.

Several types of ground ice were documented, the most abundant of which include various forms of segregated ice, ice-wedge ice, snowbank ice and pore ice. The origin of ice-wedge ice, ice layers and lenses, pore ice and, more recently, snowbank ice are well understood. However, uncertainty still remains concerning the origin of massive tabular ice bodies which are frequently several metres in thickness. Thus more studies concerning ground ice are needed in order that permafrost evolution for Herschel Island and the Yukon Coastal Plain may be better understood.

Chapter 6

Landform Process

6.1 Introduction

There are few process studies based upon the measurement or direct observation of rapid forms of mass movement. Furthermore, there are no documented cases of rapid mass movements in areas underlain by permafrost. This is mainly due to the difficulty in predicting the occurrence of slumps and slides as well as instrumenting their location. The lack of documented cases also reflects the complex nature of factors influencing slope instability at any given time. Geotechnical analysis can provide an indication of which slopes have high failure potential (low safety factor), and possibly even the type of process or event that may occur, but there is no way of knowing, *ipso facto*, when they will occur. As a result, the majority of studies have been concerned with morphometric and/or material characteristics of failed slopes in areas underlain by permafrost (Algus, 1986; Mackay, 1966; McRoberts and Morgenstern, 1974a, 1974b). The latter applies specifically to active layer slide process studies. In the case of retrogressive thaw slumps, their slow rate of growth (relative to true slumps) and their growth by thermokarst and thermal erosion lend this process to more detailed direct observations. The following is a summary of process investigations for three retrogressive thaw slump sites and four active layer slide sites on Herschel Island.

6.2 Retrogressive Thaw Slumps

6.2.1 Process Terminology

Retrogressive thaw slumps are a form of backwearing thermokarst, of which the dominant genetic processes are thaw subsidence and headwall retreat generated by the melting of massive ground ice and ice-rich permafrost. Secondary processes are block failure and flow of supersaturated sediment on the slump floor. In this context, the use of the term "slump" to define these features, although prefixed by the descriptive and genetic modifiers "retrogressive thaw" is geomorphically incorrect. Thermokarst is not recognized as a mass wasting process, even though it involves a downward movement of material under the influence of gravity. The role of runoff produced by melting ground ice as a transporting medium is generally believed to be significant enough to categorize thermokarst and mass wasting as entirely different geomorphic processes. However, as the following summary will show, this distinction is not clear when standing within an active retrogressive thaw slump. The accuracy of the term "slump" is discussed later in this chapter.

The retrogressive thaw slumps RTS 1 and RTS 2, as well as a number of other slumps observed on Herschel Island and along the Yukon coast, are polycyclic in nature. This simply means that the current slump is growing within a former, stabilized slump. In some cases, evidence of more than one previous episode of slump activity exists.

6.2.2 Slump Initiation

It is generally accepted that retrogressive thaw slumps are initiated by the erosion and mass wasting of ice-rich sediment, although there are no references to support this. On Herschel Island, coastal retreat, thermal and mechanical erosion of ice wedges, and fluvial erosion frequently produce ice-rich permafrost exposures up to 15 m high. Based on topographic position, ground ice characteristics and information obtained from 1951 and 1970 air photos (A 13470 and A21921, respectively), it is believed that RTS 1 was initiated by a combination of coastal retreat and thermal erosion of ice wedges. The 1970 air photos clearly show the polygonal pattern of ice wedges in the vicinity of RTS 1, as well as the presence of a thaw pond above the active headwall. Thaw pond occurrence may be the result of melt-out along ice-wedge troughs. In 1988, the pond was no longer present. In the case of RTS 2, initiation is believed to have been a result of coastal retreat. This is based upon ground ice characteristics (lack of ice-wedge ice), and the close proximity of the RTS 2 headwall to the coast in 1987 (Figure 4.11). The presence of fresh water shells (Naylor *et al.*,

1972) and detrital wood in thick peat deposits exposed in the walls of RTS 3, as well as a deep thaw unconformity that mirrors a surface depression, suggests that this site is a former thaw pond. The air-photos from 1970 reveal no surface expression of ice wedges surrounding the site, nor evidence of their disruption, as is the case at RTS 1. Hence, RTS 3, which is the largest slump in the Thetis Bay area, was likely triggered by coastal erosion of ice-rich sub-lake sediment.

During this study it became apparent that several other factors could contribute to the initiation of thaw slumps. For example, during 1988-1989, a small retrogressive thaw slump developed within the floor of an active layer slide (ALS 1). The thaw slump was bowl-shaped and roughly 9.8 m long, 7.5 m wide and 2.5 m high (Figure 6.1). Within the headwall, a 2-5 cm ice lens was found at a depth of 1.2 m. On another occasion, it was observed that human activity could also induce thermokarst, and in particular, reactivate thaw slump activity. In this case, regular use of a foot path to access the RTS 2 slump floor caused a stable headwall to become active. Subsidence cracks initially developed perpendicular to the headwall; they widened and gradually exposed massive segregated ground ice which began melting, eventually causing block failure and thermal erosion of exposed ground ice.

Figure 6.1 Retrogressive thaw slump development within an active layer slide at ALS 1 (July, 1988).

6.2.3 Headwall Retreat

Introduction

The bowl or amphitheatre shape that typifies retrogressive thaw slumps expands primarily by headwall retreat. The headwall of RTS 3 has relatively straight and long segments (Figure 4.10), but the semicircular headwalls of RTS 1 and 2 are more common. The headwalls of RTS 1 and 2 tend to increase in height from the ends of the semicircle towards the middle, and the most rapid retreat occurs at that portion of the slump with the highest headwall.

Survey Technique and Timing

The investigation of thaw slump processes and, in particular, headwall retreat involved installation of a grid of survey markers around the slump headwall and on the slump floor. The weekly measurement of the position of the headwall relative to survey markers provided the basis for headwall retreat rates and retreat direction. Survey markers were installed in June 1988, at RTS 1-3, as well as at two coastal bluff locations for comparison with the retrogressive thaw slumps sites. RTS 3 was monitored to test the hypothesis that larger slumps experience higher rates of headwall retreat.

Retreat

Headwall retreat at RTS 1 for a six week period in the summer of 1988 (June-August), averaged 2.3 m with a maximum of 7.8 m. Retreat for the same period at RTS 2 averaged 3.5 m, with a maximum of 12.2 m, and at RTS 3 retreat averaged 3.4 m with a maximum value of 4.0 m. Detailed surveys in April 1989 allowed computation of an annual retreat rate for 1988. Annual retreat for the three sites averaged 6.5 m. At RTS 1 annual retreat ranged between 0 to 10.7 m, with an average of 5.8 m. Up to 19.5 m of retreat occurred for a 12-15 m long segment of the headwall at RTS 2, and averaged 10 m for the site. Figure 6.2 presents plotted survey data for RTS 1 and 2 and shows how differential retreat can modify slump shape. It was anticipated that RTS 3, which is the largest of the thaw slumps, would experience the highest annual headwall retreat rates; however, this was not found to be the case. Annual retreat rates at RTS 3 were up to 10.7 m, with an average of 9.3 m. The retreat rates recorded at Herschel Island are amongst the highest recorded in the literature for the development of retrogressive thaw slumps (Table 6.1).

Maximum and minimum air temperatures were recorded both inside and outside the slump bowl at RTS 1 (Figure 6.3). Over a six week period, average daily maximum temperature inside the slump was 4°C warmer on clear days (approximately 75% of the time) than outside the slump. Figure 6.4 illustrates the cumulative headwall retreat at RTS 1 and 2 for the period of June to August, 1988. Highest rates of headwall retreat were documented from July to early August, when daily air temperatures are at their warmest. It was found that retreat

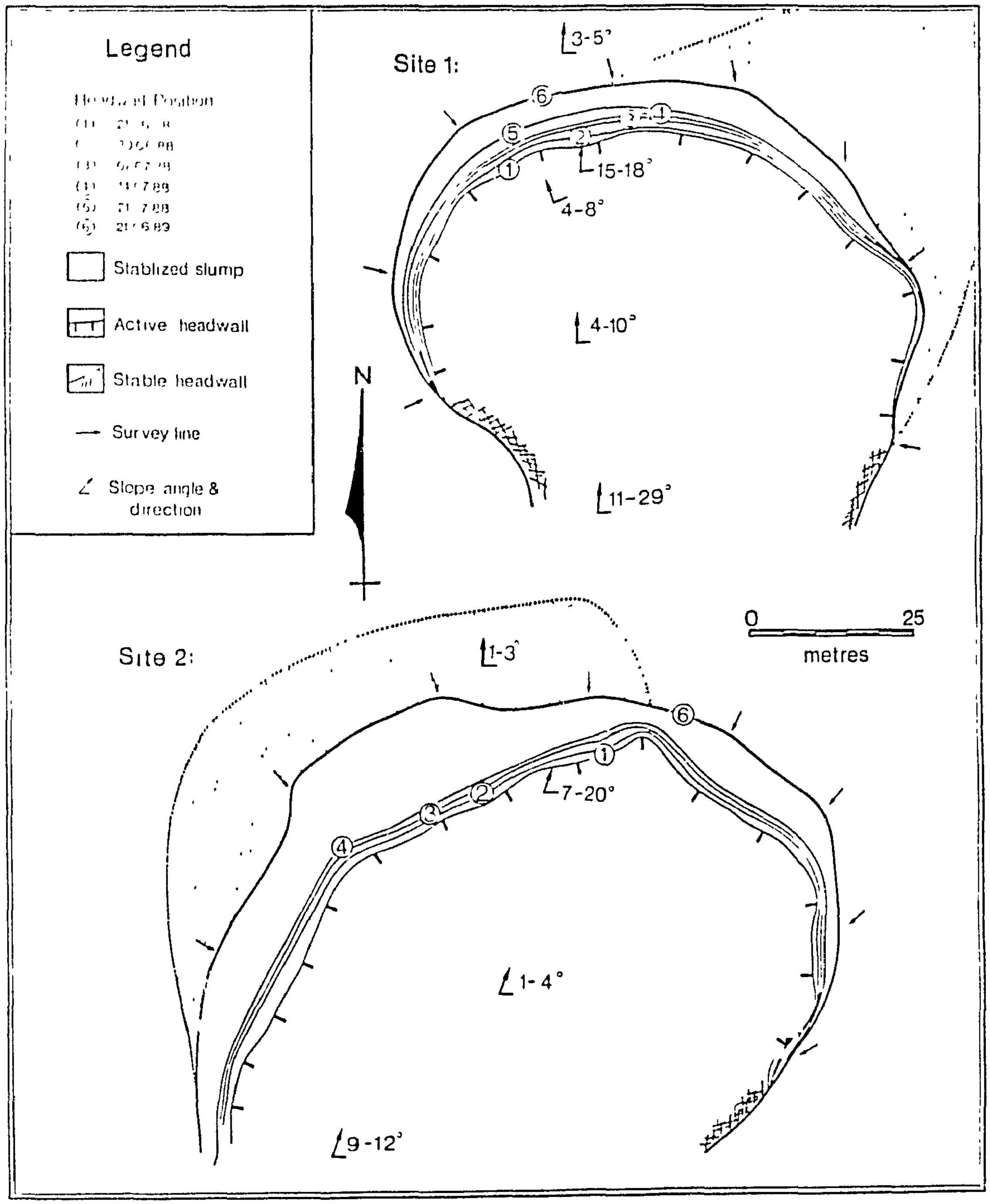
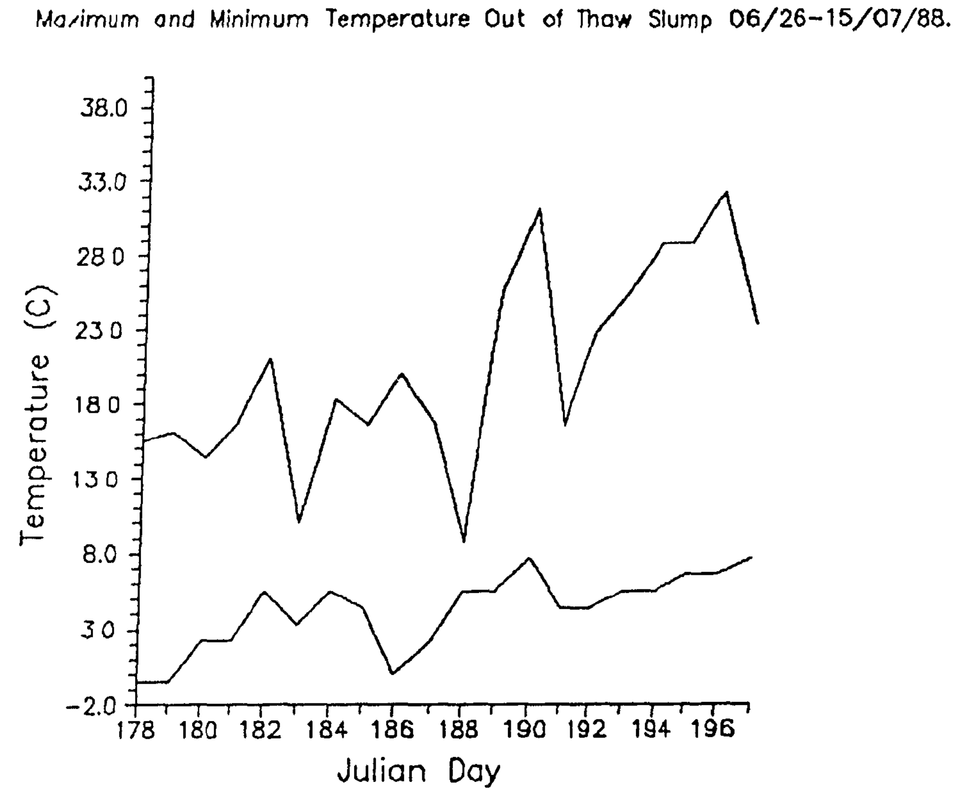


Figure 6.2. Diagram illustrating headwall retreat at RTS 1 and RTS 2 from June 1988 to June 1989.

Location	Rate (m yr^{-1})	Source
Arctic coast, U.S.S.R.	10	Are (1983)
Central Yakutia, U.S.S.R	7-9.5	Are (1978)
Ellet Rignes Island, N.W.T	7-10	Lamonthe and St Onge (1961)
E. Melville Island, N.W.T.	7-8	Hegmbottom (1984)
E. Banks Island, N.W.T.	7-10	French and Egginton (1973)
E. Banks Island, N.W.T.	6-8	French (1974)
S. Banks Island, N.W.T	14	Lewkowicz (1987)
Tuk. Peninsula, N.W.T.	7	Mackay (1986)
Garry Island, N.W.T.	7.2	Kerfoot and Mackay (1972)
Mackenzie Delta, N.W.T.	1.5-4.5	Mackay (1966)
Mayo, Yukon Territory	14-16	Burn (1989)
Colville Delta, Alaska	1-3	Walker (1983)
Macleod Point, Alaska	10-18	Black (1983)

Table 6.1. Documented annual retreat rates for retrogressive thaw slumps for various locations (Burns and Friele, 1989).



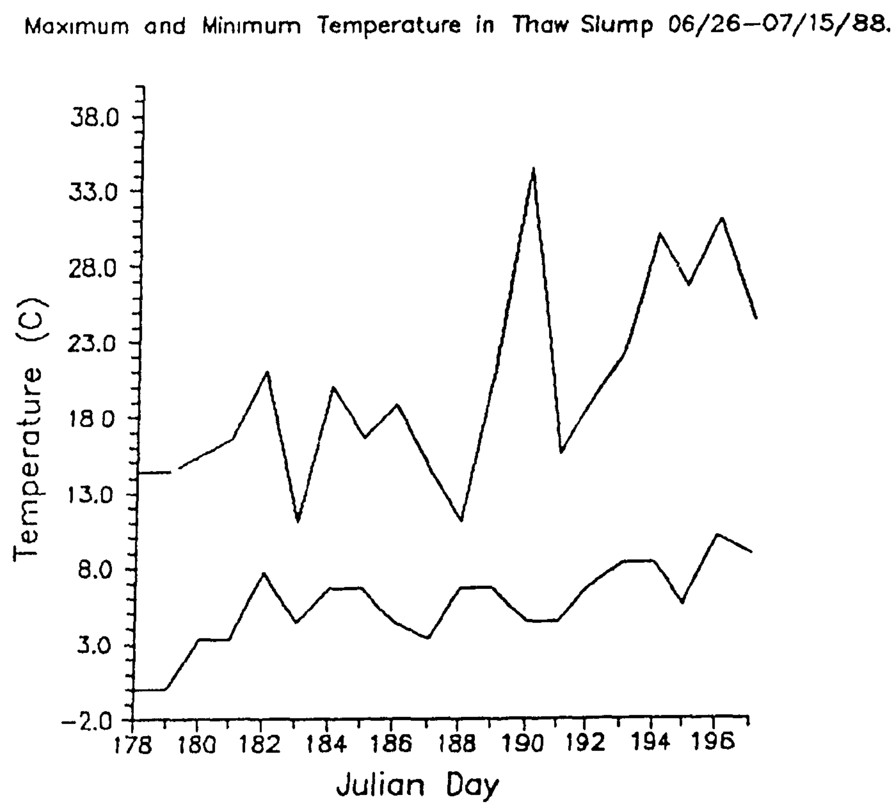
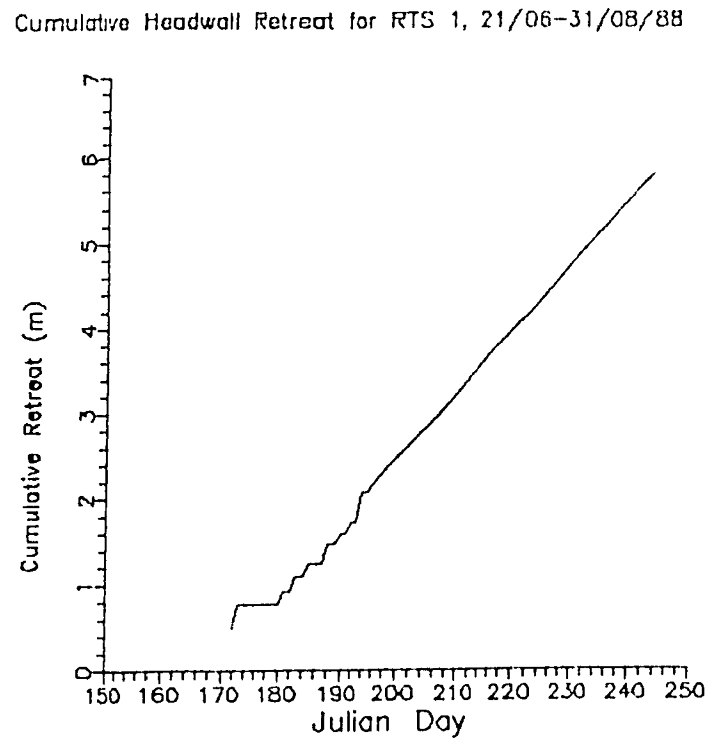


Figure 6.3. Maximum and minimum air temperatures recorded inside and outside the slump bowl at RTS 1.

A



B

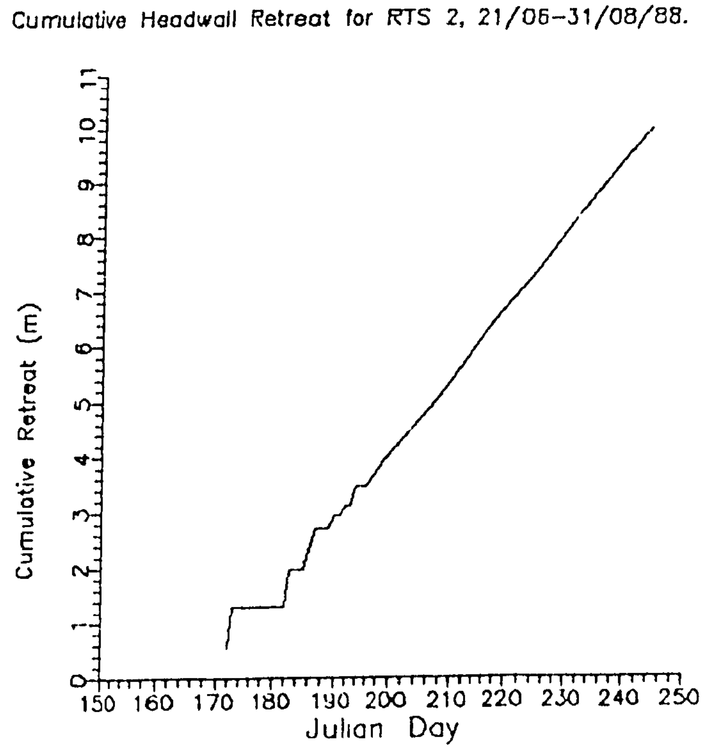


Figure 6.4. Cumulative headwall retreat (average retreat for the site) for (a) RTS 1 and (b) RTS 2.

rates diminished, or the section of the headwall stabilized quickly, when ground ice contents fell below 150%. Sidewall retreat is particularly slow because of low ice content, the rapid burial by mudflow from the headwall, and/or because the gradient of the slump floor perpendicular to the sidewall is low. Kerfoot (1969) notes that the critical factor influencing the longevity of thaw slumps is the maintenance of fresh exposures of substrata that have a high ice content. In comparison to slump headwalls, coastal bluff retreat over a distance of 100 m along Thetis Bay averaged only 30-50 cm per year, with a maximum retreat of 1 m.

Evolution of Thaw

In early June of 1988, the headwall of RTS 1-3, with the exception of the upper 1-2 m at each site were snow covered (Figure 6.5). Headwall retreat of the exposed portion of the face (primarily the active layer) began on June 9-10, approximately 16 days after mean daily temperatures were consistently above 0°C. Retreat at this time ranged between 0-15 cm/week. By mid-June slumped blocks of organics and sediment from the active layer accounted for up to 50-70 cm of weekly retreat at each of the sites. The highest rates of retreat in the early thaw season occurred along the south - facing headwalls at RTS 1 and 3, and the southeast - facing headwall at RTS 2. These segments were the highest parts of the ice faces, cleared of snow and debris, and contained

Figure 6.5. Photograph of snow covered headwall in late June 1988. Snow remained within slump bowls until the first week of July.

massive segregated ice and ice-rich permafrost. Even though active headwall retreat was taking place, thawed debris from the exposed portion of the face and active layer toppled onto snowbanks within the slump bowl. This acted as insulation which, in turn, reduced rates of ablation of the lower 4-5 m of the ice face.

By early July, warmer air temperatures (average temperature at Shingle Point for the first week in July was 13.7°C) caused higher rates of snowmelt and increased exposure of the headwall. This resulted in a corresponding increase in ablation rates of the ice face and headwall retreat. By July 6-7 all snowbanks within slump bowls had melted. At this time the south - facing headwall of RTS 3 was very active with recorded retreat rates of 10-45 cm/day. However, as the thaw season progressed the rapid accumulation of slumped material buried the face and stabilized most of the south - facing headwall (Figure 6.6). Simultaneously, the west - and southwest - facing segments of the headwall became very active and experienced retreat rates of up to 50-80 cm/day. This portion of the headwall displayed massive ground ice in the form of layers and ice-wedge ice. In the case of RTS 1 and 2, retreat rates were highest on the south - facing portions of the headwall. Retreat along the west - and east - facing headwall of RTS 1 was very limited and only localized. At RTS 2, retreat for a period of approximately six weeks was as high

Figure 6.6. Slumped material caused stabilization of the south - facing headwall at RTS 3 (July 1988).

as 6.7 m for the east - facing headwall and 9.8 m for the west - facing headwall. The highest retreat rates at all sites occurred from mid-July to early August, during which time warm temperatures caused high ablation rates and the collapse of large blocks of thawed material up to 1.5 m in length from the overlying active layer. Based on survey data and the analysis of slump shape change, it is suggested that the relatively high rates of retreat, particularly at RTS 2, are due to the combined influence of high ice content, southerly aspect and above average air temperatures within the slump bowl.

6.2.4 Runoff

Drainage channels develop on the slump floor both as meltwater from the ice face and residual snowbanks and water from supersaturated mud lobes drain. A dendritic pattern of drainage channels deeply incise stable parts of the bare slump floor at all sites (Figure 4.4). The largest channel found at RTS 1 had a depth of 1.0-1.5 m. By July, with headwall retreat very active, the floor at the base of the headwall, and an area up to 5 m away, was covered by a mass of supernatant mud. A dam of plastic mud at the base of the headwall maintained this condition until a lobe of fluid mud either breached or overflowed the obstruction. These lobes move downslope and are bordered by mud levees left by previous flows. The lobes at the terminus of the mud stream develop transverse ridges and both radial and transverse cracks after hardening. Flow at the base of the slump is concentrated into streams which carry dirty water and liquid mud. Thus, there is a succession from less stable material near the headscarp to stable material away from the headwall (Figure 5.7). Survey markers were installed in the slump floor to monitor movement of material. Although measured at weekly intervals for the greater part a season it was found that the floor at all sites was not moving, but that material was sliding over the existing slump floor. This resulted in the burial of survey markers. As an alternative, measurements were taken in order to monitor the growth of mud lobes on the floor at RTS 1 and 2. Over a period of four weeks mud lobes grew 16 m and 20 m in length at RTS 1 and 2, respectively.

6.3 Active Layer Slides

A major problem confronting researchers attempting to document active layer slide processes is predicting where they will occur and observing and measuring the process taking place. Active layer slides are a thaw-induced rapid mass movement involving failure and mass displacement of material by a single event. This differs significantly from the gradual growth of retrogressive thaw slumps. Consequently, the analysis of active layer slide occurrence is based mostly on *a posteriori* evidence and on a combination of material properties, timing of the event and meteorological data.

During the period of this study, several active layer slides are known to have occurred (in the late summer of 1987, one slide was witnessed by a biologist training park rangers). Thus, this discussion is based on a combination of inferred and observed events.

Physical Characteristics

Air photo interpretation and field reconnaissance by foot and helicopter revealed that more than 50 active layer slides have occurred on Herschel Island. Based on the degree of revegetation, and comparison between 1970 air photos and the present, more than 40% of the 30 slides in the Thetis Bay area may be considered to be recent. It is apparent that active layer slides on Herschel Island have occurred without a preferred orientation. Furthermore, they occur on both gentle ($< 10^{\circ}$) and steep ($> 30^{\circ}$) slopes and on bare to well-vegetated surfaces (Figure 4.8).

Slides ALS 1-4 occurred during a two week period beginning in mid-August, 1987. Of particular interest is an eyewitness report of the largest of five slides that developed in this period at ALS 1. The slide happened over a period of twenty minutes on August 13, 1987 (personal communication, Dr. D. Pattie, Northern Alberta Institute of Technology, and Mr. F. Elanjk, Yukon Territorial Government). According to these reports, "movement was initially slow, but velocity increased as material progressed downslope. Failure and sliding of a large section of slope consisting of pieces of the vegetation mat and nearly one metre of underlying sediments exposed liquified mud and ground ice in the slide floor". Subsequently, the slide floor and sidewalls were covered by a layer of smoothed, grooved wet clay forming a slickenslide (Figure 6.7). Slickenslides are common in this form of slope failure, and are generally indicative that the type of mass wasting process taking place is that of sliding along a translational failure plane.

Meteorological Observations

To better explain the slide activity in August, 1987, meteorological records for Shingle Point, as well as fragments of data for Herschel Base Camp, were analysed. These data reveal that in July 1987, temperatures were 2-3 $^{\circ}$C warmer than the average 11°C at Shingle Point. Meteorological data recorded at base camp revealed a mean average monthly air temperature of 14°C for July, and temperature extremes of 24.6°C recorded at Shingle Point. Warm temperatures would initiate early snowmelt and enhance thawing of the active layer. Rainfall for the Yukon Coastal Plain and Herschel Island is generally low. Total annual precipitation for the arctic slope averages less than 200 mm, most of which falls as rain or drizzle during the summer months. According to territorial park records, the last heavy rainfall recorded at Herschel Island was on August 8, five days prior to the

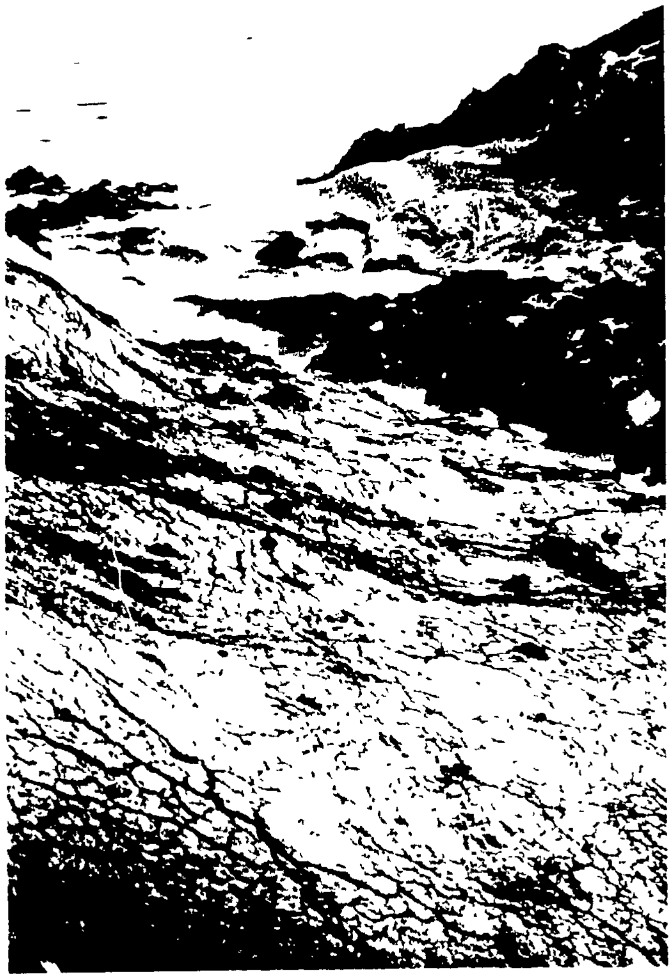


Figure 6.7. Slickenslides representative of sliding having taken place.

occurrence of the active layer slide on August 13. Although heavy rainfall is frequently a triggering mechanism of conventional slides, they do not seem to be directly responsible in this case. A model for active layer slide occurrence is proposed later in this chapter.

Field Observations

In the summer of 1988 survey markers were installed above the slide headwalls and within the floors in order to monitor subsequent change in form. No further movement was recorded on slide floors, but headwall retreat of up to 1.0 metre at ALS 1 and 1.5 m at ALS 2 was documented. As no icy permafrost was exposed in the headscarp, ablation *sensu stricto* was not the mechanism of retreat. A network of tension cracks oriented perpendicular and parallel to the headwall widened, and blocks of material broke away from the headwall as the season progressed. These blocks tended to tumble only a few metres from the face. Slide floors were dry, except for areas immediately below the headwall or at the slide base which were characterized by damp spots and ponding of water. The source of water was either from the melting of snow that naturally collects in these depressions or from the melting of aggradational ice exposed in some of the headwalls. Water pooled at the base of the headwall and eventually flowed into drainage channels that had developed in the slide floors.

Water also pooled at the lower end of the scar because the displaced material blocked further drainage, particularly where material developed a ridged corrugated appearance (Figure 4.7.b). Each ridge is the width of the slide, and therefore acts as a barrier to the movement of water downslope. Ponding also occurred because of the nearly impermeable nature of the underlying frozen sediment at the base of the active layer.

Failure planes occur at depths of 60-100 cm, with an average of 70 cm. In 1988, active layer development which was monitored at ALS 1, 2, and 3 was found to range from 20 cm to a maximum of 65 cm in depth by August. It was also observed that active layer depths tended to increase in the upslope direction and were deepest in areas at, and immediately above, the slide headwall. Active layer slides occur along seepage zones and on relatively steep slopes. Slope angles at the point of detachment (failure) range between 10-30°. Areas above slide headwalls recorded lower slope angles of up to 8° (Figure 4.9). During excavation of the slide headwall at ALS 2 in 1988, a 2-5 cm thick ice lens was exposed at a depth of 70-80 cm. Furthermore, ice-rich sediments and semi-continuous ice lenses and layers up to 5 cm in thickness were also observed at the base of the active layer in the exposed sections of thaw slumps. Given the general lack of precipitation, moisture essential for active layer slide occurrence was probably provided by the melting of subsurface aggradational ground ice, caused by the downward penetration of the active layer.

6.4 Discussion

The occurrence of retrogressive thaw slumps and active layer slides on Herschel Island has resulted in extensive terrain disturbance. Observation and measurement of retreat revealed that retrogressive thaw slump growth is the result of five interacting processes, including (1) detachment and freefall, or sliding, of blocks of sediment and organics from the active layer, (2) thaw of ice and ice-rich material from and immediately below the active layer, (3) ablation and thermal erosion of the ice face, (4) sloughing of material off the ice face, and (5) mudflow at the base of the thaw face and within the slump floor. Each process contributes, within a broad polygenetic framework, at different stages in the seasonal development of the retrogressive thaw slumps. Collapse of large blocks of sediment in response to ablation did not occur on a daily basis, but increased in frequency as the season progressed. Where ice wedges penetrated massive ground ice bodies or ice-rich sediments, they tended to melt slower than the enclosing ground ice. This may be due to differences in crystal size, shape, dimensional orientation, c-axis orientation characteristics and sediment associated with different ice types (Pollard, 1989; 1990). Differential melt rates resulted in the formation of protruding ice masses, overhangs and uneven retreat across the headwall. As thawing progressed the overhangs were undercut and unstable material either slid down the ice face, or toppled "catastrophically" from a height of several metres to the base of the ice face. Where closely spaced ice wedges truncate relatively ice-poor sediment their ablation also caused the intervening sediment to be isolated as separate blocks which also collapse. Failure of blocks of thawed and frozen material up to 50 m^3 from the slump headwalls was documented. Based on regular measurement, an estimated 10-15% of retreat was attributed to this mechanism. These processes are self-perpetuating until stabilization occurs by one or several mechanisms, for example when the ice content of the frozen ground changes or the ice body completely melts (Lewkowicz, 1988).

It is difficult to fully comprehend the cyclic nature of retrogressive thaw slumps. However, in order for retrogressive thaw slumps to be polycylic no one thermokarst event can be responsible for complete melting of a ground ice body, and a disturbance is needed to reactivate retrogressive thaw slump activity. Areas containing massive ground ice which is several metres in depth and several metres in horizontal extent may require two or three cycles of thermokarst over the course of many years (e.g., RTS 1 and 2). Retrogressive thaw slumps which are not cyclic (RTS 4 is not situated within a former thermokarst scar) develop in areas of limited ground ice which seems to melt completely within the course of a thermokarst event.

Active Layer Slide Model

Based upon evidence provided by stratigraphic and material analysis together with observations on topography and active layer development, the following genetic model for active layer slide occurrence on Herschel Island is proposed.

Active layer slides develop during summers experiencing above average temperatures which produce deeper than average seasonal thaw. As the active layer propagates beyond the normal range of development it begins to melt a zone of aggradational ice which often occurs near the top of permafrost. In areas of fine-grained sediment, like the marine silts and clays which make up this part of the island, the supernatant water released by the melting of aggradational ice does not readily drain, creating a zone of high pore water pressure. Furthermore, the material may be slightly thixotrophic due to the marine clays that make up 30% of its volume. Since the active layer tends to mirror the surface topography, the shallow valleys that characterize the seepage lines where active layer slides occur experience a slightly greater accumulation of supernatant water and thus are weakened further. A critical situation develops where the shear stress of the increased active layer mass (produced by the increase active layer depth) and the slope angle will exceed the cohesive strength of the sediment. Strength associated with the effective normal stress is diminished by increased pore water pressure. Failure proceeds along a plane a few centimetres above the top of the permafrost table. Frictional resistance along this plane would be reduced by the presence of lenses and layers of ice that are concentrated in the aggradational ice zone. Thus, rather than precipitation acting as a triggering mechanism, a period of warm temperatures may actually precipitate the occurrence of active layer slides. Failure appears to begin at the headwall, and the entire failed mass moves downslope intact. This is clearly reflected in the displaced material where a ridged or corrugated appearance is common. Where the slide tunnels into a narrow drainage channel it will accelerate and cause greater deformation of the failed mass (ALS 1).

The slickenslide surface which characterizes slide floors, and its often grooved appearance clearly suggest that failure occurred in either thawed or plastic material. Given the possible saline fine-grained nature of the sediment, this zone of failure may actually exist a few centimetres deeper than the position of the 0°C isotherm. The slickenslides rarely exceed 1-2 cm in depth; thus, it is assumed that failure occurs very close to the boundary between the aggradational ice and underlying plastic sediment.

Terminological Alternatives

The general term "active layer failure" has been adopted by the Permafrost Subcommittee (1988) to include all slope failures which occur in arctic environments. The term "active layer slide", which is frequently used in the permafrost literature, describes movement of material along a failure plane. Since sliding is the dominant genetic process and the failed mass is the active layer, the term is properly used. However, a problem arises with respect to use of the term "retrogressive thaw slump", more specifically, use of the term "slump" to describe this thermokarst landform. Since retrogressive thaw slumps are a distinctive form of thermokarst involving thaw subsidence of ice-rich permafrost through the backwearing of an exposed ice face, then the dominant genetic processes are melting and thermal erosion of exposed ice and icy sediments. Secondary processes include small-scale slumping and block failure as surficial material with low ice contents are undercut, and flow of supersaturated sediments in the slump floor. Therefore, the use of the term "slump" to define these features is geomorphically incorrect, because slumping processes are not dominant. Furthermore, morphometric indices developed for true slumps in nonpermafrost areas (Crozier, 1973), although similar for retrogressive thaw slumps, do not apply because retrogressive thaw slumps form by very different processes. Unfortunately, the misuse of the term slump with reference to these forms is perpetuated by the current literature (Permafrost Subcommittee, 1988). It seems quite clear that the terminology referring to this process should be revised. A new term should have both clear genetic and morphologic connotations. The terms "thermokarst basin" or "thermokarst sink" may more readily fulfill this requirement. The new term could be supported by descriptive modifiers such as retrogressive and/or polycyclic.

Chapter 7

Summary and Conclusions

7.1 Introduction

Ground ice is an important component of permafrost sediments and plays a major role in the evolution of periglacial landscapes. Considerable research has been undertaken to investigate the nature and origin of ground ice, as well as landforms associated with its occurrence. However, in view of the predicted warming of arctic environments associated with global climate change, the study of thermokarst and thaw-induced processes and landforms will become increasingly important.

This thesis summarizes the investigation of retrogressive thaw slumps and active layer slides on Herschel Island, Yukon Territory. Over the course of this study it became apparent that these are two of the most dynamic landforms modifying the terrain in this part of the western Arctic. This study focused on three aspects of retrogressive thaw slumps and active layer slides, including (1) morphology and setting, (2) material properties and ground ice characteristics, and (3) process.

7.2 Retrogressive Thaw Slumps

Most of the retrogressive thaw slumps on Herschel Island, including those studied for this research, are situated along the shoreline. These areas are geomorphically very dynamic, experiencing fluvial erosion, gully formation and coastal retreat processes which are perpetuated by the melting of snow and exposed ground ice during the thaw season. Furthermore, ground ice is widespread on Herschel Island, and is mapped as "massive ground ice" in areas of retrogressive thaw slump activity (Bouchard, 1974; Pollard, 1989). Thus, the potential for retrogressive thaw slump occurrence is high since these thermokarst landforms require ground ice and a disturbance (e.g., fluvial erosion, thermal erosion of ice wedges) for their initiation and growth.

Retrogressive thaw slump morphology is a result of several interacting processes and cryostratigraphic characteristics which vary within and between slumps. RTS 1 and 2 display the classic amphitheatre shape. This is a result of relatively homogeneous ground ice conditions across the slump headwall. However, the morphology of RTS 3 is influenced by at least twenty-three ice wedges exposed in the headwall. These ice wedges measured up to 10 m in depth and 2-3 m in width. Ice wedges melted slower than the enclosed massive ground ice and ice-rich sediment, resulting in irregular rates of retreat across the slump headwall, and a jagged or serrated headwall appearance. It is clear that ground ice is needed for retrogressive thaw slump occurrence; however, slump growth is the result of five processes: (1) detachment, followed by falling or sliding, of blocks of sediment from the active layer, (2) thaw of ice and ice-rich material immediately below the active layer, (3) ablation and thermal erosion of the ice face, (4) sloughing of material off the ice face, and (5) mudflow within the slump floor. Each process contributes to slump growth during the thaw season. The most important process is ablation and backwearing of the ice face.

7.3 Active Layer Slides

Active layer slides on Herschel Island are situated on a variety of slopes displaying a range of aspects, slope angles and material properties. Unlike retrogressive thaw slumps, the active layer slides on the island are not limited to the vicinity of the shoreline, but are widespread on the island. However, many do occur along seepage lines, and those at ALS 1 are upslope from ephemeral drainage channels.

Active layer slides have a general appearance similar to mudslides of the temperate environments, except that they are shallow in nature. On Herschel Island active layer slide headwalls reached maximum depths of 1.5 m. The sites examined differ in dimension and slope characteristics; these differences are best explained as a result of varying topographic conditions. For example, slides not restricted by topographic conditions such as narrow channels or steep ridges tend to cover an extensive area (e.g., ALS 4 covered an area of 9,860 m^2 as compared to a slide scar at ALS 1 which measured 275 m^2).

Since active layer slides are a form of rapid mass movement, it is very difficult to obtain process measurements. However, based upon personal communication (Dr. Pattie, Northern Alberta Institute of Technology), it is known that the slides examined (ALS 1-4) occurred over a two week period in August 1987. In July 1987, temperatures were 2-3°C warmer than the average 11°C. These warm temperatures would have caused a deeper than average active layer. As the active layer reached depths of 70-80 cm a zone of aggradational ice found at the top of permafrost melted, which, combined with material properties and the increased load of a deepening active layer, caused failure along the zone of aggradational ice.

7.4 Conclusions

From the field studies undertaken at Herschel Island, and laboratory analysis the following conclusions can be drawn:

(1) Retrogressive thaw slumps and active layers slides are currently some of the most active degradational geomorphic processes occurring on the island. In 1989, an aerial survey of the northern part of the island revealed slump and slide frequencies of 6-7 per km^2 and 10 per km^2, respectively. Furthermore headwall retreat rates documented at Herschel Island are the highest for this part of the Arctic.

(2) Five processes contribute, within a broad polygenetic framework, to different stages in the development of retrogressive thaw slumps. The most important process is backwearing of an exposed ice face. Retrogressive thaw slumps are most active (e.g., block failure), and have the highest retreat rates, late in the thaw season, averaging up to 0.80 cm of retreat per day.

(3) Retrogressive thaw slump development is influenced primarily by ground ice conditions. Firstly, thermal erosion along ice wedges is a main triggering mechanism in slump initiation at Herschel Island. Secondly, RTS 1 and 2 have the classic amphitheatre shape, which reflects relatively homogeneous ground ice conditions across the headwall. However, RTS 3 has varying ground ice conditions. The ice wedges exposed in the headwall melted slower than enclosing massive ice and ice rich sediments. This resulted in irregular ablation rates across the headwall, which in turn influenced the overall shape of the slump.

(4) Active layer slides appear related to warm temperatures and not to precipitation events. Warm temperatures result in deep active layer development and increased pore water pressure associated with the melting of aggradational ice in marine sediments. This, combined with the fact that slides occur along seepage lines and have steep slopes at their point of detachment,

contributes to unstable slope conditions. Failure results in the form of sliding along a planar surface at the top of permafrost.

(5) The term retrogressive thaw slump is geomorphically inaccurate because thermokarst is not a gravity induced process. On the basis of observations at Herschel Island it is proposed that the terms "thermokarst basin" or "thermokarst sink" be adopted.

(6) Based upon air photo analysis and helicopter reconnaissance it can be concluded that retrogressive thaw slumps and active layer slides are key degradational landforms. There is evidence of a long history of thermokarst (e.g., the presence of numerous slump scars) in this part of the arctic, hence, one should be careful when attaching a global climatic significance to the current trend in these phenomena. However, without a doubt, areas of ice-rich permafrost will experience more thermokarst if current annual air temperatures increase.

7.5 Recommendations for Future Research

During the course of this research it became apparent that four avenues of research should be given further consideration:

(1) During this study no one direct observation was obtained to explain the polycyclic nature of retrogressive thaw slumps. A long-term study may provide insight into the

cyclic nature of retrogressive thaw slumps.

(2) Although logistical support was unavailable for a detailed micrometeorological study it was observed that the microclimate within a thaw slump is very different from the local and regional climates. A detailed microclimatic study could provide more information on ice ablation, rates of headwall retreat, slump stabilization and subsequent revegetation.

(3) To investigate further the ground ice petrography of various ice types and its influence on ablation of the ice face.

(4) A geotechnically oriented study on active layer slides is needed focusing on the relationship between material properties and thawing in order that slide occurrence may be better understood.

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