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LA THÈSE A ÉTÉ
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ASPECTS OF THE PERMAFROST GEOMORPHOLOGY
OF SOUTHWEST BANKS ISLAND
WESTERN CANADIAN ARCTIC

by
David G. Harry

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

UNIVERSITY OF OTTAWA
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'In Nature's infinite book of secrecy
A little I can read'

William Shakespeare

'To spend a summer on Banks Island as
we did that one was a delight . . .
Here was a beautiful country of valleys
everywhere gold and white with flowers
or green with grass . . .'

Vilhjalmur Stefansson

CHAPTER ONE

INTRODUCTION

1.1 Permafrost geomorphology

Permafrost is defined as a ground temperature condition in which earth materials remain below 0°C for two or more years (Muller, 1947, p. 3). Permafrost geomorphology thus comprises the study of landforms and processes related to the existence of perennially frozen ground. Early studies in this field consisted primarily of paleogeographic reconstructions within areas of Europe which were peripheral to the zone of Pleistocene glaciation. Geomorphological phenomena associated with this environment were first described as 'periglacial' by W. Lozinski in 1909 (Dylikowa, 1962; Jahn, 1954). However, subsequent usage of this term quickly expanded to include all processes and landforms believed to be associated with cold, non-glacial climates. The apparent relationship between cold climate conditions and certain geomorphological processes was emphasized by many workers (e.g. Büdel, 1953; Troll, 1944), and is implicit in Peltier's (1950) definition of a 'periglacial morphogenetic zone'. Moreover, the widespread application of this concept focused research upon uniquely periglacial phenomena, such as patterned ground and freeze-thaw processes, and discouraged analysis of the arctic landscape as an integrated system.

In recent years, the relatively narrow approach to arctic terrain science imposed by this framework of climatic geomorphology has been replaced by one stressing the importance of frozen ground phenomena. In addition, the necessity of undertaking long-term process studies has been recognized. It is now generally accepted that, on the macro-scale, arctic landscapes primarily reflect the action of azonal processes, particularly fluvial erosion and deposition, and mass wasting. On a more local scale, the importance of permafrost as an initiator of landforms, particularly in areas underlain by ice-rich unconsolidated sediments, is well established (e.g. Brown, 1974; Mackay, 1963a; Rampton, 1974).

Some researchers have equated the periglacial domain with the extent of permafrost (e.g. Péwé, 1969, p. 4), however the relationship is complex since not all periglacial processes and phenomena are restricted to the permafrost zone (e.g. Embleton and King, 1975, p. 3; French, 1976a, pp. 2-4; Karte, 1981, p. 46; Tricart, 1970, p. 58). The term 'periglacial' is thus both anachronistic and imprecise when applied to many geomorphological studies in high latitudes. For this reason, it has been suggested that 'permafrost geomorphology' (French, 1979) or 'geocryology' (Washburn, 1979) more appropriately describe modern arctic terrain investigations.

In the Soviet Union, geocryology, referring to a rather broader general permafrost science, is a

well-established discipline (e.g. Kudryavstev, 1978; Tsytoich, 1975). This reflects in part the long history of settlement and scientific research in northern regions of the U.S.S.R. Studies concentrating entirely upon the characteristics of permafrost terrain and associated cold climate processes are common (e.g. Grigor'yev, 1976; Kudryavstev et al., 1977; Romanovskii, 1977; Vt~~o~~pin, 1975; Vtyurina, 1974; Vyalov et al., 1973; Zhestkova et al., 1969). The rapid development of this branch of earth sciences in North America and the U.S.S.R. is best illustrated by the proceedings volumes of the Second and Third International Conferences on Permafrost (National Academy of Sciences, 1973; 1978; National Research Council of Canada, 1978; 1979). Permafrost research has also commenced in the Peoples Republic of China, which contains the greatest extent of permafrost within its territory after Canada and the U.S.S.R. (Academia Sinica, 1980).

1.2 Permafrost and ground ice terminology

Permafrost is defined by its temperature, rather than by the amount or phase status of its moisture. Thus, if the freezing point of groundwater is depressed below 0°C, permafrost soils may contain significant quantities of unfrozen moisture (e.g. Williams, 1967). It is necessary, therefore, to define permafrost terms unambiguously (Brown and Kupsch, 1974). Following Van Everdingen (1976), the term 'frozen' is used in the present study to imply cryotic

(i.e. subzero degrees Celsius) temperature conditions. The term 'talik' is used in the sense of a thermal (non-cryotic) talik, within which ground temperatures remain perennially above 0°C.

Most moisture within permafrost exists in the form of ground ice, as coatings on sediment particles, within pore spaces, as veins and lenses, or as massive bodies of segregated ice (Gell, 1976; Mackay, 1972a). Permafrost may contain excess ice (i.e. a water equivalent ice volume greater than available pore volume in the thawed sediment). If this material is unconsolidated, it may be classified as 'thaw-sensitive' (Van Everdingen, 1979), since permafrost degradation would result in surface subsidence and the development of thermokarst terrain (e.g. Czudek and Demek, 1970; Kachurin, 1962).

The distribution of permafrost in the northern hemisphere is illustrated in Figure 1.1 (Péwé, 1981). Within the continuous zone, permafrost is ubiquitous, except beneath deep water bodies and in areas of recent sedimentation, and may extend to depths of 500-1,000 m (Baranov, 1964, pp. 15-19; Brown, 1972; Grave, 1968). Permafrost may also exist beneath arctic oceans, either in equilibrium with negative sea bottom temperatures or in relation to former surface conditions (Lewellen, 1973; Mackay, 1972b). Since permafrost underlies nearly 50% of Canada and 80% of Alaska (Brown, 1978; Ferriáns, 1965), problems associated with frozen ground are a major factor

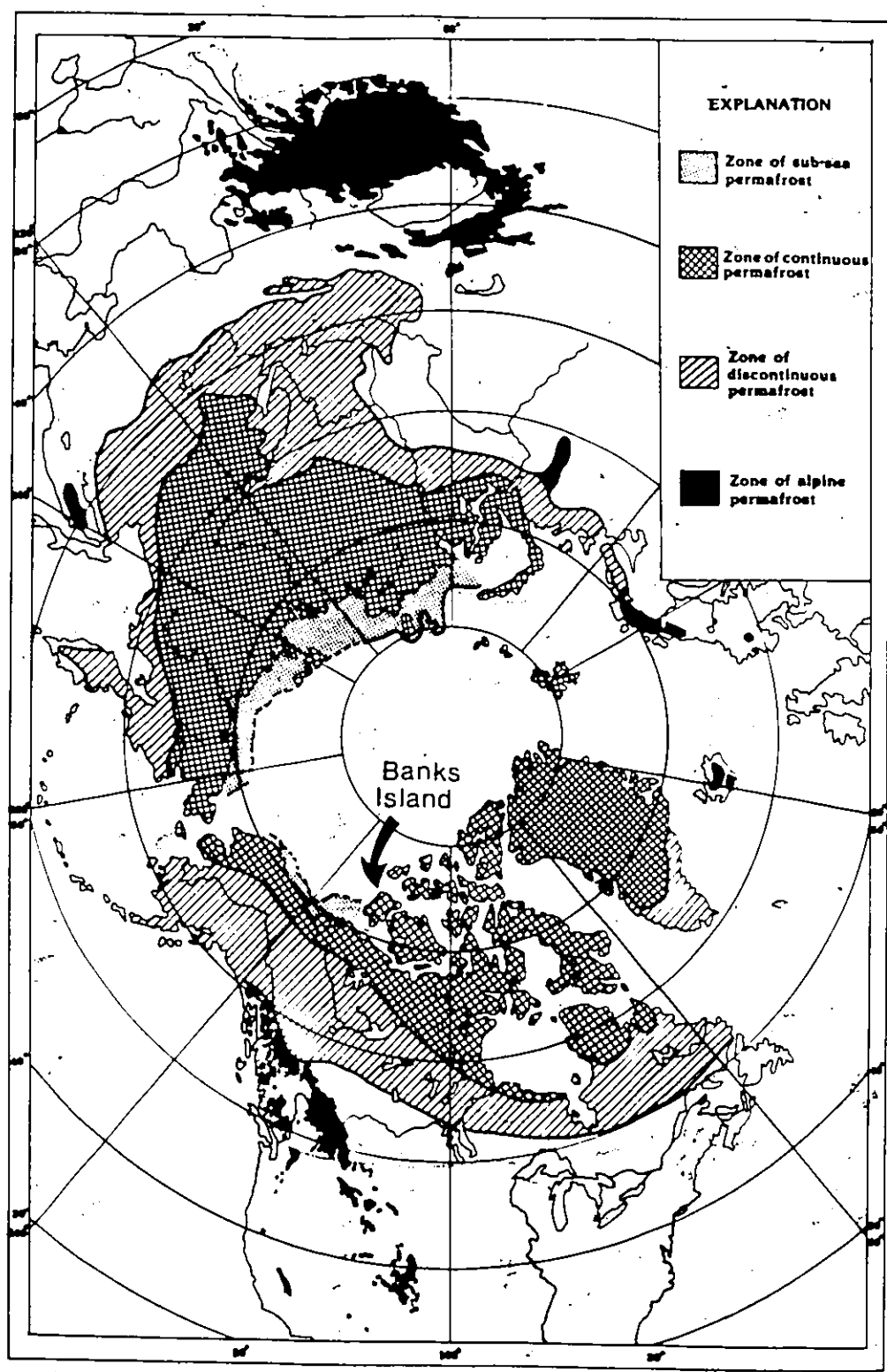


Figure 1.1 Permafrost distribution in the Northern Hemisphere (after Péwé, 1981).

influencing northern development (see e.g. Brown, 1970; Ferrians et al., 1969; Muller, 1947). Interest in the permafrost region of Arctic Canada has increased considerably in the last thirty years, stimulated to a great extent by the search for hydrocarbon reserves. Detailed studies of permafrost terrain form a necessary prerequisite both to its efficient utilization and protection from irreversible damage (see e.g. Brown and Grave, 1979; French, 1978; 1980; French and Smith, 1980; Lawson et al., 1978).

1.3 Research aims and strategies

The present study investigates the relationship between permafrost conditions, geomorphic processes and landscape evolution within one area of the Western Canadian Arctic. It thus falls directly within the field of permafrost geomorphology. The area chosen for study lies within a 50 km radius of Sachs Harbour (71°59'N, 125°17'W), located at the southwest extremity of Banks Island on the coast of Thesiger Bay (Figure 1.2). Reconnaissance fieldwork suggested that permafrost has a direct influence on three aspects of meso-scale landform on southwest Banks Island.

First, ground ice aggradation in unconsolidated sediments of Quaternary age gives rise to a number of surface features, the most notable of which are pingos and ice-wedge polygons. Sections excavated through pingos and exposed in coastal cliffs enable the stratigraphic

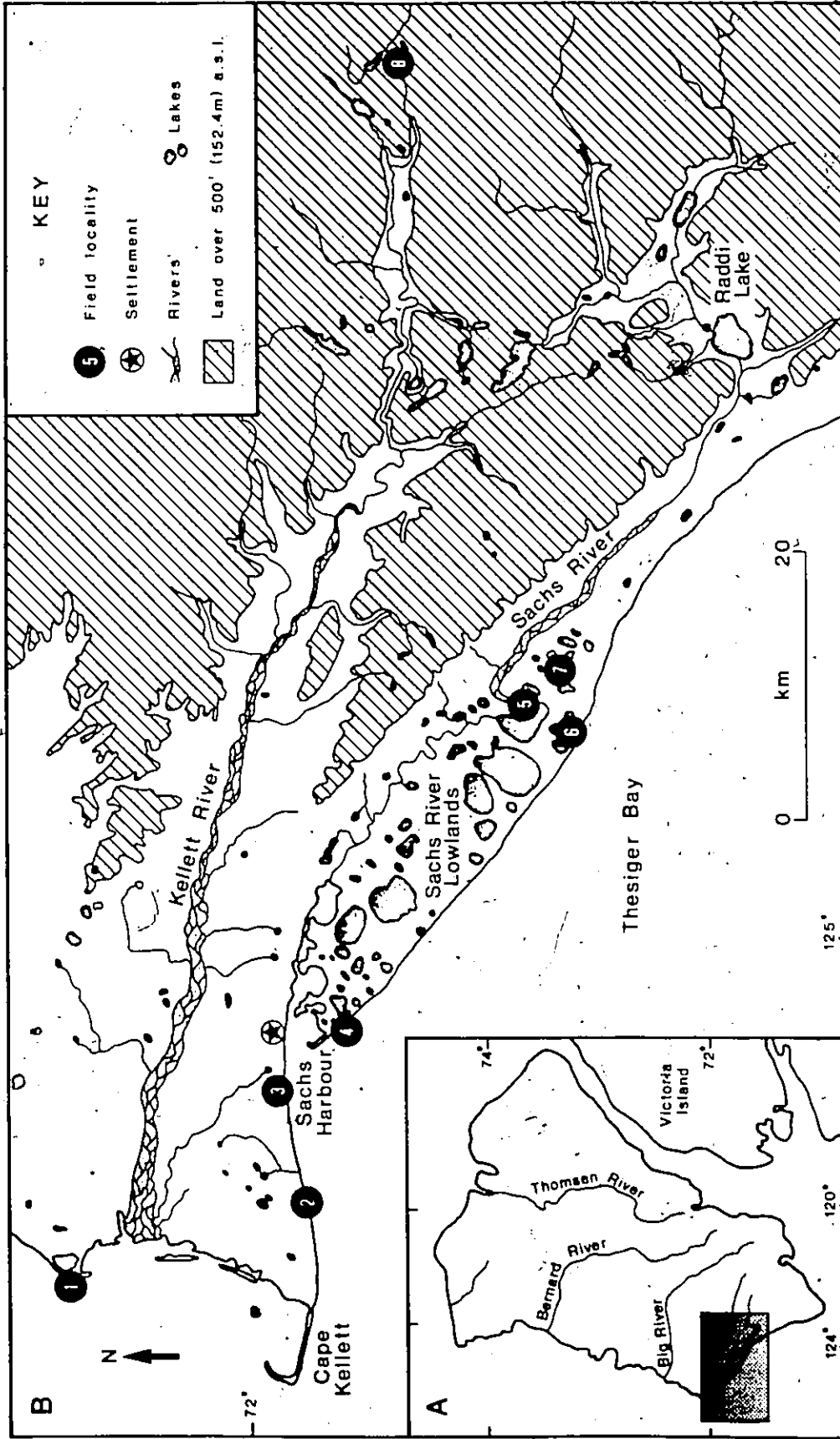


Figure 1.2 Location of the study area, southwest Banks Island. (a) Regional setting. (b) Physiography of southwest Banks Island. Key to numbered field localities: (1) Blue Fox Harbour, (2) Duck Hawk Bluff, (3) Allen Creek permafrost sections, (4) Sachs River lowlands permafrost section and truncated lake basin, (5) Carpenter Lake pingo, (6) Partially drained lake in the Sand Hills area of ice-cored terrain, (7) Sand Hills ice- ablation feature, (8) Upper Kellelt River pingo.

analysis of ground ice bodies, as they relate to terrain characteristics. Second, the melt-out of ice within permafrost has led to the formation of numerous thermokarst lakes and tundra ponds. Third, coastal evolution has been strongly influenced by thermal and mechanical erosion of ice-cemented sediments. This has resulted in rapid cliff retreat and, in some areas, the truncation and drainage of thaw lakes. A number of rapidly changing constructional shoreline features further testify to the dynamic nature of the coastal system. These observations suggested that a significant contribution to knowledge would result from an understanding of the relationships between contemporary permafrost conditions and terrain characteristics. Furthermore, the study of landscape evolution, in relation to episodes of permafrost aggradation and degradation, is central to the geomorphology of ice-rich permafrost terrain. Within this context, three specific aims are identified:

- (1) To investigate the evolution of permafrost and ground ice stratigraphy, within a framework of Quaternary paleoenvironmental reconstruction.
- (2) To explain the origin of thaw lake terrain, with particular emphasis upon the subsequent development of an equilibrium basin morphology.

- (3) To study the nature and rate of coastal change in ice-rich permafrost terrain.

Consideration of these themes suggested the utilization of four distinct but complimentary research strategies: stratigraphy, morphometry, thermal modelling and process observation. The nature of the permafrost-ground ice system is closely related to both the thermal and depositional histories of enclosing earth materials. For this reason, stratigraphic analysis is essential to any investigation of permafrost geomorphology. The study of thaw lake terrain represents a spatial problem and is thus most readily approached by morphometric analysis. However, thaw lake development is intimately related to disturbance of the subsurface temperature field. In this study, the geothermal effect of thaw lakes is simulated by computer models, based on heat conduction theory. The coastal zone in ice-rich permafrost terrain is characterized by rapid change. It is thus possible to identify the character of short-term shoreline evolution by direct observation, supplemented by the analysis of sequential air photo coverage.

1.4 Fieldwork and logistics

The study area was selected for several reasons. First, Banks Island has been the focus of systematic geomorphic and Quaternary research during the last decade by H. M. French and associates (Egginton, 1976; French,

1970a, b; 1971a, b; 1973; 1974a, b; 1975a, b; 1976b; French and Dutkiewicz, 1976; French and Egginton, 1973; French and Lewkowitz, 1981; Lewkowitz, 1978; 1981; Lewkowitz and French, 1982a, b; Lewkowitz et al., 1978; Pissart and French, 1976), and by officers of the Geological Survey of Canada (Pissart et al., 1977; Vincent, 1978a, b; 1979; 1980). Thus a broad framework exists, within which a more localized study of permafrost geomorphology may be undertaken.

Second, Sachs Harbour provides a good base for such investigations, since it is accessible by regular air service from the mainland. This increases the flexibility of fieldwork scheduling and reduces logistical costs. Third, the coastal lowlands of southwest Banks Island contain excellent examples of ground ice landforms, thaw lake terrain and a dynamic coastal system. These features have not previously been examined in detail and together they offer a unique opportunity to study the relationship between permafrost, geomorphic processes and landforms. Fourth, a number of sources contribute to a high level of base-line data availability. These include a Department of Transport Upper Air station providing climatic records dating from November 1955, sequential panchromatic air photo coverage at intervals from 1950 onwards, and 1:50,000 scale National Topographic Series photomaps. Finally, the Inuit inhabitants of Sachs Harbour provide useful local information. Their intimate knowledge

of terrain conditions, gained by first-hand experience and at all times of the year, forms a reservoir of data pertaining to local geomorphic phenomena.

Fieldwork on Banks Island was carried out during the periods June 15-August 22, 1979, July 14-September 04, 1980 and August 18-23, 1981. The Polar Continental Shelf Project facility at Sachs Harbour was used as a base. Access to thaw lake localities in the Kellett River valley and Sachs River lowlands, and to coastal permafrost exposures southeast and west of Sachs Harbour, was by Honda ATC-90 vehicle and by Inuit boat rental. Between July 10 and July 26, 1979, a camp was established near Carpenter Lake, approximately 30 km southeast of Sachs Harbour, to investigate a collapsed pingo and other ice-ablation features. On August 06, 1980, several sites near Fish lake, in the upper Kellett River valley, and at Blue Fox Harbour, were visited by helicopter.

Comparative studies of permafrost and terrain conditions in the Tuktoyaktuk-Mackenzie Delta area were carried out between June 06 and June 14, 1979. Visits were made to a number of sites on Richards Island, including Illisarvik and the Ya Ya borrow pits. Ground ice sections were examined at Peninsular Point and on the northwest coast of Pelly Island. A two-week reconnaissance program in the central and high Arctic Islands in August 1978, together with visits to Ellef Ringnes and Victoria Islands in the summer and fall of 1981, have also proved invaluable

in placing the present study within a broader arctic perspective.

CHAPTER TWO

SOUTHWEST BANKS ISLAND

2.1 Introduction

Banks Island lies in the Western Canadian Arctic between latitudes 71°N and 75°N, and its coastline flanks the western approaches to the Northwest Passage. With an area of 60,165 km², it forms the fifth largest island in Canada. Archaeological evidence indicates the presence of Inuit hunting parties at least 3,600 years ago (Arnold, 1980; Müller-Beck, 1977). In more recent times, the island was sighted by Lieutenant Beechey on Parry's 1820 expedition and named 'Banks Land', but was not visited by Europeans until 1850. At present, the only permanent settlement is Sachs Harbour (1982 population circa 150). This Inuit trapping community originated as a summer hunting camp, and year-round occupation dates only from the 1940s (Usher, 1971).

Early observations of geology and geomorphology were made by Washburn (1947) and also by Manning (1953; 1956), following a circumnavigation of the island by canoe. The first systematic survey was carried out by the Geological Survey of Canada, in 1959 (Thorsteinsson and Tozer, 1962). During the early 1970's, seismic surveys were undertaken by a number of oil companies and exploration drilling commenced in 1971. To date, eleven wildcat wells have been drilled. Hydrocarbons have not been found in commercial

quantities and all wells are currently abandoned. However, oil and gas exploration activities continue, highlighting the need for detailed terrain investigations in this area.

2.2 Climatic conditions

The only long-term climatic data available for southern Banks Island are from the Department of Transport Upper Air station at Sachs Harbour, where records have been kept since November 1955. These data are invaluable to the present study. The station is located on a ridge north of the settlement at an elevation of 94 m a.s.l. and thus wind data are unbiased by local topographic effects. Comparison of station records in July-August, 1979, with data collected by an M.R.I. automatic weather station, located on Sachs Harbour spit, supports a hypothesis that D.O.T. wind data are representative of the area.

Southern Banks Island experiences a cold, arid climate, dominated by the influence of maritime arctic and continental arctic air masses (Burns, 1973; Maxwell, 1980, p. 25; Figure 2.1). The mean annual air temperature is -14.1°C . Mean monthly temperatures fall below -30°C in winter and range from 2°C to 6°C during the summer thaw period, which extends from early-June to early-September. Mean annual precipitation totals only 11.4 cm, of which approximately 50% falls as snow. Although snow depths rarely exceed 30 cm on level surfaces, deep snowbanks develop in lee-slope positions.

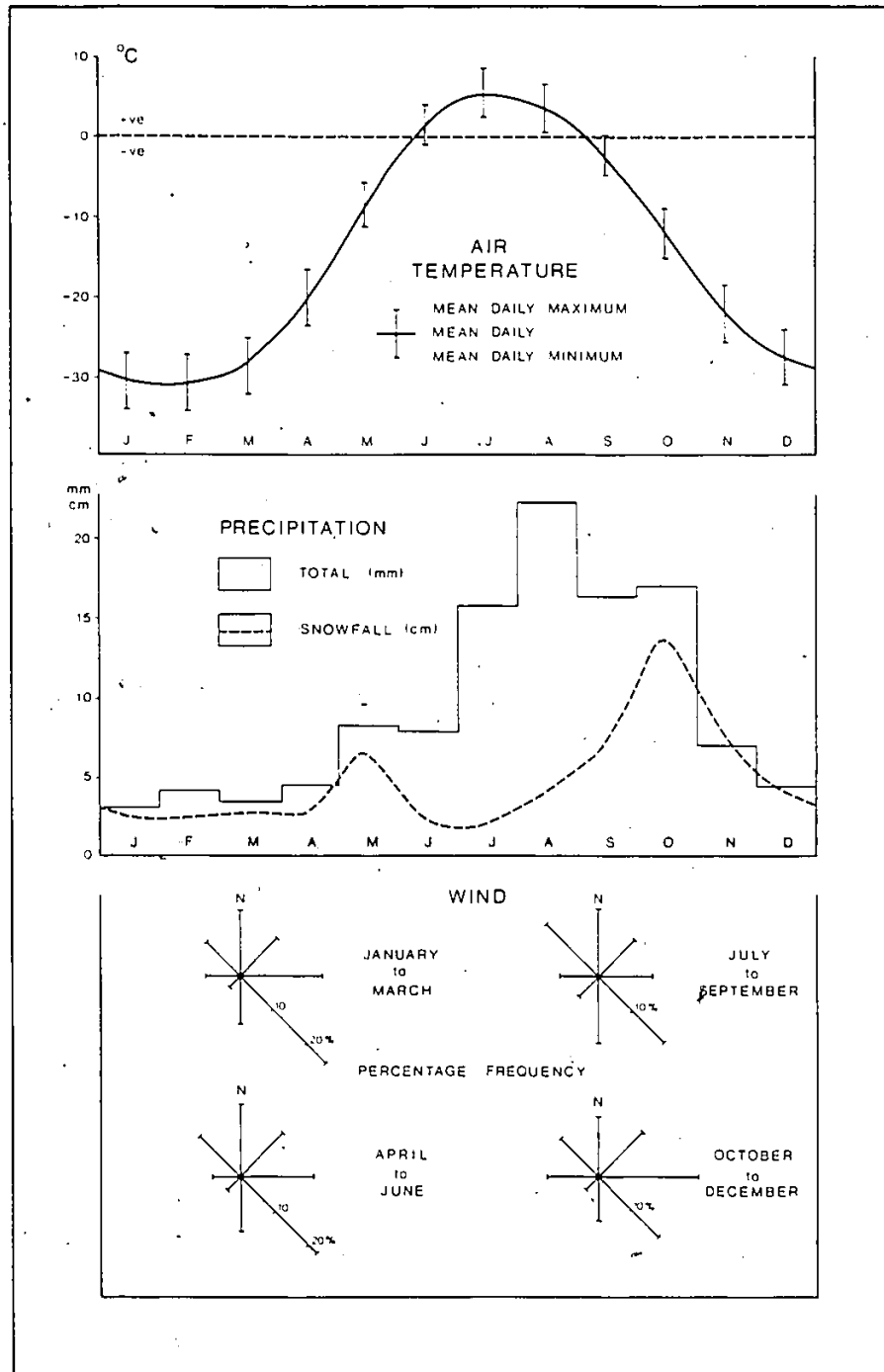


Figure 2.1 Selected climatological data, D.O.T. Sachs Harbour. Air temperature and precipitation 1955-1981; Wind 1955-1977.

Annual wind data for the period 1955-1977 indicate that prevailing winds are from the southeast, with north and northwest winds forming an opposed sub-prevalent component. Wind speed also exhibits little seasonal variation; for example, in all seasons there is approximately 10% probability of hourly wind speed exceeding 30 km/h (Table 2.1). The maximum hourly wind speed recorded at Sachs Harbour is 97 km/h and frequency analysis suggests that maximum wind speeds in excess of 90 km/h have a return period of about 10 years (Figure 2.2).

Of particular interest to the understanding of thaw lake evolution and coastal dynamics is the wind regime during the period of open water, at which time littoral currents and wave action are effective. At Sachs Harbour, open water conditions usually exist from July to September (see p. 42). Directional wind data recorded during this period in 1971-77 display a bimodal frequency distribution, with 33.9% and 31.8% originating in opposed WNW-N and ESE-S quadrants respectively (Table 2.2a). It is necessary also to consider the frequency and direction of high-magnitude storm winds. Storm events are defined, following McCann (1972), as periods during which, (a) hourly wind speed is 20 knots (37 km/h) for at least three consecutive hours, (b) hourly wind does not fall below 20 knots (37 km/h) for more than two consecutive hours and, (c) wind direction does not vary by more than 90°. The threshold condition of a 20 knot (37 km/h) wind with a duration of three hours

Table 2.1
 Values of Hourly Wind Speed (km/h) Equalled or Exceeded at Given
 Probability Levels, Sachs Harbour, 1955-1972.

Period	Percentage Probability								
	1	5	10	20	50	70	90	98	
Dec-Jan-Feb	56	37	31	24	15	10	5	2	
Mar-Apr-May	52	39	34	27	16	11	6	2	
June-Jul-Aug	47	35	31	26	18	13	6	2	
Sept-Oct-Nov	44	35	32	29	18	13	8	2	

Source: Maxwell, 1980.

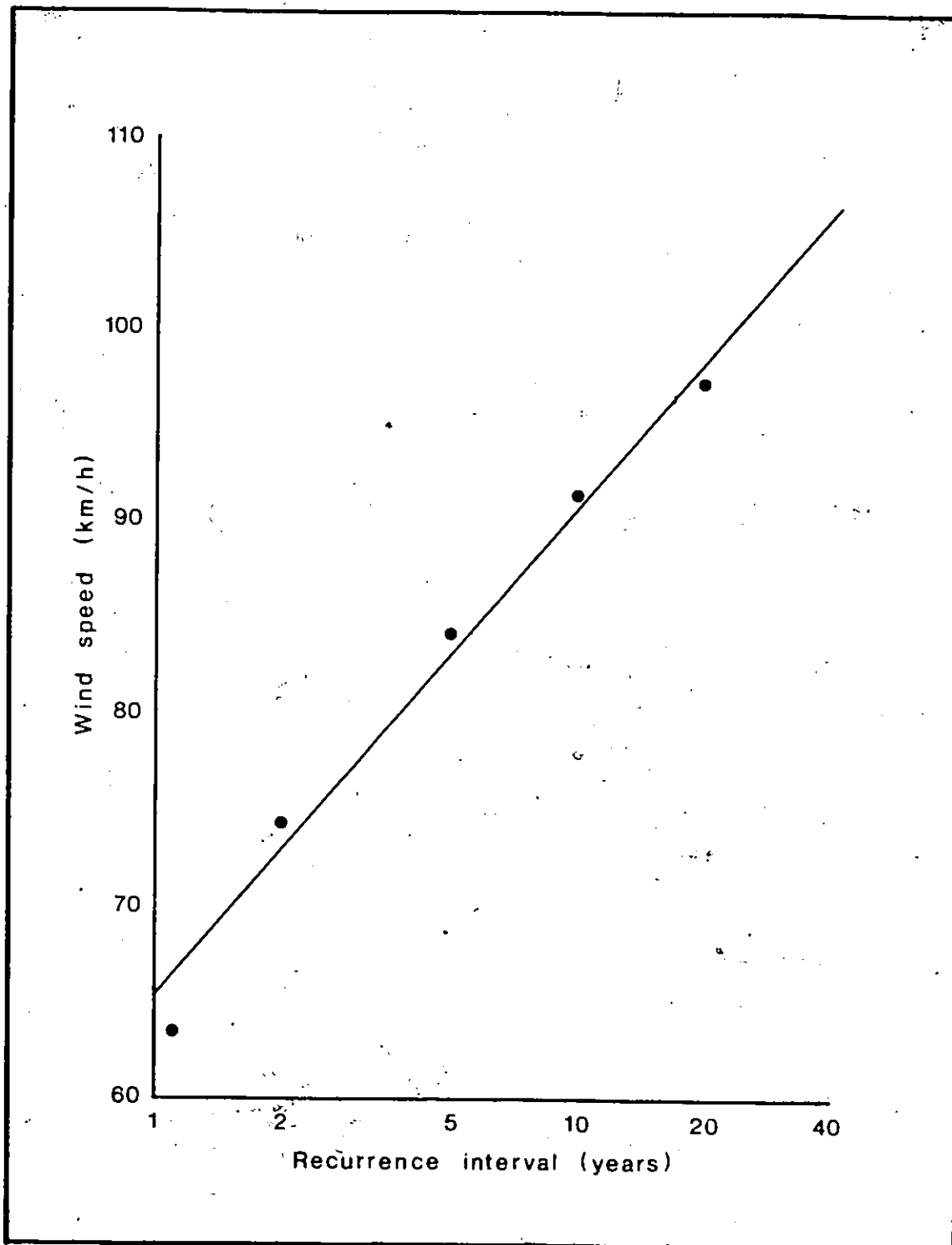


Figure 2.2 Recurrence interval of annual maximum hourly wind speed, Sachs Harbour 1955-1972. (Data abstracted from Maxwell, 1980).

Table 2.2
Wind Frequency, by Direction, Sachs Harbour, July-September 1971-1977.

(a) All Winds.

Sector	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW	N	Calm
July	7.6	3.2	3.6	2.4	2.8	7.9	10.9	8.6	4.7	4.1	3.3	3.0	3.8	9.2	8.4	14.1	2.2
Aug.	6.1	3.6	4.4	2.8	6.3	10.0	8.9	6.4	2.8	2.8	4.1	3.7	6.2	9.7	9.7	10.3	2.2
Sept.	7.8	4.6	4.7	4.2	9.2	11.7	8.5	4.3	2.1	1.9	2.7	4.2	4.4	6.7	9.2	10.2	3.5
July-Sept.	7.2	3.8	4.2	3.1	6.1	9.9	9.4	6.4	3.2	2.9	3.4	3.6	4.8	8.5	9.1	11.5	2.6
Mean by Quadrant	18.3	31.8	13.1	33.9	2.6												

(b) Storm Winds.

Sector	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW	N	Of Total
July	6.8	0.0	0.0	0.0	1.5	13.6	12.3	3.2	3.0	0.8	0.4	1.7	6.6	19.4	11.1	19.6	9.0
Aug.	3.4	1.4	1.4	1.2	3.0	15.7	7.1	2.6	2.0	1.6	5.9	5.3	12.4	15.1	13.1	8.8	17.0
Sept.	4.5	1.5	1.5	2.5	10.7	19.3	10.1	3.9	0.8	0.6	2.9	3.9	3.7	8.5	11.5	14.2	17.1
July-Sept.	4.6	1.1	1.1	1.5	5.7	16.7	9.4	3.2	1.8	1.5	3.6	4.0	7.8	13.4	12.0	13.2	14.4
Mean by Quadrant	8.3	35.0	10.9	46.4													

Source: Derived from Station Daily Record, D.O.T. Sachs Harbour.

is sufficient to generate 1.0 m waves in a fetch-limited deepwater environment and 10-15 cm waves in a shallow, lacustrine environment (United States Army C.E.R.C., 1973, pp. 3.36, 3.47).

Wind directions during summer storm events also have a strongly bimodal frequency distribution, with 46.4% and 35.0% originating in the WNW-N and ESE-S quadrants respectively (Table 2.2b). On average, storm conditions prevail for 10-15% of the open water season. Storm occurrence is highest in September and least in July. While geomorphic processes operating on thaw lakes are affected by winds from all directions, the coastal environment is affected primarily by the action of onshore winds, from the SE-W sector. Onshore storm winds account for 40% of the total and prevail during approximately 5% of the open water season in Thesiger Bay.

The majority of summer storm events last less than 12 hours and have maximum wind speeds of less than 45 km/h. However, although storm events of long duration account for less than 10% of the total, they are likely to be significant geomorphically as a result of their ability to generate large storm waves. The annual maximum summer storm for each year from 1971 to 1980 was identified on the basis of duration and kilometres of wind run, the latter providing an indication of storm magnitude (McCann, 1972). These data were used to generate a storm frequency curve (Figure 2.3) from which the recurrence intervals,

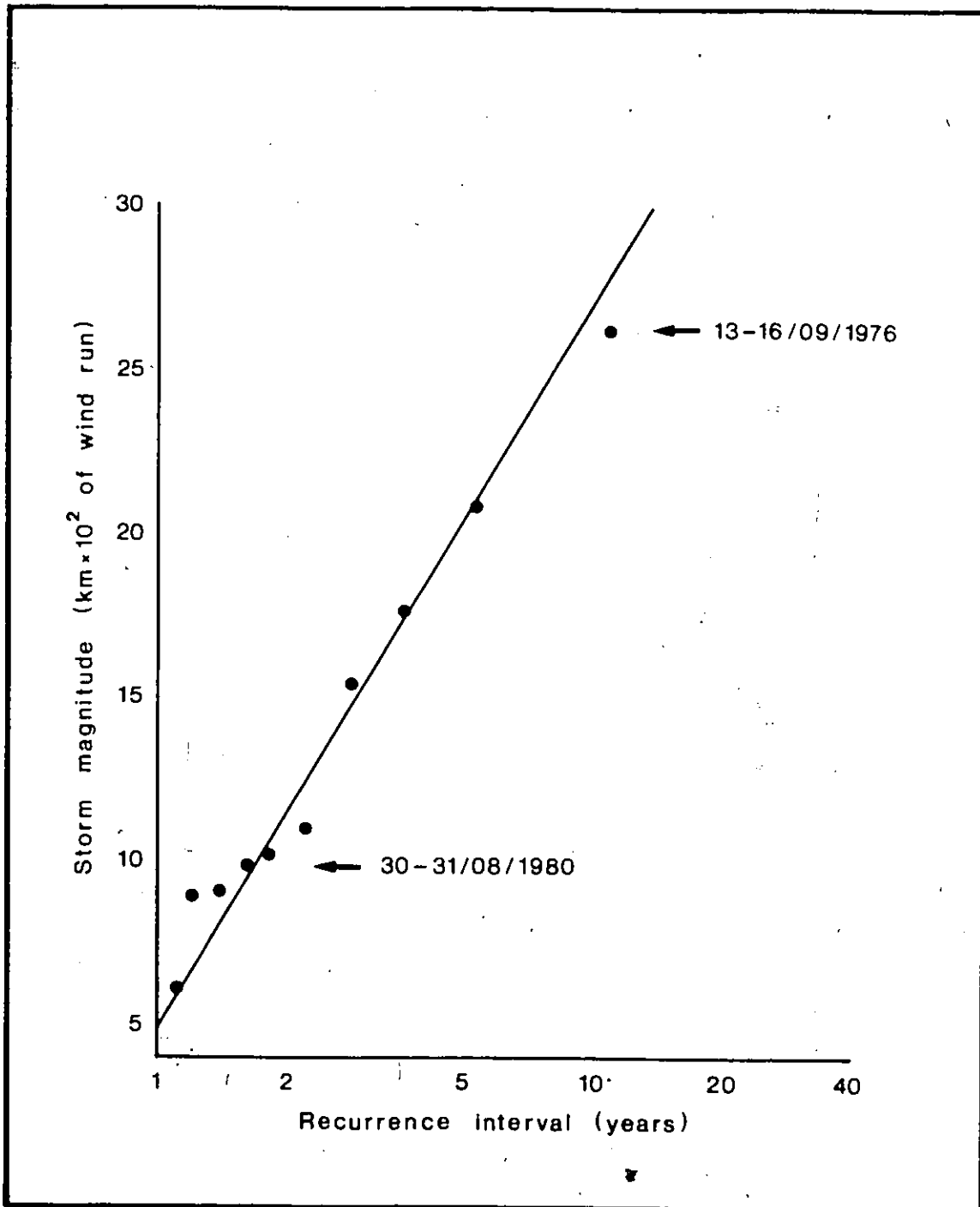


Figure 2.3 Recurrence interval of annual maximum onshore storm event during open water conditions in Thesiger Bay, 1971-1980. The magnitude of a storm observed in August 1980 is shown in relation to that of the maximum storm during the ten year period of record. (Wind data from Station Records, D.O.T. Sachs Harbour).

and thus geomorphic significance, of storm events observed between 1979-81 were determined.

2.3 The terrestrial environment

2.3.1 Pre-Quaternary geology

Southwest Banks Island is underlain by poorly consolidated marine and fluvial sediments of Mesozoic and Tertiary age (Miall, 1979). However, an extensive cover of Quaternary surficial sediments limits their surface outcrop (Figure 2.4). Five major lithostratigraphic units are recognized; in order from oldest to youngest, these are the Isachsen, Christopher, Kanguk, Eureka Sound and Beaufort Formations. The Beaufort Formation has an unconformable contact with earlier units and overlies progressively older strata from north to south. For example, north of the Kellett River, Beaufort sediments are underlain by sands of the Eureka Sound Formation, while at Duck Hawk Bluff, 8 km west of Sachs Harbour, Beaufort Formation sands lie directly upon silty shale of the Kanguk Formation (Vincent, 1980, pp. 139-40). Topographic sections illustrating inferred stratigraphic relationships within the study area are presented in Figure 2.5.

2.3.2 Quaternary geology and geomorphic history

Although Banks Island lies within the zone affected by Pleistocene fluctuations of the Laurentide ice sheet, large areas remained unglaciated during the Wisconsin

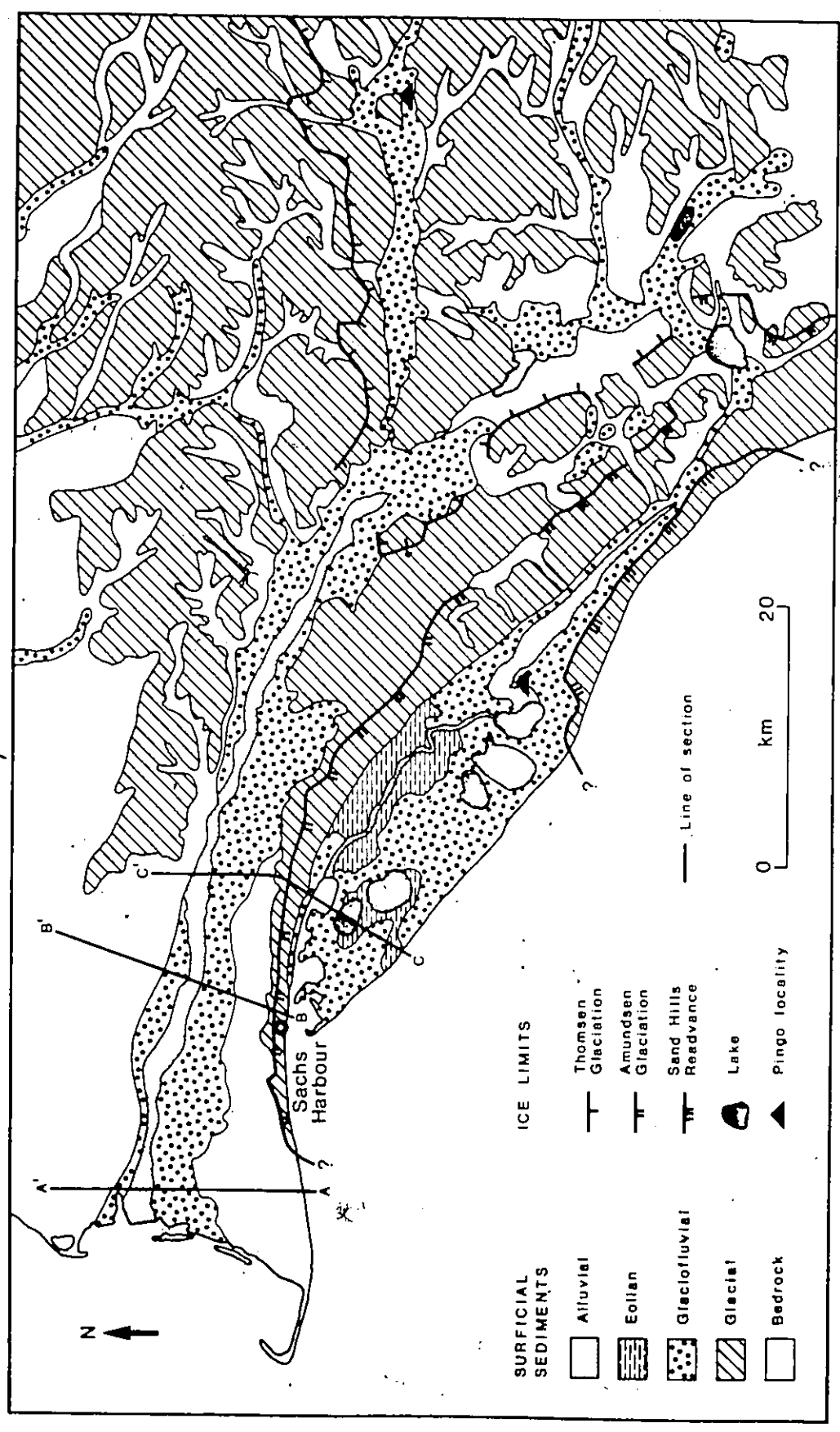


Figure 2.4 Surficial geology of southwest Banks Island. (Modified from Vincent, 1979).

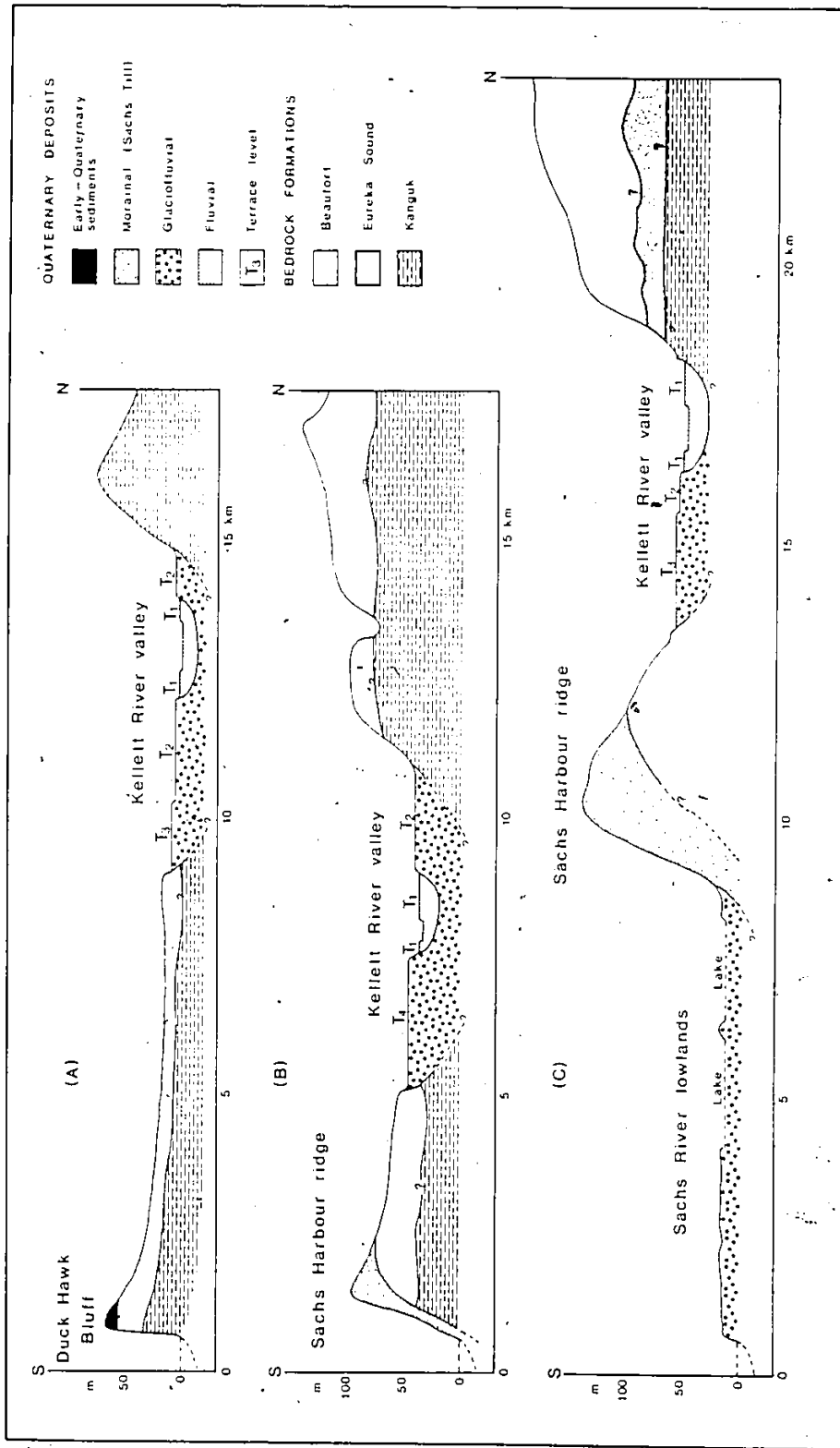


Figure 2.5 Topographic sections illustrating inferred stratigraphic relationships, southwest Banks Island.


or last major glacial period. Early studies recognized a major ice limit, marked by glacial lakes and moraine systems, extending from Parker Point in the north to Thesiger Bay in the southwest (Hobbs, 1945; Jenness, 1952). Thus, the island was divided into a western, unglaciated zone and an eastern zone affected by continental, probably Wisconsin age, glaciation (e.g. Wilson et al., 1958). However, subsequent field investigations indicated that the area outside the moraine belt, with the exception of the Beaufort Plain in the northwest, has been affected by earlier glaciations (Craig and Fyles, 1960; Fyles, 1962). These were interpreted as pre-Wisconsin in age, while the morainal limit was attributed to the 'classical Wisconsin' ice advance (e.g. Prest et al., 1968).

More recently, Vincent (1978a, b; 1979; 1980) has suggested that at least three glacial episodes, each associated with marine, glaciolacustrine and glaciofluvial phases, may be recognized on Banks Island. The Banks Glaciation is the oldest and most extensive, and covered all but the northwestern part of the island. There is little evidence of this glaciation within the study area, although the associated Bernard Till is exposed at Duck Hawk Bluff (Vincent, 1980, pp. 139-40). During the subsequent Thomsen Glaciation, which overrode only southern and eastern Banks Island, large areas in the west were isostatically depressed by as much as 60 m., and may have been submerged by the Big Sea. Radiocarbon dating of

peat overlying marine sediments on southeastern Banks Island suggests a minimum age of 61,000 years B.P. for this event (Vincent, 1980, p. 186).

The Amundsen Glaciation impinged upon Banks Island in the form of two lobes, one flowing northwards along Prince of Wales Strait and one northwestwards into Thesiger Bay, reaching a limit marked by the Sachs Harbour ridge. The Sachs Till, deposited by this ice, has a pinkish sand-silt matrix and contains mainland erratics. Southeast of Sachs Harbour, an area of 'fresh' morainic topography extends parallel to Thesiger Bay. This marks the limit of the Sand Hills Readvance, following retreat of the Thesiger lobe from its maximum position. On the basis of radiocarbon and amino-acid shell dating, Vincent (1980, pp. 208-9) attributes the Amundsen Glaciation to early or mid-Wisconsin time, with an age of greater than 41,000 years B.P. This implies that Banks Island was not affected by late-Wisconsin glaciation and suggests that the island may have been ice-free for 20,000 to 30,000 years.

During periods in which the glacial ice margin lay to the east of the study area, considerable volumes of sediment were discharged into proglacial rivers and carried westwards to be deposited as sandur plains. Both the proto-Sachs and proto-Kellett River valleys formed major spillways, eroded into unlithified sediment by glacial meltwater. As the Amundsen ice margin receded from the Sachs Harbour ridge, outwash sediment was deposited in



a shallow marine environment, possibly correlative with the Meek Point Sea on the west coast of the island (Vincent, 1980, p. 103). These sediments now underlie the outwash plain of the Sachs River lowlands.

Palynological analysis of late-Quaternary sediments in the Mackenzie Delta area (Mackay and Terasmae, 1963; Ritchie and Hare, 1971) indicates a complex sequence of climatic change which may also be applicable to southwest Banks Island. The Pleistocene-Holocene transition is marked by a climatic amelioration, which culminated between 8,500 B.P. and 4,000 B.P. in a period when mean temperatures were appreciably higher than at present. A subsequent climatic deterioration, between 5,000 B.P. and 2,500 B.P., is thought to have caused the refreezing of taliks beneath some lakes and river channels on Banks Island, initiating the growth of pingos (French, 1975b; Pissart and French, 1976). The climate may also have become more arid in late-Holocene times; it has been suggested that eolian activity increased in importance after about 4,000 B.P. (Pissart et al., 1977).

2.3.3 Permafrost and terrain conditions

Southwest Banks Island lies entirely within the zone of continuous permafrost (Brown, 1972; 1978; Figure 1.1), although it is likely that taliks exist beneath the larger lakes and principal river channels. Deep temperature measurements have been made at only one locality on Banks Island, the Storkerson Bay A-15 wellsite, located

approximately 110 km north of Sachs Harbour (Taylor and Judge, 1974, p. 47). These show permafrost thickness to be at least 430 m at this site (Judge, 1973, p. 38) and possibly as great as 550-600 m.

In summer, mean daily air temperature at Sachs Harbour remains above 0°C for 91 days on average, between June 09 and September 07, and a seasonally thawed layer, the active layer, develops. Consideration of heat conduction theory suggests that, as a first approximation, the maximum depth of thaw, z (cm), is given by:

$$z = \sqrt{\alpha P / \pi} \cdot \log_e \left| A_o / T_o \right| \quad [1]$$

where T_o and A_o are the annual surface temperature mean and amplitude respectively, α is the soil thermal diffusivity and P is the period of the temperature cycle (Gold and Lachenbruch, 1973, Equation 11b). Assuming values of $T_o = -10.8^\circ\text{C}$, $A_o = 18^\circ\text{C}$, $\alpha = 0.01 \text{ cm}^2 \cdot \text{s}^{-1}$, $P = 1$ year, the predicted active layer thickness at Sachs Harbour is 160 cm. Actual thaw depths vary considerably with site conditions but are generally less than one metre. The disparity reflects the simplicity of this model, which neglects the latent heat effects of water-ice phase changes.

A more sophisticated simulation model has been developed to predict active layer thermal regimes (Pollard, 1980; Smith, 1975; 1977). Equilibrium surface temperature is derived by iterative solution of the surface energy

balance equation and subsurface thermal regime is calculated on the basis of soil thermal properties. Thus, by varying input data (see Appendix A), site-specific ground temperature conditions may be computed. The program was run for surface conditions simulating (a) dry tundra underlain by sand and (b) wet meadow tundra underlain by a thick organic horizon.

Predicted ground thermal regimes are illustrated in Figure 2.6. Results suggest that on dry tundra surfaces, ground temperature rises above 0°C in mid-June and a maximum thaw depth of 60 cm is attained by mid-August. In areas of wet meadow tundra, thaw penetration is retarded until late-July, as a result of the loss in heat energy associated with evaporation from the saturated surface layer. Maximum active layer development is less than 30 cm, reflecting the low thermal conductivity of unfrozen organic sediment. In both cases, freeze-back commences in early-September. These predictions are in good agreement with observed active layer conditions. In 1979 and 1980, active layer thickness in mid-August varied from 20-30 cm in wet meadow tundra terrain and from 50-80 cm in dry tundra terrain. At well-drained gravel ridge sites, active layer thickness varied from 100-150 cm. The significantly greater thaw depth reflects the high thermal capacity of surficial materials. For example, near-surface temperatures as high as 15°C have been recorded in similar terrain on northern Banks Island (French, 1970a).

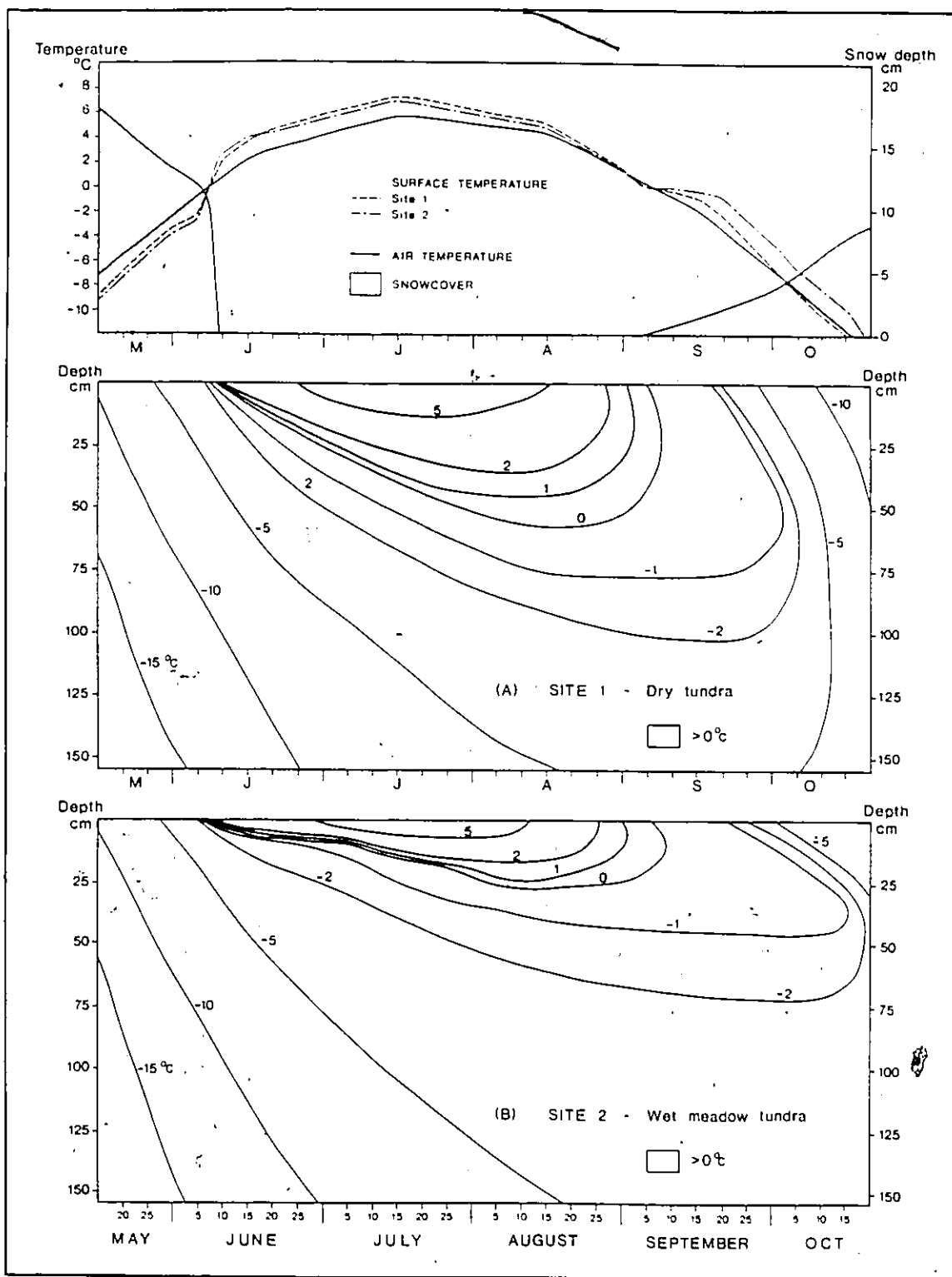
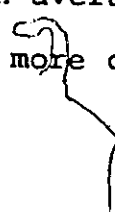


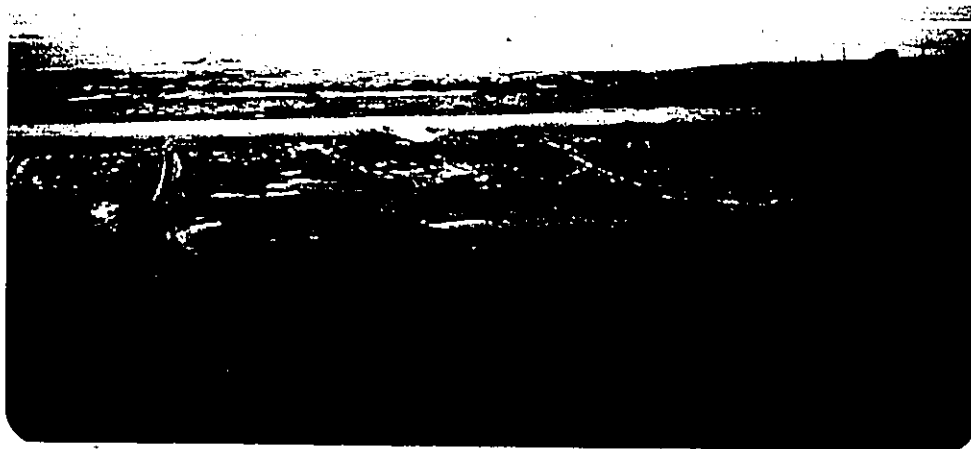
Figure 2.6 Simulated active layer regimes, southwest Banks Island.
 (a) Dry tundra. (b) Wet meadow tundra.

In the vicinity of Sachs Harbour, terrain disturbance associated with construction activity and vehicle movement has resulted in enhanced seasonal thaw and the development of man-induced thermokarst (Dwyer, 1981, pp. 85-113; Figure 2.7). During the construction of the airstrip between 1959 and 1962, a 1-2 m deep layer was bulldozed from the adjacent tundra surface for use as fill. These borrow areas are now characterized by hummocky topography, resulting from melt-out of the underlying ice-wedge network. Thermokarst was initiated within three years of initial disturbance and by late summer, 1973 had not completely stabilized (French, 1975a). The development of large-scale gullies at the townsite may similarly be related to terrain disturbance caused by off-road vehicles and snowmobiles, the use of which commenced in 1970 (French, 1975a).

Three major physiographic units may be recognized in the vicinity of Sachs Harbour. To the north of the settlement, the Sachs Harbour ridge forms an asymmetrical upland area, separating the drainage basins of the Sachs and Kellett Rivers. The east-west trending ridge has an elevation of 90 m a.s.l. north of Sachs Harbour, and rises gradually eastwards to over 250 m a.s.l. (Figure 2.8a). The northern flank slopes at an angle of 2-5° towards the Kellett River. The southern flank is much steeper, with an average slope of 15-20°, and is also morphologically more complex. Superimposed upon the main



(a)



(b)

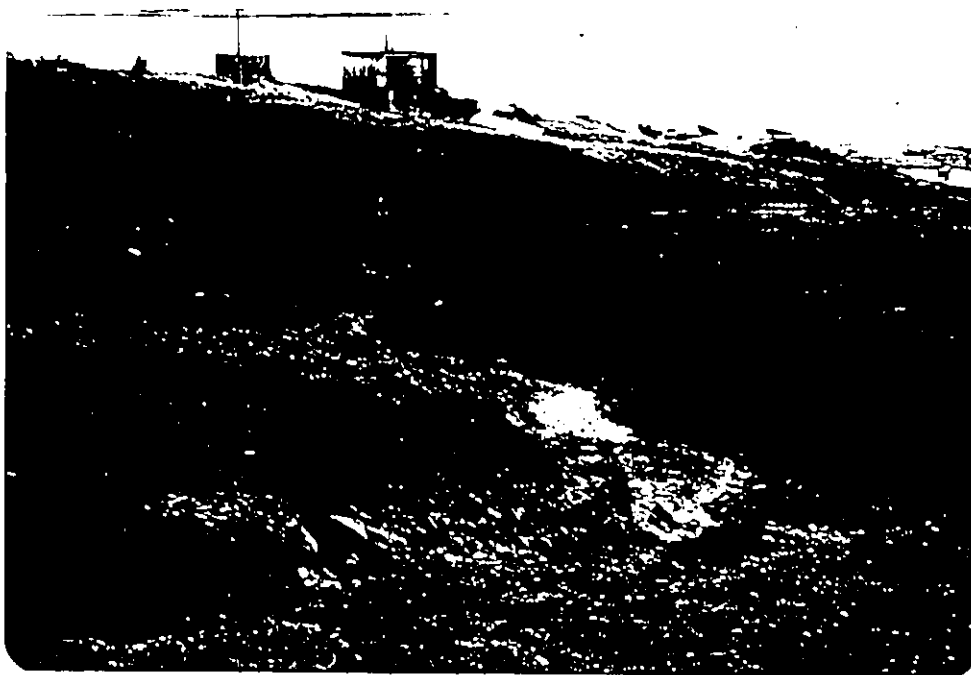


Figure 2.7 Man-induced thermokarst at Sachs Harbour.

- (a) Thermokarst terrain developed by ice-wedge degradation in borrow pits adjacent to the airstrip. Note ice wedge polygons in left foreground. August 06, 1980.
- (b) Thermokarst gully at Sachs Harbour townsite, resulting from terrain disturbance by off-road vehicles. August 18, 1981.



(a)



(b)

Figure 2.8

Terrain conditions, southwest Banks Island.

- (a) Sachs Harbour Ridge, 5 km west of Sachs Harbour. Note the bare gravel surface and steep northern flank, rising from the coastal lowlands. July 08, 1979.
- (b) Kellett River valley, 6 km northeast of Sachs Harbour. Note the unvegetated alluvial floodplain (right); low fluvial terrace with tundra ponds (centre); glaciofluvial terrace bluff (foreground and left). August 17, 1980.

ridge are a number of discontinuous ridges and knolls, composed primarily of sand and gravel, which may represent glacial ice-contact deposits.

To the north of the ridge, the Kellett River valley trends east-west and is also asymmetrical in cross-profile, with northern bluffs rising rapidly to over 200 m a.s.l. South of the ridge, the Sachs River lowlands comprise a low fluvial and glaciofluvial outwash plain, bounded by the coast of Thesiger Bay. The area is lenticulate in shape, reaching a maximum width of 10 km and extending 35 km southeastwards from Sachs Harbour. The Kellett and Sachs Rivers rise in the moraine belt of eastern Banks Island and flow westwards to discharge into the Beaufort Sea. Variation in late-Quaternary discharge levels is marked by a sequence of terraces, on which are developed many hundreds of tundra ponds and lakes ranging in diameter from tens of metres to several kilometres (Figure 2.8b).

In postglacial times, this landscape has been modified primarily by mass wasting on slopes, and by fluvial erosion and deposition within river valleys. The efficacy of these processes is a function of both the periglacial climate and the unconsolidated nature of surficial materials. Solifluction occurs in this area on slopes as low as 2-4° (Figure 2.9); for example, in a study of mass wasting east of Sachs Harbour, a mean downslope movement of 2.0 cm.yr⁻¹ was recorded (French, 1974b). Significant quantities of sediment may also be transported in suspension or

(a)



(b)



Figure 2.9

Mass wasting east of Sachs Harbour.

- (a) Air photograph illustrating mass wasting on the south flank of the Sachs Harbour ridge. (Part of A 22953-129). Note the unvegetated gravel knolls and nonsorted stripes developed on the colluvial sheet.
- (b) Detail of nonsorted stripes in the area marked 'A' above. Note the Dryas/Salix vegetation in the shallow depressions between the regolith stripes. July 29, 1979.

solution by overland flow and subsurface wash. For example, a recent study in central Banks Island concluded that slopewash may form the most important denudational process in that area, primarily by solutional surface lowering (Lewkowitz, 1981, p. 265). Slopewash processes are particularly effective at snowbank sites, where water is supplied to the slope along a broad watershed.

Rivers on southwest Banks Island are characterized by nival regimes, in which the majority of annual runoff occurs during the period of spring snowmelt. Discharge data are not available for the Sachs and Kellett Rivers, however the Big River, at a point 65 km north of Sachs Harbour, has a flow period which extends on average from early-June to early-September (Environment Canada, 1976-1980). The Kellett River has a drainage basin area of 2,641 km², and a mean channel gradient of 2.8 m.km⁻¹. The lower 50 km of its course are braided, reflecting both the high sediment load derived from unconsolidated surface material and the seasonal nature of the flow regime. The Sachs River has a drainage basin area of 1,214 km² and a mean channel gradient of 2.4 m.km⁻¹. The basin shape is complex, probably reflecting glacial disruption of an initial drainage network. Below Raddi Lake, the river flows across the northern margin of the Sachs lowlands, which form an elongate lower basin. Above Fish Lake the channel is braided, while below the lake outlet the channel meanders and has a low gradient (0.3 m.km⁻¹).

Terrain microrelief is closely related to soil and vegetation characteristics. The soils of Banks Island are classified as cryic regosols, developed under arid, low-temperature conditions which inhibit pedogenic processes (Tedrow, 1977). The area lies approximately 350 km north of the treeline and is transitional between low-arctic and high-arctic tundra vegetation zones. Porsild (1955) recognized 172 species of vascular plants on Banks Island; although this represents a rather poor variant of low-arctic vegetation (Kuc, 1974), it is richer than that of most western arctic islands. It has thus been speculated that part of the island may have formed a Wisconsin refugium (Maher, 1968).

Three soil series may be recognized on southwest Banks Island, each related to a distinct vegetation-landform assemblage (French, 1970b; Tedrow and Douglas, 1964). These frequently occur in the form of a soil catena (Figure 2.10a). The Storkerson Series is a polar desert soil, occurring extensively on upland ridges and plateaus. It is characterized by regolith surfaces with a 25-35% cover of lichen-moss tundra. The Bernard Series is associated with the dry earth hummocks of sloping land; soil texture is that of a sandy loam, with a hummocky Dryas sp. microrelief of 10-12 cm (Figure 2.10b). Kellett Series soils occur in areas of lowland meadow tundra, beneath a continuous vegetation mat of Carex sp., Salix sp. and mosses. Poorly drained areas support a marsh vegetation dominated



(a)



(b)

Figure 2.10 Typical soil and vegetation relationships, southwest Banks Island.

- (a) Soil catena developed on the northern flank of Sachs Harbour Ridge. 1. Storkerson Series - developed on gravel ridges (opposite and foreground). 2. Bernard Series - Developed on sloping valley sides. 3. Kellett Series - Developed in poorly drained valley bottom. August 18, 1981.
- (b) Detail of Bernard series soil, showing hummocky Dryas microrelief (ice axe is 60 cm long). August 07, 1979.

by Eriophorum sp. and Sphagnum sp.

2.4 The coastal environment

Within arctic regions, the coastal environment is characterized by a short summer period of open water conditions, during which wave action is effective. The frequency and magnitude of onshore storms during this period are major factors controlling processes and patterns of coastal change.

2.4.1 Open water conditions

Open water conditions are defined as occurring when the pack ice concentration is less than 1/10 (Taylor and McCann, 1976). On this basis, open water conditions in Thesiger Bay prevail on average for 81 days, or approximately 22% of the year (Table 2.3). The mid-July to late-September ice-free period may be considerably reduced in heavy pack ice years; for example, in 1964 the shoreline was only ice-free between August 28 and September 09. The longest open water season recorded is that of 1973, when ice-free conditions existed for 119 days, between June 26 and October 23.

Ice breakup in the eastern Beaufort Sea normally commences in early-June, with the development of a major lead off the west coast of Banks Island (Dey, Moore and Gregory, 1979). Data recorded in the period 1959-74 (Lindsay, 1975; 1977), have been compiled in the form

Table 2.3

Breakup and Freezeup Data, Thesiger Bay, 1956-1981.

	Earliest	Latest	Mean	Standard Deviation (Days)	Number of Years
First ice deterioration	May 04	July 08	June 10	17.4	21
Bay clear of ice	June 26	Aug. 28	July 12	13.3	22
First ice formation	Sept. 02	Oct. 23	Sept. 30	12.8	24
Complete ice cover	Sept. 27	Nov. 11	Oct. 18	14.1	24

Open Water Conditions	Minimum	Maximum	Mean	Standard Deviation (Days)
Duration (days)	5	119	80.8	22.7
\bar{x} of year	1.4	32.6	22.1	-

Source: Allen, 1964 and Station Records, D.O.T. Sachs Harbour.

of maps showing median pack ice conditions (Markham, 1981). These may be used to illustrate the pattern of sea ice breakup and freezeup adjacent to southwest Banks Island (Figure 2.11).

In most years, an extensive area of open water develops to the south and southeast of Banks Island by late-July. In August and September, the ice limit is defined primarily by the margin of the permanent polar pack, which extends in an arc from northern Banks Island to north of Herschel Island. In October, growth of nilas (new ice) results in the progressive restriction of open water to an area south of Banks Island. However, shorefast ice normally develops by mid-October, effectively isolating the coastal zone from this influence.

During the open water season, the effectiveness of storms of given magnitude is determined by the ice-free fetch available for wave generation. Weekly ice observations in the period 1971-80 (personal communication, D. Mudry, Ice Climatology and Applications Division, Atmospheric Environment Service) were analyzed to determine the frequency of ice-free fetch conditions by sector through the open water season (Table 2.4). It should be noted that the western fetch is never ice-free, since it intersects the permanent polar pack. In this case, maximum fetch is assumed to exist when the pack ice limit lies 100 km or more west of Sachs Harbour. The frequency of ice-free fetch conditions to the southeast is 50% or

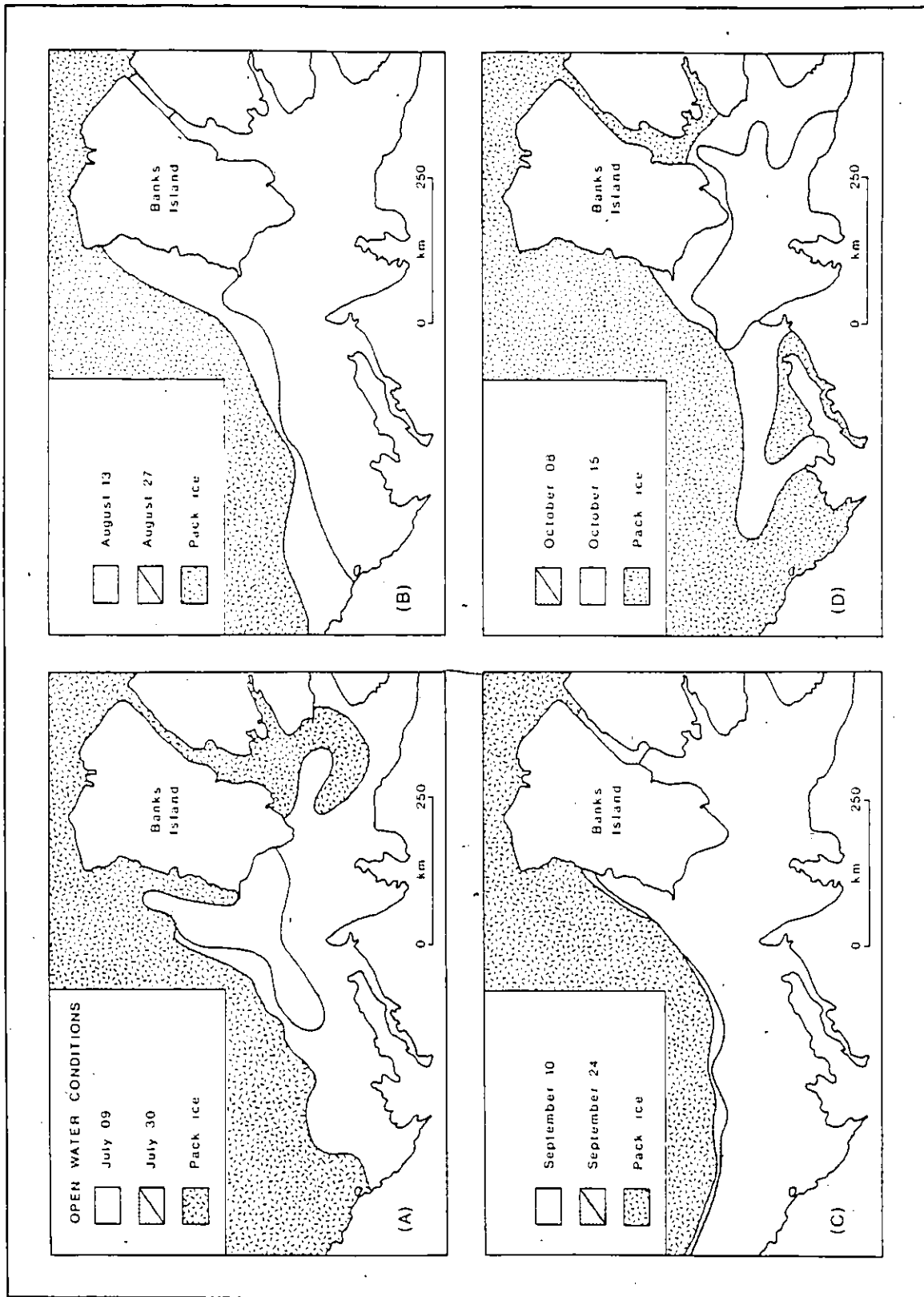


Figure 2.11 Median open water conditions, Beaufort Sea - Amundsen Gulf, 1959-1974. (a) July. (b) August. (c) September. (d) October.

Table 2.4
 Occurrence of Ice-Free Fetch Conditions, Southwest Banks Island, 1971-1980.

Period	% Occurrence of Open Water	% Occurrence of Ice-Free Fetch by Sector										ALL			
		SE	SSE	S	SSW	SW	WSW	W							
June 17-23	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
24-30	20	-	-	-	-	-	-	-	-	-	-	-	-	-	-
July 01-07	60	-	-	-	-	-	-	-	-	-	-	20	-	-	-
08-14	60	-	-	-	-	-	-	-	-	-	-	20	-	-	-
15-21	80	10	20	20	30	30	30	30	30	10	10	20	20	20	10
22-28	80	30	50	40	30	30	30	30	30	10	10	20	20	20	10
29-04	90	40	50	40	40	20	20	20	20	10	10	20	20	20	10
Aug. 05-11	80	50	40	40	30	30	30	30	30	10	10	20	20	20	10
12-18	80	50	50	40	30	30	30	30	30	20	20	20	20	20	10
19-25	80	60	70	50	40	40	40	40	40	30	30	30	30	30	20
26-01	80	70	60	60	50	30	30	30	30	20	20	20	20	20	20
Sept. 02-08	80	60	70	70	60	50	60	50	50	40	40	30	30	30	10
09-15	80	70	70	70	60	60	60	60	60	20	20	40	40	40	20
16-22	100	80	80	60	70	70	70	70	70	60	60	50	50	50	50
23-29	70	70	70	70	70	70	70	70	70	70	70	60	60	60	60
30-06	60	60	60	60	50	40	40	40	40	30	30	30	30	30	30
Oct. 07-13	30	20	20	20	20	10	10	10	10	10	10	-	-	-	-
14-20	30	30	30	10	10	-	-	-	-	-	-	-	-	-	-
21-27	20	10	10	-	-	-	-	-	-	-	-	-	-	-	-
28-03	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-

Source: Weekly ice observations, Ice Climatology, Environment Canada.

greater from early-August to late-September. Maximum fetch conditions to the south and west tend to develop later in the season, reflecting the northwestward retreat of the pack ice limit. During the last two weeks of September, all sectors have a 50% or greater frequency of ice-free fetch conditions.

Periods of maximum wave generation are likely to occur during storm events which coincide with the existence of ice-free fetch. These conditions are satisfied during less than 4% of the open water season and are most frequently associated with southeasterly storms (Table 2.5). In the period 1971-77, no storms have acted over a western fetch greater than 100 km and the long-term probability of such an event is less than 0.1%. Wave climate may be predicted using the model of Sverdrup, Munk and Bretschneider (United States Army C.E.R.C., 1973, pp. 3-35). Wind records suggest that maximum fetch-limited waves are rarely generated, due to insufficient storm duration. However, of the 47 storm events with ice-free fetch recorded in the period 1971-77, approximately 40% have predicted significant wave heights greater than 2.0 m (Figure 2.12). The effect of storm waves may be enhanced if they occur in conjunction with a storm surge, during which water levels may rise appreciably (e.g. Henry and Heaps, 1976; Hume and Schalk, 1967; Reimnitz and Maurer, 1979). Since this is a microtidal environment, with a mean tidal amplitude of only 0.1-0.2 m (Canadian

Table 2.5

Frequency of Ice-Free Fetch and Onshore Storm Conditions, During the Open Water Season in Thesiger Bay 1971-1977.

Sector	Ice-Free Fetch (km)	% Occurrence of Ice-Free Fetch Conditions	% Occurrence of Onshore Storm Winds	% Occurrence of Storm Winds and Ice-Free Fetch Conditions
SE	399	45.4	2.56	2.15
SSE	288	48.5	1.75	0.84
S	283	43.1	0.28	0.18
SSW	172	39.2	0.21	0.12
SW	328	34.6	0.25	0.09
WSW	541	23.1	0.40	0.19
W	100 ¹	28.5	0.27	0.00
Total	-	-	5.71	3.57

¹ Assumed value.

Source: Weekly ice observations and Station Records, D.O.T. Sachs Harbour.

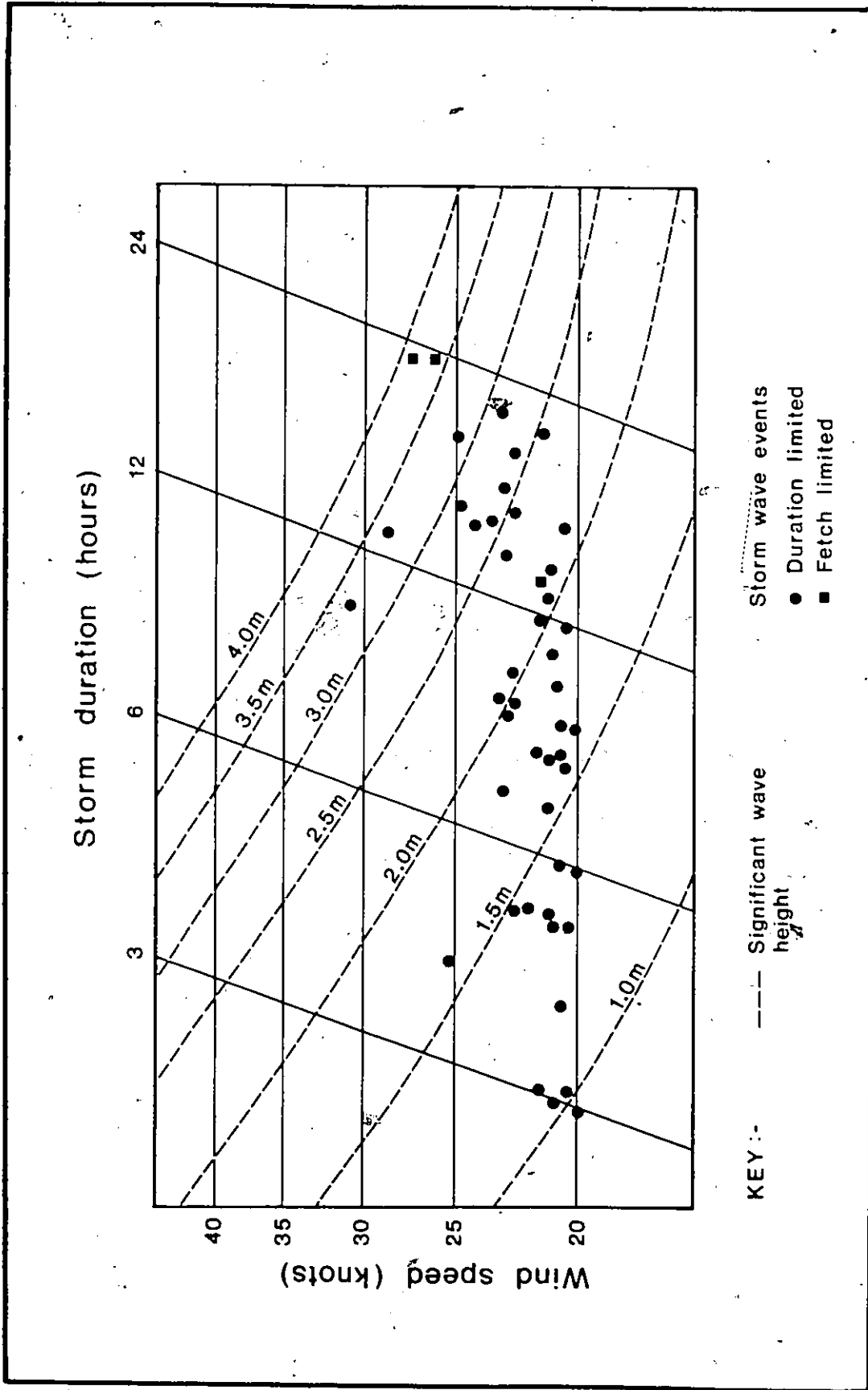


Figure 2.12 Hindcast deepwater wave climate during onshore storms associated with ice-free fetch conditions, 1971-1977.

Hydrographic Service, 1979), storm surges may be of particular geomorphic significance.

2.4.2 Ice conditions and wave climate, 1979-81

Coastal processes were monitored directly during the periods June-August 1979, July-September 1980 and August 1981. The 1979 season may be strongly contrasted to the 1980 and 1981 seasons in terms of both ice conditions and wave climate. In 1979, ice-free fetch conditions did not develop in Amundsen Gulf until early-September. Although Sachs Harbour was ice-free from July 15, effective fetch was less than 50 km in all sectors and frequently less than 10 km. As a result, very low-energy wave conditions were observed, with wave heights less than 20 cm at all times.

By contrast, in 1980 pack ice in Sachs Harbour broke up by July 7 and ice-free fetch conditions existed in most sectors by mid-August. As a result, strong onshore winds between August 20 and September 4, acting over fetches of 200-400 km, were able to generate waves greater than 0.5 m in height. On August 31, 20-30 knot (37-55 km/h) SSW-W winds generated waves which attained heights of 1.0-1.5 m within the harbour (Figure 2.13a). Two further high-magnitude storm events were monitored during ice-free fetch conditions in 1981. On August 17-18 and again on August 22-23, 20-25 knot (37-46 km/h) winds generated waves greater than 1.0 m in height (Figure 2.13b). On

(a)



(b)



Figure 2.13 High-energy wave conditions, southwest Banks Island.
(a) Storm waves at Allen Creek, August 31, 1980. Note undercutting of cliffs resulting in the formation of a thermo-erosional niche.
(b) Storm waves at the base of Cape Kellett, August 23, 1981. Wave height was estimated at 2.5 m and overwash of the spit occurred.

the latter occasion, estimated wave height at the base of Cape Kellett exceeded 2.0 m, and a storm surge resulted in almost total inundation of Sachs Harbour spit. Analysis of 1971-1977 wind records suggests that these conditions are unexceptional, and probably have a return interval of 1-2 years (see Figure 2.3).

The initiation and duration of sea ice freezeup is a function of both temperature and wind conditions (McCann and Taylor, 1975; Short and Wiseman, 1972). For example, in 1979 the freezeup period at Sachs Harbour was characterized by strong (30-50 km/h) winds, alternating between onshore and offshore sectors (Figure 2.14). This resulted in repeated disintegration and movement of the nilas sheet. Consequently, complete ice cover was not reported until 35 days after initial ice formation. By contrast, in 1980 maximum daily winds during freezeup were relatively light (20-30 km/h) and uniformly offshore, greatly reducing the probability of significant wave action. As a result, freezeup was completed within 18 days of initial nilas formation.

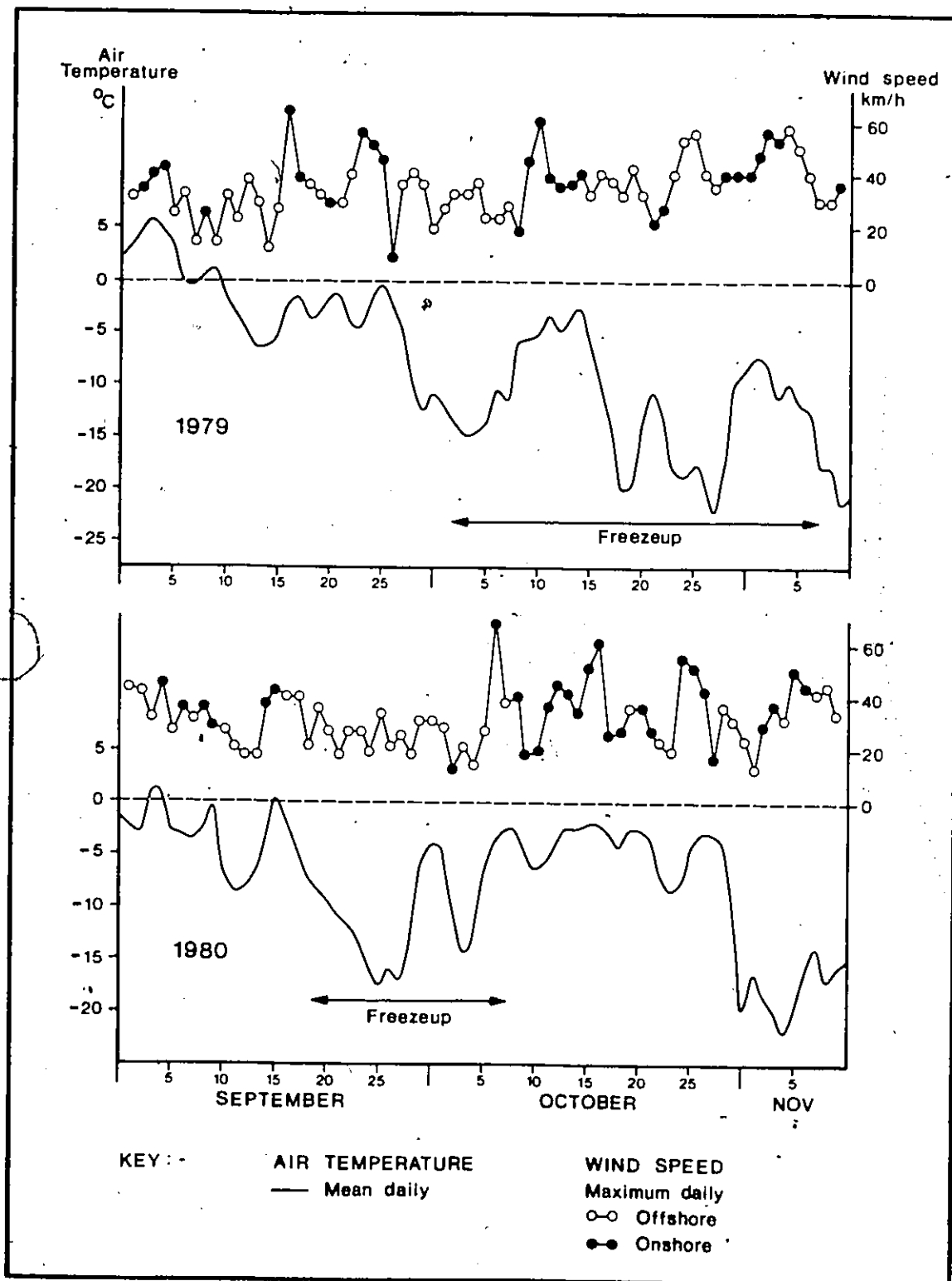


Figure 2.14 Air temperature and wind conditions during freezeup at Sachs Harbour, 1979 and 1980.

CHAPTER THREE

PERMAFROST STRATIGRAPHY

3.1 Introduction

The nature and occurrence of ground ice bodies within permafrost sections can be used to infer past geomorphic processes and may assist in paleoenvironmental reconstruction. A stratigraphic approach to geocryology has been applied widely in the unglaciated areas of central and eastern Siberia (e.g. Katasonov, 1975; 1978; Katasonov and Ivanov, 1973; Mackay, Konishchev and Popov, 1979, pp. 10-11; Popov, 1969; Sher and Kaplina, 1979) and, to a lesser extent, in Alaska (e.g. Péwé, 1975; 1977; Sellmann, 1967; Sellmann and Brown, 1973). In many areas of arctic Canada, the relatively short postglacial history of permafrost has limited the use of this approach. Only in parts of the Pleistocene Mackenzie Delta, which were ice-free during the Wisconsin glaciation, have stratigraphic studies of permafrost been successfully undertaken (e.g. Mackay, 1975; 1976; 1978a).

Although a summary paper is available (French, Harry and Clark, 1982), this chapter describes in more detail stratigraphic investigations of permafrost, undertaken on southwest Banks Island in 1979 and 1980. The objectives are, first, to illustrate how permafrost stratigraphy provides information relating to the formation of permafrost landforms and, second, to use the evidence

of ground ice bodies together with their enclosing sediments to reconstruct the late-Quaternary history of the region. Specific attention is focused upon the nature of ice wedge systems and pingos. The former were examined in natural coastal exposures west and southeast of Sachs Harbour. Sections through pingos and other ice-cored features were produced artificially using a Wajax high-pressure fire pump, which was also used to remove slumped debris from coastal cliffs.

3.2 Ice wedges

3.2.1 Nature and occurrence

Ice wedges are vertically foliated bodies of massive ground ice formed along the axes of thermal contraction cracks. Their mode of formation has been described by Lachenbruch (1962; 1966), Black (1963), and Romanovskii (1973), following early research in Arctic Alaska (Leffingwell, 1915; 1919, pp. 205-14) and Siberia (see e.g. Shumskii, 1964, pp. 5-9). The orientation and spacing of thermal contraction cracks varies considerably, reflecting contrasts in the thermal and geotechnical properties of underlying sediments and, possibly, variation in the climatic regime (Dostovalov, 1960; Dostovalov and Popov, 1966; Kerfoot, 1972, Lachenbruch, 1961). Natural exposures of ice wedges are rare and usually short-lived, since melt of the ice body is accompanied by slumping of adjacent sediments. For this reason, ice wedges may normally only be examined

in actively eroding sea cliff and riverbank sections.

Throughout southwest Banks Island, ice-wedge polygons occur widely wherever horizontal or low-angle surfaces are underlain by fine-grained alluvial and outwash sediments. Within the Kellett River valley, most polygons consist of upthrust rims surrounding a central, often water-filled depression. On the lowlands west and southeast of Sachs Harbour, high-centred polygons bounded by ice-wedge troughs predominate. Tundra polygons are poorly developed on the sand and gravel surfaces of the Sachs Harbour ridge; however, wedge ice was encountered in two out of seven boreholes drilled in 1973 near the airstrip (French, 1975a). Within the Sachs River lowlands, polygon diameter ranges from 8-20 m, with a mean value of approximately 15 m. Following the terminology of Lachenbruch (1966), contraction cracks in this area form a predominantly random orthogonal or hexagonal network (Figure 3.1). The relationship between frost cracks and active ice wedges may be examined in coastal sections (Figure 3.2).

The size of ice wedges may provide an approximate indication of their age. In studies of Alaskan and Antarctic ice wedges, Black (1952; 1963; 1973) concluded that growth rates of $0.5-1.0 \text{ mm.yr}^{-1}$ are typical, suggesting that a 1.0 m wide wedge may grow in 1,000-2,000 years. However, following detailed field observations on Garry Island, Mackay (1974a) concluded that, on average, only 40% of contraction cracks open in a given year. Thus, mean ice



Figure 3.1 Intersection of 3 ice-wedge troughs at angles of approximately 120° , within an area of high-centred polygons 4 km west of Sachs Harbour. August 18, 1981.



Figure 3.2 Relationship between frost crack and active ice wedge, exposed in a coastal section 4 km southeast of Sachs Harbour. June 25, 1979.

wedge growth rates on Garry Island may actually be less than 0.2 m, yr^{-1} . These results imply that the size of an ice wedge is not directly related to its age.

In general, ice wedges on southwest Banks Island are relatively small, being 1-2 m wide at the frost table and extending downwards 3-5 m into permafrost (Figure 3.3). They may thus be contrasted to the large ice wedges observed, for example, in the Siberian coastal lowlands. In that area, ice wedges apparently exceed 5 m in width and 40-50 m in depth (e.g. Dostovalov and Popov, 1966; Shumskii, 1964, p. 36). This disparity may be partly a function of growth mechanism, since many Siberian wedges are believed to have formed syngenetically, in relation to aggrading alluvial surface (e.g. Popov, 1962; Shumskii and Vtyurin, 1966). In the North American Arctic, few drainage basins have long histories of cold, nonglacial climate combined with floodplain development. Thus syngenetic wedges are rare (Mackay and Black, 1973). Although a number of small Pleistocene syngenetic wedges have been identified on Hooper Island (Mackay, 1976; Mackay, Konishchev and Popov, 1978, p. 6), the majority of North American ice wedges appear to be epigenetic in origin.

This conclusion is supported by observations on southwest Banks Island. In this area, epigenetic wedges are most common, although small syngenetic ice veins are also recognized. Some wedges appear inactive and are truncated below the present frost table, while others possess



Figure 3.3 Typical epigenetic ice wedge developed in ice-rich silt, 4 km west of Sachs Harbour. Note deformation of strata adjacent to the wedge. August 07, 1980.

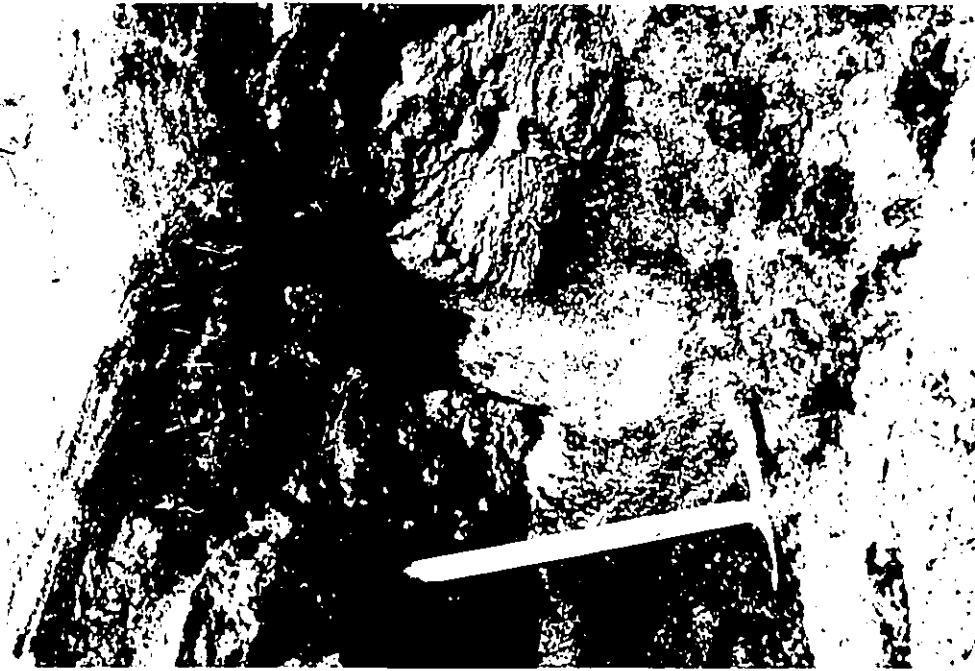


Figure 3.4 Two-stage epigenetic ice wedge, 4 km west of Sachs Harbour. (Photograph by H. M. French, August 04, 1978).

a complex structure indicating thaw degradation and subsequent wedge rejuvenation (Figure 3.4). The latter are particularly interesting since they provide additional evidence of a regional thaw unconformity, existing at depths of 1.5-1.7 m below surface in the Western Canadian Arctic (e.g. Mackay, 1975; 1978a), and at rather deeper levels in central Alaska and the interior northern Yukon Territory (e.g. Campbell, 1952, p. 60; Naldrett, 1981, p. 134, Péwé, 1965, pp. 10, 32).

Localized thaw unconformities may be formed by permafrost degradation beneath transient water bodies. For example, a coastal section located 4.0 km west of Sachs Harbour illustrates the permafrost stratigraphy resulting from the growth and subsequent drainage of a shallow lake (Figure 3.5). Following lake growth, the ice wedge exposed in this section was truncated by enhanced thaw to depths of 60-100 cm below surface. At a later date, lake drainage occurred, probably as a result of coastal cliff recession. During freeze-back of the saturated thaw zone, layers of clear pond ice formed (Figure 3.6). These are characterized by elongate ice crystals, oriented parallel to the freezing direction (Gell, 1976, pp. 224-35), which in this case appears to have been from above. Subsequent subaerial exposure of the lake bottom led to renewed thermal contraction cracking of the surface and the growth of a second-stage ice wedge, extending upwards to the present frost table.



Figure 3.5 Permafrost stratigraphy associated with the growth and drainage of shallow lakes, 4 km west of Sachs Harbour. This ice wedge was truncated by thaw beneath a shallow lake. Following lake drainage, the wedge was rejuvenated and layers of pond ice developed. August 17, 1979.

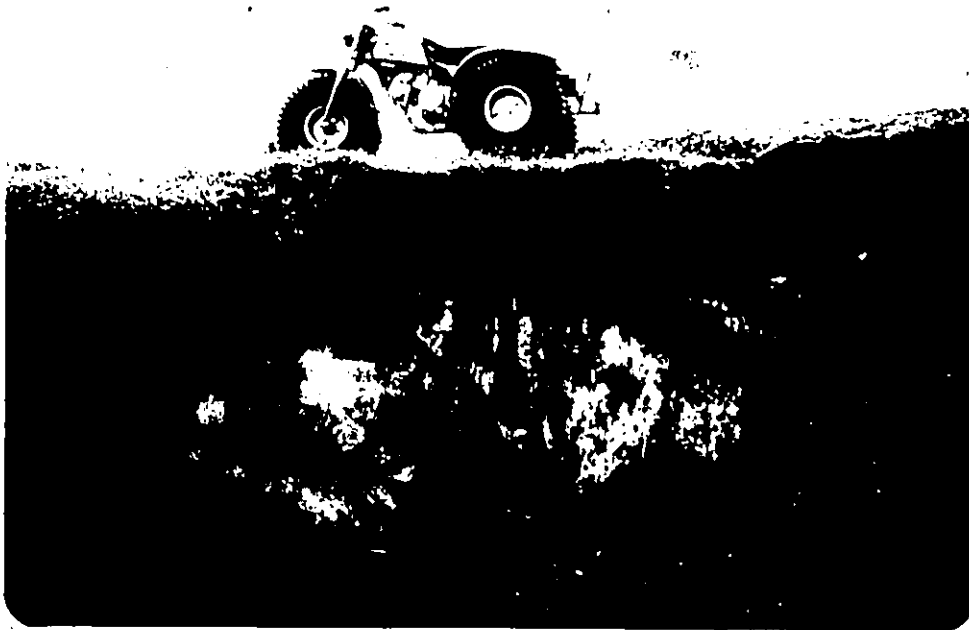


Figure 3.6 Tabular sheets of pond ice beneath the present active layer, exposed in an adjacent permafrost section. August 21, 1981.

Occasionally, spring snowmelt runoff erodes preferentially along ice wedge axes to form small, steep-sided thermokarst ravines. These develop periodically on the lower slopes of the Sachs Harbour ridge, downslope of large snowbanks (Figure 3.7). Both Mackay (1974b) and French (1975a) have commented upon the rapidity of ice wedge destruction by thermal and mechanical erosion.

3.2.2 Stratigraphy of coastal sections

Detailed examination of coastal sections in the Sachs River lowlands, during August of '79 and '80, suggests that six lithostratigraphic units may be recognized in this area. All but the lowest are illustrated in Figure 3.8, which is a measured section of the coastal cliffs 2.5 km southeast of Sachs Harbour (71°57'N, 125°14'W). The basal unit, not exposed at this locality, is a blue-grey till, containing striated Paleozoic erratics within a sand-silt matrix. This is exposed near sea level in the vicinity of the Sachs Harbour townsite, and its presence elsewhere is frequently indicated by a beach armouring of exhumed igneous erratics. The till is overlain by 1-5 m of laminated grey silt and silty sand with poorly defined and irregular cross-bedding. The contact between this unit and the overlying 1-4 m of yellow-brown medium and well-sorted sand is gradational; within the transition zone is a thin (1-6 cm) horizon of detrital willow roots and stems (Figure 3.9). A number of small frost fissures, characterized by downturned sediment laminae, extend downwards into the grey silty sand from



(a)



(b)

Figure 3.7

Terrain conditions associated with ice wedge thermokarst.

- (a) Incipient collapse along a gully formed by melt-out of an ice wedge, 5 km east of Sachs Harbour. July 31, 1979.
- (b) Thermokarst ravine 2 km east of Sachs Harbour, formed by nival runoff along an ice-wedge trough. July 29, 1979.

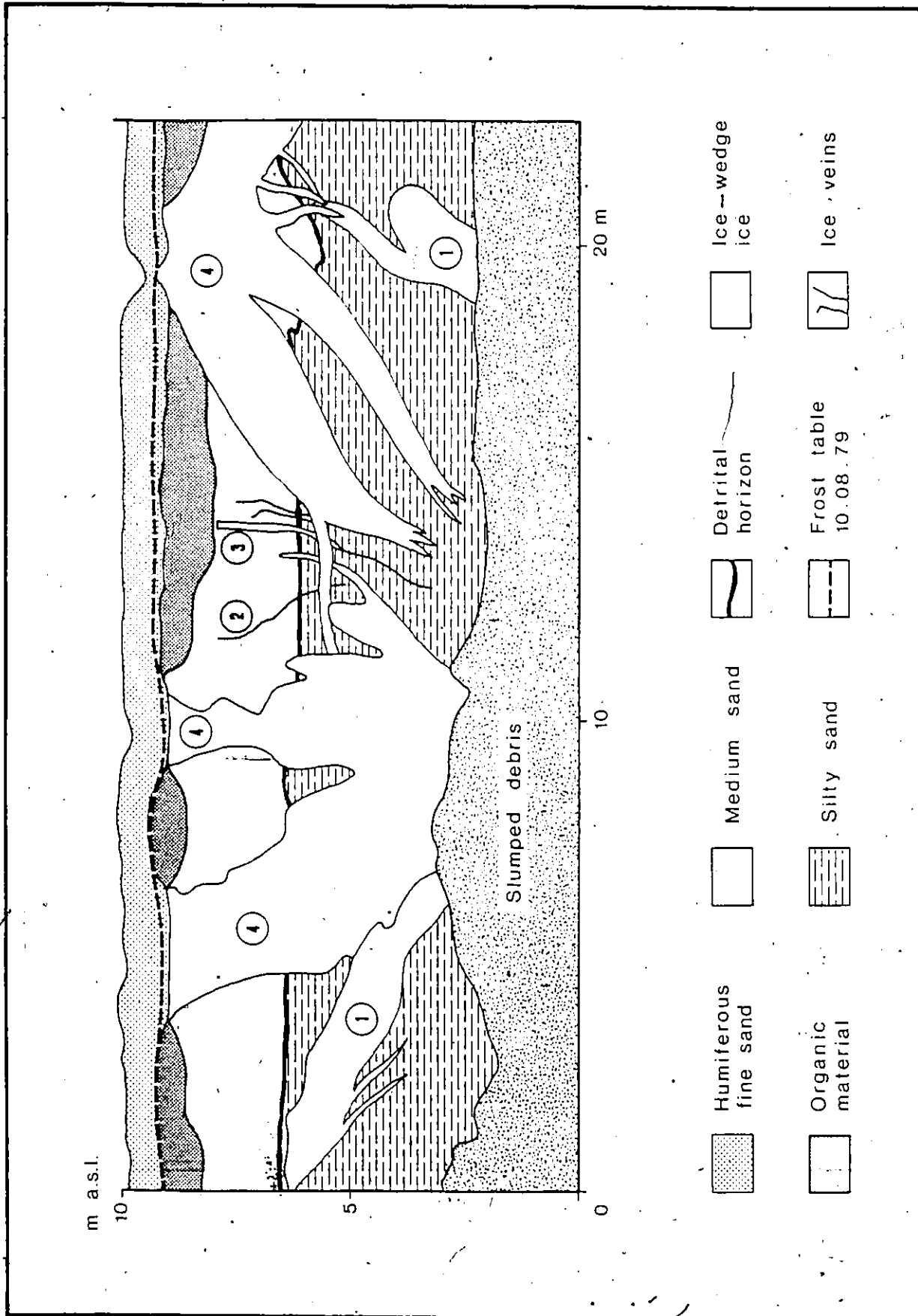


Figure 3.8 Section illustrating permafrost stratigraphy exposed in coastal bluffs of the Sachs River lowlands, 2-5 km southeast of Sachs Harbour. Surveyed August 10, 1979 and August 12, 1980.



Figure 3.9 Contact between silty sand and overlying sand, Sach's River lowlands permafrost section. Note detrital organic material in contact zone. August 08, 1980.

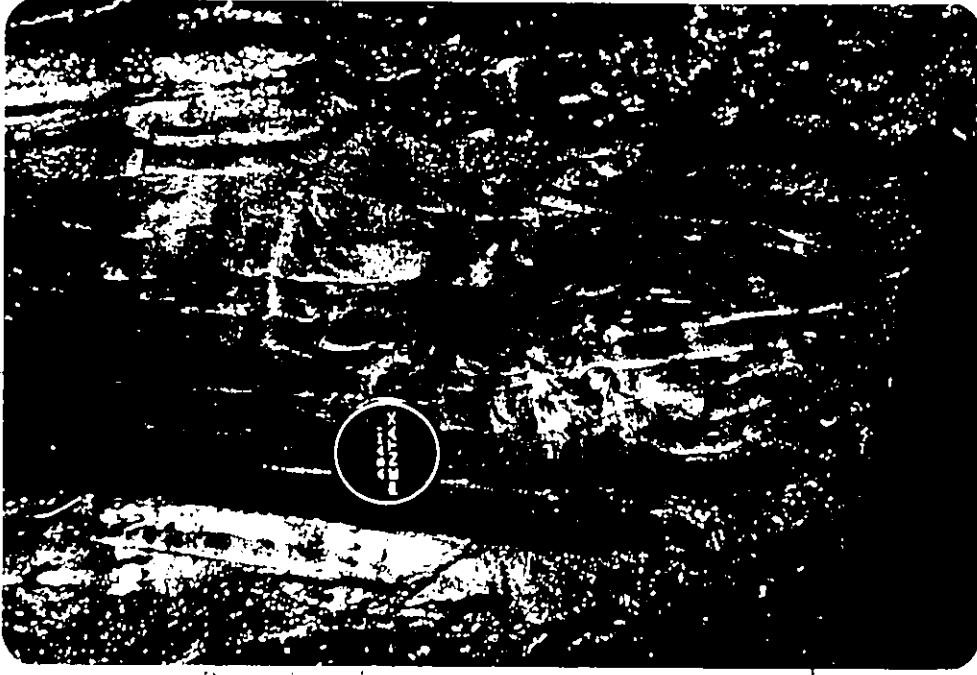


Figure 3.10 Small frost fissure extending downwards from the detrital organic horizon. August 12, 1980.

this horizon (Figure 3.10). The yellow-brown sand is overlain by a peaty organic unit, which attains 1.0 m in thickness between ice wedges and is absent beneath ice wedge troughs. The peat unit grades upwards into approximately 0.25 m of humiferous eolian sand.

Ground ice content of the two major stratigraphic units differs considerably. The grey silty sand is ice-rich, containing 50-250% ice by weight and 20-75% excess ice by volume. Cryotextures within this unit consist of horizontal and inclined segregated ice lenses. Near the upper boundary of the unit, the ice structure is more granular and may indicate episodic thawing and recrystallization. By contrast, the overlying yellow-brown sand contains approximately 20% ice by weight and less than 10% excess ice by volume. Few ice structures are visible within this unit.

Radiocarbon age determinations provide an absolute chronology for this sedimentary sequence (Table 3.1). Material collected from the detrital willow horizon yields a ^{14}C date of $10,600 \pm 130$ years B.P. (GSC-3229). This provides a maximum age for the overlying yellow-brown sand and a minimum age for the underlying grey silty sand. A second ^{14}C date of $6,490 \pm 60$ years B.P. (GSC-3216) was obtained from material collected at the base of the peaty organic layer, at a depth of 1.45 m below surface. This provides a maximum age for the initiation of peat accumulation. At a second locality, approximately 500 m to the southeast,

Table 3.1

Radiocarbon Dates Pertaining to Late-Quaternary Stratigraphy,
Sachs River Lowlands, Banks Island.

Locality	Laboratory Dating Number	Date (years B.P.)	Stratigraphic Position	Material	Significance
Coastal bluffs 3.5 km SSE of Sachs Harbour	GSC-3229	10,600±130	Detrital organic horizon, under- lying outwash sand and overlying del- taic silty-sand.	<u>Salix</u> sp.	Min. age for surface marked by frost fissures, max. age for deposition of outwash sand.
	GSC-3216	6,490±60	Base of peaty organic layer, overlying out- wash sand.	<u>Carex</u> sp.	Max. age for initia- tion of peat accumulation.
Drained lake basin 4.0 km SSE of Sachs Harbour	GSC-3292	8,560±210	Peat at base of lacustrine sedi- ment sequence, over- lying outwash sand.	<u>Salix</u> sp.	Min. age of outwash sediment.
Small kettle lake 40 km SE of Sachs Harbour	GSC-2200 ¹	9,870±340	Base of lake sedi- ments overlying ice contact depo- sits associated with Carpenter Till	Moss	Min. age for Sand Hills readvance.
Gully in escarpment 10.0 km E of Sachs Harbour	GSC-2419 ²	8,430±120	Peat below eolian sediments, over- lying outwash sand.	Peat	Max. age for initiation of eolian activity.

¹ Collected by R. J. Mott (see Vincent, 1980, p. 237).² Collected by J-S. Vincent (see Pissart et al., 1977, p. 2478).

a ^{14}C date of $8,560 \pm 210$ years B.P. (GSC-3292) was obtained from willow fragments, contained within lacustrine sediments overlying the yellow-brown sand. This provides a minimum age for the sand unit and is consistent with the date of $8,430 \pm 120$ years B.P. (GSC-2419), obtained by Vincent (1980, p. 236), for peat overlying outwash sand east of Sachs Harbour. Few data are available with respect to the age of the Sand Hills Readvance. However, a ^{14}C date of $9,870 \pm 340$ years B.P. (GSC-2260) has been obtained for organic material, collected by R. J. Mott, from a small lake basin within the glacial limit (Vincent, 1980, p. 237). This material overlies ice-contact sediments and thus the ^{14}C date provides a minimum age for the Sand Hills event.

Four distinct ice-wedge systems may be recognized within the section illustrated in Figure 3.8. The oldest set of wedges consists of large ice bodies, 3-5 m deep and 0.5-1.5 m wide, which occur within the grey silty sand and appear to be truncated at or near the detrital willow horizon (Figure 3.11). As such, they are unrelated to the present ground surface and are assumed to be inactive. The ice wedges are highly foliated and vein-like, and contain grey silt. In some cases, 'fingers' of ice extend laterally for some distance from the main body of the wedge (Figure 3.12). This is considered characteristic of epigenetic rather than syngenetic ice wedge growth (Shumskii, 1964, p. 37).



Figure 3.11 Ice wedges exposed in the Sachs River lowlands permafrost section. Two systems of epigenetic wedges are visible. The oldest wedges are truncated at or near the upper boundary of the grey silty-sand. August 06, 1979.



Figure 3.12 Detail of the older epigenetic ice wedge seen at bottom left in Figure 3.11. Note the branching veins and grey silt contained within foliations. August 12, 1980.

A second system of wedges consists of small multiple ice veins, 1-2 m in length (Figure 3.13). These occur predominantly within the yellow-brown sand unit, but in some cases extend downwards into the grey silty sand. The width of these ice bodies varies vertically; at some levels they attain a thickness of 3-6 cm while at other levels they are represented by only a single fissure, marked by downturned sediment laminae. For this reason, these ice bodies are believed to be syngenetic in origin; the width of the ice vein being inversely proportional to the rate of sedimentation at any given level within the sedimentary unit.

A third system of wedges consists of epigenetic ice bodies, 1-3 m deep and up to 0.5 m wide (Figure 3.14). These occur within the yellow-brown sand and are truncated at the base of the overlying organic unit, approximately 1.5-1.7 m below the present ground surface. These wedges appear inactive as they are unrelated to the surface network of frost cracks. A fourth system of wedges consists of epigenetic ice bodies, 4-5 m deep and up to 2 m wide, which extend upwards to the present frost table (see Figure 3.11). Many of these wedges penetrate the underlying grey silty sand, but contain yellow-brown sand within their foliations at all levels. This set of wedges is clearly related to the surface distribution of thermal contraction cracks.



Figure 3.13

Small syngenetic ice wedge developed in the medium sand unit, Sachs River lowlands permafrost section. Note the downward extension of this wedge, as a frost fissure, into the underlying silty sand. August 12, 1980.



Figure 3.14

Epigenetic ice wedge developed in the sand unit and truncated at the base of organic sediments, at a depth of 170 cm below the present ground surface. August 06, 1979.

3.2.3 Discussion

The recognition of four ice wedge systems, together with interpretation of their enclosing sediments, permits a reconstruction of late-Quaternary paleoenvironments in this part of the Sachs River lowlands. This model may be applicable to other areas of southwest Banks Island.

The basal blue-grey till is probably equivalent to the Sachs Till, deposited by the Thesiger ice lobe during the main advance of the Amundsen Glaciation (Vincent, 1980, pp. 64-66). This represents the penultimate glacial event to affect southwest Banks Island. The overlying grey silty sand and yellow-brown sand units are glaciofluvial outwash sediments, possibly associated with the Sand Hills glacial readvance. At that time an ice limit, marked by fresh morainic topography, was established approximately 30 km southeast of Sachs Harbour. Outwash deposition appears initially to have been subaqueous, into a shallow marine embayment possibly correlative with the Meek Point Sea (Vincent, 1980, p. 103). During this time the fine grained silty sand unit was deposited. As the delta front prograded to the northwest, shallow marine conditions were progressively replaced by a subaerial, fluvial depositional environment, in which the yellow-brown sand accumulated as a sandur plain.

Aggradation of permafrost into the freshly exposed deltaic sediments was accompanied by the development of thermal contraction cracks, marked stratigraphically by

the presence of small frost fissures near the upper boundary of the silty sand. It is likely that the earliest set of ice wedges grew in relation to a ground surface, marked by the detrital organic horizon. A minimum age for this surface of $10,600 \pm 130$ years B.P. is provided by the ^{14}C date (GSC-3229) of the willow fragments. The dimensions of these wedges suggest that the surface may have remained relatively stable for a period of time following emergence. Subsequently, rapid surface aggradation, associated with deposition of the yellow-brown sand unit, buried this set of ice wedges and favoured the growth of small syngenetic ice veins.

Following stabilization of the surface in early-Holocene times, epigenetic ice wedges developed within the yellow-brown sand. Some were truncated at depths of 1.5-1.7 m below surface, during the mid-Holocene climatic optimum of 8,500-5,500 years B.P. Subsequently, a number of wedges continued to grow and are currently active, reflecting modern cold climate conditions. The warm, moist conditions of the climatic optimum favoured accumulation of a thick organic layer. Variation in the thickness of this unit may well reflect the infilling of a series of low-centered polygons by organic material. A similar origin has been proposed for buried lenticulate peat bodies which are frequently observed in northern Siberia (Shumskii, 1964, p. 37).

The ^{14}C date of 6,490 \pm 60 years B.P. (GSC-3216) provides a minimum age for the early-Holocene surface on which this organic material accumulated. Paleoecological analysis of a sample of this material indicates a primary composition of sedge or grass stem/rhizome fragments, with a minor component of moss fragments (personal communication, J. V. Matthews, Geological Survey of Canada). A fossil coleoptera assemblage identified within the sample includes Pterostichus (Cryobius) sp., Hydrosporus sp., and Rhynchaenus (Isochnus) sp. The latter appears to feed on Salix sp., and probably occurs on Banks Island today. The presence of this species suggests, therefore, an environment similar to that of the present. In late-Holocene times, a return to colder climate conditions resulted in either rejuvenation or continued growth of epigenetic ice wedges. The accumulation of a surficial layer of humiferous wind-blown sand suggests that the climate also became more arid. This is consistent with the conclusion of Pissart et al. (1977) that eolian activity on Banks Island was initiated approximately 4,000 years B.P.

A number of problems remain in the interpretation of this stratigraphic sequence. Following the chronology of Vincent (1980), the Amundsen Glaciation is thought to be of mid-Wisconsin age, probably greater than 40,000 years B.P., while the late-Wisconsin ice limit is placed well to the east of Banks Island. The ^{14}C date of 10,600 \pm 130 years B.P. (GSC-3229) obtained from the detrital willow

horizon suggests, therefore, that either a period of much more recent glacial activity has affected southwest Banks Island, or that sediments underlying the Sachs River lowlands are nonglaciogenic. Since the sedimentology and spatial distribution of these materials are strongly indicative of a glaciofluvial origin, it would appear likely that at least part of southwest Banks Island received outwash from a late-Wisconsin ice sheet. The resolution of this problem, which has implications for Quaternary reconstructions elsewhere in the Western Arctic, can only be achieved by additional and more detailed stratigraphic investigations. In particular, the age of the Sand Hills event is crucial, since this clearly represents the last glacial episode on southern Banks Island. Vincent (1980, p. 207) argues that it is unlikely to represent a major glaciation, more recent than the Amundsen, since there is no evidence of such an event elsewhere on the island. At the same time, however, the problem remains of reconciling the fresh morainic topography of the Sand Hills ridges with the more subdued morphology of deposits attributed to the main Amundsen glacial advance. This contrast strongly suggests that a significant time interval separated the two events.

3.3 Pingos

3.3.1 Nature and occurrence

Pingos are defined as intrapermafrost ice-cored mounds. Investigation of their morphology, structure and

origin has formed a major component of permafrost geomorphology (e.g. Mackay, 1962; 1973a; 1977a; 1978b, c; 1979; Muller, 1947; Müller 1959; Porsild, 1938; Shumskii, 1964, pp. 21-27; Soloviev, 1952; 1973). This has contributed to a general assumption that pingos are a common permafrost landform. In reality, however, they develop in response to a narrow range of local and limiting hydrological conditions and thus are relatively rare within most permafrost regions.

The growth of pingos occurs in association with the aggradation of permafrost into taliks. This may occur either as a result of local geomorphic events, for example lake drainage and channel abandonment, or as a result of a regional climatic deterioration. Mackay (1973a) has shown that an increase in pore water pressure with respect to overburden pressure during talik freeze-back will theoretically result in a four-stage sequence of pingo growth. Under initial conditions of low pore water pressure, pore ice forms within the talik sediments. As pore pressure increases to a value equal to or greater than the overburden pressure, a lens of segregated ice begins to grow and lake bottom heave occurs. If pore pressure increases still further, water may be intruded into the overburden faster than it can freeze and tongues of injection ice are formed. Finally, when pore pressure exceeds the yield point of overburden sediments, the pingo ruptures. This cycle may be repeated several times during pingo growth.

It is also clear that expulsion of pore water from saturated sediments subjected to freezing within a closed system can produce a continuum of ground ice features. These range from tabular, sill-like sheets of injection ice, through conical pingos, to flat areas underlain by massive segregated ice (Mackay, 1972a; 1978b). It has been suggested, therefore, that the distinctive characteristic of a pingo is its discrete mound form, rather than the nature of its ice core (Mackay, 1979, p. 52).

On Banks Island, a number of pingos and pingo-like features have been recognized (French, 1975b; French and Dutkiewicz, 1976; Pissart and French, 1976). Although many lack the classical hemispherical morphology of Mackenzie Delta type pingos, they appear to be formed by essentially similar processes. A major difference, however, relates to their geomorphic setting. In contrast to Mackenzie Delta pingos, the majority of pingos on Banks Island occur on low fluvial terraces within major river valleys. Furthermore, many are in varying states of collapse. Melt-out of the ice core and slumping of overburden has typically resulted in the formation of either an annular ridge, surrounding a central depression, or an irregular hummocky topography. Pissart and French (1976, pp. 943-45) conclude that pingos in the Thomsen River area of northern Banks Island formed during a period of climatic deterioration, approximately 5,000-3,500 years B.P. At that time, permafrost aggraded into taliks previously developed beneath

river channels during the mid-Holocene climatic optimum.

Within this context, a typical range of pingos and pingo-like forms occur on southwest Banks Island. They are located predominantly on terraces within the Sachs and Kellett River drainage systems, although a few occur in drained thermokarst lake basins in the Sachs River lowlands (French, 1975b, p. 461). Morphologically, these features range from upstanding or partially collapsed conical mounds, 5-10 m in height, to elongate ridges often several hundreds of metres long. The former are smaller both in dimension and total number than their counterparts in the Mackenzie Delta area. This probably reflects the more rigorous climatic environment of Banks Island, which limits talik size and thus potential pingo development.

3.3.2 Pingo stratigraphy

Stratigraphic investigations of pingos provide an opportunity to deduce both the nature of pingo growth mechanisms and the paleoenvironmental conditions which led to their development (e.g. Mackay and Stager, 1966; Pihlainen et al., 1956; Pissart, 1967). During July 1980, a large collapsed pingo located near Carpenter Lake in the Sachs River lowlands (71°50'N, 124°28'W) was studied in detail. The specific objectives of this work were to verify Mackay's (1973a) model of pingo growth and to establish the age of pingos on southern Banks Island.

The Carpenter Lake pingo is located in an abandoned meltwater channel associated with the Sachs River drainage

system (Figure 3.15). The feature is 110 m in length, 55 m in width and the annular rampart rises to an elevation of 7.5 m above the central depression, which contains a small pond. These dimensions suggest that, at its maximum development, the pingo may have attained an elevation in excess of 20 m above surrounding tundra. A 20 m long section was excavated through the southeastern corner of the rampart to investigate pingo structure (Figure 3.16).

The central depression is underlain by a core of dark blue-grey massive ice, interpreted as segregation ice. The frequent mineral inclusions are inclined near-vertically; this orientation may reflect deformation of the ice core during pingo growth (Mackay, 1979, p. 42). Within the rampart, the sediments overlying the ice core are penetrated by several tongues of pure white ice, containing numerous gas inclusions (Figure 3.17). This ice has an unconformable contact with the main body of segregated ice (Figure 3.18). The lack of mineral inclusions, distinctive shape and location relative to the main ice core, all suggest that the white ice formed by injection of free water from within or beneath the growing pingo. These observations are in accordance with other studies in Arctic Canada (Mackay, 1973a; Pissart and French, 1976) and the Soviet Union (Baulin et al., 1978), which conclude that pingo ice cores may comprise both segregated and injection ice.



Figure 3.15

Oblique air view of the Carpenter Lake pingo, Sachs River lowlands ($71^{\circ}50'N$, $124^{\circ}28'W$). Plan dimensions are approximately 55 m x 110 m, and the rampart rises to a maximum of 7.5 m above surrounding tundra. Note the section excavated through the southwest rampart. July 11, 1979.

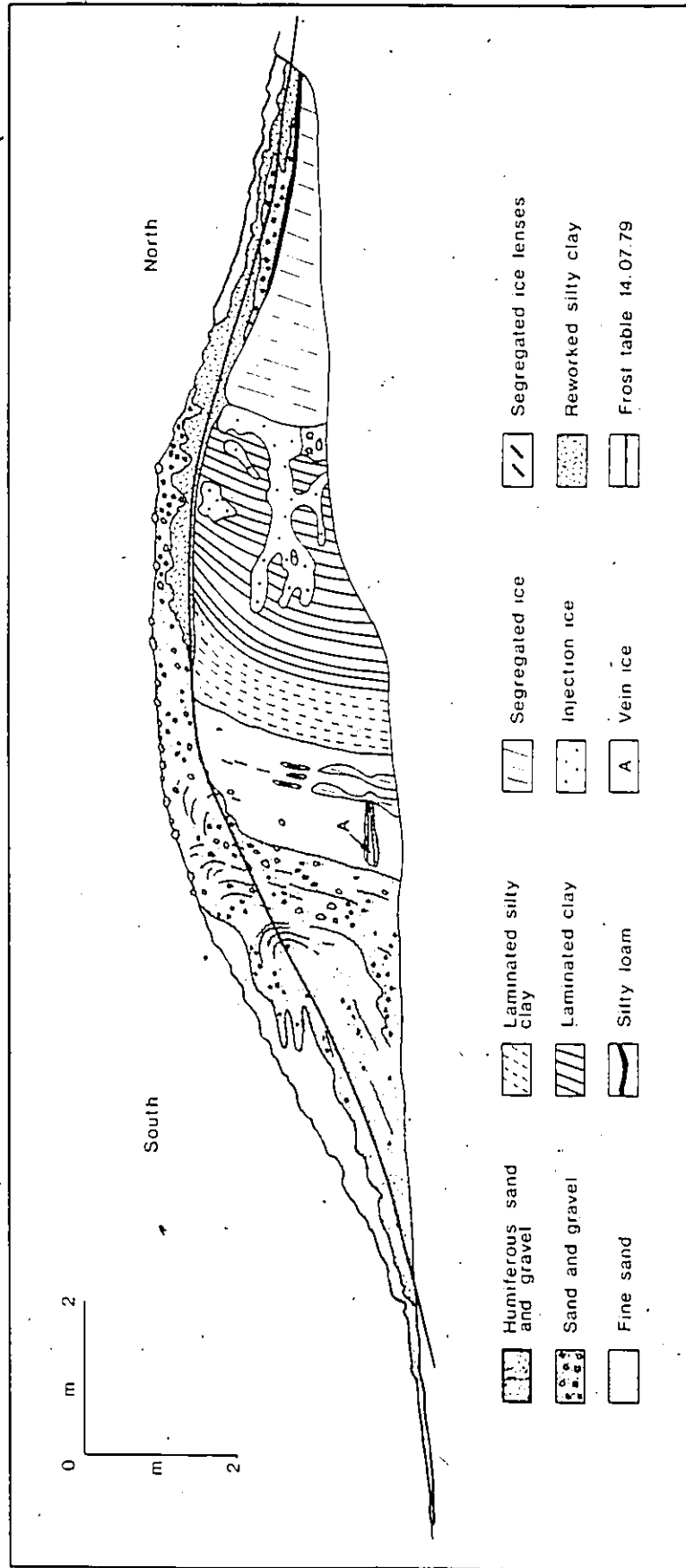


Figure 3.16 Section illustrating permafrost stratigraphy exposed in the southern rampart of Carpenter Lake pingo. Surveyed July 14, 1979.



Figure 3.17 Ice core and sediments exposed in the southern rampart of the Carpenter Lake pingo. Note tongues of white injection ice extending into the rampart and massive segregated ice underlying the central depression. (Photograph by H. M. French, July 14, 1979).

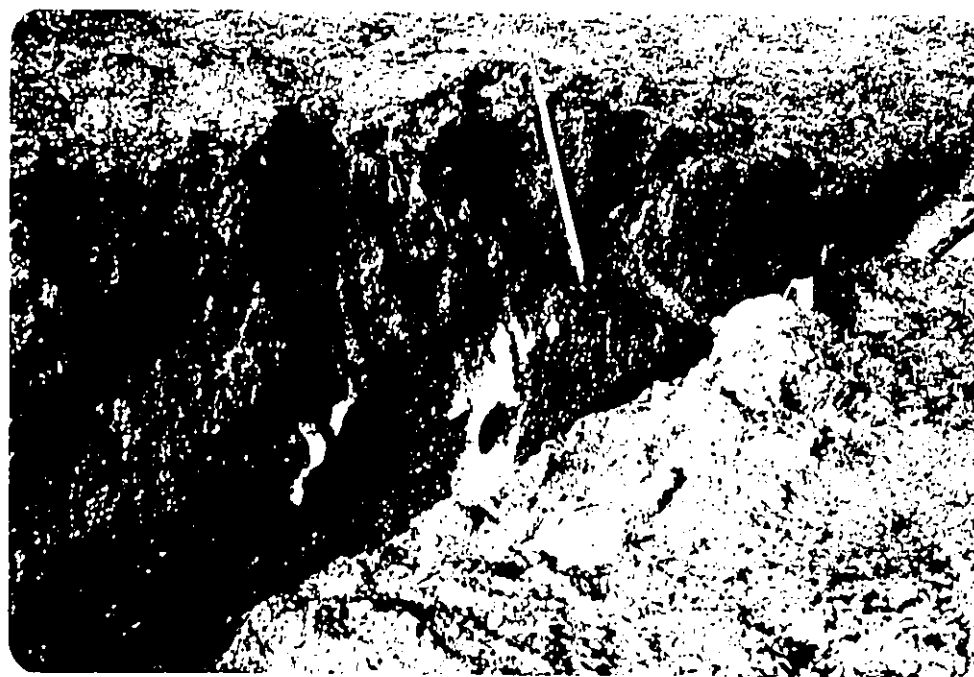


Figure 3.18 Detail of the contact between injection ice and segregated ice core within the Carpenter Lake pingo section. (Photograph by H. M. French, July 14, 1979).

A sequence of steeply dipping sedimentary units may be recognized within the pingo rampart. Immediately overlying the ice core is a 2.5 m thick unit of blocky, laminated clay. This material is ice-rich (greater than 60% excess ice content) and contains numerous small segregated ice lenses up to 5 cm in diameter. These sediments are flanked by approximately 1 m of laminated silty clay, containing 20-30% excess ice by volume. The outermost unit in the rampart sequence consists of at least 3 m of fine to medium ripple cross-laminated sand, grading upwards into gravel. Within the sand unit, a small, near-horizontal ice vein penetrates several large segregated ice lenses. The presence within the vein of gravel derived from a stratigraphically higher horizon demonstrates the epigenetic growth of this ice body.

3.3.3 Pingo growth

The sequence of ice bodies and enclosing sediments exposed within the pingo rampart may be used to reconstruct depositional paleoenvironments and permafrost history at this locality. The silt and clay were probably deposited in a low-energy, deep water environment, predating the meltwater channel in which the pingo is located. The lack of organic material in these sediments, which precluded the establishment of a radiocarbon-based chronology, may in part be a result of deposition in an ice-marginal zone; the Carpenter Lake pingo is located only 5 km from the Sand Hills glacial limit. The sand and gravel flanking

this material may be related to the establishment of fluvial conditions, associated with the development of proglacial and ice-marginal drainage channels. During the postglacial climatic optimum, taliks may have formed beneath the deeper channels on this sandur plain. Subsequently, a return to colder conditions, accompanied by abandonment of the meltwater channel, resulted in permafrost aggradation.

The pattern of permafrost aggradation into the underlying sediments is indicated by the distribution of ice bodies within the pingo rampart. The presence of the small epigenetic ice vein demonstrates the existence of permafrost prior to pingo growth. At least 2 m of frozen sand and gravel overlie the upper limit of the vein, which thus represents a thaw unconformity, probably related to temporary occupation of the river channel. The ice vein extends approximately 3-4 m inwards towards the pingo centre, indicating the probable extent of permafrost aggradation prior to initiation of ice core growth. This is consistent with the suggestion by Mackay (1978b, p. 139) that during permafrost aggradation the 'shut-off pressure', at which pore water is expelled from saturated sediments, is usually exceeded within a depth of several metres.

Four stages of pingo evolution may be deduced from the observations described above. First, it can be hypothesized that freezing of saturated sediments beneath the abandoned river channel resulted in pore water expulsion, and the growth of segregated ice lenses. Second, further

growth arched the overlying sediments upwards, to produce the sequence of steeply dipping units observed in the pingo rampart, and simultaneously reoriented the ice vein from a vertical to near-horizontal disposition. Third, at a later stage, increasing pore water pressure resulted in the intrusion of water into these sediments to form tongues of injection ice. Fourth, dilation cracking at the summit, together with solifluction and/or permafrost creep on the oversteepened pingo flanks, led to degradation of the ice core and pingo collapse to its present form. This reconstruction is consistent with Mackay's (1973a) theoretical model of pingo evolution.

3.3.4 Pingo age

Since no organic materials were found in sections excavated through the Carpenter Lake pingo, it was impossible to determine its period of growth. For this reason, attention was directed to a second pingo, located in the upper Kellett River basin approximately 65 km east of Sachs Harbour ($71^{\circ}54'N$, $123^{\circ}05'W$). This feature had been examined previously by H. M. French and D. N. Mottershead in July 1976, and was visited again by H. M. French and D. G. Harry in August 1980. It was thought that if the age of this pingo could be determined, then a similar age could be suggested for the Carpenter Lake feature.

Morphologically, this pingo consists of an elongate gravel ridge, 5-8 m in height, the southern end of which has partially collapsed to form a ramparted depression

(Figure 3.19). Within the central section of the ridge, part of the western flank is being eroded by an adjacent tundra lake. The coarse nature of surficial sediments precluded the excavation of a good section in 1976 (H. M. French, personal communication), and thus the presence of an ice core remains conjectural. However, the steep dip of exposed sediments, together with the collapsed central depression and evidence of active thermokarst processes, strongly suggest the melt of a substantial ice core. Re-examination of this feature in 1980 did not provide any cause to change this opinion.

Two radiocarbon dates, obtained from material collected by H. M. French and D. N. Mottershead in 1976, relate to the growth of the pingo. First, willow fragments contained within frozen gravel below the surface of the terrace on which the pingo is located provide a ^{14}C date of $3,920 \pm 80$ years B.P. (GSC-2395). Since this sample was collected approximately 2.5 m above river level (H. M. French, personal communication), there is little chance of contamination during flood events. The organic material occurred 1.4 m below the surface of the terrace, at a distance of approximately 400 m from the pingo. Thus, it probably provides a maximum age for the terrace and hence for the pingo itself. A second sample of detrital organic material, collected from the collapsed summit depression of the pingo, yields a ^{14}C date of $2,480 \pm 50$ years B.P. (GSC-2397). Since most field observations indicate that



Figure 3.19 Oblique air view of the Upper Kellett River pingo ($71^{\circ}54'N$, $123^{\circ}05'W$). Note the central summit depression and erosion by an adjacent thaw lake. August 06, 1980.

pingos grow most rapidly in their early stages of development (e.g. Mackay, 1973a, pp. 996-98), it is likely that material which accumulated in the summit depression post-dates the majority of pingo growth. Thus, available evidence suggests that this pingo developed during the time period separated by the two age determinations.

It seems reasonable to assume that, with the exception of those formed by more recent drainage of thermokarst lakes, the majority of pingos on southern Banks Island, including the Carpenter Lake feature, grew during this period. There is also broad synchronicity between the age of the Kellett River pingo and the age suggested for pingos in the Thomsen River area of northern Banks Island (Table 3.2). Pingo growth appears to have occurred rather earlier in the latter region, probably reflecting more severe climatic conditions.

3.4 Ice-ablation phenomena

3.4.1 Occurrence

A number of ice-ablation phenomena occur close to and within the Sand Hills glacial limit, approximately 40 km southeast of Sachs Harbour (Figure 3.20). Morphologically, they appear similar to collapsed pingos, since they consist of elongate or irregular ramparted depressions. In all cases, the features show evidence of thermokarst subsidence and are thus inferred to be a form of ground ice depression. Their wide distribution suggests that

Table 3.2
Radiocarbon Dates Pertaining to Pingo Growth,
Banks Island.

Locality	Laboratory Dating Number	Date (years B.P.)	Stratigraphic Position	Material	Significance
Northern Banks Island, Thomsen River and Able Creek ¹	GSC-2117	4,990±90	Sand overlying pingo ice core, Able Creek.	<u>Salix</u> sp.	Max. age for pingo growth.
	GSC-2124	3,460±80	Thomsen River terrace beneath eolian sand.	<u>Salix</u> sp.	Min. age for abandon- ment of pingo terrace.
Southern Banks Island, Upper Kellett Valley ²	GSC-2395	3,920±80	River terrace upon which pingo is located.	<u>Salix</u> sp.	Max. age for pingo terrace.
	GSC-2397	2,480±50	Summit depres- sion of pingo.	Fine or- ganic detritus.	Probable cessation of pingo growth.

¹ See Pissart and French, 1976.

² See French, Harry and Clark, 1982.



Figure 3.20 Air photograph illustrating the distribution of ice-ablation features in the Sand Hills area. (Part of A 15980-92). Note: Feature 'A' was examined in detail. Locality 'B' is the Carpenter Lake pingo.

much of the Sand Hills moraine belt may be ice cored.

One feature, located approximately 4 km southeast of Carpenter Lake (71°48'N, 124°26'W) was studied in detail in July 1979. The dimensions and morphology of this feature are similar to those of the Carpenter Lake pingo, since it consists of an elongate ramparted depression, 125 m long and 35 m wide (Figure 3.21). However, the depression trends directly downslope at an angle of 5-10°, and terminates in a thermokarst or kettle lake. It is difficult to reconcile this topographic situation with the limiting hydrological conditions required for pingo growth. The field objective was to investigate, therefore, the internal structure of the ice-ablation feature in order to determine its mode of origin.

3.4.2 Stratigraphy

Two 15 m long sections were cut through the northern and southern ramparts of the feature described above (Figure 3.22). In both sections, massive ice was encountered at a depth of approximately 1.0 m beneath the inner margin of the ramparts (Figure 3.23). The ice contains clay-silt foliations, which are contorted into isoclinal folds with an amplitude of 2-5 cm. Numerous mineral inclusions are present, including small pebbles up to 50 cm in diameter. Insufficient large clasts were seen to determine their fabric; however, several discoid pebbles were oriented with their maximum projection plane parallel to the surface of the ice body. The contact between the ice core and



Figure 3.21 Oblique air view of an ice-ablation feature, Sand Hills area ($71^{\circ}48'N$, $124^{\circ}26'W$). Note the downslope alignment of this feature, and the section cut through the southern rampart. August 06, 1980.

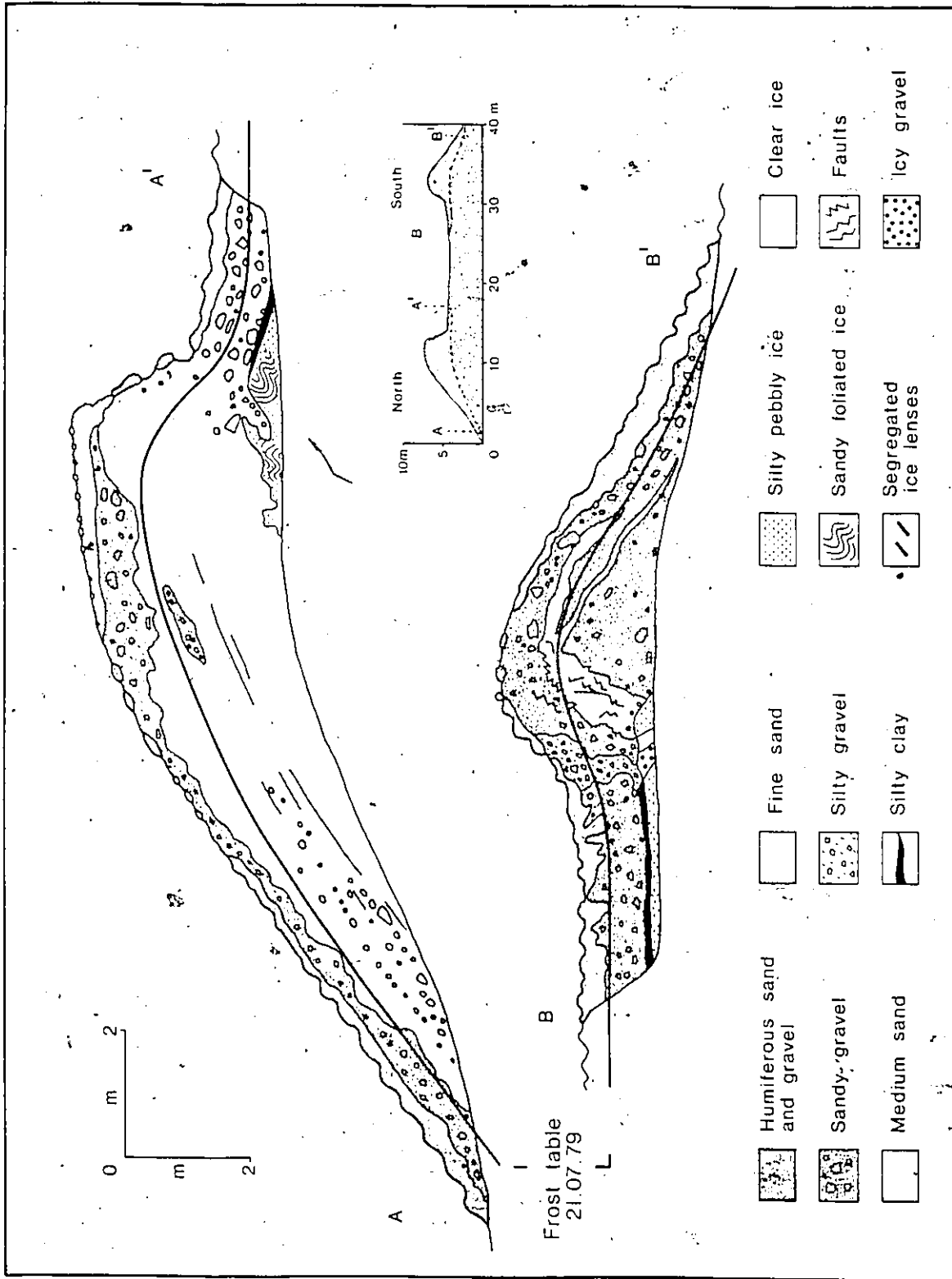


Figure 3.22 Section illustrating permafrost stratigraphy exposed in ramparts of the ice ablation feature shown in Figure 3.21. Surveyed July 21, 1979.



Figure 3.23 Detail of ice core exposed in the rampart of the feature shown in Figure 3.21. Note contorted foliations with mineral inclusions. July 21, 1979.

overlying sediments was determined to be irregular and frequently interpenetrating.

Within the central depression, a thin (0.5 cm) horizon of clay overlies the ice core, however this material is not present within the ramparts, where the ice is overlain by medium sand containing small ice segregations. The remainder of the rampart is composed primarily of coarse sand and clast-supported sandy gravel. These sediments are stratified and dip gently outwards, approximately parallel to the outer surface of the ramparts. No organic materials were found to enable radiocarbon dating of either the sediments or ice core.

3.4.3 Origin

Within areas of glaciated permafrost terrain, the origin of ice-cored mounds may be related to either periglacial or glacial processes. First, they may result from processes of ice segregation and intrusion, during periods of either local or regional permafrost aggradation as described earlier in this chapter. Second, they may reflect the presence of buried relict glacier or snowbank-derived ice, which has been preserved by the permafrost regime.

A variety of mounds may be formed by frost action. In particular, mounds may be either short-term, annual features (i.e. frost blisters, icing blisters or icing mounds), or long long-term, perennial phenomena (i.e. palsas or pingos). Most short-term frost mounds are relatively small, as their ephemeral nature precludes the development

of a large ice core (e.g. Van Everdingen, 1978). As such, they are unlikely to account for the Sand Hills features.

Palsas are peaty mounds, containing thin lenses of segregated ice, but consisting primarily of mineral soil (e.g. Seppälä, 1972; Zoltai and Tarnocai, 1971). For this reason, their pattern of thermokarst collapse is quite different to that of pingos, and rarely results in the formation of a ramparted depression.

The possibility exists that the Sand Hills features are hydrostatic pingos, similar to the Carpenter Lake and Kellett River features described above. However, such features (grow during permafrost aggradation into taliks, formed beneath water bodies. They are unlikely, therefore, to occur on sloping ground. On the other hand, hydraulic system pingos (e.g. Holmes et al., 1968; Hughes, 1969) grow as a result of movement of water to a freezing plane under gravity from a distant elevated source. They commonly occur in areas of discontinuous permafrost, where subpermafrost water circulates freely. It is possible that such conditions existed immediately subsequent to deglaciation of the Sachs River lowlands, as permafrost aggraded into unfrozen glacial sediments. Alternatively, the hydraulic gradient may have resulted from pore water expulsion beneath a fluctuating glacier margin (Mackay, 1959; Mathews and Mackay, 1960).

Stratigraphic evidence does not, however, support an hydraulic origin for the Sand Hills features. First,

the two rampart sections do not reveal symmetrical, steeply dipping stratigraphic sequences, as would be expected if the overburden had been uparched during ice core growth. Rather, they give the appearance of having been draped over a pre-existing ice body. Second, the nature of the ice core, particularly the intense deformation, presence of large mineral inclusions and irregular contact with overlying sediments, is typical of neither intrusive nor segregated ice. Moreover, the evidence of ice wedges (see pp. 72-74) indicates permafrost aggradation to have been continuous on southwest Banks Island in late-Pleistocene and Holocene times, with little evidence for a prolonged period of discontinuous permafrost, as would be required to form such large hydraulic system pingos.

For the above reasons, it is necessary to consider the alternative, i.e. glacial, origin for the Sand Hills ice-ablation phenomena. Air photo evidence suggests that the feature examined may be the western extension of a linear ridge, mapped by Vincent (1979) as a morainic crest. Although the two features are separated by a kettle lake, they are joined by a shallow subaqueous bar which may represent a remnant of the intervening section of ridge.

Ice-cored glaciogenic ridges may form initially as eskers, moraines, or during general ice disintegration.

Ice-cored eskers are known from modern glacierized areas in North America, Iceland, and Norway (e.g. Howarth, 1971; Petrie and Price, 1966; Price, 1969; Stokes, 1958).

Upon ablation of the ice sheet, englacial stream deposits may protect a remnant ice core. The structure of ice-cored moraines has been studied in Canada and Scandinavia (e.g. Østrem, 1964; Østrem and Arnold, 1970; Souchez, 1971). Johnson (1971) has shown that contemporary ice-cored moraines, associated with the Donjek glacier, form by burial of stagnant ice beneath ablation debris and outwash sediments. Slow degradation of either type of ice body results in the formation of a double-crested ridge, morphologically similar to many of the Sand Hills ablation features (e.g. Price, 1966).

The process of ice disintegration beneath a debris mantle may result in the formation of hummocky, 'uncontrolled', topography. However, the final stages of ice sheet movement may impose an alignment on the zones of weakness within the stagnant ice mass, resulting in the formation of 'controlled' disintegration topography (Gravenor and Kupsch, 1959). A range of ice-contact rings and ridges may form during controlled ice disintegration (e.g. Parizek, 1969). These are most frequently composed of till, although some contain stratified material. Stalker (1960) attributes the 'plains plateaus' and 'moraine plateaus' of Alberta to melt-out of ice pressed or squeezed into tunnels, holes and crevasses during ice sheet disintegration. These features consist of a near-circular till rim, surrounding a central fill of water-lain deposits. Some features have no central fill and thus form ramparted depressions. In

cross-section, many ice disintegration features (e.g. Winters, 1961) bear a striking resemblance to the Sand Hills ice-ablation phenomena.

It is difficult to discriminate with certainty between these possible modes of origin for the Sand Hills ice-ablation features. The presence of stratified sediments within the sectioned ramparts indicates an aqueous depositional environment. However, this may well be a secondary structure, developed during degradation of the ice core. The Carpenter Till, associated with the Sand Hills glacial advance, has a coarse, sandy gravel texture (Vincent, 1980, p. 71) and downslope reworking of this material by relatively low volumes of meltwater could readily account for the observed stratigraphy. Moreover, the presence of significant quantities of silt and clay within the rampart does not support a purely glaciofluvial origin for this material.

In summary, it seems most likely that the feature investigated represents a partially degraded ice-cored moraine. It may be hypothesized further that many, or all, of the morainic ridges within the Sand Hills glacial limit are ice-cored.

3.4.4 Implications for permafrost research

Two major implications arise from this discussion of the origin of the Sand Hills ablation features. First, it is clearly necessary to recognize the possible occurrence of glacier ice-cored topography in recently glaciated permafrost regions. Hughes (1973) has mapped the potential

distribution of 'glacier permafrost', although few documented instances of this phenomenon are known. Mackay (1971; 1973b) has argued convincingly against a glacial origin for the massive icy bodies which characterize coastal plains in Arctic Canada. However, Mackay et al. (1979) note the probable preservation of Wisconsin age ice in some Western Arctic localities. It is likely that detailed field investigations elsewhere on Banks Island and in other areas well within the late-Wisconsin ice limit, for example Victoria Island, would provide new evidence for this debate.

Second, the recognition within the same area of morphologically similar collapsed pingos and glacial ice-ablation features has significance for research in temperate regions which experienced Pleistocene cold climate conditions. Relict pingo scars are conclusive evidence of former permafrost conditions and many paleoenvironmental reconstructions are predicated upon correct identification of these features (Flemal, 1976). For example, a range of forms believed to be pingo remnants have been identified in Western Europe (e.g. Cailleux, 1956; Mitchell, 1973; Pissart, 1963; Sparks et al., 1972; Watson, 1971; Watson and Watson, 1974; Wiegand, 1965). Many of these features occur at typical hydraulic system pingo sites, for example on lower valley-side slopes and at springlines. They frequently possess a complex morphology, consisting of linear ridges and open or closed mutually interfering ramparts.

Relatively few pingo remnants have been identified within the North American zone of Pleistocene permafrost, although a periglacial origin has been proposed for both the prairie mounds of western Canada (Bik, 1969) and the de Kalb mounds of Illinois (Flemal et al., 1973). Both types of feature consist of a conical mound rather than a ramparted depression, suggesting that material moved into the mound rather than out from it, as would be expected during pingo collapse. Flemal (1976) explains this surplus of material in terms of relief inversion during degradation of surrounding permafrost. However, since no modern analogue of this process has been observed, it seems equally reasonable to attribute a glacial origin to these mounds.

Identification of Pleistocene pingo scars has been facilitated by the rapid growth in knowledge relating to contemporary pingos. In particular, recognition of possible variation in pingo form permits a more liberal interpretation of relict features. However, three serious problems remain in the use of pingo scars in Pleistocene paleoenvironmental reconstructions. First, Pleistocene permafrost conditions probably have no exact modern analogue; thus it is impossible to deduce the hydrologic conditions under which pingos may have developed. Second, regional degradation of ice-rich permafrost may result in the development of chaotic thermokarst terrain, in which pingo remnants are likely to be unrecognizable. Third, as comparison of the Carpenter Lake pingo and Sand Hills ice-ablation feature suggests,

morphologically similar landforms may result from melt-out of ice cores of contrasting origin. It is possible, therefore, that many features identified as Pleistocene pingo scars and occurring in previously glaciated terrain may have developed by ablation of buried glacier ice.

CHAPTER FOUR

THAW LAKES

4.1 Introduction

Thaw lakes develop by thermokarst subsidence of ice-rich permafrost terrain. Although their occurrence, particularly on arctic coastal plains, has been reported from both northern Alaska (e.g. Anderson and Hussey, 1963; Cabot, 1947; Hopkins, 1949; Muller, 1947, p. 84; Wallace, 1948) and Siberia (e.g. Bojcov, 1965; Litvinov, 1960; Orlov, 1960; Pavlov, 1965; Romanovskii, 1961; Simova, 1964), the conditions under which thaw lakes evolve remain poorly understood. Much research has focused upon thaw lake morphology and the development of a preferred long-axis orientation (e.g. Black and Barksdale, 1949; Carson and Hussey, 1960a, b; 1962; Kuznecova, 1961; Livingstone, 1954; Price, 1963; Rex, 1961; Rosenfeld and Hussey, 1958; Stremyakov, 1963).

In northern Canada, oriented thaw lakes occur in several areas, including the Tuktoyaktuk Peninsula, Old Crow Basin, and Great Plain of the Koukdjuak (Bird, 1967, pp. 211-216; Mackay, 1956a, b; 1957; Price, 1968). In addition, they are a major terrain component on southwest Banks Island, occurring extensively in the Sachs River lowlands and Kellett River valley. Their initiation, morphological development and thermal influence constitute significant problems within the field of permafrost

geomorphology (Black, 1969)).

This chapter describes four aspects of the thaw lake problem with respect to southwest Banks Island. First, the geologic and thermal controls over thaw lake growth are established. Second, stratigraphic sections, exposed beneath drained lake basins, are examined to deduce paleoenvironments associated with thaw lake development. Third, lake morphology is analyzed to investigate possible equilibrium relationships between lake form and geomorphic processes. Fourth, a number of lake drainage mechanisms are discussed, with reference to field observations and air photo interpretation.

To place the thaw lakes of this area into a regional perspective, some discussion of lake development on Banks Island in general is appropriate. Four major lake types may be recognized (Figure 4.1). Within the eastern moraine belt, a number of large kettle lakes of irregular outline exist, formed by the melt-out of buried glacier ice. Smaller kettle lakes also occur within the area of the Sand Hills Readvance on southwest Banks Island. The western and central lowlands are characterized by numerous small lakes which belong to one of two categories. The first type consists of isolated basins occurring within dissected interfluvial zones. These have been termed 'ground moraine ponds' (Fyles, 1962, p. 14), and are contained within basins formed by topographic irregularity of the surficial sediment cover. The second and more important type occurs within concentrated

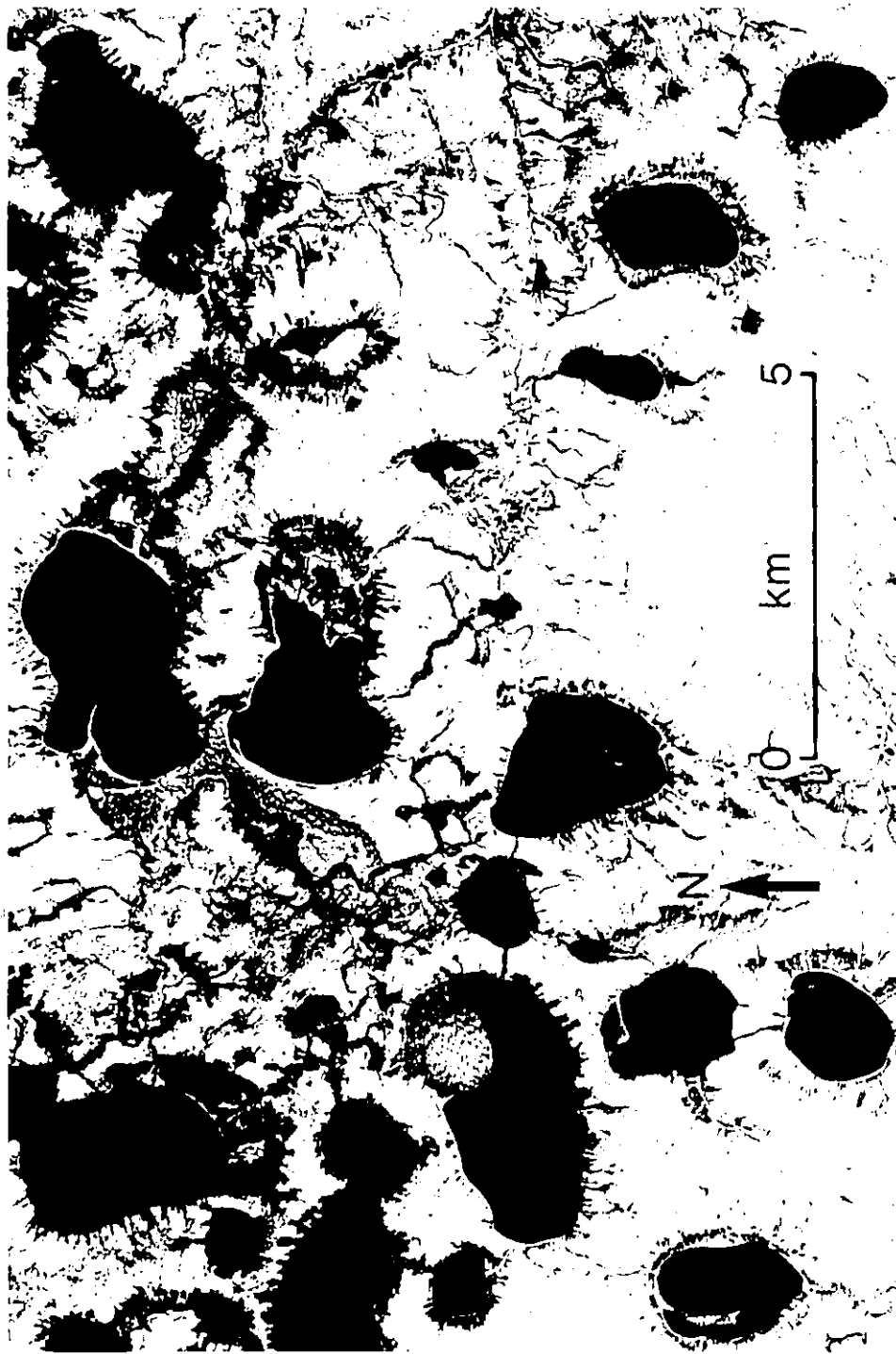


Figure 4.1a Air photographs illustrating lake development on Banks Island. Large kettle lakes located in the moraine belt of eastern Banks Island, approximately 20 km north of Johnson Point. (Part of A 16829-67).

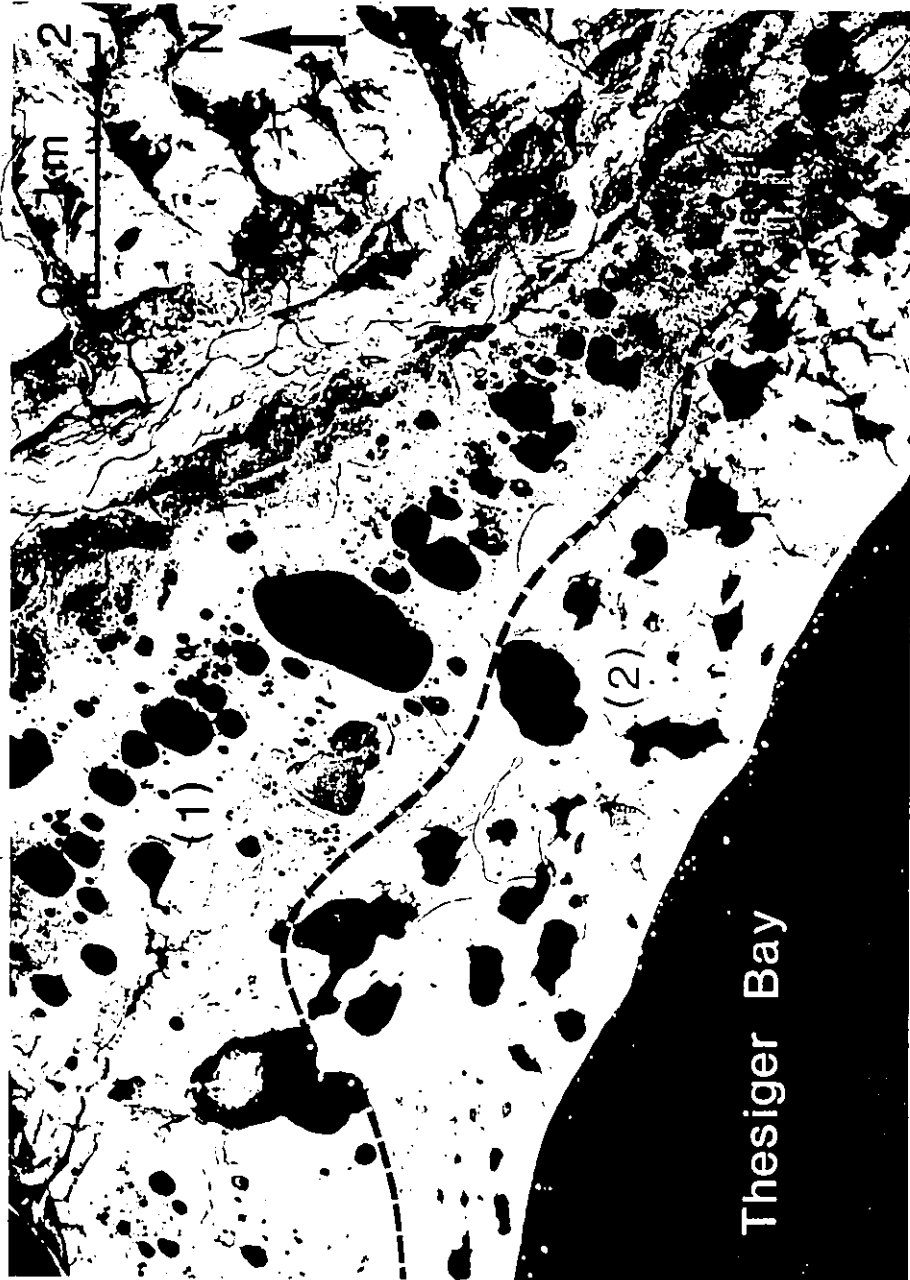


Figure 4.1b Air photographs illustrating lake development on Banks Island. Lake development 40 km southeast of Sachs Harbour. (Part of A 15980-91). Note the contrast between (1) Oriented thaw lakes, developed on a terrace of the Sachs River and, (2) Small kettle lakes developed within the limit of the Sand Hills Readvance.



Figure 4.1c

Air photographs illustrating lake development on Banks Island. Lake development near the Big River delta, approximately 55 km north of Sachs Harbour. (Part of A 17053-11). Note the contrast between (1) Numerous tundra ponds (thaw lakes) developed on terraces of the Big River and, (2) Isolated ground moraine ponds north of the river valley.

zones, containing tens to hundreds of members, developed upon fluvial and glaciofluvial terraces and outwash plains. These have been termed 'tundra ponds' (Fyles, 1962, p. 14), and are believed to originate primarily by the melt-out of ground ice. Although it is likely that all lake basins developed in ice-rich permafrost owe their origin at least in part to thermokarst processes, it is to this latter group that the term 'thaw lake' is restricted in the following discussion.

4.2 Geologic and thermal controls

4.2.1 Ground ice and potential thaw subsidence

The primary factor influencing thaw lake occurrence is the presence of excess ice within unconsolidated sediment. If the volume of ice within frozen ground is known, then the potential thaw subsidence resulting from a given depth of thaw may be calculated (e.g. Hussey and Michelson, 1966; Sellmann et al., 1975). Observations of ground ice distribution in the Sachs River lowlands enable calculation of ground ice volume, based on a methodology developed by Brown (1967) in northern Alaska and subsequently applied by Pollard and French (1980) in the Mackenzie Delta area. Following these authors, the dominant components of total ice volume are believed to be wedge ice, segregated ice and pore ice.

To determine the contribution of pore and segregated ice, gravimetric ice content (W) and excess ice

content (E) of sediments were sampled at 50 cm vertical intervals within three coastal permafrost sections, located approximately 4.0 km southeast of Sachs Harbour. Volumetric ice content (V_i) was calculated following the methodology of Pollard and French (1980, p. 511):

$$V_i (\%) = V_i \times 100 / (V_i - V_s) \quad [1]$$

where, $V_i = W + (0.009W) \quad [2]$

and $V_s = Y_d / P_s \quad [3]$

(where: V_i = volume of ice, V_s = volume of soil solids, Y_d = dry soil weight, P_s = particle density). The contribution of wedge ice to the ground ice profile is complex, since both active and inactive (i.e. buried), ice wedges occur in this area. Analysis of ice wedge distribution in permafrost sections, together with measurement of polygon diameters along a 2.0 km coastal transect, suggests that active epigenetic wedges account for approximately 15% by volume of the upper 7.0 m of permafrost. Smaller wedges, truncated at the base of the organic horizon 1.5-2.0 m below surface, account for a further 5% by volume. At depths between 5.0 m and 10.0 m, inactive epigenetic wedges developed within the grey silty sand, together with basal sections of active wedges, occupy approximately 11% of permafrost.

Typical values of V_i and E for the upper 8.0 m of permafrost in the Sachs River lowlands are shown in Table 4.1. It is likely that these data represent maximum values, as ice wedge volume may have been overestimated. Mackay (1977b) notes that, where ice wedges are measured in linear exposures, apparent width is greater and apparent spacing less than true values. As a first approximation, however, the present analysis suggests that ice may constitute 58.9% of the upper 8.0 m of permafrost. Total thermal degradation of this material may result in thermokarst subsidence of at least 3.0 m. This is consistent with field observations which suggest that most thaw lakes on southwest Banks Island range from 1-3 m in depth. However, several large lakes in the Sachs River lowlands may be considerably deeper. For example, the estuary of the Sachs River consists in part of drowned lake basins which are between 30 m and 40 m in depth (Canadian Hydrographic Service, 1972). It is unlikely that they could have formed solely by melt-out of ground ice and may instead represent kettle lakes, formed by degradation of buried glacier ice.

4.2.2 Thermal effects

The development of a thaw lake produces a heat reservoir which substantially disturbs equilibrium geothermal conditions (Chekovskii and Shamanov, 1976; Chizov, 1972; Grigor'yev, 1959; Mackay, 1979, pp. 29-30). An important distinction exists between shallow lakes, with depths less than the maximum thickness of winter ice cover, and deeper

Table 4.1

Representative Values of Volumetric Ice Content, Excess Ice Content and Potential Thaw Subsidence, Sachs River Lowlands.

Depth (m)	Material	Mean Porosity ¹	Volumetric Ice Content V_i (%)			Excess Ice Content E (%)	Total Cumulative Thaw Subsidence (cm)
			Pore and Segregated Ice (V_{PS})	Wedge Ice (V_W)	Total Ice (V_T) ²		
0.5 - 1.0	Peat	N/A	50.0 ³	28.8	64.4	80.0 ³	40.0
1.0 - 2.0	Sand	0.30	36.4	25.3	52.5	32.1	72.1
2.0 - 3.0	Sand	0.30	36.7	20.8	49.9	28.4	100.5
3.0 - 4.0	Sand	0.30	37.2	16.3	47.4	24.9	125.4
4.0 - 5.0	Silty Sand	0.45	65.8	14.0	70.6	46.5	171.9
5.0 - 6.0	Silty Sand	0.45	75.4	16.0	79.3	62.4	234.3
6.0 - 7.0	Silty Sand	0.45	60.8	15.8	67.0	40.0	274.3
7.0 - 8.0	Silty Sand	0.45	55.2	11.3	60.3	27.8	302.1

¹ Porosity calculated from known values of V_i and E.

² Total ice volume calculated from $V_T = V_W + (V_{PS} \times (100 - V_W))$.

³ Ice contents of peat represent visual estimates only.

lakes. Since the former freeze completely in winter, the mean annual lake bottom temperature may be almost as low as that of the surrounding tundra. A talik cannot develop beneath such lakes. If, however, a lake remains unfrozen at depth throughout the winter, mean annual lake bottom temperature is positive and a talik forms. Sellmann et al. (1975) suggest that, as a thaw basin deepens, it is in the central zone that water depth first exceeds maximum winter ice thickness. Thus, the talik forms first beneath the centre of the basin, promoting rapid thaw subsidence, while the sublittoral zones of the lake, which continue to freeze in winter, remain as shallow shelves. This provides a possible explanation for the bipartite structure of lake basins observed in Alaska and the Mackenzie Delta (Carson and Hussey, 1962; Mackay, 1963a, p. 45), and also visible on air photos of the Sachs River lowlands (Figure 4.2). If thermal disturbance is sufficient, a through-going talik may develop, completely perforating the permafrost. Mackay (1979, p. 30) suggests that, as a first approximation, lakes with minimum widths of twice the undisturbed permafrost thickness will be underlain by throughgoing taliks.

The chief factor determining the thermal influence of lakes on underlying permafrost is the mean annual temperature of bottom sediments. This is related to the lake thermal regime, itself a function of regional climate. Although no systematic records of lake temperature are

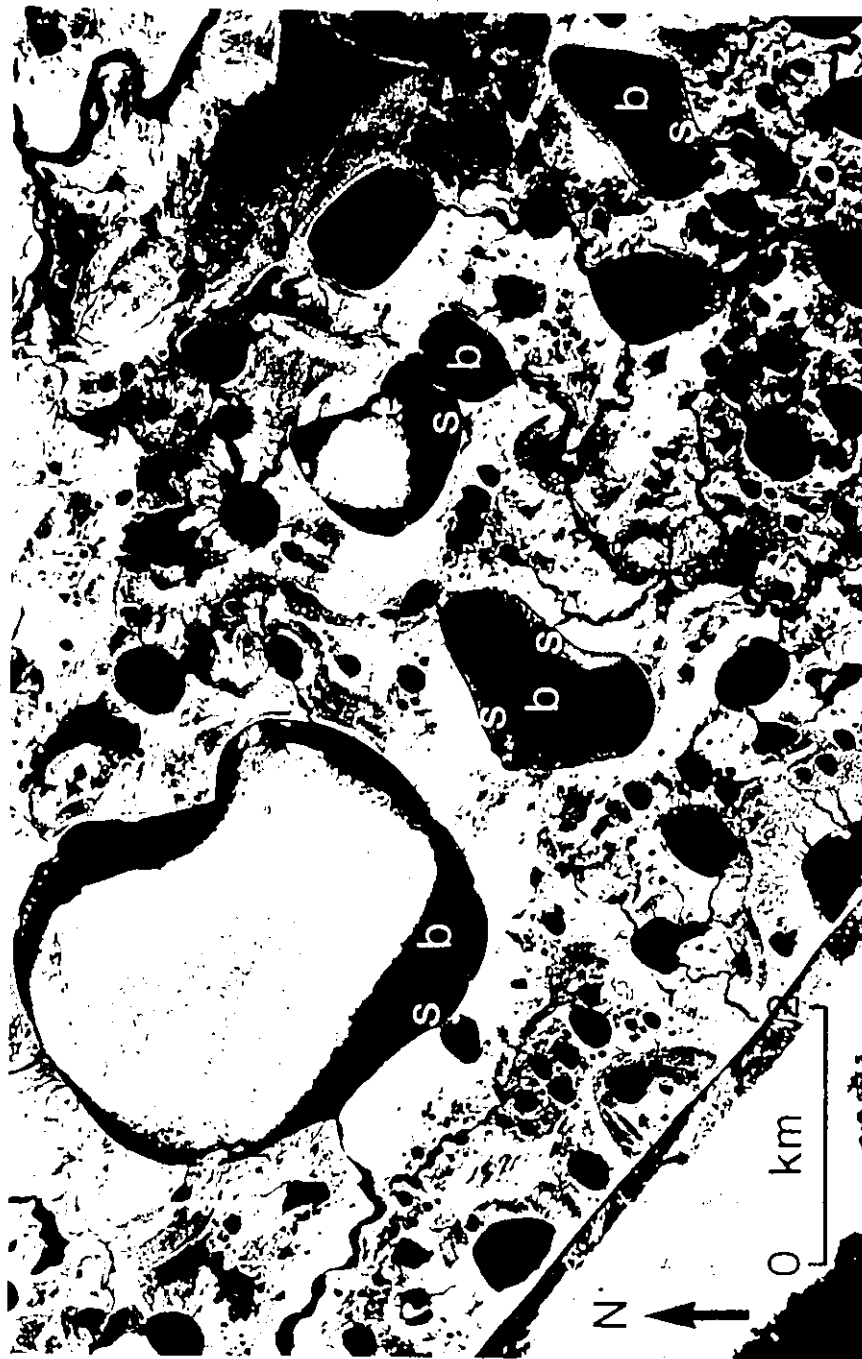


Figure 4.2

Air photograph illustrating typical thaw lake terrain in the Sachs River lowlands, 10-20 km southeast of Sachs Harbour (Part of A 15980-29, dated July 14, 1958). Note: (1) Development of deep central basins 'b' and shallow sublittoral shelves 's'; (2) Truncation of several basins by coastal retreat; (3) Presence of residual ice cover on some deep lakes.

available on southwest Banks Island, ice cover has been monitored on a small lake (approximately 500 m in radius), located close to the Department of Transportation facility on the Sachs Harbour ridge. During the period 1958-1981, open water conditions have existed, on average, for 70 days between July 07 and September 16 (Table 4.2). Ice breakup is initiated on average a full month before the lake is completely open. A maximum ice thickness of approximately 2.0 m is attained shortly prior to breakup. Although ice cover is likely to persist later on larger lakes, it may be assumed that these data are representative of most lakes on southwest Banks Island.

Few comprehensive thermal surveys of arctic lakes have been undertaken, however some data are available which allow estimation of mean lake temperatures, given a knowledge of ice cover. For example, Brewer (1958) measured water and bottom sediment temperatures in Imikpuk Lake, Alaska. This lake, which is 1.1 km long and 0.7 km wide, has a maximum water depth of 2.8 m. Measurements indicate that, during the open water period, the lake is essentially isothermal. During the period of ice cover, the lake bottom at depths greater than 1.98 m (the maximum winter ice thickness) remains at a temperature of between 0.5°C and 1.5°C. However, beneath shallow areas of the lake, which are completely frozen in winter, bottom sediment temperatures fall below 0°C. Thus, the mean annual temperature of bottom sediments may range from -5°C or less close to the shore,

Table 4.2

Breakup and Freezeup Data, Unnamed Lake, Sachs Harbour, 1958-1981.

	Earliest	Latest	Mean	Standard Deviation (Days)	Number of Years
First ice deterioration	May 15	June 27	June 08	15.1	16
Lake clear of ice	June 22	Aug. 04	July 07	11.2	17
First ice formation	Aug. 28	Sept. 30	Sept. 16	9.7	16
Complete ice cover	Sept. 07	Oct. 10	Sept. 26	8.0	18

Open Water Conditions	Minimum	Maximum	Mean	Standard Deviation (Days)
Duration (days)	32	95	70.9	14.5
% of year	8.8	26.0	19.4	-

Source: Allen, 1964 and Station Records, D.O.T. Sachs Harbour.

to between 1.5°C and 2.0°C near the lake centre.

A similar thermal model is probably applicable to lakes on southwest Banks Island. Within the unfrozen portion of deep lakes, bottom sediment temperature is likely to range from zero to 1.5°C during the winter period of ice cover. During the open water period, mean lake temperature is closely related to mean air temperature, and probably ranges from 3.0°C to 5.0°C. Mean annual bottom sediment temperature is thus likely to lie between 0.5°C and 1.5°C. Shallower lakes will have correspondingly lower mean annual bottom temperatures, depending primarily on the duration of bottom-fast ice.

These data may be used to predict the impact which thaw lakes have upon the ground thermal regime. Equilibrium geothermal conditions may be computed using numerical simulations based on heat conduction theory (e.g. Smith, 1973; 1977; Smith and Hwang, 1973). For the purpose of geothermal analysis, thaw lake terrain may be divided into two regions, lake and tundra. Subsurface temperature at any point is then determined by the extent to which the lake component modifies the tundra ground thermal regime. Ground temperature, T , at depth, z , may thus be calculated from the summation of three factors:

$$T_z = T_g + z \cdot Q + \theta \quad [4]$$

where:

- T_g = undisturbed equilibrium surface temperature;
 $z \cdot Q$ = the increase in temperature with depth due to the geothermal gradient, Q ;
 θ = the increase in temperature due to the presence of lakes.

Lachenbruch (1957a) has developed a solution for θ , which may be applied directly to the thaw lake problem. Under long-term equilibrium conditions, the solution may be stated as:

$$\theta_{(x,y,z)} = (T_g - T_1) \sum_0^{2\pi} \frac{\lambda}{360} \left[\frac{1}{1 + (R_1/z)^2} - \frac{1}{1 + (R_2/z)^2} \right] \quad [5]$$

where:

T_1 = mean annual lake bottom temperature.

The first term in this expression represents the difference between equilibrium lake bottom and tundra surface temperatures. The second term represents a summation of lake temperature influence at the common apex of sections of a circular annulus of central angle, λ , and inner and outer radii, R_1 and R_2 (see Brown, 1963; Smith, 1973, p. 59).

On southwest Banks Island, ground temperatures are further modified by proximity to the Beaufort Sea. Under equilibrium conditions, Lachenbruch (1957b) showed that the thermal effect of an ocean at a distance, d , from the shoreline

and at a depth, z , below surface, is approximated by:

$$a = (T_s - T_g) \left[\frac{\tan^{-1} z/d}{180} \right] \quad [6]$$

where:

T_s = mean annual sea bottom temperature.

Two models of lake thermal influence may be derived from equation [5] based on previous analyses by Mackay (1963a, pp. 78-80) and Smith (1973, pp. 65-72). The simplest application is to consider the thermal disturbance beneath the centre of a circular lake of radius R . Then, $R_1 = 0$, $R_2 = R$, and equation [5] reduces to:

$$\theta(x,y,z) = (T_g - T_1) \left[1 - \frac{1}{1 + (R/z)^2} \right] \quad [7]$$

The first model uses this relationship to calculate equilibrium thermal conditions beneath thaw lakes and is, for example, primarily of use in estimating maximum talik depths associated with lakes of varying dimensions. The second model uses equation [5] to calculate the thermal contribution of lakes to the ground temperature field at any point. Program listings for both models are presented in Appendix B.

The following assumptions are implicit in both models: First, heat transfer takes place by conduction alone; thus for example, convective heat transfer within lakes is not considered. Second, earth materials are homogeneous

in both space and time with respect to thermal properties. Thus, for example, the latent heat effects associated with water-ice phase changes are not considered. Third, ground temperatures are related to current climatic conditions, which have not changed significantly during the period required to establish an equilibrium geothermal regime.

The first model was tested by comparison of observed and predicted talik development, in the Mackenzie Delta and Alaskan coastal plain regions. Assumed values of thermal parameters are listed in Table 4.3. In the Mackenzie Delta area, lakes greater than 85 m in radius are predicted to possess throughgoing taliks, which perforate the estimated 100-150 m of permafrost. These results are consistent with field observations in this area (Brown et al., 1964; Johnston and Brown, 1961; 1964; 1966; Smith, 1973). In northern Alaska, the model predicts that a critical lake radius of 300 m is required to generate a talik which perforates the estimated 300-400 m of permafrost. This suggests that Imikpuk Lake should possess a throughgoing talik, rather than an open talik as indicated by Brewer (1958). This is not thought to be a fault of the model, but rather may indicate that permafrost temperatures in this region are in disequilibrium with the present climate (see, for example, Lachenbruch and Marshall, 1969). When applied to southwest Banks Island, the model indicates that deep lakes which are greater than 700-800 m in diameter are probably underlain by throughgoing taliks. Lakes greater than 2.0 m in depth, but below

Table 4.3

Representative Temperature Values Used in Geothermal Models of Thaw Lake Terrain.

Parameter	Southwest Banks Island		Rationale	Source
	Mackenzie ¹ Delta	Northern ² Alaska		
Mean annual surface temperature T_G (°C)	-4.2	-9.0	-10.8	Mean annual air temperature, $T_A = -14.1^\circ\text{C}$ (Personal commun.) $T_G = T_A + 3.3^\circ\text{C}$ Brown (1966, 1973)
Mean annual lake bottom temperature T_L (°C)	3.2	1.8	0.5 to 1.5	T_L (winter) = 0.0°C to 1.0°C T_L (summer) = 3.0°C to 5.0°C Based on thermal regime of Imikpuk Lake, Alaska, modified with respect to Sachs Harbour climate. Brewer (1958)
Mean annual sea bottom temperature T_S (°C)	-	-	-1.3	Mean annual Beaufort Sea temperature in Alaskan nearshore zone. Lewellen (1973, pp. 131-132)
Mean geothermal gradient Q (mK/m)	25.0	22.2	22.2	Mean geothermal gradient measured at Storkerson Bay A-15 wellsite. Taylor and Judge (1974, p. 122)

¹ After Smith (1973).

² After Brewer (1958) and Péwé (1975).

this threshold diameter, are underlain by open taliks of varying dimensions (Figure 4.3).

The second model was used to construct a geothermal transect across the Sachs River lowlands, to illustrate the considerable thermal influence of thaw lakes in this area (Figure 4.4). Maximum predicted permafrost thickness is 450-500 m beneath land and 50-60 m offshore. However, both the upper and lower permafrost boundaries vary greatly in depth, due to the influence of water bodies. Beneath lakes approximately 50 m in radius (for example lakes A and C on Figure 4.4), the thermal disturbance extends downwards for 100-150 m and shallow taliks are developed. Beneath lake B, which has a radius of approximately 250 m, the upper permafrost boundary is depressed to form a 40 m deep talik. The thermal influence of this lake is also sufficient to raise the lower permafrost boundary. If lake B were 50% larger, the permafrost boundaries would meet to form a narrow 'hour-glass' talik. Lakes greater than 1.0 km in radius (for example lakes D and E in Figure 4.4), may be expected to possess throughgoing taliks under equilibrium conditions. The thermal influence of such lakes is very extensive; for example, although permafrost exists beneath the narrow isthmus which separates lakes D and E, it is predicted to extend only 250 m below surface.

Models of this type are limited by their inability to predict disequilibrium permafrost conditions. In many regions, permafrost temperatures are related to past

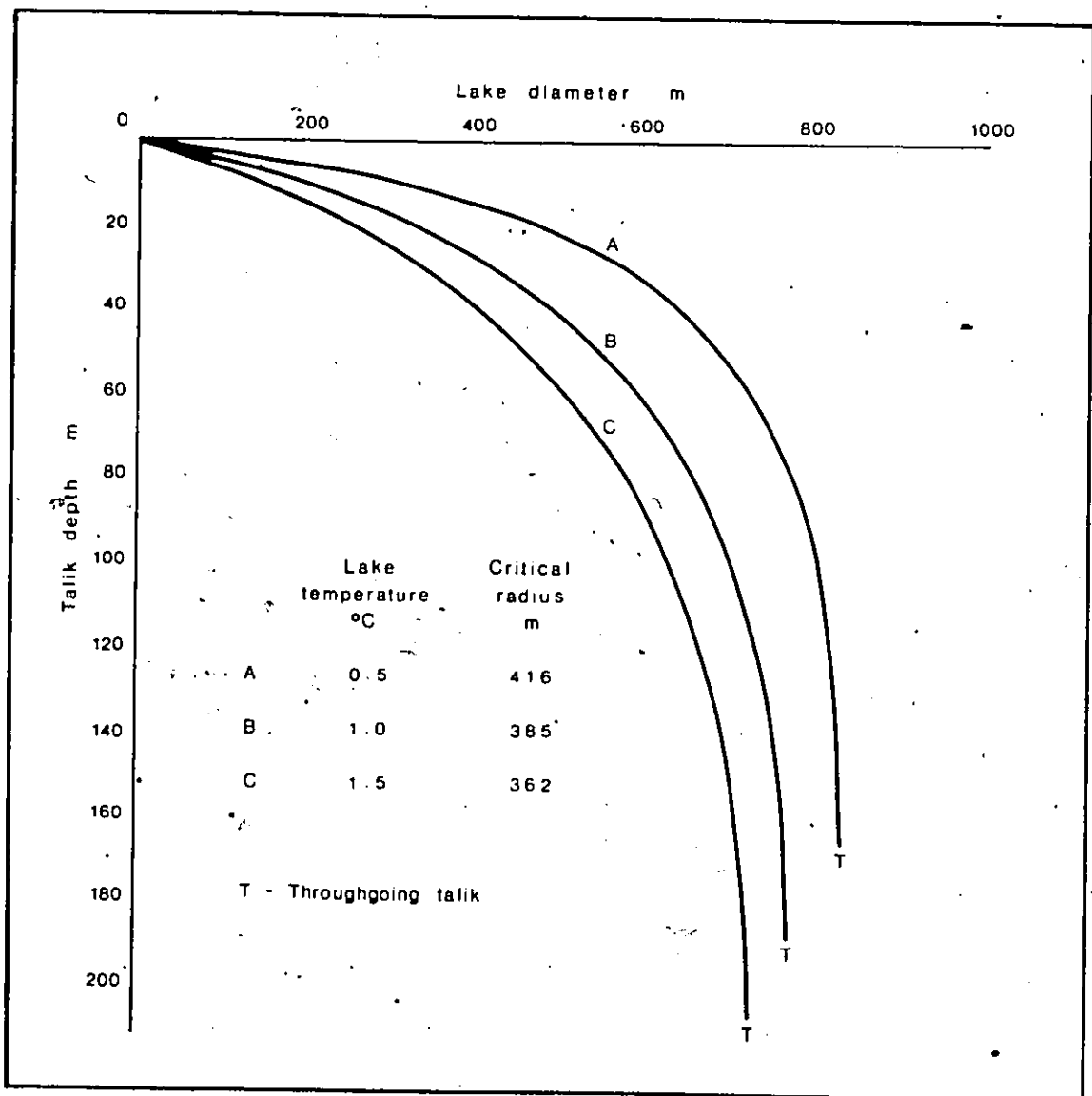


Figure 4.3

Predicted equilibrium talik dimensions beneath circular thaw lakes on southwest Banks Island.

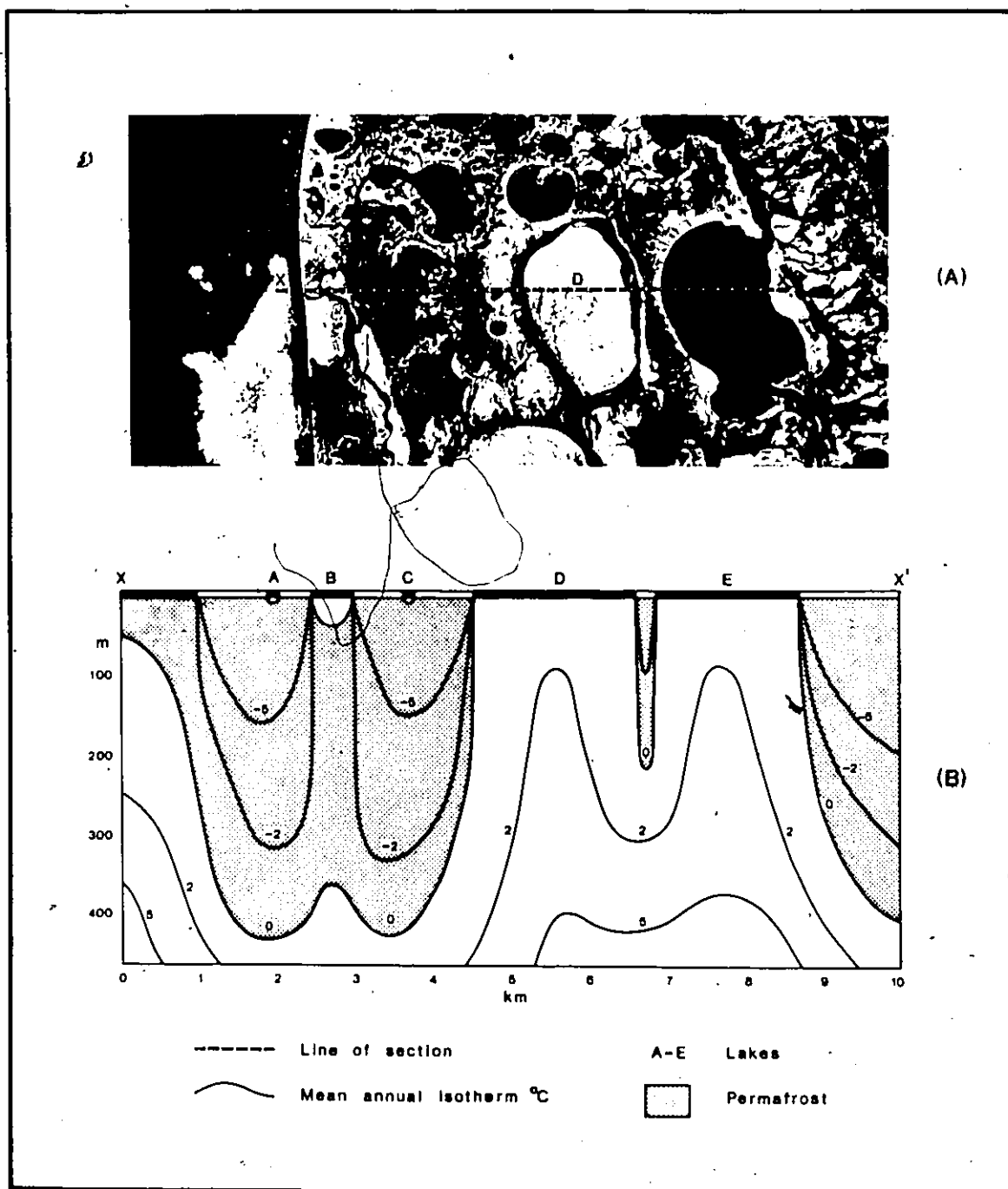


Figure 4.4 Geothermal transect across a part of the Sachs River lowlands.
 (a) Air photograph showing terrain conditions along the line of section (parts of A 15980-31, 66).
 (b) Predicted equilibrium mean annual ground temperature field.

climatic conditions. The downward penetration of a thermal wave is such that a lag develops progressively with respect to surface temperature. For example, the depth of permafrost aggradation into unfrozen sediments is approximated by:

$$z_p = b \sqrt{t} \quad [8]$$

where:

- z_p = permafrost thickness;
- b = the constant in Stefan's heat conduction equation;
- t = elapsed time since initiation of permafrost growth.

In areas of thaw lake terrain, disequilibrium thermal conditions may be related to both lake development and drainage. For example, Mackay (1979, p. 30) estimates that a thaw period of 10,000-20,000 years would be required to form a throughgoing talik in 500 m of permafrost with a mean surface temperature of -10.0°C . In the Mackenzie Delta area, temperature anomalies of $2-3^{\circ}\text{C}$ commonly exist beneath lake basins more than 100 years after drainage. Substituting values of $b = 2.0 - 3.0$ in equation [8], it follows that a 100 m talik beneath a drained lake may require 1,000-2,000 years to freeze completely. In spite of these limitations, thermal modelling provides a useful tool in the study of

thaw lake terrain. For example, predicted talik dimensions may be used to estimate the effect of thaw lake development on regional permafrost stratigraphy.

4.3 Basin stratigraphy and age

Coastal truncation of thaw lakes in the vicinity of Sachs Harbour provides a unique opportunity to examine permafrost stratigraphy beneath drained basins. This may be interpreted to reconstruct paleoenvironments associated with thaw lake evolution. One section, located approximately 4.0 km southeast of Sachs Harbour (71°57'N, 125°14'W) was examined in detail in August 1980. At this locality, an elongate basin has been truncated across its short axis and is now drained completely, with the exception of a small residual pond approximately 100 m from the coast. Along the line of the coastal section, the basin is 420 m wide and has a maximum depth of 5.0 m. A number of stratigraphic sections, spaced at irregular intervals across the basin, were examined in the 3.0-9.0 m high sea-cliffs (Figure 4.5).

On either side of the basin, the stratigraphic sequence is similar to that described previously (see pp. 62-66). At least 5.0 m of grey silty sand is overlain by 1.0-2.0 m of yellow-brown medium sand, with a detrital organic horizon within the contact zone. Surface sediments consist of between 0.2 m and 0.4 m of humiferous eolian sand. Beneath the central part of the basin, a sequence of stratified lacustrine sediments is exposed (Figure 4.6a). These

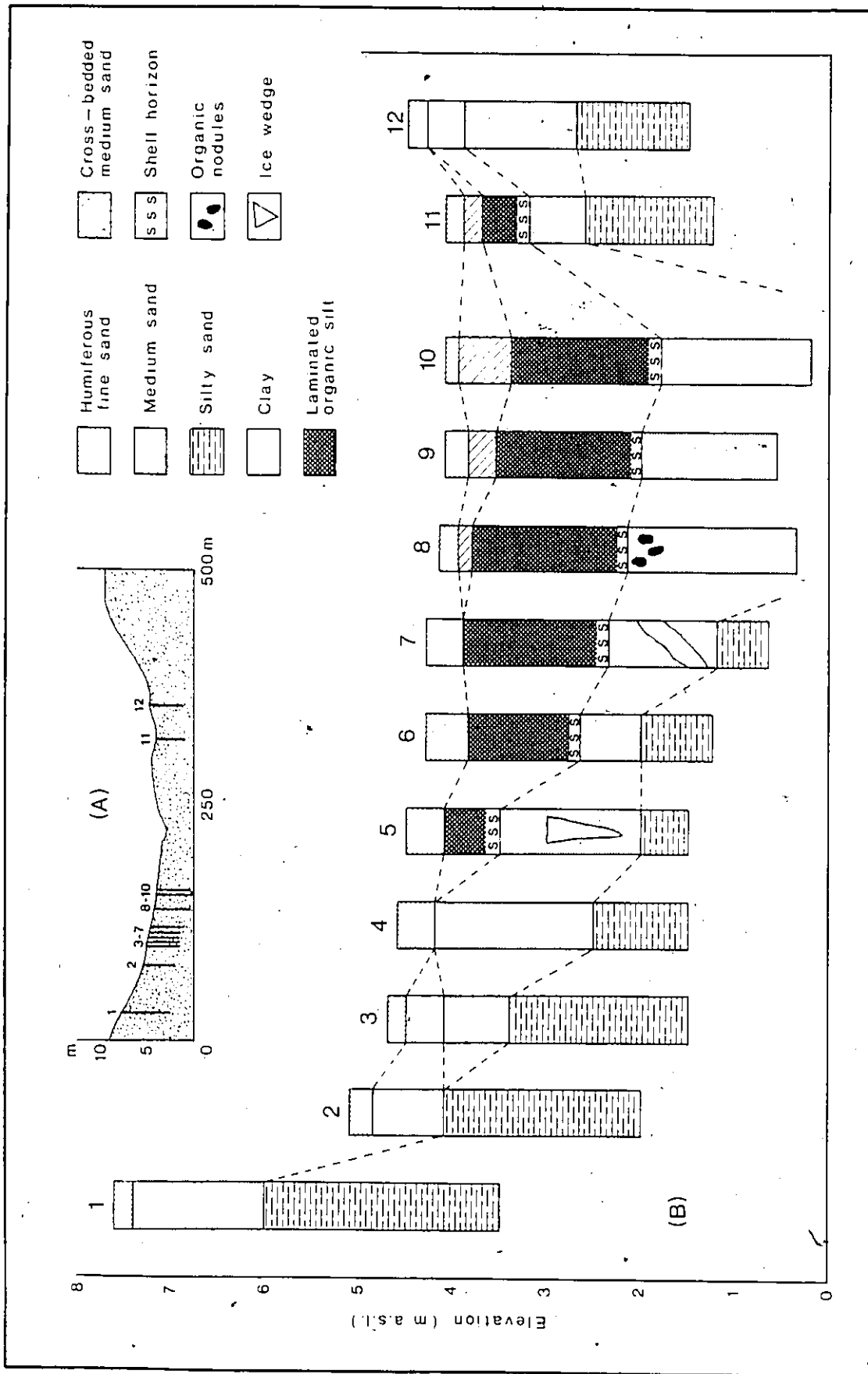
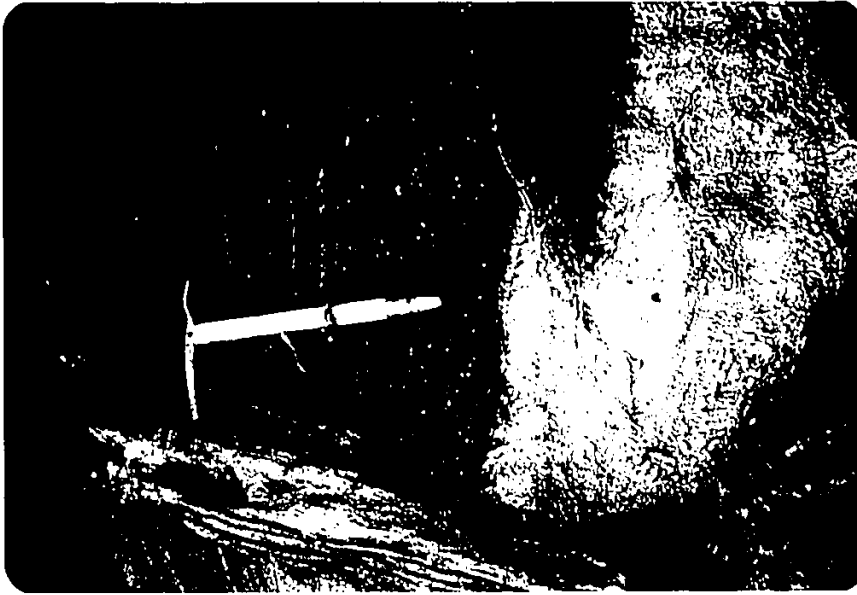
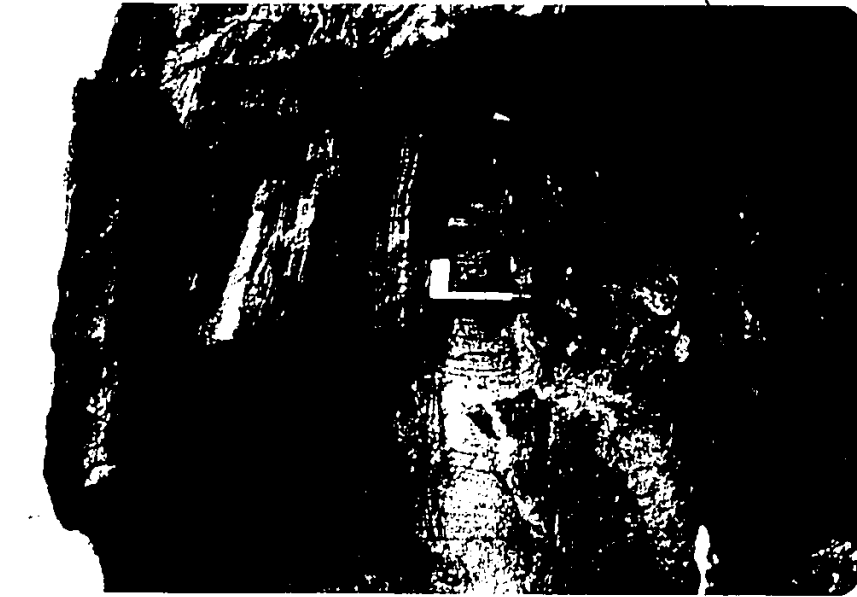


Figure 4.5 Sections illustrating permafrost stratigraphy exposed beneath a drained lake basin in the Sachs River lowlands, 4 km southeast of Sachs Harbour. Surveyed August 20-26, 1980.



(a)



(b)

Figure 4.6 Permafrost stratigraphy exposed beneath a drained lake basin, 4 km southeast of Sachs Harbour, August 20, 1980.

- (a) Section beneath the northwest flank of the basin (see Figure 4.5, section 7) comprises:
1. Grey non-oxidized sand (but note oxidized zone along ice wedge).
 2. Stratified lacustrine sediments, with a basal shell horizon (at axe head).
 3. Humiferous eolian sand, lying unconformably upon lacustrine sediments.
- (b) Detail of basal lacustrine sediments. Note intact shells near base of axe and small ice wedge penetrating underlying sand.

reach a maximum thickness of 1.6 m and consist of laminated organic silts with a basal horizon containing intact gastropod and pelycypod shells (Figure 4.6b). This unit is overlain in places by up to 0.6 m of fine to medium ripple cross-laminated sand. Both units dip inward towards the basin centre and are overlain unconformably by eolian sand. Lenses of blocky blue-grey clay, up to 0.8 m in thickness, flank the laminated organic sediments on each side of the basin. Beneath the lacustrine sediments, the stratigraphy consists of grey non-oxidized sand overlying grey silty sand. These sediments contain 40-50% less excess ice than equivalent materials outside the lake basin.

This lithostratigraphy illustrates several stages of thaw lake development and drainage. A thermokarst origin for this basin is indicated by the vertical compression of the stratigraphy beneath the basin and by the reduced ice content of these sediments. For example, the detrital organic horizon, which occurs normally at an elevation of approximately 6.0 m a.s.l., dips to 0.5 m a.s.l. beneath the basin flank and is probably beneath sea level in the central part of the basin. Since this lake may have been up to 5.0 m in depth and 200 m in radius, it almost certainly possessed a talik (see Figure 4.3). The stratigraphic evidence suggests thaw subsidence of at least 5.0-6.0 m; given the high initial ice content of the silty sand (see Table 4.1), this is certainly feasible.

Beneath the central part of the basin, deep water sedimentation resulted in accumulation of a sequence of laminated organic silt. The clay flanking these sediments may represent marginal mud-flat deposits. The ripple cross-laminated sand, which overlies the basin sediments, indicates an abrupt change to a higher energy environment, characterized by bottom currents and/or wave action. These sediments probably represent a transition to shallow water conditions, during or subsequent to lake drainage. Since this basin was truncated by coastal retreat, it is possible that some of this sediment represents an overwash deposit, transported during storm events into a shallow lagoonal environment. Following emergence, a thin layer of eolian sand was deposited on the lake floor. The unconformable contact between eolian and lacustrine sediments suggests that the latter were subject to a period of deflation prior to the onset of net deposition. This process may be observed at other sites of more recent lake drainage in the Sachs River lowlands.

A number of radiocarbon age determinations relate to thaw lake development on southwest Banks Island (Table 4.4). Organic material collected from the basal horizon of lacustrine sediments in the section described above yields a ^{14}C date of $8,560 \pm 210$ years B.P. (GSC-3292). This provides a maximum age for the initiation of lake sedimentation at this site. At a second locality in the Sachs River lowlands, Vincent (1980, p. 237) obtained a ^{14}C date of

Table 4.4

Radiocarbon Dates Pertaining to Thaw Lake Development,
Southwest Banks Island.

Locality	Laboratory Dating Number	Date (years B.P.)	Stratigraphic Position	Material	Significance
Drained lake basin 4.0 km S-SE of Sachs Harbour	GSC-3292	8,560±210	Peat at base of lacustrine sedi- ment sequence, overlying outwash sand.	<u>Salix</u> sp.	Max. age for initia- tion of lake sedimentation.
Drained lake basin 15.0 km SE of Sachs Harbour	GSC-2246 ¹	8,280±140	Peat below lake sediments, over- lying outwash sand.	<u>Salix</u> sp.	Max. age for initia- tion of lake sedimentation.
Coastal bluffs 4.0 km W of Sachs Harbour	GSC-2364 ²	9,490±80	Slumped blocks of organic material contained within frozen grey silt.	? <u>Salix</u> sp.	Provides a maximum age for talik development.

¹ Collected by J-S. Vincent (see Vincent, 1980, p. 237).² Collected by H. M. French (H. M. French, personal communication).

8,280±140 years B.P. (GSC-2246) from peat underlying lacustrine sediments within a drained lake basin. In 1974, a coastal exposure approximately 4.0 km west of Sachs Harbour allowed examination of frozen silt beneath a drained lake basin (H. M. French, personal communication). The silt contained blocks of fibrous organic material, which yielded a ^{14}C date of 9,490±80 years B.P. (GSC-2364). This may indicate a maximum age of talik development at this locality.

It seems likely, therefore, that many thaw basins located in the lowlands west and southeast of Sachs Harbour were initiated approximately 8,000-9,000 years B.P. This period corresponds to an early phase of the mid-Holocene climatic optimum, during which warmer air temperatures may have triggered the melt-out of buried ice. This conclusion is consistent with that of Rampton and Bouchard (1975; Rampton, 1974), with respect to the origin of thaw lakes on the Tuktoyaktuk Peninsula.

By contrast, many thaw lakes within the Kellett valley are developed upon river terraces which probably post-date the climatic optimum. It follows, therefore, that thaw lake development in this area must have occurred during a more recent period of cold climate conditions. At that time a network of ice-wedge polygons may have formed on the terrace surface, facilitating the ponding of water bodies and initiation of self-sustaining thermokarst. It is significant in this respect that the majority of small lakes in this area are located wholly within low-centred

polygons. A similar mechanism has been invoked to explain the formation of many small thaw lakes in the Mackenzie Delta area (Mackay, 1979, p. 31), while Gravis (1978) also suggests that periods of thaw lake development in the northern U.S.S.R. may be correlated with cold climate conditions.

4.4 Lake morphology

4.4.1 Distribution and size

On southwest Banks Island, high concentrations of lakes occur in two major areas (Figure 4.7). To the north of Sachs Harbour, lake cover exceeds 20% in some parts of the Kellett River valley, with highest concentrations occurring on low terraces above the present floodplain (Figure 4.8). Numerous small ponds occupy low-centred polygon sites, while larger lakes have formed by basin coalescence following breaching of the polygon ramparts. Southeast of Sachs Harbour, the Sachs River lowlands are characterized by a number of large lakes, greater than 1.0 km in diameter, together with numerous smaller tundra ponds. Most of this area has a lake cover in excess of 20%, rising in some places to over 50%. While both areas contain a similar proportion of small lakes, the Sachs River lowlands also contain a small percentage of lakes which are a full order of magnitude larger than those occurring within the Kellett River valley (Figure 4.9).

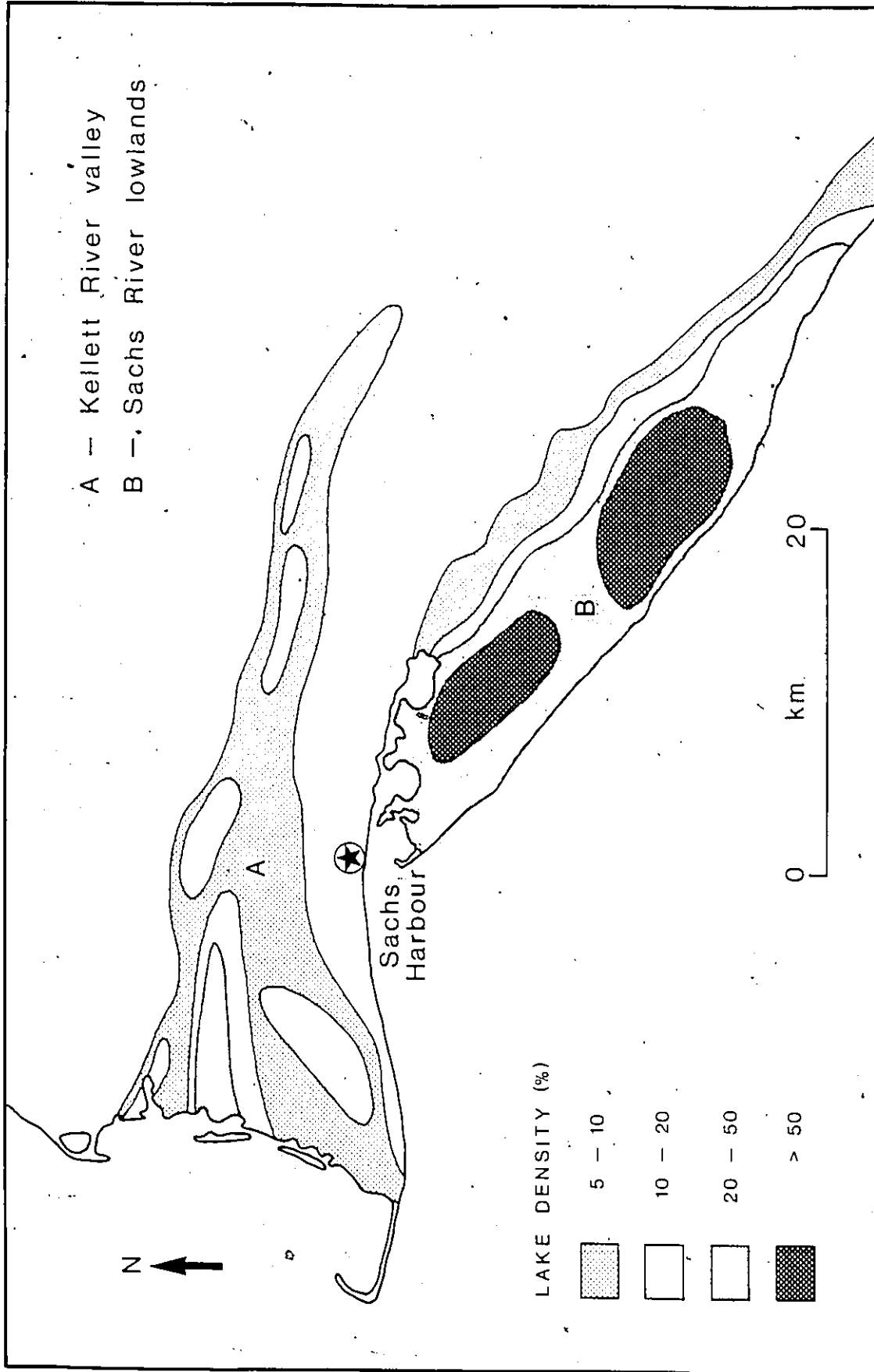


Figure 4.7 Lake density, southwest Banks Island. Note: Density is expressed as percentage lake cover per 25 km².



Figure 4.8 Air photograph illustrating typical thaw lake terrain in the Kellett River valley, 10 km northwest of Sachs Harbour (Part of A 15980-72). Note: (1) Occurrence of D-shaped lakes (shoreline characteristics of basin 'A' are described in the text). (2) Frequency of basins which have coalesced along a NE-SW axis. (3) Linear zone of drained basins related to an ephemeral stream channel.

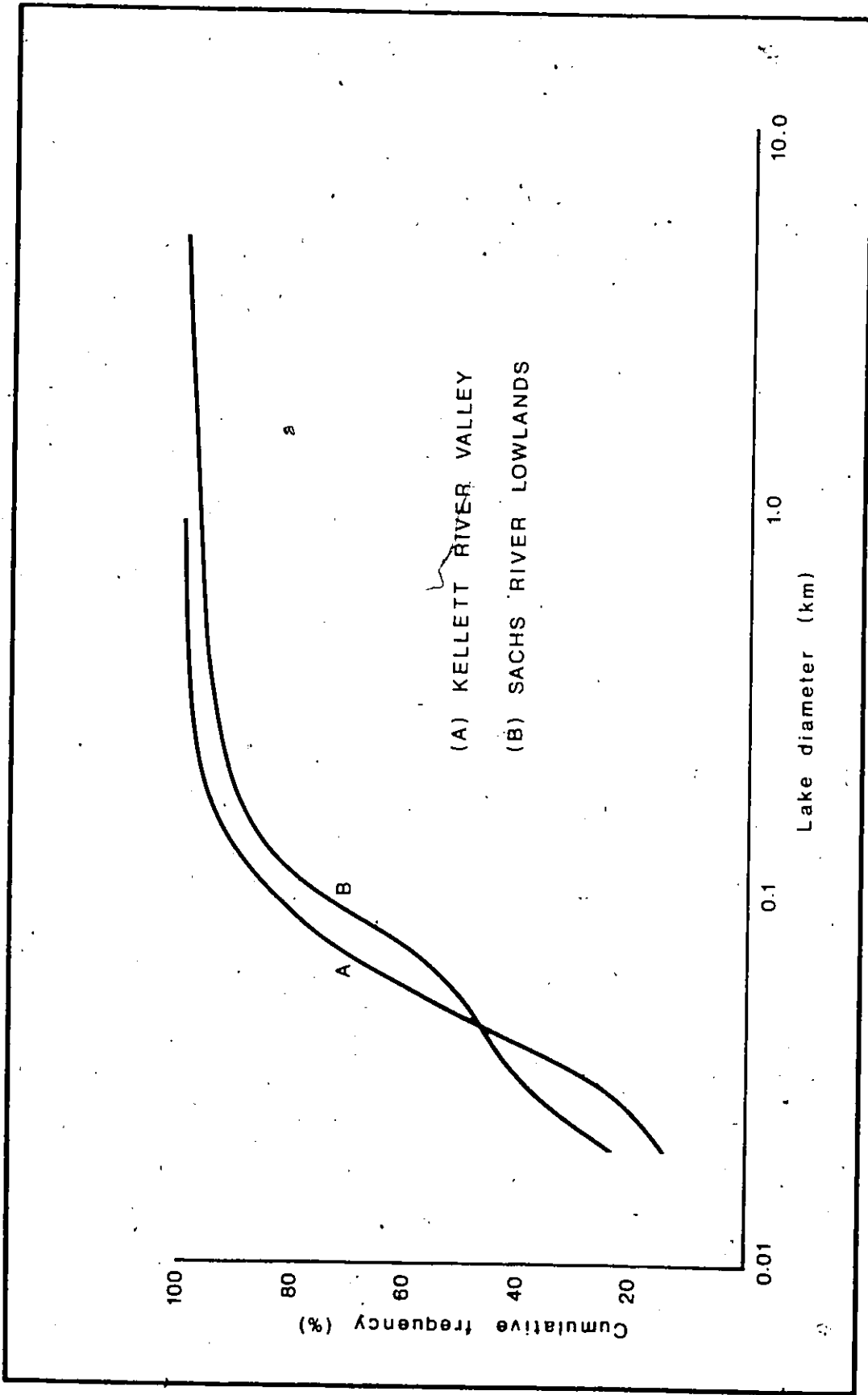


Figure 4.9 Lake size distribution, southwest Banks Island.
Note: Minimum diameter discriminated is 20 m.

Few comparative data exist with respect to thaw lake morphometry. This is surprising in view of the widespread application of morphometric analysis to the thaw lake problem. Sellmann et al. (1975) present histograms illustrating lake dimensions within a number of terrain subdivisions of the Alaskan coastal plain, and it is interesting to compare these to the Banks Island data. As these authors utilized satellite imagery in their analysis, the minimum lake diameter discriminated is correspondingly large, 500 m. On southwest Banks Island, only 55 lakes exceed this diameter, all located within the Sachs River lowlands. Comparison suggests that, first, the Banks Island lakes are close to the small (L3) Alaskan lakes in size and, second, the range of lake size is much greater on Banks Island (Figure 4.10).

4.4.2 Equilibrium morphology

While the initiation of thaw basins is a function of subsurface thermal and ground ice conditions, the subsequent development of basin morphology is controlled by geomorphic processes. In the absence of pronounced topographic control, basin morphology may be regarded as representing a quasi-equilibrium geomorphic condition.

Many thaw lakes possess an equilibrium shape which, in its simplest form, is expressed as a preferred long-axis orientation (e.g. Black and Barksdale, 1949; Carson and Hussey, 1962; Mackay, 1956). Random sampling indicates that 84% of lakes in the Sachs River lowlands and 97% of

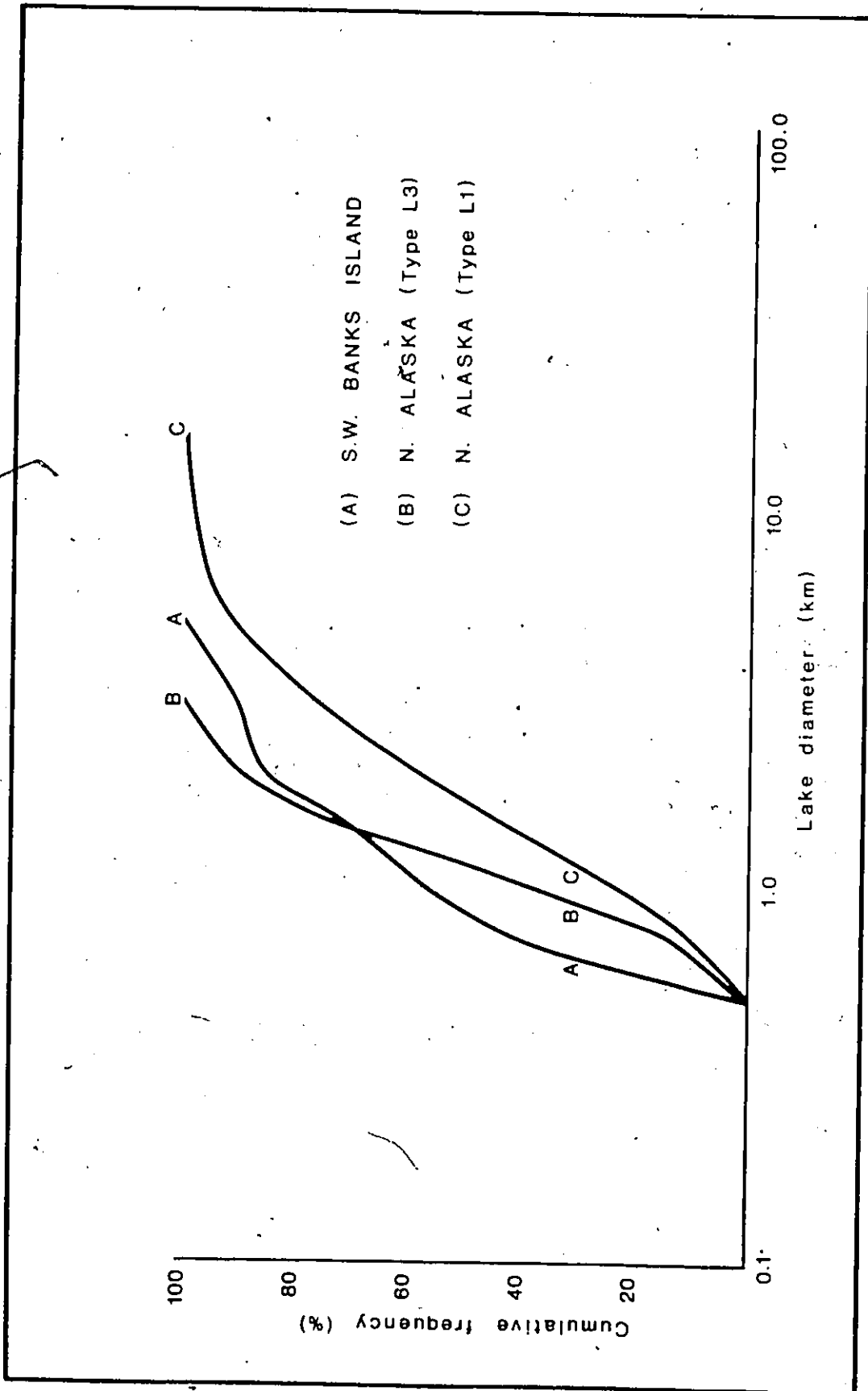


Figure 4.10 A comparison of lake size on southwest Banks Island and in northern Alaska. (Alaskan data after Sellmann et al., 1975). Note: Minimum diameter discriminated is 500 m.

lakes in the Kellett River valley have a measurable long-axis orientation (defined by a length-width ratio greater than 1.1). Mean lake elongation is 1.53 in the Sachs River lowlands as compared to 1.92 in the Kellett River valley (Figure 4.11). This disparity is partly accounted for by the prevalence of basins in the latter area which have coalesced in the direction of elongation. Lake orientation patterns are similar in the two areas, with a major northeast-southwest component and a minor perpendicular component (Figure 4.12). The Kellett River valley lakes display a very low degree of variance in orientation, with approximately 60% of long axes possessing azimuths between 230° and 240° .

Detailed morphological analysis requires the definition of shape categories. Since some lakes have a simple shape, while others possess intermediate characteristics, a two-level shape classification was devised to describe lakes on southwest Banks Island. First, a number of primary morphologies were identified from air photographs. These were defined as circle, oval, ellipse and rectangle, based in part on shape categories used by Mackay (1963, pp. 47-50) to describe lakes in the Mackenzie Delta area. A D-shaped class was added to describe highly asymmetrical elliptical lakes and a 'complex' category was established to include irregular or multiple basins. Second, where lake shape is intermediate between two categories, the next closest class was used to define a secondary morphology. Thus, for example, a circular lake is defined as

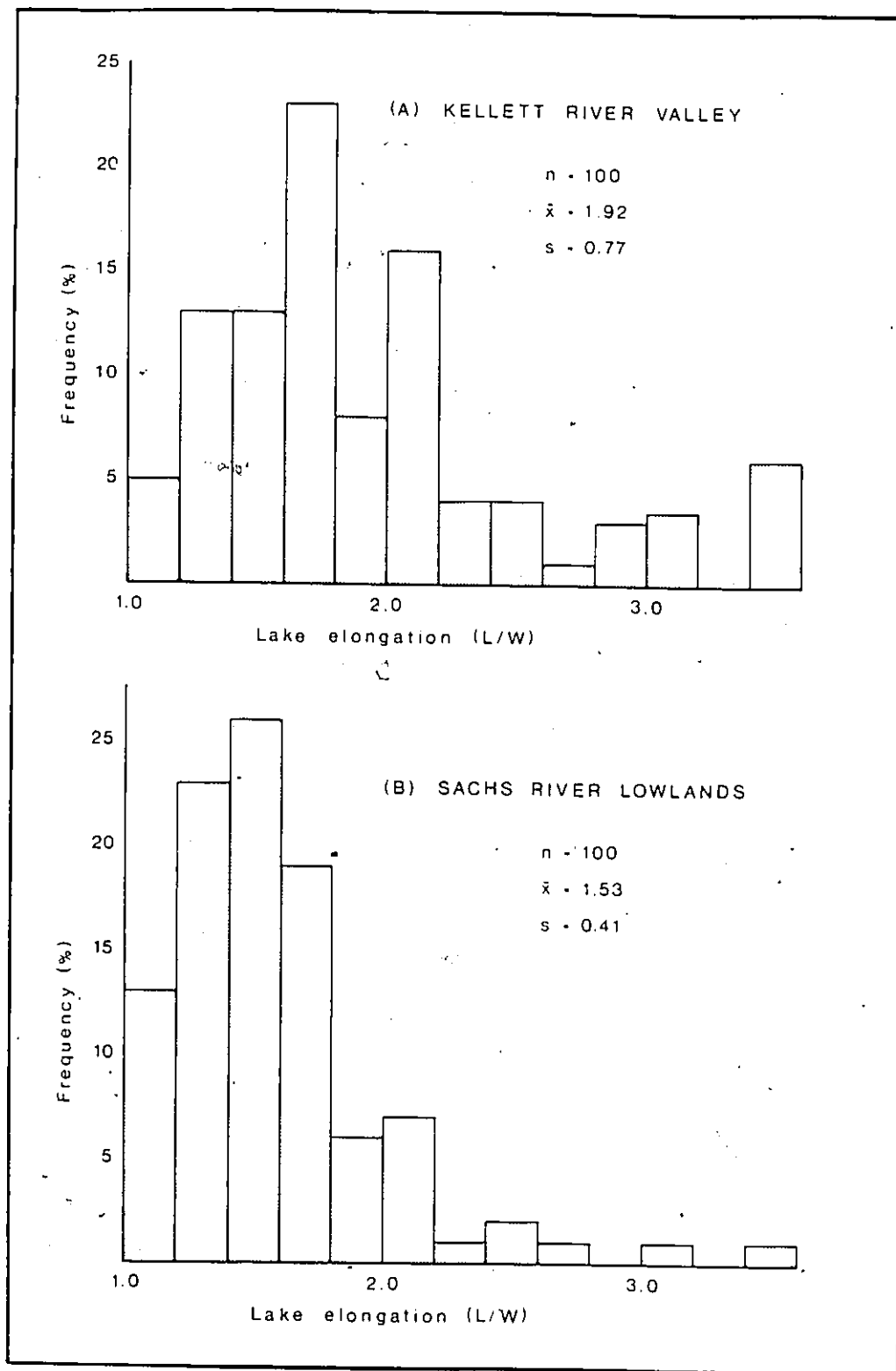


Figure 4.11 Thaw lake elongation, southwest Banks Island.

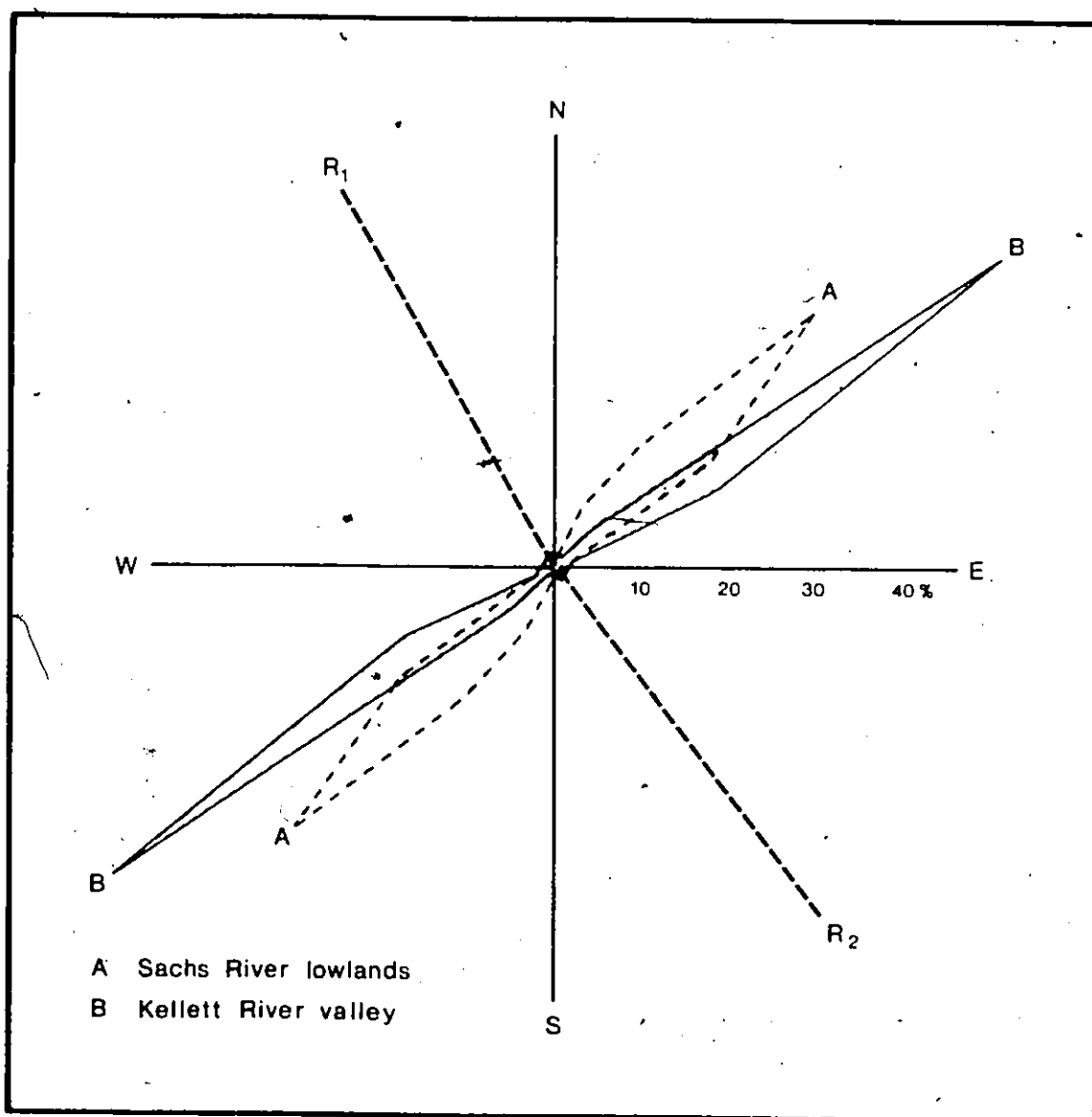


Figure 4.12 Thaw lake orientation, southwest Banks Island.
Note: R_1 and R_2 are opposed vector resultants of the summer storm wind regime at Sachs Harbour.

circular/simple, while an equidimensional sub-angular lake may be classified as circular/rectangular. This procedure establishes a matrix of 36 shape categories. Analysis of random samples of 100 lakes suggests that, in the Sachs River lowlands, complex, elliptical or D-shaped morphological types are most common (Table 4.5a). In the Kellett River valley, primary morphology is predominantly complex, reflecting the high frequency of coalescent basins. However, over 50% of lakes possess a D-shaped element (Table 4.5b).

Under equilibrium conditions, it is probable that a lake basin will develop an outline consisting of a number of smoothly curved bays (Mackay, 1963a, p. 52). This form should be widespread and developed upon a range of surface materials. On southwest Banks Island, lakes of the D-shape category most consistently satisfy these criteria. The D-shape is, in fact, curvilinear; the 'straight' segment forms approximately half of an extremely flattened ellipse, while the opposing 'bow' segment forms a broad semi-ellipse or semicircle. Non-dimensional analysis of D-shaped lakes of varying size in the Kellett valley indicates only slight variation in shape (Figure 4.13a). It is suggested, therefore, that this shape represents the equilibrium form of thaw lakes on southwest Banks Island.

The development of equilibrium morphology and orientation is not restricted to arctic thaw lakes (see e.g. Kaczorowski, 1977; Price, 1968), and it follows that the presence of permafrost serves only to modify the action

Table 4.5
Lake Shape Classification, Southwest Banks Island.

Primary Morphology	Simple	D-Shape	Ellipse	Secondary Morphology			Circle	Complex	Total
				Rectangle	Oval	Oval			
<u>(a) Kellett Valley</u>									
D-Shape	16	-	1	1	-	-	-	34	52
Ellipse	-	-	-	1	-	-	-	3	4
Rectangle	3	1	-	-	-	-	-	14	18
Oval	3	-	1	1	-	-	-	1	6
Circle	1	-	-	-	-	-	-	1	2
Complex	18	-	-	-	-	-	-	-	18
Total	41	1	2	3	0	0	0	53	100
<u>(b) Sachs River Lowlands</u>									
D-Shape	15	-	3	1	1	1	1	1	22
Ellipse	16	6	-	4	-	-	-	4	30
Rectangle	7	2	-	-	1	-	-	2	12
Oval	9	3	-	-	-	1	-	-	13
Circle	3	1	1	-	-	-	-	1	6
Complex	17	-	-	-	-	-	-	-	17
Total	67	12	4	5	2	2	2	8	100

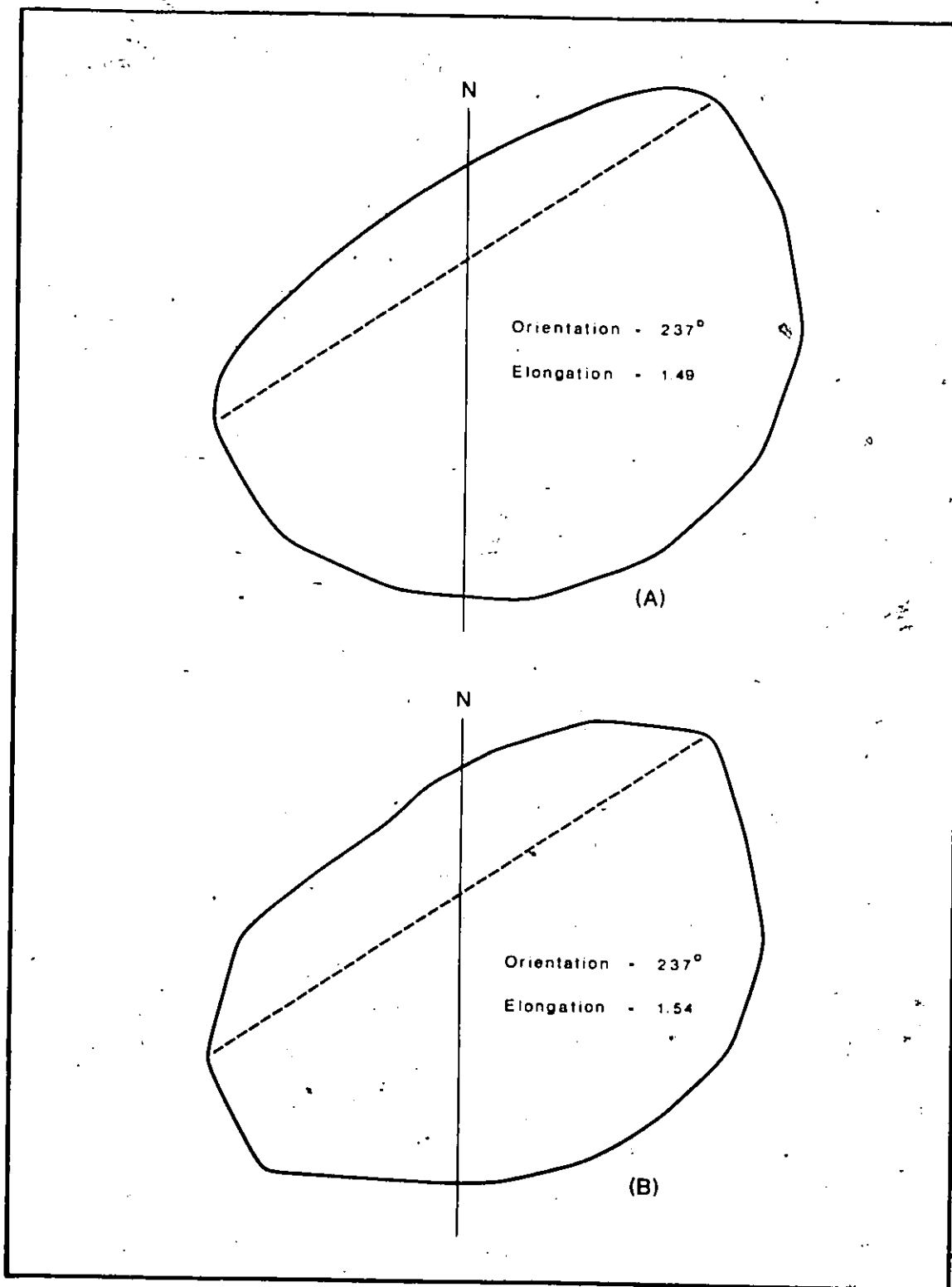


Figure 4.13 Equilibrium thaw lake morphology, southwest Banks Island.
(a) Observed morphology, based on analysis of five D-shaped lakes in the Kellett River valley.
(b) Predicted morphology, based on calculation of equilibrium bays related to the summer storm wind regime at Sachs Harbour.

of essentially azonal processes. A number of possible orienting mechanisms have been suggested which fall into two broad categories, related either to the geologic setting of lakes or to surface and atmospheric processes operating upon them. Bedrock structure is believed to control the shape of some oriented lakes (Plafker, 1964; Short and Wright, 1974); however, this factor may be discounted on southwest Banks Island where lakes are developed upon a thick sequence of unconsolidated sediments. Thaw lakes which are initiated as ice wedge junction ponds, or tundra ponds located within low-centred polygons, may possess an orientation related to a former ice wedge network (Carson and Hussey, 1959; Tolstov, 1966). However, it is likely that this orientation would be substantially modified during basin expansion and coalescence.

In all probability, therefore, the preferred orientation of thaw lakes on southwest Banks Island results from the operation of geomorphic processes. This is true, for example, of thaw lakes in northern Alaska and the Mackenzie Delta area, where the majority of lake basins are oriented perpendicular to prevailing wind direction (e.g. Black and Barksdale, 1949; Carson and Hussey, 1962; Mackay, 1956a). It has been shown that this relationship is a function of wind-generated lake circulation (Livingstone, 1954; Rex, 1961). On southwest Banks Island, the mean long-axis orientation of thaw lakes bisects the vector resultants of the opposed summer wind regime (see Figure 4.12). In

particular, the long axes of D-shaped lakes, which extend parallel to the 'straight' shoreline segment, have an orientation almost exactly perpendicular to the northwest-southeast wind resultants.

4.4.3 A process-response model of lake morphology

Since lake orientation appears to be systematically related to wind direction, an understanding of equilibrium morphology may be based on models of wind-generated wave and current distributions. It has been shown that the equilibrium form of a shoreline developed in unconsolidated sediment is a cycloid, within which erosion is greatest in zones oriented at 50° to wave approach (Bruun, 1953). Mackay (1963a, pp. 53-54) proposed a simple process-response model of lake morphology, based on the assumption that winds from each compass point tend to develop cycloidal bays. In this way, equilibrium morphology may be regarded as the integrated form of a number of cycloids of different size. Assuming that the diameter of the cycloid generating circle is proportional to the wind resultant, R , in that direction, it follows that the vector lengths of the cycloid at angles of 22.5° , 45.0° , 67.5° and 90.0° are approximately $1.04R$, $1.16R$, $1.36R$, and $1.56R$ respectively. Lake radius in any direction may thus be calculated by summation of nine individual vector components.

This model was tested with respect to thaw lake morphology on southwest Banks Island, using seasonal wind

data from Sachs Harbour (see Table 2.2). Predicted lake morphology is elliptical, with a length-width ratio of 1.23 and a long-axis azimuth of approximately 247° . Although this provides a good first approximation to observed lake morphology, a much closer correspondence is obtained if only storm wind data are used. In this case, the model predicts a D-shaped lake, with a length-width ratio of 1.54 and a long-axis azimuth of approximately 237° (Figure 4.13b). The significantly better fit suggests that storm events play a major role in shoreline evolution.

Field observations made at a typical D-shaped lake in the Kellett River valley ($72^\circ 04'N$, $125^\circ 33'W$) tend to support this model. Shoreline characteristics are closely related to the inferred pattern of wind-generated littoral drift (Figure 4.14). At the northeast and southwest corners of the lake, and along the southeastern 'bow' segment, shoreline morphology is predominantly erosional. The 10-50 cm high shoreline bluff is undercut and terrace gravels are exposed in the lake bottom (Figure 4.15a). By contrast, the 'straight' shoreline is characterized by a broad, low angle depositional flat (Figure 4.15b). According to Rex's (1961) model of lake circulation, this is expected to be a zone of minimum littoral drift under prevailing wind conditions.

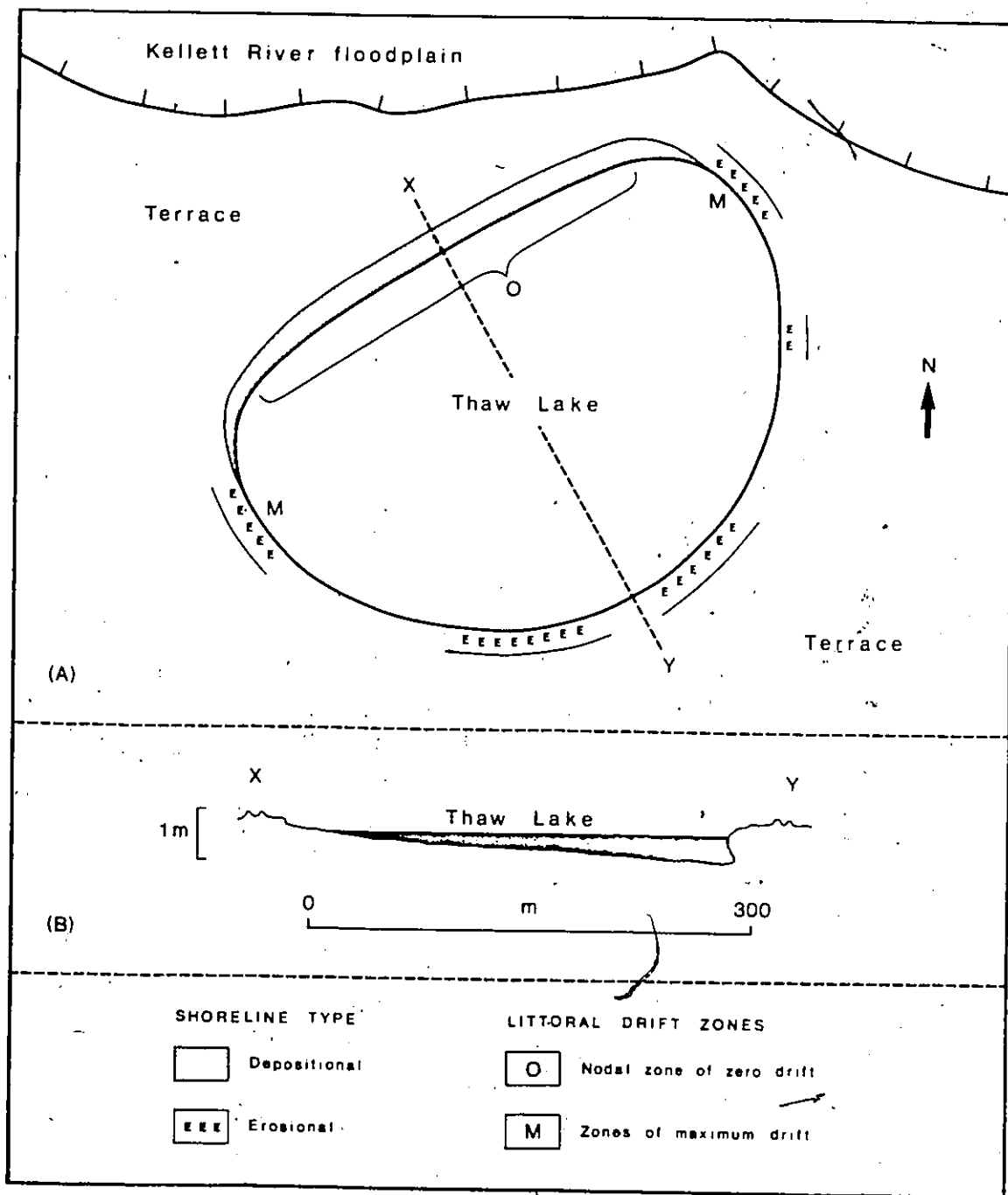


Figure 4.14 Geomorphology of a D-shaped lake, Kellett River valley. Note: The location of this lake is indicated on Figure 4.8.



(a)



(b)

Figure 4.15

Shoreline characteristics of a D-shaped lake, Kellett River valley. August 04, 1980.

(a) Predominantly erosional 'bow' shoreline segment. Note the undercut bluff and terrace gravel exposed in lake bottom.

(b) Predominantly depositional 'straight' shoreline segment. Note the broad mudflat colonized by Eriophorum sp.

4.5 Lake drainage

Within areas of thaw lake terrain on southwest Banks Island, many basins are either partly or completely drained. The mechanism of thaw lake drainage therefore represents an important aspect of the more general problem of thaw lake evolution. Two models of lake drainage have been proposed. The first consists of slow infilling, segmentation and revegetation (e.g. Kaczorowski, 1977, p. 111), while the second consists of catastrophic outflow, following lake tapping or truncation by coastal retreat (e.g. Mackay, 1979, p. 31; Walker, 1979; Weller and Derksen, 1979). On southwest Banks Island, evidence points strongly to the importance of the latter model.

4.5.1 Lake tapping

Catastrophic lake drainage may result from tapping by headward erosion of streams or by coalescence with a basin at a lower elevation. Once flow is initiated through the new outlet, drainage may be extremely rapid. Although no measurements have been made of a lake draining under natural conditions, discharge rates of $10 - 20 \text{ m}^3 \cdot \text{s}^{-1}$ were recorded during the artificial drainage of Illisarvik Lake in the Mackenzie Delta (Mackay, 1981). This results in the formation of box-canyon outlet channels, characterized by steep bluffs and flat floors, which may be incised into the former lake bed. Following lake drainage, these channels may be either dry or occupied by misfit streams.

In the Sachs River lowlands, a number of lakes appear to have been drained by this mechanism. For example, to the west of Fish Lake, approximately 30 km southeast of Sachs Harbour, a series of small lakes have drained by a combination of stream tapping and basin coalescence (Figure 4.16). The lake adjacent to Fish Lake is particularly interesting, since it appears to have been affected by two distinct drainage events, indicated by the occurrence of box-canyon drainage outlets. One channel, which is deeply incised into the former lake floor (Figure 4.17a), may be associated with basin capture by Fish Lake, the level of which is at present approximately 20 m beneath the drained lake floor. A second, shallower drainage outlet is located at the southeast end of the basin (Figure 4.17b). This channel links the basin to a second drained lake and forms part of an integrated drainage network, discharging into Middle Lake 2 km to the northwest. The sequence of drainage events at this locality is not known.

4.5.2 Truncation by coastal retreat

Truncation of thaw lake basins by coastal retreat also results in catastrophic lake drainage, accompanied in some cases by marine inundation of all or part of the basin. For example, in the Sachs River lowlands immediately southeast of Sachs Harbour, a complex sequence of drainage events is evidenced by the distribution of drained or inundated lake basins, former lake bluffs and box canyon drainage outlets (Figure 4.18). Analysis of air photographs

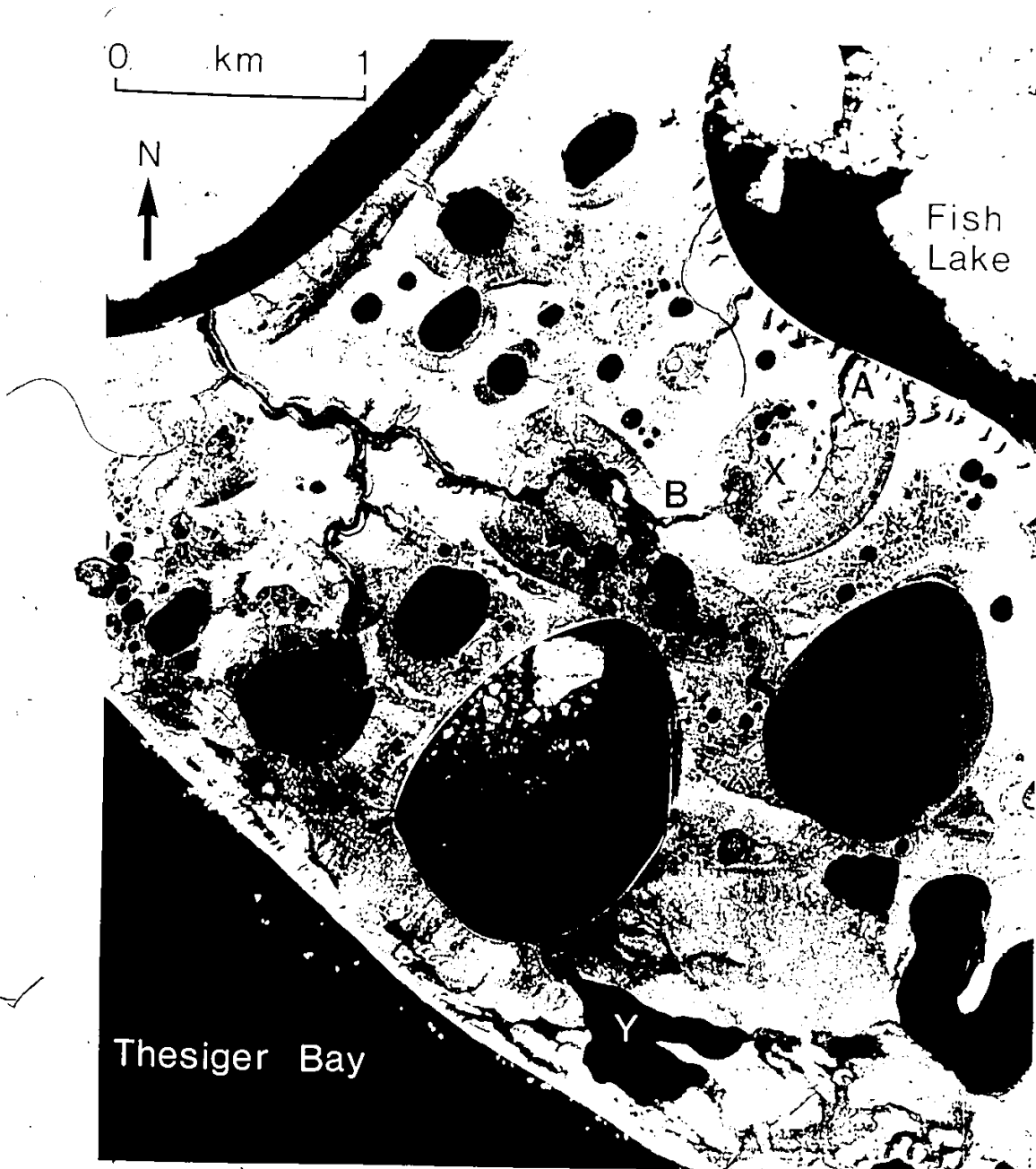


Figure 4.16

Air photograph illustrating lake drainage by basin capture and stream tapping, Sachs River lowlands (Part of A 15980-25). Note northern 'A' and western 'B' outlets from basin 'X', described in the text. Lake 'Y' drained catastrophically in January 1977 (see pp. 155-157)..

(a)



(b)



Figure 4.17 Outlet channels associated with the drainage of a small basin near Fish Lake, Sachs River lowlands. July 22, 1979.

- (a) Drainage channel eroded into the floor of the basin, probably as a result of capture by Fish Lake (in distance). Note figures for scale.
- (b) Shallow box-canyon located at the southeast end of the basin. This forms part of an integrated drainage network, discharging into Middle Lake.

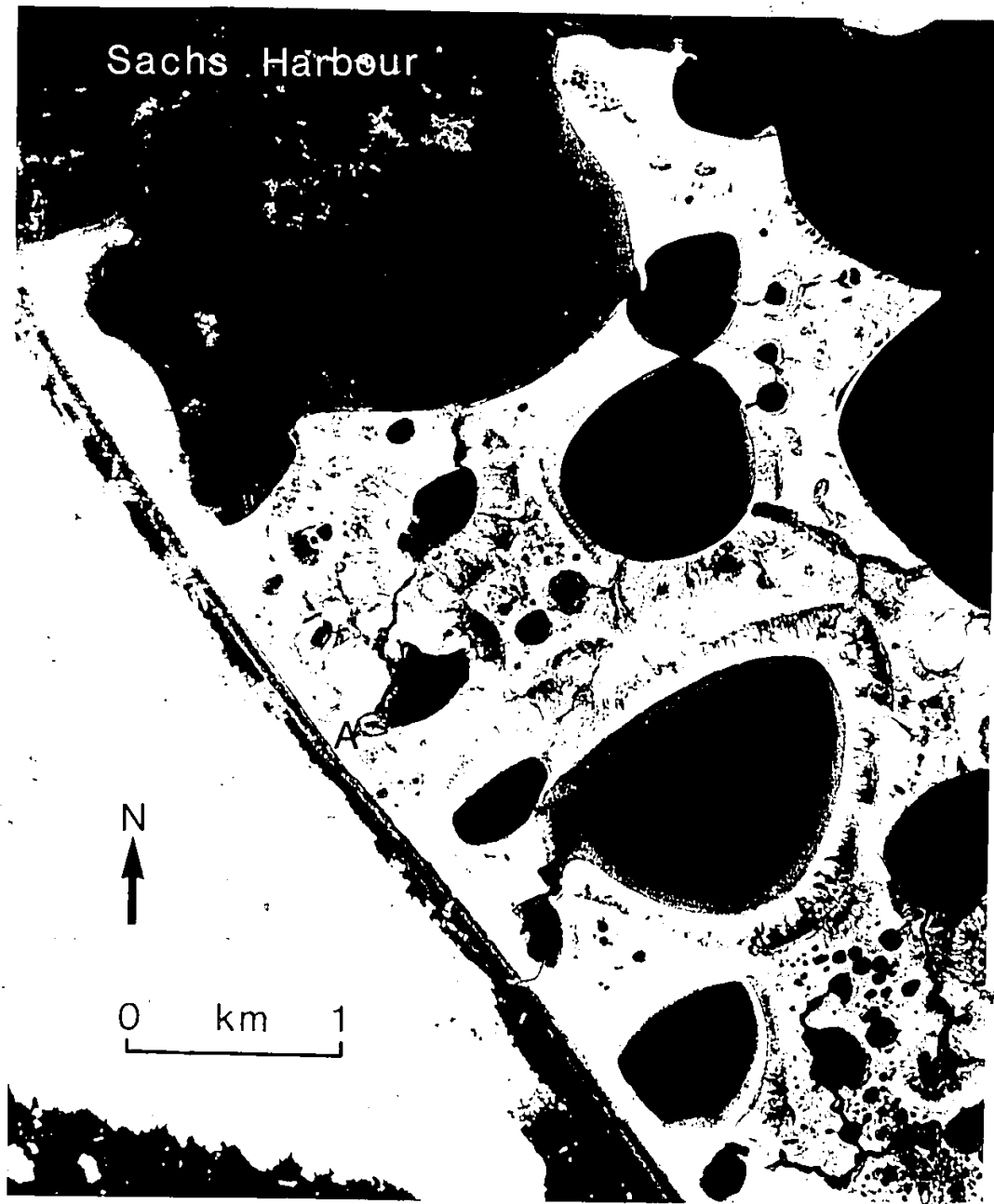


Figure 4.18 Air photograph illustrating lake drainage by coastal truncation and basin capture, Sachs River lowlands (Part of A 15980-65). Note: Permafrost stratigraphy exposed beneath truncated basin 'A' is described in the text (see section 4.3).

indicates that at least fifteen lake basins greater than 20 m in diameter are at present truncated by the coastline of Thesiger Bay.

Few basins located within 50 m of the coastline contain lakes and it is likely that drainage occurs prior to coastal breaching, through melt-out of ice bodies (Figure 4.19a). For example, two small lake basins, located approximately 5 km west of Sachs Harbour, have drained through gullies formed by erosion along ice-wedge troughs. Analysis of sequential air photo coverage indicates that the western basin drained prior to 1950, at which time the lake margin was at least 45 m from the coast. Ground survey in 1979 showed that during drainage the ice-wedge trough was enlarged into a thermokarst ravine, 5.0 m in width and 2.0 m in depth (Figure 4.19b). The eastern basin was drained at some time between 1972 and 1979, when the shoreline was 20-30 m distant. This lake also discharged along the line of an ice-wedge trough.

4.5.3 Lake drainage and ice-cored terrain

Within areas of ice-cored terrain, other mechanisms of lake drainage may exist, related to the distribution of ice bodies. On southwest Banks Island, extensive areas within the limit of the Sand Hills Readvance may be underlain by buried glacier ice (see pp. 97-99). Under certain circumstances, this may provide a subsurface outlet for lake drainage. For example, one lake located approximately 40 km southeast of Sachs Harbour (71°48'N, 124°35'W), was

(a)



(b)

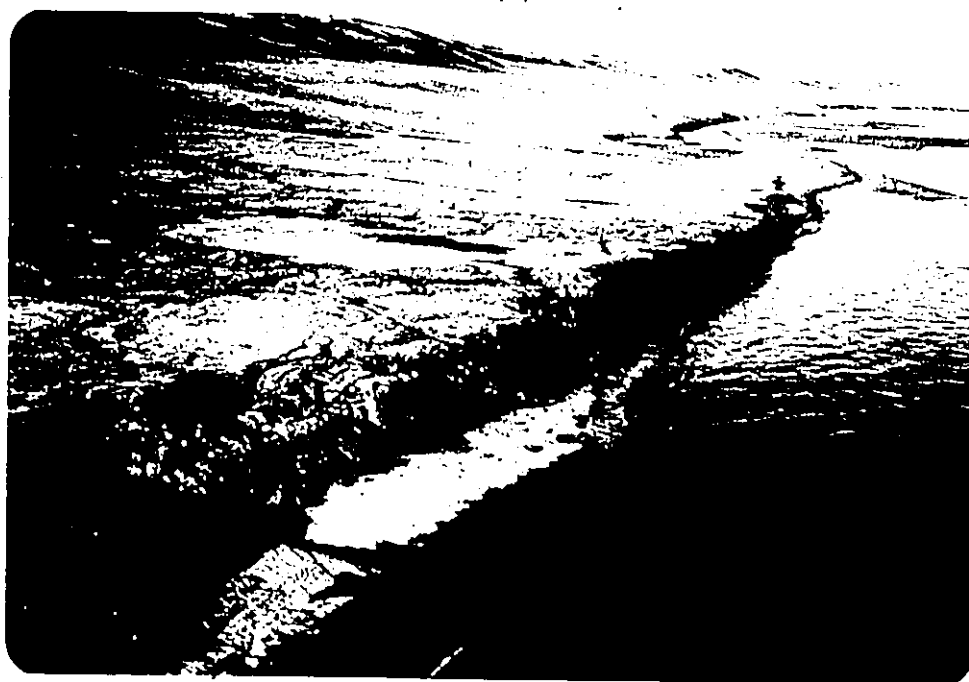


Figure 4.19 Examples of drainage by coastal truncation, August 06, 1980.

- (a) Drained basin truncated by sea cliffs, approximately 30 km southeast of Sachs Harbour. Note the deep gully eroded during lake drainage.
- (b) Small drained lake basins, located approximately 4 km west of Sachs Harbour. Note the rectilinear outlet channels developed along ice wedge troughs.

observed by an Inuit during a period of catastrophic drainage in early-January 1977 (David Nasogaluak, personal communication). At least partial lake drainage was indicated by collapse of the ice cover. Although this was accompanied by considerable noise, no surface flow was observed. The lake was visited by D. G. Harry and H. M. French in July 1979, at which time the basin was partly filled (Figure 4.20a). The minimum volume of water lost, given by the product of basin area and observed drop in lake level, is $6.6 \times 10^5 \text{ m}^3$. Although the lake is located within 100 m of the coast, no evidence of previous surface flow was visible. Instead, at the southeast corner of the lake, evidence of recent collapse (Figure 4.20b) suggests the possibility of subterranean drainage. While this may be common in areas of thin or discontinuous permafrost, resulting in the formation of thaw sinks (e.g. Hopkins, 1949; Tedrow, 1969), it is rarely reported from regions underlain by thick, continuous permafrost. It is thus hypothesized that outflow from the lake was caused by thermal contraction cracking, allowing escape of water into an underlying cavity within buried glacier ice.

4.6 Summary

A number of researchers have advanced the concept of a thaw lake cycle (e.g. Billings and Peterson, 1980; Britton, 1966; Everett, 1981; Tedrow, 1969), consisting of sequential stages of initiation, expansion and drainage.



(a)



(b)

Figure 4.20 Partially drained lake within the Sand Hills area of ice-cored terrain. July 22, 1979.

(a) Lake shoreline photographed July 22, 1979. Note minimum 3.7 m drop in lake level.

(b) Evidence of recent collapse at the southeast corner of the lake. During catastrophic outflow in January 1977, water may have drained into a cavity within buried glacier ice. (See Figure 4.16 for vertical air view of this area in July, 1958).

However, observations made on southwest Banks' Island suggest that such models are by no means universally applicable. It seems, rather, that a number of alternative models of thaw lake development are possible, in response to a delicate equilibrium between process, materials and morphology.

On southwest Banks Island, thaw lakes appear to have originated by melt-out of ground ice during a mid-Holocene period of climatic amelioration, or by ponding within ice-wedge polygons developed during subsequent cold climate conditions. It is also possible that some of the large, deep basins within the Sachs River lowlands originated by melt-out of buried glacier ice. Once a lake basin forms, the thermokarst process may become self-sustaining, since each increment in basin size increases the geothermal disturbance associated with the lake. The maximum size of basins is thus limited only by the potential thaw subsidence of underlying sediments.

In the absence of topographic control, lake expansion is strongly influenced by wind-related patterns of wave and current erosion, resulting in the development of a preferred long-axis orientation. The sensitivity of this process is demonstrated by the relative rarity of oriented lakes elsewhere on Banks Island. For example, thaw lakes developed upon terraces within the Big River valley, only 50 km north of Sachs Harbour, show no evidence of preferred orientation (see Figure 4.1c). It would appear that the development of preferred orientation is dependent on the

existence of an opposed summer wind regime, unmodified by local topographic effects. For example, within the Kellett valley, a marked eastward reduction in the occurrence of preferred orientation occurs, presumably as a result of the progressive intrenchment of the valley and consequent disturbance of the wind regime.

It has been suggested that the orienting mechanism may also result in migration of thaw lakes across the tundra surface (e.g. Sukhodrovskii, 1960; Tedrow, 1969; Tomirdiaro and Ryabchun, 1978). No evidence of this phenomenon was found on southwest Banks Island. Wind-generated patterns of shoreline erosion have resulted in asymmetrical expansion to form D-shaped lakes, rather than causing an overall translocation of the basin. Thaw lakes in this area appear, therefore, to represent quasi-equilibrium landscape elements.

CHAPTER FIVE

COASTAL CHANGE IN ICE-RICH

PERMAFROST TERRAIN

5.1 Introduction

Much of the coastline of southwest Banks Island is developed in ice-rich unconsolidated deposits of Tertiary or Quaternary age. Studies elsewhere have shown that thaw degradation of ice-rich permafrost cliffs, subject to wave attack during the open water season, results in rapid coastal retreat (Are, 1972; 1978; Grigor'yev, 1976, pp. 75-94; Harper, 1978; Hume et al., 1972; Lewellen, 1970; MacCarthy, 1953; Mackay, 1963b; 1972a, p. 9). At the same time, mobilization of sediment provides a source for the development of constructional shoreline features, for example spits and bars (e.g. Lewis and Forbes, 1974; McDonald and Lewis, 1973).

Several reconnaissance studies of coastal conditions have been undertaken on Banks Island (Manning, 1954; Miles, 1976; 1977; Stephen, 1976), however the coastline adjacent to Sachs Harbour has not yet been described in detail. Although this coastal zone is classified as primarily wave-erosional (Beak Consultants Ltd., 1978), areas exist where depositional processes clearly dominate, partly as a result of the shelter afforded by the Sachs River estuary. In other areas, however, the occurrence of truncated thaw lakes and absence of cliff-foot debris accumulation, provide evidence of active coastal erosion.

This chapter describes coastal morphology in the vicinity of Sachs Harbour, with particular respect to the influence of permafrost conditions. The objectives are, first, to determine the nature and rate of coastal erosion in an area of ice-rich permafrost and, second, to document the evolution of depositional coastal landforms, which result from the release of ice-cemented cliff sediments. Detailed investigations were restricted to the coastline between Duck Hawk Bluff and the Sachs River lowlands. However, a number of visits were made to Cape Kellett in order to monitor high-magnitude processes, which attain their maximum level on this exposed coastline.

5.2 Coastal erosion

5.2.1 Cliff materials and profiles

Sea cliffs in the vicinity of Sachs Harbour are developed in poorly consolidated and frequently ice-rich permafrost sediments. Except at Duck Hawk Bluff, where Tertiary age sediments outcrop, the cliffs are rarely more than 6.0 m to 8.0 m in height. To the southeast of Sachs Harbour, cliffs are developed in unconsolidated fine-medium sand, the rapid erosion of which provides a major source for beach and spit nourishment. Coastal profiles, measured in August 1979, indicate a close relationship between cliff and beach morphology (Figure 5.1). For example, cliffs behind Allen Creek spit are progressively less degraded in

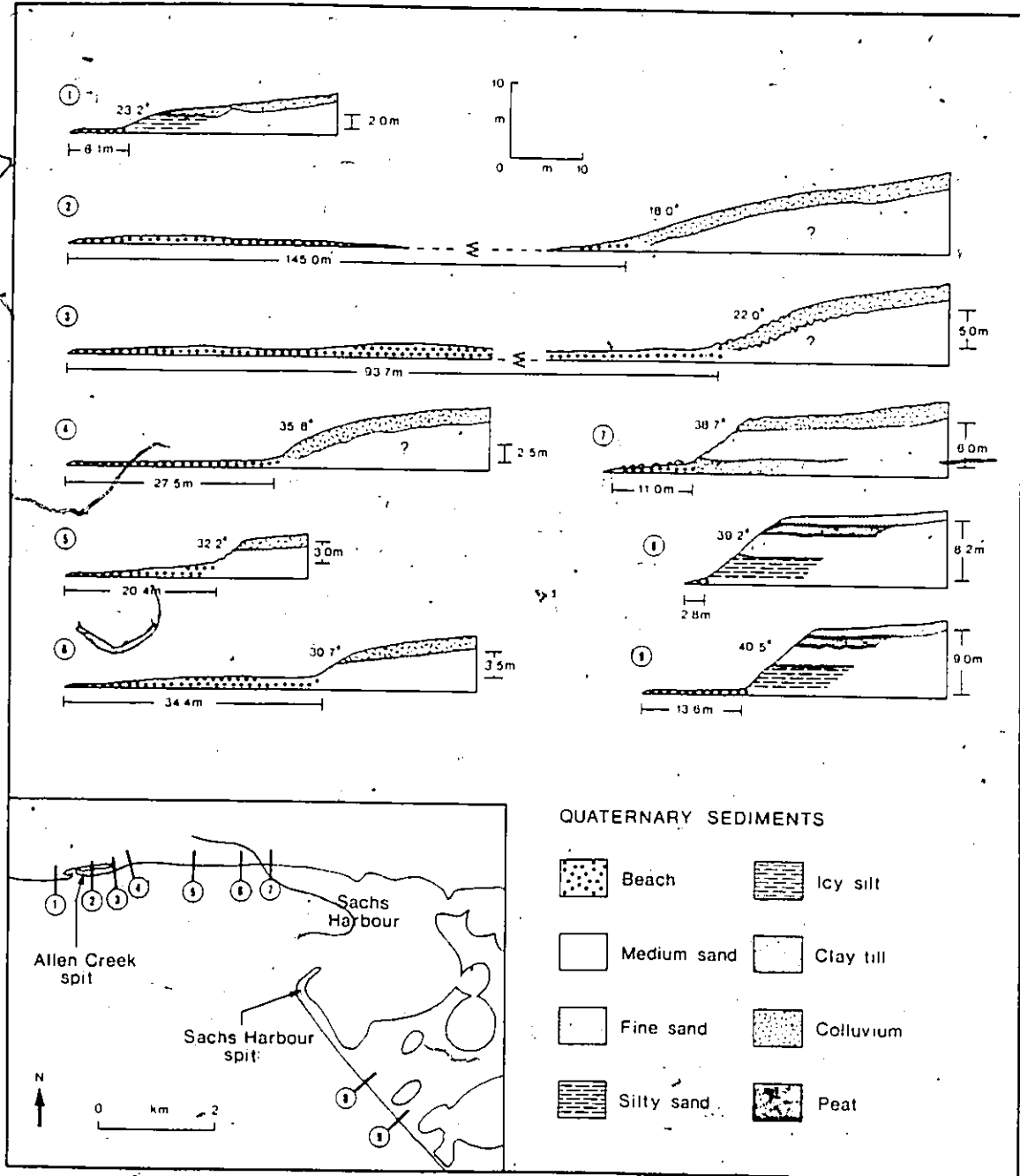


Figure 5.1 Coastal profiles in the vicinity of Sachs Harbour, showing generalized Quaternary stratigraphy.
 (a) Location of profiles.
 (b) Beach and cliff profiles (surveyed by M. J. Clark and W. H. Pollard, August 1-4, 1979).

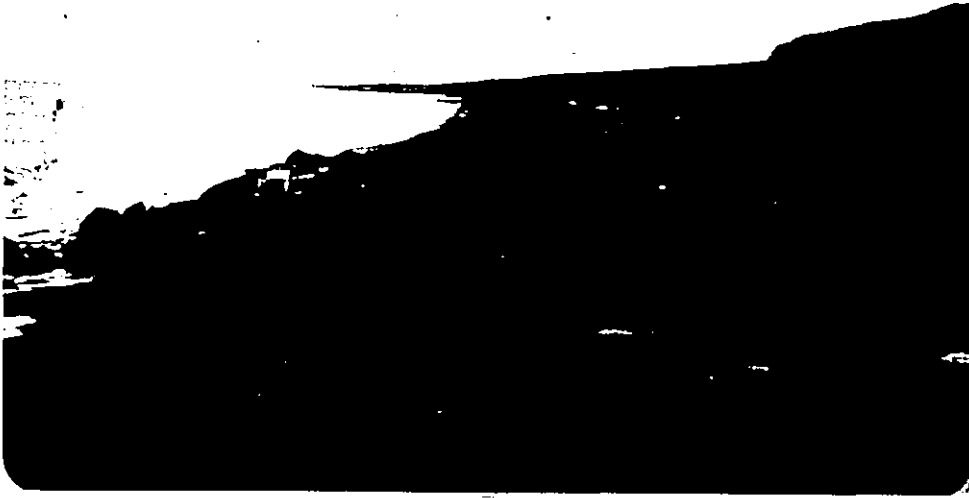
the direction of spit progradation (profiles, 2, 3, 4). This reflects variation in the time since accumulation of a beach armour and cessation of cliff-foot erosion.

The nature of cliff and beach material is also an important control on morphology. For example, cliffs west of Allen Creek, developed in ice-rich organic silts and peat, are subject to frequent slumps and flows. Thus cliff angle is frequently low, despite periodic rapid erosion (profile 1). West of the Sachs Harbour townsite, the beach is mantled by exhumed igneous erratics which are derived from a basal till exposed near the cliff-foot (Figure 5.2a). The narrow beach zone is backed by steep, actively eroding cliffs (profile 7). By contrast, the beach zone immediately to the west is considerably wider and is backed by low, degraded cliffs, consisting primarily of colluvium (profile 6).

5.2.2 Processes of cliff degradation

Wave erosion of sea cliffs is achieved primarily during storm events, through the hydraulic pressure exerted by wave impact and by the abrasive action of entrained sediment (Komar, 1976, p. 15). Processes and rates of cliff degradation are thus normally related to the structure and lithology of cliff materials, and to the frequency and magnitude of storm-wave attack. In permafrost regions, sediment cohesiveness is partly a function of temperature, since frozen ice-cemented sediments possess a strength many times that of similar materials in an unfrozen state (see e.g.

(a)



(b)



Figure 5.2

Examples of cliff materials and morphology, in the vicinity of Sachs Harbour.

- (a) Cliff 1.0 km west of Sachs Harbour, developed in clay till overlain by colluvium. Note the erratic blocks armouring the narrow beach zone. The wide beach in background is backed by degraded cliffs developed entirely in colluvium (see Figure 5.1, profiles 6 and 7). August 20, 1981.
- (b) Cliff 4.0 km southeast of Sachs Harbour, developed in fine-medium glaciofluvial sand. Note the accumulation of a talus rampart (see Figure 4.1, profile 9). August 13, 1979.

Johnston et al., 1981, pp. 73-147; Tsyrovich, 1975, pp. 138-78, 213-56). As a result, it seems likely that actively eroding arctic sea cliffs follow an annual temperature-related morphological cycle. A further constraint on this cycle is provided by the seasonal regime of pack ice-limited wave action.

In winter and early-spring, cliff materials are frozen and almost inactive. The onset of thaw may be retarded by the presence of an insulating snow ramp, formed by winter snowdrifts extending upwards from the beach. As the upper section of cliff thaws, mobilized sediment may be transported onto the snow ramp and, in some cases, may flow across the ice-foot covering the beach. This material may then protect part of the beach ice and snowcover from thaw.

As summer progresses, the thawing front penetrates further into the cliff face, releasing ice-cemented sediment which accumulates as a talus apron at the cliff foot. This process continues until an equilibrium profile is developed, related to the angle of repose of cliff sediments and the thickness of slumped debris required to protect the cliff from further thaw (Figure 5.2b). Observations suggest that most of the slumped debris is removed during late-season storms, so that actively eroding cliffs probably attain their steepest profile just prior to freezeup.

Cliffs developed in ice-rich sediments may degrade very rapidly by a number of failure mechanisms, including flows, slides and falls (McRoberts and Morgenstern, 1973,

1974). These processes can supply large quantities of sediment to the beach system (Figure 5.3). For example, in August 1981, a bimodal flow was observed extending across the beach east of Mary Sachs Creek. This was related to an active layer detachment slide, which extended approximately 10 m inland from the cliff edge. In many areas, the surficial 20-30 cm layer of cliff material is bound by a vegetal mat and, when undercut, fails in coherent blocks up to 1.0 m in diameter which slide down the debris slope to the beach. Where massive ground ice or very icy sediment occurs, the cliff face may form a low-friction surface across which detached blocks readily move (Figure 5.4).

In many areas, cliff degradation is controlled by ice-wedge distribution (e.g. Walker and Arnborg, 1966). Where ice wedges are oriented normal to the coast, preferential thaw results in the formation of a crenulated cliff-line (Figure 5.5). Ice wedges oriented parallel to the coast act as natural lines of weakness and facilitate cliff failure by block detachment (Figure 5.6). During high-magnitude storm events, thermal erosion at the cliff foot may form a thermo-erosional niche, extending up to 5 m beneath the cliff (Figure 5.7). If the niche intersects a shore-parallel ice wedge, massive cliff failure may occur during a single storm event. For example in 1981, failure of a polygonal block during the storm of August 18, was followed by its complete destruction on August 22 (Figure 5.8).



(a)



(b)

Figure 5.3 Examples of cliff failure mechanisms in ice-rich permafrost.

(a) Ground ice slumps in ice-rich silt, approximately 20 km southeast of Sachs Harbour. Note debris flow across the talus slope to the beach zone. August 06, 1980.

(b) Mudflow extending across the beach approximately 9 km west of Sachs Harbour. This bimodal flow is fed by an active layer detachment slide on sloping ground above the cliff face. August 18, 1981.

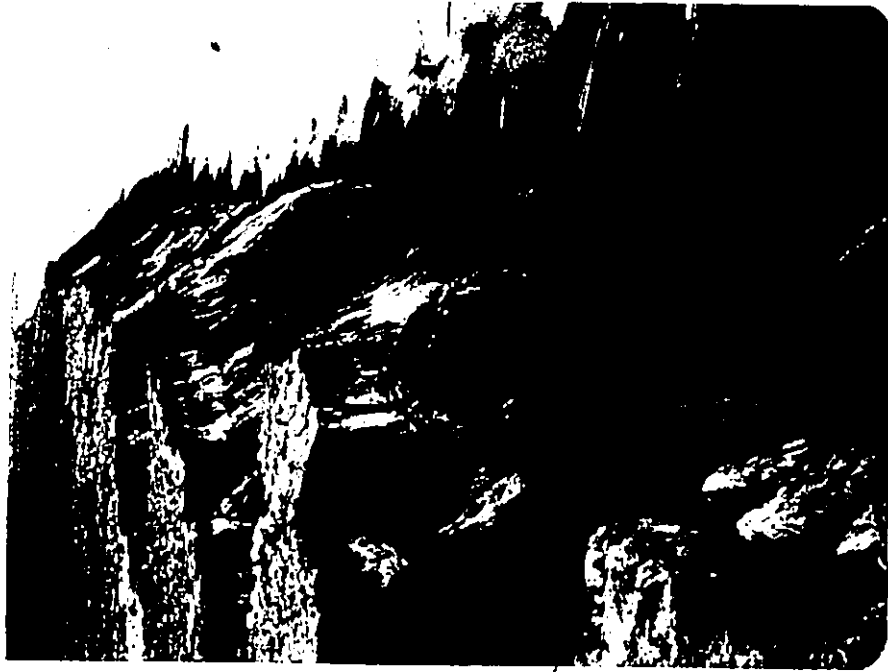


Figure 5.5 Crenulated cliff 4 km south-east of Sachs Harbour, formed by thaw of ice wedges oriented normal to the cliffline. August 19, 1981.



Figure 5.4 Undercut blocks of vegetal mat sliding across icy sediments, exposed in cliffs 4 km west of Sachs Harbour. August 21, 1981.



Figure 5.6 Cliff failure by block detachment along ice wedges oriented parallel to the cliffline 4 km west of Sachs Harbour. (Photograph by H. M. French, July 08, 1976).

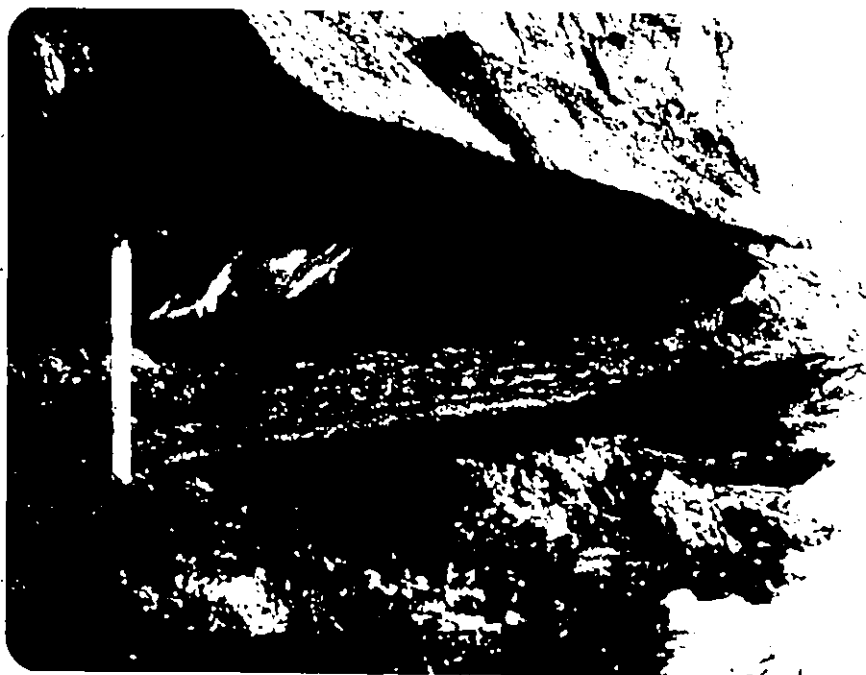


Figure 5.7 Thermo-erosional niche beneath cliffs 4 km west of Sachs Harbour. The niche undercuts cliffs by up to 5.0 m at this locality. August 07, 1980.

(a)



(b)



Figure 5.8

Destruction of a tundra polygon, located 4 km west of Sachs Harbour, by storm wave attack in August 1981.
(a) Failure of polygon along a shore-parallel ice wedge, following undercutting during the storm of August 18, 1981. (Photograph taken August 19, 1981).
(b) Destruction of polygon by wave attack, during the storm of August 22, 1981.

Considerable volumes of sediment may be entrained during storm events. For example, during the storms of August 31, 1980 and August 18, 1981, the debris rampart along the coast of the Sachs River lowlands was completely removed. Assuming a mean cliff height of 8.0 m and a cliff angle of 40°, this represents a loss of approximately $3.8 \times 10^4 \text{ m}^3$ of sediment per kilometre of cliff in the space of 1-2 days. Following both storms, the newly-formed cliff face was either vertical or overhanging by up to 2.0 m (Figure 5.9a). As a result, the navigation beacon located southeast of Sachs Harbour is now seriously threatened.

Accumulation of debris at the cliff foot begins immediately following cessation of wave erosion (Figure 5.9b). For example, in August of 1979 and 1980, coastal sections were cleared of debris using a Wajax high-pressure fire pump. Subsequent recovery to an equilibrium profile was extremely rapid, usually in the order of 7-10 days. In the vicinity of Sachs Harbour, this process is favoured by the south-facing aspect of most sea cliffs, which intercept maximum solar radiation.

5.2.3 Coastal change 1950-1981

The spatial pattern and rate of coastal change in the vicinity of Sachs Harbour was investigated by comparison of air photo coverage dated 1950, 1962, and 1972, and by field survey in the period 1979-81. Net accretion has occurred on the Allen Creek and Sachs Harbour spits (see pp. 177-184), while between Allen Creek and Duck Hawk Bluff,



(a)



(b)

Figure 5.9

Removal and replenishment of cliff-foot talus rampart, Sachs River lowlands.

- (a) Near-vertical cliff profile developed following removal of unfrozen talus during the storm of August 31, 1980. Note the massive cliff failure, threatening stability of the navigation beacon. September 01, 1980.
- (b) Replenishment of the talus slope, following the storm of August 18, 1981. August 19, 1981.

and in the Sachs River lowlands, coastal retreat has averaged 1.0-3.0 m.yr⁻¹. Particularly high rates of cliff recession have occurred in the area immediately west of Allen Creek spit. Detailed analysis of coastal change in this area is facilitated by the availability of 1972 air photo coverage at a scale of 1:7,600. An accurate record of coastal recession since this date may be determined with reference to the erosion of ice-wedge polygons (Figure 5.10). Between 1972 and 1979, the coastline has receded by up to 35 m and two small thaw lake basins have been drained and truncated (see pp. 155-156).

5.3 Development of depositional coastal landforms

Depositional features are characteristic of arctic coasts developed in ice-rich unconsolidated sediment and subject to a significant period of open water conditions. Offshore bars and spits, which typify extensive areas of the Yukon and northern Alaskan coastal plains, also commonly occur on the west coast of Banks Island. For example, approximately 15 km west of Sachs Harbour, Cape Kellett spit forms one of the largest depositional coastal landforms in the western arctic. In the immediate vicinity of Sachs Harbour, two smaller depositional features exist, referred to as the Sachs Harbour spit and Allen Creek spit. Analysis of the evolution of these features provides useful data regarding rates of coastal change in this environment.

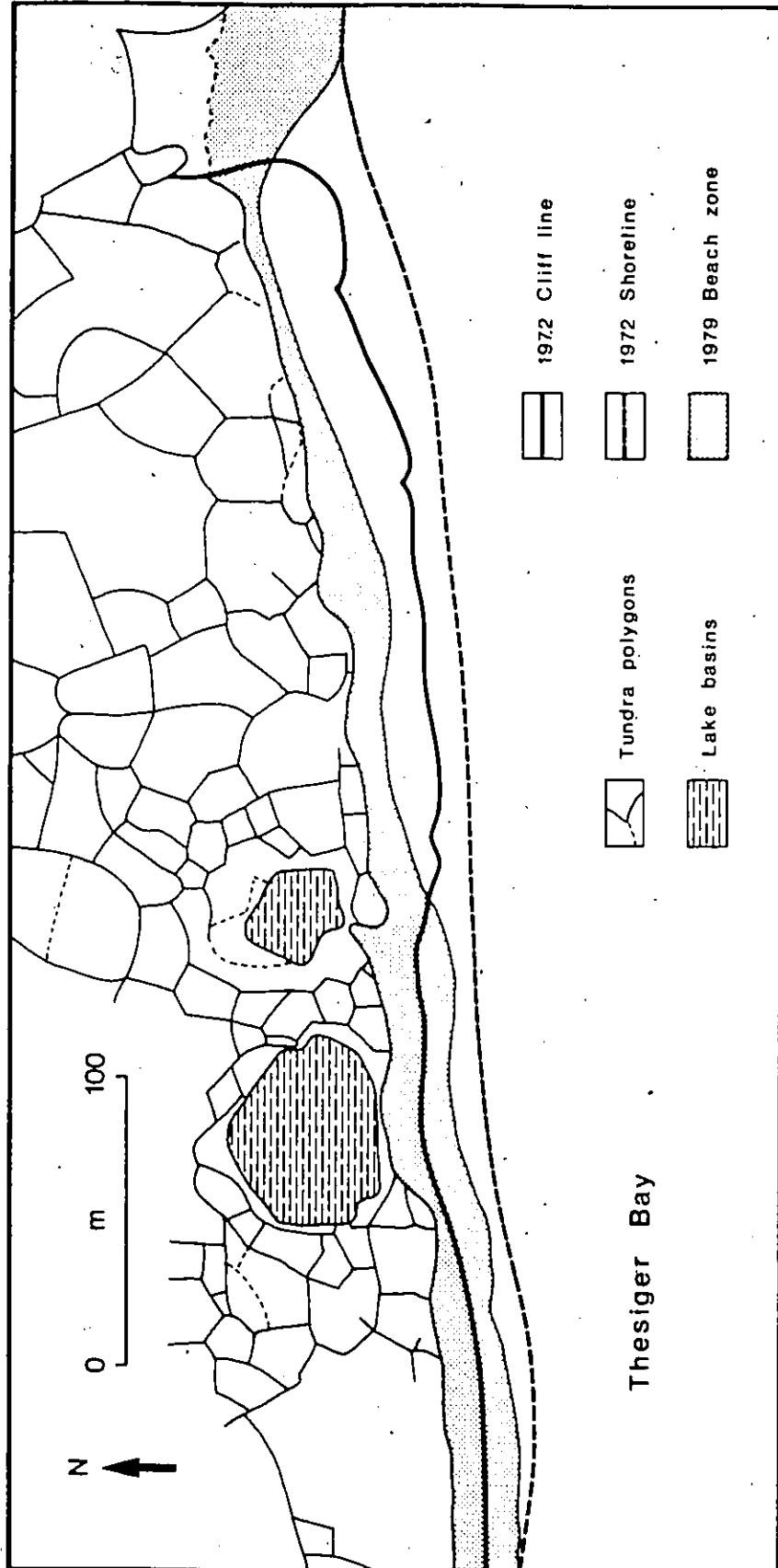
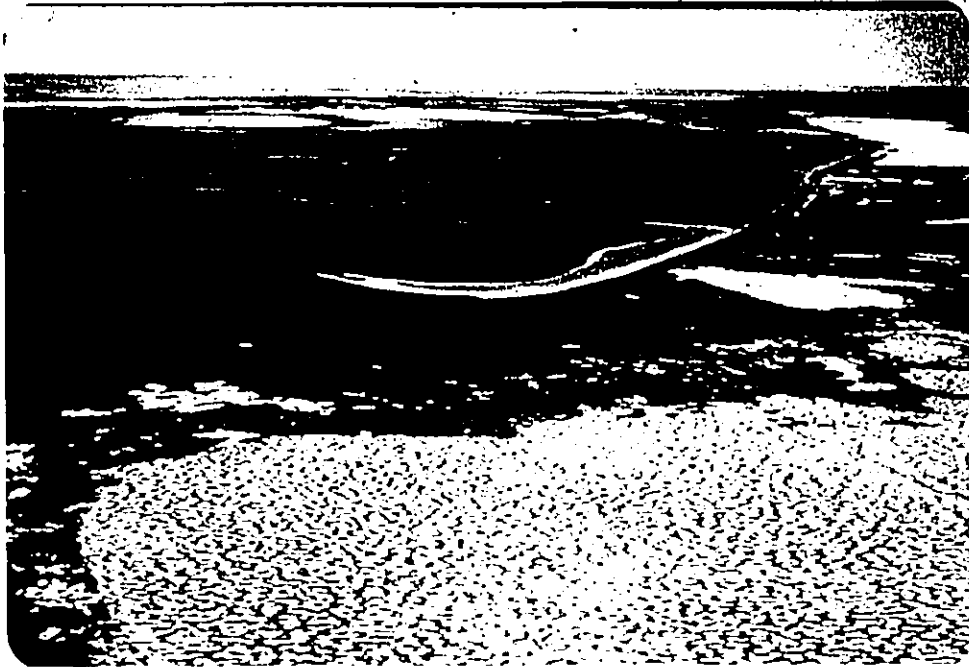


Figure 5.10 Coastal change west of Allen Creek spit, 1972-1979. Note: The 1979 beach zone was surveyed August 14, 1979; the 1972 cliff and shoreline is as shown on air photos A 22953-145, 146, dated August 04, 1972. See Figure 4.19b for oblique air view of this area in August 1980.

5.3.1 Sachs Harbour spit

This feature consists of a recurved sand spit with a 1.6 km linear trunk and a single 0.7 km hook (Figure 5.11). The trunk axis is oriented northwest-southeast, parallel to the coastline of the Sachs lowlands. The spit is asymmetric in cross-profile and the main ridge crest lies close to the southern shoreline. This zone is characterized by a clean, dry sand surface, while central and northern areas of the spit surface consist of organic silt and a remnant, wave-washed vegetal mat. At the base of the recurve, the vegetal mat is intact and surmounted by a number of small, well-vegetated sand dunes.

A survey of the supëraqueous portion of the spit was undertaken in July 1979, and further topographic information was derived from air photo coverage dated 1950, 1958 and 1961. These data indicate that the present simplicity of form conceals a considerably more complex developmental history (Figure 5.12). The 1950 outline shows not only a major northward displacement of the main trunk's central portion, but also evidence of a complete breach 30-40 m in width. The concave indentation thus formed on the spit's southward margin had an area of over 50,000 m², yet had been completely infilled by 1958. A crude estimate of the volume of material involved may be calculated using an empirical relationship between beach area and sediment volume (U.S. Army C.E.R.C., 1973, p. 5.8). This suggests that a 0.09 m² increase in beach area is equivalent to a 0.76 m³



(a)



(b)

Figure 5.11 Sachs Harbour spit.
 (a) Oblique air view of Sachs Harbour spit. (Photograph by H. M. French, July 16, 1972).
 (b) View northwest along the spit axis. Note the asymmetrical cross-profile with a clean sand surface south of the ridge crest. The broad area north of the axis has a surficial cover of organic silt. August 19, 1981.

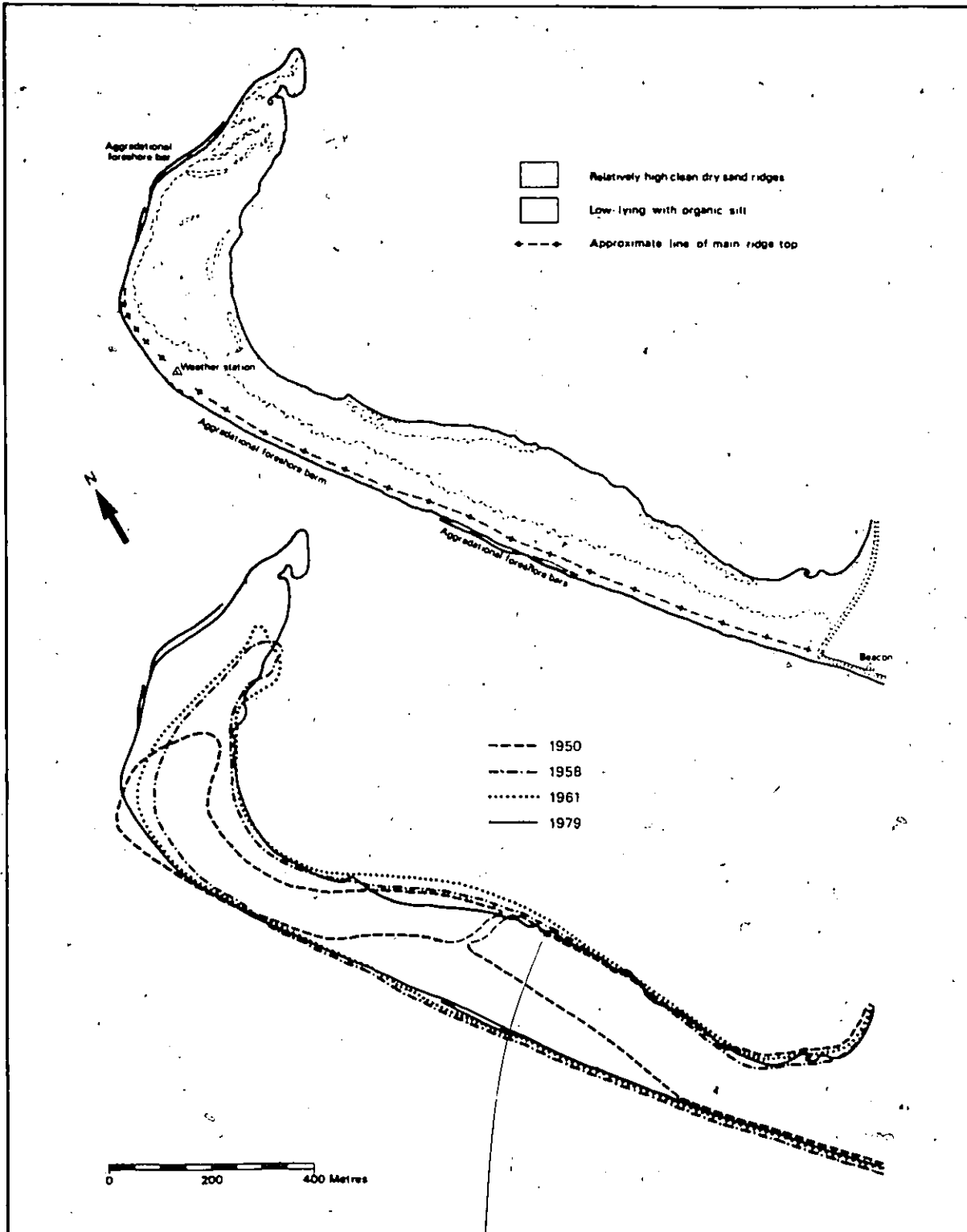


Figure 5.12 Topography and sequential development of Sachs Harbour spit, 1950-1979. Note: Map derived from air photo coverage dated 1950, 1958, and 1961, and from survey by M. J. Clark and J. A. Baker, July 29-30, 1979.

gain in sediment volume. If this assumption is valid, it is likely that approximately $4.6 \times 10^5 \text{ m}^3$ of material was emplaced between 1950 and 1958, a net gain of $5.7 \times 10^4 \text{ m}^3 \text{ yr}^{-1}$ above sea level.

A temporary shoreline indentation of this type forms a very efficient sediment trap within the littoral drift system. Thus the overall rate of spit growth probably offers a more representative indication of sediment accumulation rate. Between 1950 and 1979 the distal end of the spit prograded approximately 400 m, reflecting an annual areal increase of almost $2.5 \times 10^3 \text{ m}^2$. This suggests an annual distal sediment gain of $2.1 \times 10^4 \text{ m}^3$, despite the fact that breach infilling was also in progress at the beginning of this period. If similar growth rates have prevailed in the past, then the entire superaqueous portion of the spit could have been deposited in a period estimated at between 70 and 230 years, depending on the time period selected for rate calculation, with a mean age estimate of 135 years.

5.3.2 Allen Creek spit

This feature extends for approximately 1.5 km parallel to the coastline west of Sachs Harbour, and encloses a small lagoon. The spit is constructed primarily from sand, although medium-coarse gravel is also present. Plant colonization has commenced in some areas of the backshore. Analysis of sequential air photo coverage suggests that, between 1950 and 1972, the primary morphological change consisted of eastward progradation of the spit (Figure 5.13).

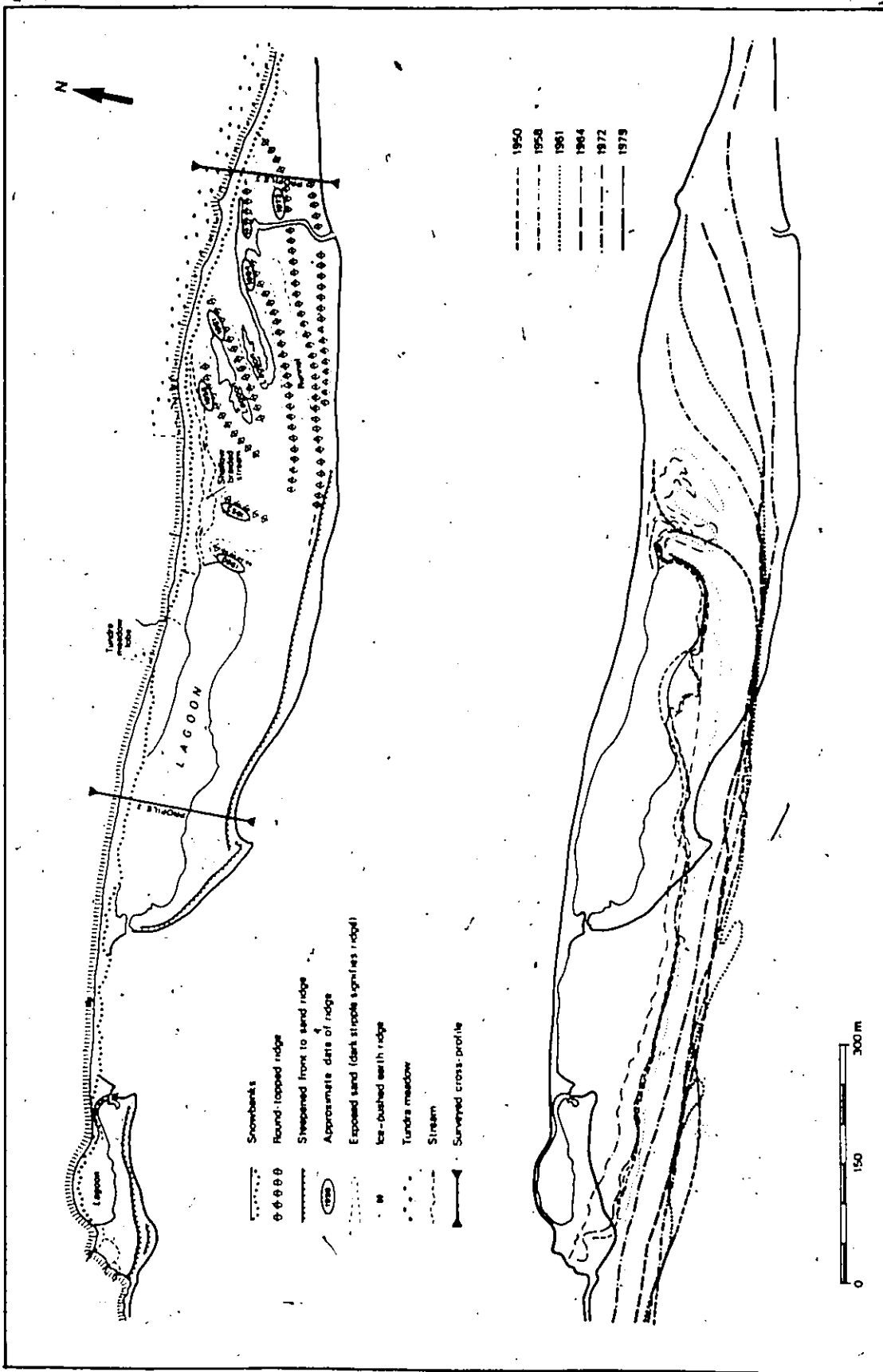
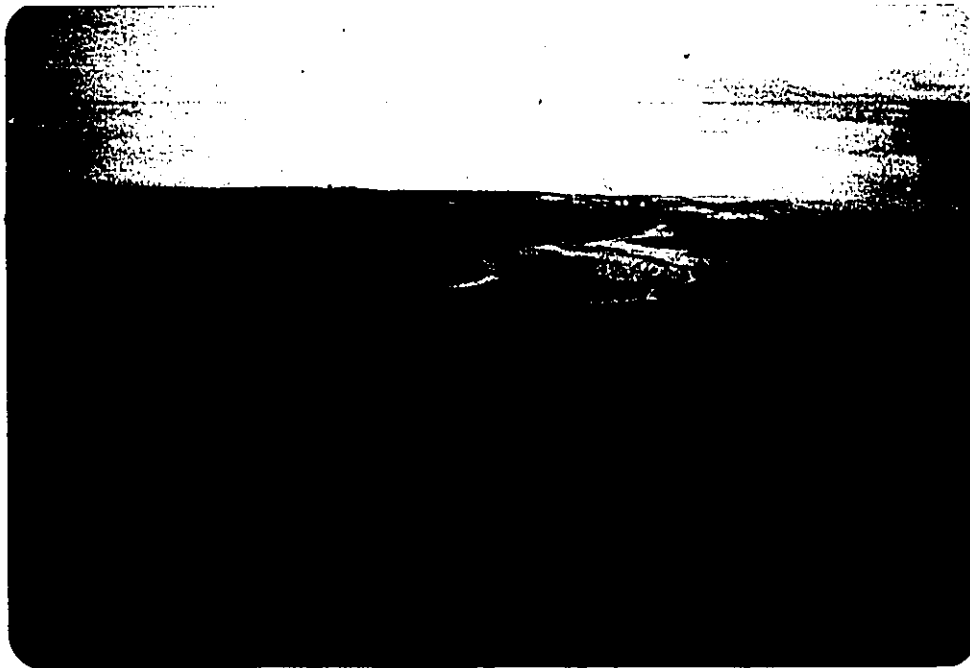


Figure 5.13 Topography and sequential development of Allen Creek spit, 1950-1979.
 Note: Map derived from air photo coverage dated 1950, 1958, 1961, 1964 and 1972, and survey by M. J. Clark and J. A. Baker, July 8-9, 1979.

Successive stages in distal extension may be correlated to clearly defined ridges, mapped in July 1979. Distal extension since 1950 exceeds 600 m, representing an annual foreshore growth of $3.3 \times 10^3 \text{ m}^3$, in close accord with the equivalent rate calculated for the Sachs Harbour spit. Growth rates of this order suggest a total growth period of between 30 and 70 years for the Allen Creek spit, with a mean age estimate of 45 years.

The development of this spit was interrupted by a major breach, which was first noticed in early-July 1978 (personal communication, H. M. French). The breach represents a loss in area of approximately $5.5 \times 10^4 \text{ m}^2$, equivalent to about $4.6 \times 10^5 \text{ m}^3$ of sediment. The rates of distal accretion prevailing on this spit indicate that such a breach could be healed within 6-10 years. This is confirmed by the substantial morphological adjustment observed between July 1978 and August 1981. Beach recovery in the breach zone has resulted in the development of a smoothly concave beach plan (Figure 5.14a).

The breach may have been formed by wave action during a late-season storm in 1976 or 1977. Alternatively, it may have resulted from ice pile-up and overriding in the winter of 1977/78. Available evidence supports the former explanation. Although a low ice-pushed ridge was observed in July 1979, surmounting the 3.0 m high cliffs inshore from the breach zone, neither its composition nor its volume suggested that it represented material lost from the spit.



(a)



(b)

Figure 5.14 Allen Creek spit.
(a) Oblique air view, August 06, 1980. Note beach recovery to a smooth plan outline, following breaching in 1976 or 1977.
(b) Overwash of spit during the storm of August 22, 1981. This event did not result in major structural damage to the spit.

In general, breaching by ice-push is unlikely since, prior to breakup, beach sediment is probably ice-bonded and thus resistant to erosion (e.g. Harper et al., 1978; Owens and Harper, 1977).

Analysis of pack ice and wind records indicates that two major onshore storm events occurred in association with ice-free fetch conditions, on September 13-16 and September 21-22, 1976. These represent the highest magnitude storms in the 10 year period of record, in terms of both duration and kilometres of wind run (see Figure 2.3). They are also associated with the largest hindcast deepwater waves, in excess of 3.5 m (see Figure 2.12). Two lesser storms occurred in September 1977, similar in magnitude to storms observed directly on August 30-31, 1980 and August 22, 1981. On both the latter occasions, 1.0-1.5 m waves were generated which overwashed the spit (Figure 5.14b), eroding shallow channels through the main trunk. However, overall structural damage was slight. It is most probable, therefore, that the breach formed by wave attack during the high-magnitude storms of September 1976.

5.4 Discussion

Observations of coastal environments in the vicinity of Sachs Harbour suggest that rates and patterns of coastal change are controlled by the duration of open water, frequency and magnitude of storm conditions and nature of shoreline materials, with particular respect to their ice content.

In process terms, the area appears to be intermediate between southern storm wave environments and the High Arctic coastal environments described by S. B. McCann and associates (e.g. McCann, 1973; Owens and McCann, 1970; Taylor, 1978; Taylor and McCann, 1976). Storm wave generation in the latter area is considerably restricted by heavy pack ice conditions and limited inter-island fetch during brief periods of open water. On the other hand, it appears that the significance of sea ice to beach development is far greater than on southwest Banks Island.

The closest analogue to this area is provided by the Beaufort Sea coast of the Yukon Territory. Similarities in sea ice conditions, storm wave generation and cliff materials results in a general equivalency of coastal processes and morphology. A recent survey has demonstrated the effects of storm wave erosion of ice-rich cliffs, and abundant nearshore sediment supply resulting in development of constructional morphology (McDonald and Lewis, 1973). These authors conclude that: (1) Some coastal areas have undergone rapid retreat of up to 88 m in the period 1952-70 (an average rate of $4.9 \text{ m} \cdot \text{yr}^{-1}$); (2) Sediment derived from coastal erosion and fluvial discharge is dispersed along the coast by a well-developed system of longshore currents; (3) Longshore sediment drift is responsible for hundreds of metres of spit extension between 1952 and 1970 and; (4) Ice-push did not exert a major influence on the beach zone in 1972. These conclusions are

substantially in agreement with those of the present study.

A major problem in arctic coastal analysis is related to the importance attached to low-frequency, high-magnitude storm events (e.g. Hume and Schalk, 1967; McCann, 1972; Reimnitz and Maurer, 1979). Although this is a general problem of coastal geomorphology, it is highlighted in arctic regions by the predominance of ice-limited fetch conditions, which have led many researchers to characterize the shoreline as a low-energy environment. On southwest Banks Island, rapid coastal erosion and dynamic growth of depositional features suggest that this classification may be inappropriate.

Previous studies have indicated that sediment transport along arctic coasts is a storm-dominated process and this assumption appears justified in the present study. It is possible that the importance of a storm event such as that of August 23, 1980, in supplying sediment for coastal transport, is as great as the annual total supplied by normal summer low-energy wave action. During a major storm, coastal morphology may evolve catastrophically, by accelerated erosion, removal of cliff-foot debris and breaching of spits. At the same time, sediment transport to beach, spit and offshore sinks is probably at a maximum. During intervening periods of low-energy wave attack, coastal morphology remains relatively stable and tends towards an equilibrium status.

CHAPTER SIX

SUMMARY AND CONCLUSIONS

6.1 Introduction

The conclusions reached in this study are presented on two levels. First, specific conclusions are summarized in relation to the permafrost geomorphology of southwest Banks Island. These contribute to an understanding of the complex relationships between permafrost stratigraphy, geomorphic processes and landforms which exist in this area of the Western Canadian Arctic. Second, general conclusions relate to the use of permafrost geomorphology and stratigraphy as conceptual frameworks for arctic terrain analysis. In addition, it is possible to make a number of recommendations with respect to future research in this field.

6.2 Summary of results

6.2.1 Permafrost stratigraphy

The study of ice wedges and their enclosing sediments, exposed in coastal sections on southwest Banks Island, suggests that permafrost aggradation has been continuous in late-Quaternary times. A temporary period of deeper seasonal thaw occurred during the mid-Holocene climatic optimum. Stratigraphic analyses of two pingos support this interpretation of permafrost history.

In the Sachs River lowlands, four systems of ice wedges are recognized within a sequence of outwash sediments, attributed to the (Wisconsin) Amundsen Glaciation. The first system consists of inactive epigenetic wedges, related to a paleo-surface dated at greater than $10,600 \pm 130$ years B.P. (GSC-3229). The second system consists of small syngenetic ice veins, developed during a subsequent period of rapid surface aggradation. These are of particular interest since, in contrast to observations in the Soviet Union, few syngenetic ice wedges have been recognized in North America. The third system consists of inactive epigenetic wedges, which are truncated approximately 1.5-1.7 m below surface. The level of thaw unconformity is dated at greater than $6,490 \pm 60$ years B.P. (GSC-3216), and almost certainly represents the maximum depth of seasonal thaw during the mid-Holocene climatic optimum. The fourth system consists of active epigenetic wedges, which are related to the distribution of thermal contraction cracks.

Pingos on southern Banks Island differ from those of the Mackenzie Delta area in a number of ways. First, they are both smaller and less numerous. This probably reflects the colder climate, which limits talik development and thus potential pingo dimensions. Second, most pingos are located on low fluvial terraces, rather than within drained lake basins. Third, they range morphologically from conical mounds to elongate ridges, which may be several hundreds of metres in length. Fourth, many pingos are in

varying states of collapse.

A section, excavated through the rampart of a collapsed pingo near Carpenter Lake, indicated that the remnant ice core comprises both segregation and injection ice. A small, now horizontal, ice vein extending 3-4 m inwards towards the pingo centre indicates the probable extent of permafrost aggradation before shut-off pressure was exceeded and ice core growth commenced. Based on ^{14}C age determinations, a second pingo, located in the upper Kellett River valley, is thought to have grown between $3,920 \pm 80$ years B.P. (GSC-2395) and $2,480 \pm 50$ years B.P. (GSC-2397). Pingo development probably reflects climatic deterioration and renewed permafrost aggradation in late-Holocene times. There is broad synchronicity between the age of this pingo and that suggested for pingos elsewhere on Banks Island. It is thus likely that the majority of pingos on southwest Banks Island, including the Carpenter Lake feature, grew during this period.

Near the limit of the Sand Hills glacial readvance, ice-ablation features exist which are morphologically similar to collapsed pingos. However, their geomorphic setting is not consistent with a pingo origin. Examination of an ice core and overlying sediments suggests that these features formed by ablation of buried glacier ice. Their recognition complicates the interpretation of presumed pingo scars, in regions of both contemporary and Pleistocene permafrost.

6.2.2 Thaw lakes

Thaw lakes on southwest Banks Island appear to be quasi-equilibrium landforms, and cannot readily be interpreted within the traditional thaw lake 'cycle', as proposed for example in northern Alaska. Initial basin formation, by thermokarst subsidence, has been followed by asymmetrical expansion under the influence of geomorphic processes. This has resulted in the development of a preferred long-axis orientation. There is little or no evidence of thaw lake migration across the tundra surface.

The volume of ground ice in the upper 8 m of permafrost in the Sachs River lowlands indicates that lake basins at least 3.0 m in depth could have originated by thermokarst subsidence. This conclusion is supported by the reduced ice content and stratigraphic thickness of sediments exposed beneath a drained lake basin. Basal lacustrine sediments in this section provide a ^{14}C date of $8,560 \pm 210$ years B.P. (GSC-3292). This is consistent with the age of other thaw lake sediments on southern Banks Island, and supports a conclusion that basin formation was initiated during the mid-Holocene climatic optimum. However, many smaller lakes, located on terraces of the Kellett River, probably developed at a later date by ponding within low-centred polygons.

Geothermal analysis, based on heat conduction theory, indicates that thaw lakes significantly modify the regional permafrost regime. It is concluded that, under equilibrium

conditions, lakes greater than 700-800 m in diameter are underlain by throughgoing taliks, which perforate the estimated 400-500 m thick permafrost zone.

A majority of thaw lakes on southwest Banks Island are oriented, with a mean long-axis azimuth approximately perpendicular to the opposed resultants of the summer wind regime at Sachs Harbour. The most common, smoothly curved lake outline consists of an asymmetrical ellipse or D-shape, which is assumed to represent the equilibrium lake form.

Lake morphology was simulated using a model based on the calculation of cycloidal bay radius, from input wind frequency data. The correspondence between predicted and observed morphology is close, particularly when storm wind data are used. It is concluded that lake orientation on southwest Banks Island is directly related to wind-generated geomorphic processes. The sensitivity of this relationship is demonstrated by the rarity of oriented thaw lakes elsewhere on Banks Island. Thaw lake drainage occurs primarily by catastrophic outflow, as a result of either lake tapping or truncation by coastal retreat. In areas of ice-cored terrain, subterranean drainage occurs under certain circumstances. The sequence of infilling, segmentation and revegetation, which forms an integral part of the classic thaw lake 'cycle', is rare in this area.

6.2.3 Coastal change

Rates and patterns of coastal change on southwest Banks Island are strongly influenced by the presence of unconsolidated and perennially frozen shoreline materials. Coastal processes are further controlled by the duration of open water conditions and the frequency and magnitude of onshore storm events. In general, the area is intermediate between pack ice-dominated High Arctic coastal environments and southern storm wave environments.

Actively eroding sea cliffs follow an annual, temperature-related, morphological cycle. Cliff failure mechanisms are frequently controlled by the quantity and distribution of ground ice. Cliffs developed in ice-rich, but homogeneous sediments typically fail by active layer detachment slides and bimodal flows. By contrast, cliff failure in areas of polygonal ground is controlled by ice-wedge distribution. Catastrophic block failure usually occurs along the line of ice wedges, particularly where the cliff foot is undercut by a thermo-erosional niche. During the period 1950-1979, coastal retreat between Allen Creek and Duck Hawk Bluff, and in the Sachs River lowlands, averaged $1.0-2.0 \text{ m} \cdot \text{yr}^{-1}$. Cliffs developed in ice-rich silt immediately west of Allen Creek receded by up to 35 m between 1972 and 1979.

Sediment mobilized by rapid coastal retreat is trans-

ported by littoral drift to a number of 'sinks', including the Sachs Harbour and Allen Creek spits. Between 1950 and 1979, these features prograded by approximately 400 m and 600 m respectively. The presence of well developed depositional landforms, subject to rapid morphological change, seems characteristic of arctic coastlines which consist of unconsolidated ice-rich permafrost sediments.

6.3 Conclusions

This study demonstrates that a knowledge of permafrost conditions forms an integrating factor in the study of arctic terrain. On southwest Banks Island, three aspects of meso-scale land form are shown to comprise elements of an interdependent landscape system, controlled to a considerable degree by the nature of perennially frozen ice-rich sediments.

In this area, permafrost aggradation in late-Quaternary times has resulted in the growth of pingos and the development of multiple systems of ice wedges. At various times, thermokarst processes have led to partial collapse of ice-cored features and the formation of thaw lake basins. The latter in turn significantly modify the thermal regime of underlying and adjacent permafrost. In the coastal zone, rapid cliff recession is caused by the melt of ice-rich unconsolidated sediments, especially along ice wedges. At some localities, truncation and subsequent drainage of thaw lake basins has occurred. These obser-

vations suggest that permafrost geomorphology provides a strong conceptual framework for arctic landscape investigation.

This study also shows that it is possible to deduce paleoenvironmental conditions in permafrost regions through the study of ground ice bodies and their enclosing sediments. Stratigraphic studies have traditionally formed a basic research methodology in investigations of Quaternary geomorphology in nonpermafrost regions. Within the permafrost zone, the usefulness of such an approach is further enhanced by the possibility of reconstructing not only depositional events but also paleothermal and paleo-hydrologic conditions, based on the nature and distribution of ground ice bodies. In this respect, future studies of Quaternary geomorphology in arctic regions should include detailed analysis of permafrost stratigraphy as an integral component of field investigations.

6.4 Recommendations for future research

Many opportunities exist for further work in this field, and a number of specific lines of future research may be identified.

The relatively short history of permafrost aggradation on southwest Banks Island in postglacial times represents a significant limitation of the present study. It would be instructive, therefore, to investigate permafrost geomorphology within an area such as the Beaufort Plain of northwest Banks Island, which has experienced a cold nonglacial climate since at least the end of the pre-Wisconsin interglacial period. A more complex permafrost stratigraphy might be expected, similar to that described from unglaciated areas of Siberia.

The study of permafrost stratigraphy would be strengthened by the availability of thermal data from a nearby deep borehole. This would allow verification of thermal models generated in relation to present geomorphological conditions. It may also allow correlation of thermal events recorded within the permafrost stratigraphy with temperature anomalies in the borehole. Since thermal waves penetrate earth materials at a finite velocity, it follows that ground temperatures at depth are related to former surface conditions.

With respect to the study of ground ice bodies and their enclosing sediments, particular attention should be focused upon the nature of cryotextures. These are used widely by Soviet geocryologists as indicators of freezing and thawing histories, for example in a former talik. A method of directly dating ground ice bodies, based for example on isotope ratios, would also represent a major

contribution to the field.

The relative rarity of inland permafrost sections has limited the use of a stratigraphic approach to permafrost studies in large areas of northern Canada. In the near future, extensive artificial exposures, albeit both shallow and temporary, will be produced during the construction of northern pipelines. These sections should provide unique opportunities to examine surficial permafrost stratigraphy, in relation to a wide range of arctic terrain conditions.

APPENDIX A

COMPUTER MODEL OF ACTIVE LAYER REGIME:

TABLE OF INPUT DATA

Appendix A

Computer Model of Active Layer Regime:
Input Data

	Site #1 Dry Tundra	Site #2 Wet Tundra
<u>Surface Characteristics</u>		
Snow albedo	0.82	0.82
Snow roughness	0.01	0.01
Snow density	0.18	0.18
Surface albedo	0.20	0.20
Surface roughness	1.00	2.00
Surface wetness	1.00	1.26
<u>Subsurface Characteristics</u>		
Layer 1	Sand	Peat
Depth Range (cm)	0-1500	0-50-
Kf	0.0057	0.0020
Ku	0.0040	0.0007
Cf	0.78	0.47
Cu	0.54	1.00
Layer 2	-	Silty sand
Depth Range (cm)		50-1500
Kf		0.0054
Ku		0.0034
Cf		0.73
Cu		0.47

Kf, Ku: Frozen and unfrozen thermal conductivities (cal/cm/sec/°C).
Cf, Cu: Frozen and unfrozen volumetric heat capacities (cal/cm³/°C).

Note: For full program listing, see Smith (1977, pp. 33-49).

References: Brown and Williams, 1972; Slusarchuk and Watson, 1975;
Smith, 1975; 1977.

APPENDIX B

COMPUTER MODELS OF THAW LAKE GEOTHERMAL INFLUENCE:

PROGRAM LISTINGS

```

C THIS PROGRAM CALCULATES EQUILIBRIUM TALIK DIMENSIONS
C DEVELOPED BENEATH THE CENTRE OF CIRCULAR LAKES OF
C GIVEN RADIUS.
C
C INPUT PARAMETERS :
C TL : MEAN ANNUAL LAKE BOTTOM TEMPERATURE
C TG : MEAN ANNUAL TUNDRA SURFACE TEMPERATURE
C Q : MEAN GEOTHERMAL GRADIENT (MK/M)
C
C METHOD : DESCRIBED IN MACKAY, 1962, PP.35-48.
C
C CALCULATE EQUILIBRIUM PERMAFROST THICKNESS, P,
C MAXIMUM DEPTH TO PERMAFROST BENEATH LAKE CENTRE, C,
C MINIMUM RADIUS OF LAKE WITH THROUGHGOING TALIK, M.
99 READ(5,100,END=999)TL,TG,Q
WRITE(6,101)TL,TG,Q
B=TL-TG
P=-TG/Q*1000
A=1.0/3.0
C=((B*B*TL)**A)-TL)*Q
W((((Q*B*C)**2)/(((Q*B)+(Q*TG)+C)**2))-(C*C)**0.5
WRITE(6,102)P,C,W
C CALCULATE DEPTH, D, OF TALIK BENEATH LAKE RADIUS, R.
Z=5
2 CONTINUE
R((((Q*B*Z)**2)/(((Q*B)+(Q*TG)+Z)**2))-(Z*Z)**0.5
U=R/10
IS=IFIX(U)
RS=U-IS
IF(RS.LE.0.5) GO TO 3
IS=IS+1
3 CONTINUE
WRITE(6,103)Z,R
C TEST FOR PRESENCE OF THROUGHGOING TALIK
IF(Z.GT.C) GO TO 5
Z=Z+5
GO TO 2
5 CONTINUE
WRITE(6,104)W
GO TO 99
999 CONTINUE
STOP
100 FORMAT (2F5.1,F4.1)
101 FORMAT (1H1/5X,'PERMAFROST DISTRIBUTION BENEATH THAW LAKES'///5X,
1'LAKE TEMPERATURE = ',F5.1,10X,'TUNDRA SURFACE TEMPERATURE = ',
2F5.1,///5X,'GEOTHERMAL GRADIENT = ',F6.1,' MILLIKELVINS PER METRE'//)
102 FORMAT(5X,'MAXIMUM PREDICTED PERMAFROST THICKNESS = ',F8.1,
1' METRES'//5X,'MAXIMUM DEPTH TO PERMAFROST BENEATH LAKE CENTRE =
2',F8.2,' METRES'//5X,'MINIMUM RADIUS OF LAKE POSSESSING THROUGH TA
3LIK = ',F8.2,' METRES'///1X,'TALIK DEPTH',7X,'LAKE RADIUS')
103 FORMAT(5X,F5.1,12X,F6.1)
104 FORMAT(1X,'THROUGH TALIK',8X,F6.1,'+')
END
$ENTRY

```

```

C THIS PROGRAM CALCULATES THE THERMAL EFFECT OF WATER BODIES
C ON EQUILIBRIUM PERMAFROST TEMPERATURES.
C
C INPUT PARAMETERS :
C R1 : INNER RADIUS OF RAY
C R2 : OUTER RADIUS OF RAY
C
C METHOD : DESCRIBED IN LACHENBRUCH, 1957A.
C PROGRAM MODIFIED AFTER SMITH, 1973, PP.151-169.
C
      DIMENSION R1(100),R2(100),D(20),SSTEMP(20),ASTEMP(20)
      WRITE(6,400)
C INITIALIZE DEPTHS AT WHICH THERMAL EFFECT IS CALCULATED
      D(1)=10
      D(2)=25
      D(3)=50
      D(4)=100
      D(5)=150
      D(6)=200
      D(7)=250
      D(8)=300
      D(9)=350
      D(10)=400
      D(11)=450
      D(12)=500
      D(13)=550
      D(14)=600
      KD=14
C SET INCREMENTAL ANGLE OF RAYS
      ANGLE=20.0/360.0
      J=1
      K=1
      1 READ(5,100)R1(J),R2(J)
      400 FORMAT(1H1, 4X,'THERMAL EFFECT OF LAKES ON EQUILIBRIUM GROUND TEMPERATURE FIELD',///10X,'DEPTH (M)',12X,'DISTURBANCE',13X,'TEMPERATURE')
      100 FORMAT(2F4.0)
      IF(R1(J).EQ.9999.) GO TO 5
      J=J+1
      GO TO 1
      5 CONTINUE
      KC=J-1
C CONVERT DISTANCES TO METRES
      DO 600 J=1,KC
      R1(J)=R1(J)*50.0
      600 R2(J)=R2(J)*50.0
C CALCULATE THERMAL EFFECT OF EACH WATER BODY AT GIVEN DEPTHS
      DO 710 K=1,KD
      SSTEMP(K)=0.0
      EK=1.0/D(K)
      DO 700 J=1,KC
      RATIO1=1./((SQRT(1.+(R1(J)*EK)**2))
      RATIO2=1./((SQRT(1.+(R2(J)*EK)**2))
      SSTD=RATIO1-RATIO2

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```
700 SSTEMP(K)=SSTEMP(K)+SSTD
710 CONTINUE
    DO 720 K=1,KD
720 SSTEMP(K)=SSTEMP(K)*ANGLE
    CALCULATE THERMAL DISTURBANCE AND EQUILIBRIUM TEMPERATURE
    DO 750 K=1,KD
        ASTEMP(K)=SSTEMP(K)*11.8
        TACT(K)=ASTEMP(K)+(D(K)*0.022)-10.8
750 CONTINUE
        WRITE(6,200)(D(K),ASTEMP(K),TACT(K),K=1,KD)
200 FORMAT(11X,P5.1,17X,P5.1,19X,P5.1)
    STOP
    END
$ENTRY
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ABSTRACT

This thesis explores the relationship between permafrost conditions and landscape evolution on southwest Banks Island, Western Canadian Arctic. Particular attention is focused upon the history of permafrost and ground ice aggradation, the development of thaw lake terrain, and the influence of permafrost on rates and processes of coastal change.

Stratigraphic analysis of coastal sections suggests that permafrost aggradation has been continuous in late-Quaternary times, with the exception of a short period of deeper seasonal thaw during the Holocene climatic optimum. Both epigenetic and small syngenetic ice wedges are recognized, and their development is interpreted within a framework of paleoenvironmental reconstruction. Additional evidence of permafrost aggradation is provided by the existence of pingos. The distribution of sediments and ice bodies within sections excavated through the rampart of one feature is interpreted with respect to a model of pingo growth. There is broad synchronicity between the age of a second pingo examined and that of other pingos on Banks Island, suggesting that growth may have been triggered by regional climatic deterioration in late-Holocene times.

Thaw lakes on southwest Banks Island appear to be quasi-equilibrium landforms, and cannot be interpreted within the traditional thaw lake 'cycle'. In the Sachs River lowlands, basins at least 3 m in depth probably formed by thaw subsidence, initiated during the Holocene climatic optimum. Geothermal analysis indicates that, under equilibrium conditions, lakes greater than 700-800 m in diameter are underlain by throughgoing taliks. A majority of lakes are oriented perpendicular to prevailing summer wind direction, and possess a D-shaped outline which is in equilibrium with wind-generated geomorphic processes. Lake drainage occurs primarily by catastrophic outflow, following basin capture or truncation by coastal retreat.

Rates and processes of coastal change are controlled by the nature of perennially frozen shoreline materials. In particular, patterns of cliff failure are frequently related to ice wedge distribution. During the period 1972-1979, ice-rich cliffs west of Sachs Harbour receded by up to 35 m. The high rate of sediment supply has contributed to the growth of two sand spits in the vicinity of Sachs Harbour which have prograded by 400-600 m since 1950.

The results of this study suggest that a knowledge of permafrost conditions forms an integrating factor in the analysis of arctic terrain. Moreover, a stratigraphic approach to permafrost may be used to reconstruct the evolution of present geomorphological conditions.

RESUME

Le sujet de cette thèse est l'étude des relations entre les conditions du pergélisol et l'évolution du relief dans le sud-ouest de l'Ile de Banks, Archipel Reine Elizabeth, Territoires du Nord-Ouest. Une attention particulière est accordée à l'historique du pergélisol et de l'aggradation de la glace du sol, au développement des reliefs de thermokarst, et à l'influence du pergélisol sur les taux d'activités des processus responsables des changements côtiers.

L'analyse stratigraphique, dans les coupes le long de la côte, suggère une aggradation presque continue du pergélisol depuis le Quaternaire supérieure. Cependant, lors de l'optimum climatique de l'Holocène, le mollisol semble avoir été plus épais. Des coins de glace épigénétiques et syngénétiques co-existent et leur développement est interprété dans le cadre d'une reconstruction paléoenvironnementale. La présence de pingos fournit une évidence supplémentaire de l'aggradation du pergélisol. La distribution des sédiments et des masses de glace dans des tranchées excavées dans les remparts d'une structure est interprétée selon un modèle d'un pingo en croissance. Il existe un large synchronisme entre l'âge d'un second pingo et celui des autres pingos de l'Ile de Banks, ce qui suggère que la croissance a pu être déclenchée par une détérioration

climatique à l'Holocène supérieure.

Dans le sud-ouest de l'Ile de Banks les lacs de thermokarst sont des formes qui semble en quasi équilibre et qui ne peuvent donc être interprétés dans le contexte du cycle traditionnel des lacs de thermokarst. Dans les basses terres du bassin de la rivière Sachs, des dépressions profondes de 3 m peuvent résulter de subsidences thermokarstiques débutant à l'optimum climatique de l'Holocène.

Des analyses géothermiques indiquent que, dans des conditions d'équilibre, il n'y a pas de pergélisol sous des lacs de plus de 700 m de diamètre. La majorité des lacs sont perpendiculaires aux vents dominants d'été et leur tracé en "D" est en équilibre avec les processus éoliens. Le vidange d'un lac est surtout de nature catastrophique suite à une capture ou à une récession de la côte.

Le pergélisol dans le matériel du rivage, contrôle la nature et la vitesse des processus qui affectent les côtes. En particulier, la nature des affaissements de falaise est souvent reliée à la distribution des coins de glace. Durant la période 1972-1979, le recul des falaises, riches en glace, à l'ouest de Sachs Harbour a atteint 35 m. Depuis 1950, la quantité élevée de sédiments a contribué à la croissance de 400-600 m d'une flèche littorale près de Sachs Harbour.

Les résultats de cette thèse suggèrent qu'une connaissance des conditions de pergélisol est un facteur d'intégration dans l'analyse des formes du terrain dans

l'Arctique. En plus, une étude à base stratigraphique du pergélisol peut être utilisée pour reconstruire l'évolution des conditions géomorphiques des reliefs arctiques.