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QUATERNARY GEOLOGY OF THE INMAN RIVER AREA, NORTHWEST TERRITORIES

D.A. St-Onge and I. McMartin

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Slump scar exposing buried glacier ice and overlying till in parts of Bluenose Lake Moraine. GSC 1993-311

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Preface

This report deals with the Quaternary geology of the Inman River area which, along the eastern slope of the upland west of Bluenose Lake, includes part of the Late Wisconsinan ice limit. Thus it provides data essential to the understanding of ice dynamics along part of the margin of the Late Wisconsinan Laurentide Ice Sheet. This information has ramifications far beyond the immediate study area and is a contribution to the Quaternary geology of the Canadian land mass.

Land use is an increasing concern of Arctic inhabitants. In the tradition of many similar studies by scientists of the Geological Survey of Canada, this report contributes information required for sound management practice. Knowledge of the distribution, nature, and origin of the complex suite of glacial and other associated sediments will be a major component of land-use planning in the proposed Bluenose Lake National Park and will be fundamental for other land use such as a gas pipeline corridor which has been proposed to pass through this area.

Elkanah A. Babcock
Assistant Deputy Minister
Geological Survey of Canada

Préface

Ce rapport résulte d'une étude de la géologie du Quaternaire de la région de la rivière Inman qui comprend, sur le versant est du plateau à l'ouest du lac Bluenose, une partie de la limite glaciaire du Wisconsinien supérieur. Les résultats sont donc fondamentaux à une meilleure compréhension de la dynamique glaciaire à la marge de l'Inlandsis laurentidien au Wisconsinien supérieur. L'intérêt de cette information déborde largement les limites de la région immédiate et représente une contribution à l'ensemble de la géologie du Quaternaire du territoire canadien.

L'utilisation du sol est une préoccupation croissante des habitants de l'Arctique. Dans la tradition de plusieurs études similaires effectuées par les scientifiques de la Commission géologique du Canada, ce rapport fournit de l'information qui est requise pour une gestion éclairée du territoire de la région. La distribution, la nature et l'origine du complexe de sédiments glaciaires et dépôts associés représentent des données de base pour la planification de l'utilisation des terres dans le parc national récemment proposé pour les environs du lac Bluenose. Ces informations sont également requises pour d'autres types d'aménagement telle la proposition de construction d'un gazoduc.

Elkanah A. Babcock
Sous-ministre adjoint
Commission géologique du Canada

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Abstract

The Inman River area south of Dolphin and Union Strait includes lowlands, scarps, and tableland on nearly horizontal strata of Palaeozoic carbonates, mostly dolomite, north of the Middle to Late Proterozoic Coppermine Homocline rocks. The Late Quaternary, mostly Wisconsinan, sediments which unconformably overlie the dolostone are associated with a spectacular array of landforms including several types of moraines, a wide variety of drumlins, eskers, kames, and flights of marine beaches. Varying flow patterns resulting from increased topographic control during the waning phases of glaciation are reflected in the orientation of moraine ridges, drumlins, grooves, and striae. Periglacial processes are responsible for a series of landforms including the unique Harding River rock pingo, numerous rock mounds, blowout and dune fields, as well as pattern ground.

Concurrently with ice retreat, marine incursion imposed its own constraints on sedimentation and erosion patterns. Initially, marine basins confined between ice lobes and uplands received vast quantities of sediments entrained by meltwater from stagnant ice on upland surfaces and from subglacial streams debouching at ice fronts calving into the ocean. This episode of sedimentation was of short duration; the disappearance of stagnant ice brought it to a close.

Late Wisconsinan ice started to melt back from the eastern slopes of the upland just west of Bluenose Lake around 12 000 BP. Deglaciation from this position to Coronation Gulf has been divided into four phases defined by distinct flow patterns identified from drumlins, flutings, and striae, and by ice frontal positions marked by morainic ridges and other ice contact features.

Résumé

La région de la rivière Inman au sud du détroit de Dolphin and Union comprend des basses terres, des escarpements et des plateaux sous-tendus par des couches subhorizontales de roches carbonatées du Paléozoïque, surtout des dolomies, au nord des roches de l'homoclinal de Coppermine du Paléozoïque moyen ou supérieur. Les sédiments du Quaternaire supérieur, du Wisconsinien surtout, qui recouvrent en discordence les dolomies, supportent un imposant cortège de formes de relief y inclus différents types de moraines, une grande variété de drumlins, des eskers, des kames et des crêtes de plages marines. L'orientation des crêtes morainiques, des drumlins des cannelures et des stries reflètent des directions d'écoulement de plus en plus influencées par la topographie au cours de la déglaciation. Les processus périglaciaires sont responsables d'une série de formes de relief qui incluent le spectaculaire pingo rocheux de la rivière Harding, des pustules rocheuses, des champs de déflation éolienne et de dunes, ainsi que des sols à figures géométriques.

Au cours du retrait glaciaire, l'incursion marine a imposé ses propres contraintes à l'érosion et à la sédimentation. Initialement, les eaux de fonte en provenance des masses de glace morte sur les plateaux et les cours d'eau sous-glaciaires débouchant au niveau de fronts de vêlage dans l'océan ont transporté d'énormes quantités de sédiments dans les étroits bassins marins situés entre les lobes de glace et les hautes terres. Avec la disparition de la glace morte, cette source de sédiments fut vite tarie.

Le retrait de la glace au Wisconsinien supérieur, depuis le versant est du plateau à l'ouest du lac Bluenose, débuta aux environs de 12 ka. La déglaciation, depuis cet endroit jusqu'au golfe Coronation, a été répartie en quatre phases de retrait définies en fonction de différentes configurations de l'écoulement glaciaire telles que déduites de l'orientation des drumlins, cannelures et stries, et d'après les positions du front glaciaire associées aux crêtes morainiques et aux autres structures de contact glaciaire.

SUMMARY

The Inman River area, as used in this report, is comprised between Dolphin and Union Strait and Rae River, and from Bluenose Lake to Coronation Gulf. The relief on Palaeozoic dolomite bedrock is low to moderate with generally marshy lowlands rising in stepped scarps to generally grooved, flat topped plateaus. This bedrock-controlled physiography predates the last glaciation and, in its essential configuration, is probably pre-Quaternary.

The monotonous flat-lying, thin-bedded, and buff-coloured Palaeozoic carbonates rest unconformably on the Middle to Late Proterozoic rocks of the Coppermine Homocline (Baragar and Donaldson, 1973) which outcrop along part of the southern fringe of the map-area.

The Wisconsinan drift sheet mantles the area, mostly with bouldery till, ranging in thickness from a thin, discontinuous veneer to several tens of metres. The Wisconsin Glaciation is represented by a single till sheet which is associated with several landform types. Pre-last glaciation sediments have been found only in the upstream section of Croker River where alluvial gravel and sand overlain by glaciolacustrine silty rhythmites, occur beneath the Wisconsinan till.

Using landforms and dominant sediment types, the Inman River area has been divided into nine morphosedimentary zones: A – Felsenmeer, kames, and silt mounds complex which was covered by glacial ice but not modified by glacial movement, it marks the limit of Wisconsinan ice advance in this area; B – Bluenose Lake Moraine complex is the result of stacking of drift, mostly till, over glacier ice by thrusting at the margin of the ice sheet; C – Inman River drumlin field composed of drumlins up to 8 km long is the result of expanding flow to the north-northwest to re-equilibrate the ice front profile following rapid calving of Amundsen Gulf lobe; D – Rae River glacial lake sequence, where washed till, glacial lake deltas, and rhythmites trace a sequence of glacial lakes in the basin of Rae and Richardson rivers; E – Upper Harding River moraine system is characterized by sinuous ridges with extensive areas of boulder fields and till mounds; it defines complex lobation at the margin of the retreating ice mass; F – Harding River drumlin field which comprises relatively short, in places crosscutting, drumlins and extensive surfaces of bare bedrock grooved by overriding debris laden ice; G – Rae River moraine belt is a series of massive east-west ridges built on the north side of the ice lobe which occupied the Rae and Richardson rivers basin; H – Marine limit kame complex is a massive deposit of gravel and sand that accumulated in a corridor between glacier ice and higher ground to the southwest by meltwater rushing to the encroaching sea to the northwest; I and J – Marine offlap sequence which comprises extensive areas of wave washed material, including drumlin fields and moraines. From marine limit deltas of gravel and sand, the material fines seaward eventually to silt-clay rhythmites. Beyond this sub-zone, the surface material is a bouldery lag of wave washed material and shingle beaches fringing the present shoreline.

SOMMAIRE

La région de la rivière Inman, comme elle est définie dans ce rapport, est comprise entre le détroit de Dolphin and Union au nord et la rivière Rae au sud et entre le lac Bluenose à l'ouest et le golfe Coronation à l'est. Le relief, contrôlé surtout par des roches dolomitiques du Paléozoïque, est de faible à modéré, passant, par une série d'escarpements, de basses terres généralement marécageuses à des plateaux structuraux. Les grands traits morphologiques de cette région, qui sont régis par le substratum rocheux, ont été acquis avant la dernière glaciation et la configuration d'ensemble de ce secteur est probablement antérieure au Quaternaire.

Les lits minces et subhorizontaux de la monotone série de roches carbonatées de couleur brun pâle du Paléozoïque reposent en discordance sur les roches du Protérozoïque moyen à supérieur de l'homoclinal de Coppermine (Baragar et Donaldson, 1973) qui affleurent à la bordure sud de la région cartographique.

Le drift du Wisconsinien recouvre la région d'une couche de sédiments dominée par un till rocailleux variant en épaisseur d'un mince placage discontinu à une couche de plusieurs mètres. La glaciation du Wisconsinien est représentée par une seule couche de till qui sous-tend un relief varié. Uniquement dans la partie amont de la rivière Croker a-t-on reconnu des sédiments antérieurs à la dernière glaciation : des alluvions de sable et gravier surmontées de rythmites silteuses glaciolacustres qui reposent sous le till wisconsinien.

En se basant sur les formes de relief et sur les principaux types de sédiments, la région de la rivière Inman a été divisée en neuf zones morphosédimentaires : A – le complexe de felsenmeer, de kames et de monticules de silt qui a été englacé mais non modifié par la glace active; cette zone marque la limite de l'avancée de l'inlandsis du Wisconsinien dans cette région; B – le complexe de la moraine du lac Bluenose qui résulte de l'empilement de sédiments glaciaires, principalement du till, par chevauchement à la marge de la nappe de glace; C – le champ de drumlins de la rivière Inman composé de drumlins qui atteignent 8 km de long et qui ont été sculptés par un écoulement expansif de la glace vers le nord-nord-ouest qui rééquilibra le profil du front glaciaire suite à un vêlage important dans le golfe Amundsen; D – la séquence glaciolacustre de la rivière Rae où du till délavé, des deltas glaciolacustres et des rythmites définissent une séquence de lacs glaciaires dans le bassin des rivières Rae et Richardson; E – le système morainique de l'amont de la rivière Harding caractérisé par des crêtes sinueuses et par de larges surfaces recouvertes de champs de blocs ou de monticules de till; cette zone identifie un front lobé lors de cette phase du retrait; F – le champ de drumlins de la rivière Harding composé de drumlins relativement courts et parfois superposés et de surface de roche nue à longues cannelures qui marquent l'orientation du glacier à la base chargée de débris; G – la ceinture morainique de la rivière Rae composée d'une série de crêtes massives, orientées est-ouest, construites à la bordure nord du lobe de glace qui occupait le bassin des rivières Rae et Richardson; H – le complexe de kame à la limite marine qui résulte d'une accumulation massive de sable et gravier dans un étroit corridor entre le glacier et les terres plus élevées au sud-ouest par des eaux de fonte qui s'écoulaient vers un bras de mer au nord-ouest; I et J – la séquence de régression marine qui comprend de vastes surfaces de matériaux délavés par les vagues, y inclus des champs de drumlins et des moraines. À partir des deltas de gravier et sable à la limite marine, le matériel est de plus en plus fin vers le large et aboutit à des rythmites de silt argileux. Au-delà de cette sous-zone, la surface est tapissée d'une blocaille, le résidu du délavage par les vagues, et par les dalles peu émoussées des crêtes de plage qui surplombent la plage actuelle.

Deglaciation of the Inman River area has been divided into four phases which record the history of a west-flowing ice sheet thrusting at its margin on the east slope of Melville Hills during the Late Wisconsinan. Calving in Amundsen Gulf provoked a drawdown and, as a result, an extending flow of ice to the north-northwest which moulded the Inman River drumlin field and thinned the western ice margin to stagnation. From then on the ice mass gradually assumed a two lobe configuration, one in Dolphin and Union Strait and the other in the depression of Rae and Richardson rivers, that prevailed to its final disappearance from the area. The sea penetrated in a narrow corridor along the south side of the Dolphin and Union lobe to eventually advance southwest of Cape Young into Coronation Gulf and into the valley of the Rae and Richardson rivers. Small moraine and drumlin fields record ice surges during this period.

The thrusting of till more than 5 m thick over glacier ice, drumlin fields which are often curvilinear and at times partly remoulded, and grooves and striae which record minor variations in ice flow direction constitute strong evidence for warm-base ice. This is in marked contrast with strong evidence for cold-base ice on Prince of Wales Island farther east (Dyke et al., 1992).

Flights of raised beaches form delicate scrolls along bedrock hills and drumlins fringing the coastline. Along with marine limit deltas and other sediments they record a limit of submergence ranging from 45 m a.s.l. near the mouth of Croker River to 165 m a.s.l. at Cape Kendall. Fine grained marine sediments are found only along a narrow band, generally less than 10 km wide near marine limit, supporting the view that the burst of sediment transported to the marine environment by meltwater was short-lived.

The establishment of permafrost conditions following deglaciation had several important consequences: it ensured the preservation of buried glacier ice; it created several pingos including the unique Harding River rock pingo; bedrock surfaces were pock-marked with mounds as a result of segregated ice growth; a reticulate pattern on a 10-30 m grid resulted from the expansion of joints in bedrock by ice wedge growth; it is responsible for various types of pattern ground and other minor landforms such as palsas.

After glacier ice had disappeared from the area the greatly diminished rivers and stream still managed to carve canyons in the dolomite bedrock depositing fans or deltas at their mouth.

Peat with known thickness of at least 3 m accumulated on poorly drained, fine grained marine sediments and in swales between drumlins. These are grazing areas for caribous and for muskoxen; ponds and small lakes associated with these wetlands are prime habitats for a wide variety of waterfowl including the beautiful Whistler swan.

Two elongated corridors of wind deflation and small dune construction are located along the coast of Dolphin and Union Strait, both record the predominance of winds from the northwest.

La déglaciation de la région de la rivière Inman a été répartie en quatre phases qui retracent l'histoire d'une partie de l'Inlandsis laurentidien à écoulement vers l'ouest avec chevauchement à sa marge sur le versant est des collines Melville au cours du Wisconsinien supérieur. Le vêlage dans le golfe Amundsen entraîne un déséquilibre du profil frontal provoquant un écoulement expansif vers le nord-nord-ouest qui sculpte le champ de drumlins de la rivière Inman et qui amincit cette marge glaciaire au point de stagnation. Dès lors, la calotte glaciaire adopte une configuration en deux lobes, l'un centré sur le détroit de Dolphin and Union et l'autre dans le bassin des rivières Rae et Richardson, qui sera maintenue jusqu'à la disparition de la glace de la région. L'incursion marine se fait le long d'un étroit corridor sur le côté sud du lobe de Dolphin and Union pour éventuellement progresser au sud-ouest du cap Young vers le golfe Coronation et le bassin des rivières Rae et Richardson. De petites moraines et des champs de drumlins enregistrent des crues glaciaires durant cette période.

Le chevauchement d'une couche de till de plus de cinq mètres d'épaisseur sur la glace de glacier, des champs de drumlins qui souvent dessinent de larges arcs de cercle et où, parfois, les drumlins se superposent, des stries et des cannelures qui témoignent de fluctuations mineures dans le sens d'écoulement sont des arguments en faveur d'un glacier à base en équilibre thermique. Ceci diffère de la situation plus à l'est, sur l'île Prince of Wales, où la base du glacier était froide (Dyke et al., 1992).

Des séries de crêtes de plages forment des arabesques autour des collines de roche en place et des drumlins le long de la côte. Avec les deltas à la limite marine et d'autres sédiments, ils enregistrent une émersion marine de 45 m près de l'embouchure de la rivière Coker qui atteint 165 m près du cap Kendall. Les sédiments marins à grain fin ne se trouvent que le long d'une étroite bande, généralement de moins de 10 km, près de la limite marine. Ceci suggère que la phase de sédimentation en milieu marin des débris transportés par les eaux de fonte fut de courte durée.

L'instauration du pergélisol a eu d'importantes conséquences : la préservation de la glace de glacier enfouie; la construction de plusieurs pingos dont le très particulier pingo rocheux de la rivière Harding; la présence de monticules produits par la formation de glace de ségrégation qui parsèment les surfaces du substratum rocheux; un réseau réticulé à trame de 10-30 m résultant de l'expansion de diaclases de la roche par croissance de coins de glace; et la formation de plusieurs types de sols à figure géométrique et autres formes mineures telles les palsas.

Après le retrait glaciaire, les cours d'eau conservent suffisamment d'énergie pour sculpter des canyons dans les dolomies et déposer des cônes de déjection ou des deltas à leur embouchure.

De la tourbe, avec des épaisseur connues de plus de 3 m, s'accumule dans les endroits mal drainés en particulier sur les sédiments marins à grain fin et dans les dépressions inter-drumlins. Ce sont là des aires de pâturage prisées par les caribous et les boeufs musqués; les mares et petits lacs de ces terres humides constituent un habitat de choix pour la faune aquatique en particulier pour les spectaculaires cygnes siffleurs.

Deux corridors de déflation éolienne et de petites dunes se trouvent le long de la côte du détroit Dolphin and Union, les deux indiquent des vents prédominants soufflant du nord-ouest.

INTRODUCTION

Purpose

This report and the accompanying map (1846A) constitute the final report on the Quaternary geology of the Bluenose Lake-Bernard Harbour area, herein called the Inman River area. The area is bounded to the west by 120°W, to the south by 68°N, to the east by Coronation Gulf, and to the north by Dolphin and Union Strait. The main purpose in studying the sediments and landforms of this region was to understand the Quaternary geology of the region and, particularly, the history of deglaciation, a period during which sedimentary and erosion processes left their most dramatic imprints on the landscape. The information resulting from this project, however, is useful also to engineering projects such as possible pipeline construction and it provides scientific data essential to the management of a proposed national park which would include the Bluenose Lake area.

The last major ice advance culminated on the east slope of the highlands west of Bluenose Lake. From there east to Coronation Gulf a spectacular sequence of glacial, periglacial, and marine events is recorded which allows for the confident analysis of processes. Because of the absence of a continuous vegetation cover, this region lends itself to the detailed study of landforms of all sizes and to the analysis of the stratigraphic record which chronicles their genesis. Mapping based on extensive foot traverses and air photo interpretation makes it possible to establish a sequence of events from last glacial maximum to the present. As a result, a better understanding of glacial, paraglacial, periglacial, and modern processes can be achieved which can then be applied in regions where exposures are more sparse, or to address the broader problems related to late Quaternary events in the northwestern sector of the Canadian Arctic including the Late Wisconsinan ice limit, the genesis of drumlin fields, the glaciomarine stratigraphy, the deglaciation sequence, and the origin of periglacial and modern landforms.

Fieldwork

The area was briefly visited in 1985 as part of a preliminary survey extending from Coppermine in Coronation Gulf to Paulatuk in Darnley Bay. A few key sites were visited and the feasibility of using abandoned DEW-line sites as base camps was determined.

Three seasons of field work followed in 1986, 1988, and 1989, each of six to seven weeks duration. The number of student assistants, both graduate and undergraduate, varied

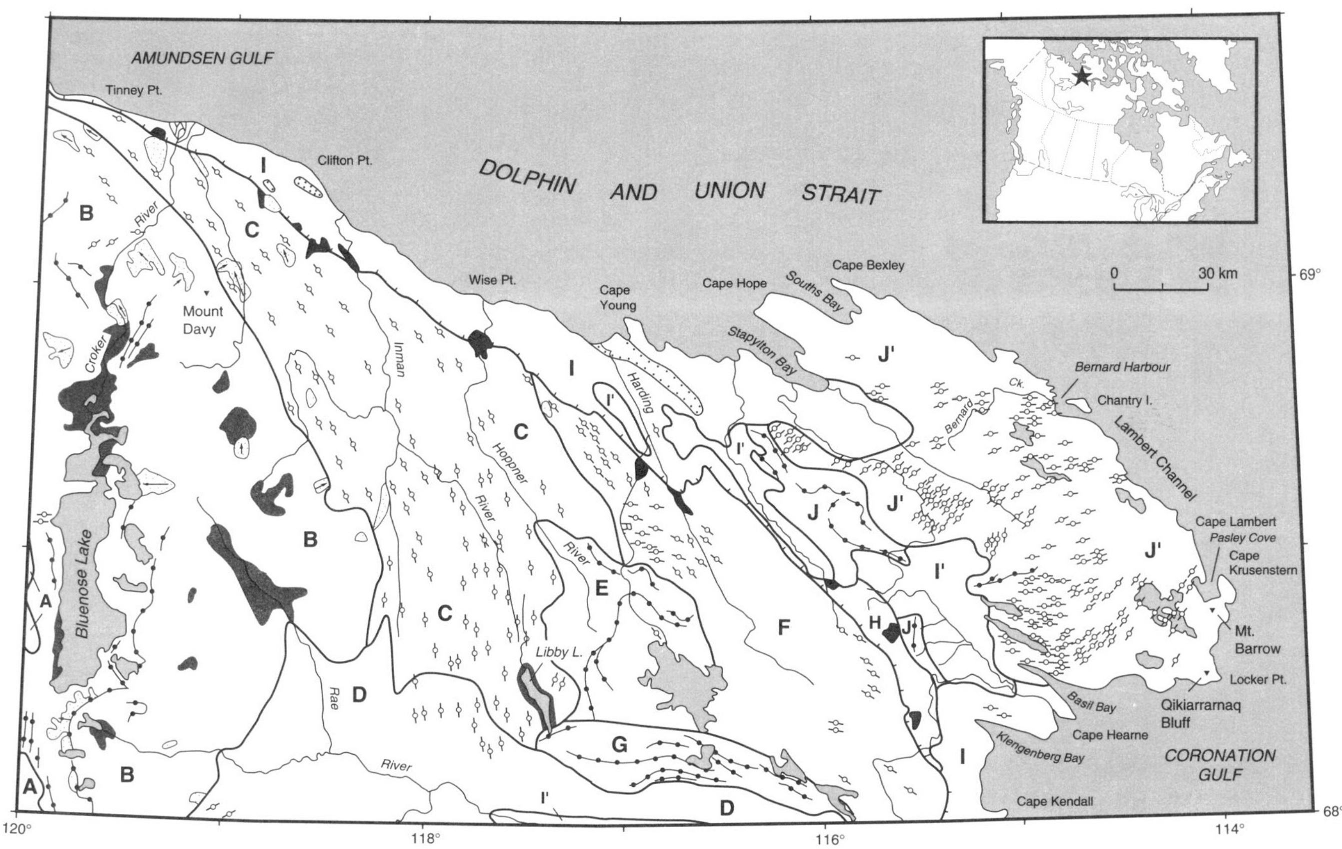

Figure 1. Inman River (Bernard Harbour-Bluenose Lake) region, location of places mentioned in text, and morphosedimentary zones.

from five to seven, including a volunteer from Germany. Their contribution to this report is enormous because of the theses they produced, from which this report borrows heavily.

Although some use was made of all-terrain vehicles (ATVs), it was realized in early 1986 that regardless where base camp was established, large rivers and lakes imposed severe restrictions on their use for systematic traversing. Instead, helicopter and float-planes were used for "set outs" and "pick ups". One or two persons were dropped off and picked up at pre-arranged spots. The traverse line based on detailed airphoto analysis usually covered 8 to 15 km depending on the complexity of the terrain. In three summers, over 100 such traverses were carried out covering virtually every major geological map unit within the study area. In addition, numerous specific features such as sections and moraines, drumlins, or marine sequences were studied in detail. As a result the confidence level in the final airphoto interpretation resulting in the surficial geology map is high.

LEGEND

Felsenmeer, kames and silt mounds complex	A
Bluenose Lake Moraine complex	B
Inman River drumlin field	C
Rae River glacial lake sequence	D
Upper Harding River moraine system	E
Harding River drumlin field	F
Rae River moraine belt	G
Marine limit kame complex	H
Marine offlap sequence	
Dolphin and Union Strait coastal area	I
Deep water facies	I'
Wave-washed morainic ridges	J
Wave-washed drumlin fields	J'
Morainic ridges	
Drumlin	
Glaciofluvial delta direction of flow	
Glaciolacustrine sediments	
Marine delta	
Eolian complex	

Legend for Figure 1.

Previous work

The Late Wisconsinan ice limit in the northwestern part of the Keewatin sector as shown on the 1987 Glacial Map of Canada (Dyke and Prest, 1987a) is still the subject of intense debate (Dyke and Dredge, 1989; Vincent, 1989, 1992). However, in the Inman River area (Fig. 1) there is strong evidence for a Late Wisconsinan age for the emplacement of sediments in the area from the Bluenose Lake Moraine to the marine offlap sequence (St-Onge and McMartin, 1987; McMartin and St-Onge, 1990). Authors do agree on the importance of the funnelling effect on ice flow imposed by the channel now occupied by Amundsen Gulf, Dolphin and Union Strait, and Coronation Gulf (Sharpe, 1992).

The coastline from the Mackenzie Delta to the mouth of Coppermine River was first explored by John Richardson, a member of Franklin's second overland expedition to the Arctic coast in 1825-27. Most of the geographical names along the coast were assigned by Richardson. "His detailed and precise notes on the flora, fauna, ice conditions, Eskimos and so forth have contributed substantially to the scientific knowledge of the north" (Mackay, 1958, p. 13).

Detailed observations of the Quaternary features of the region began nearly a century later by O'Neill (1924), a member of the Canadian Arctic Expedition in 1913-18. He collected numerous marine fossils, measured the orientation of striae, and suggested that: "At several localities along the coast in favourable places, terraces indicate that the emergence after maximum submergence of over 500 feet, was practically continuous and not intermittent, because the beach lines are so closely spaced vertically as to be almost continuous." (O'Neill, 1924, p. 33A).

The 1958 edition of the Glacial Map of Canada (Wilson et al., 1958) shows sketchy drumlin trends, some eskers, and moraine ridges east of Harding River, marine deposits at 170 feet (52 m) near Clifton Point and at 30 feet (9 m) near Bernard Harbour, and strandlines up to 180 feet (55 m) south of Dolphin and Union Strait.

Craig (1960) was the first to map glacial landforms such as drumlins, eskers, and morainic ridges. He also measured striae and determined that marine limit dips westerly from "480 feet above sea level just east of the Coppermine River, to 300 feet above sea level at long. 118°, and to 150 feet at long. 120°." (Craig, 1960, p. 6). Craig proposed that during early deglaciation the ice separated in two lobes; one from Victoria Island flowed to the southwest and deposited a series of end moraines in the Stapylton Bay area, the second flowing towards the northwest partly overrode and obliterated forms associated with the first lobe. He used crosscutting striae to support his hypothesis. Craig and Fyles (1960) likewise suggested that a small ice frontal moraine formed by a lobe from Victoria Island was overridden and fluted by the northwest flowing ice. Craig further suggested that the lobe, flowing out of Dolphin and Union Strait, coalesced with northwest flowing ice north of Klengenberg Bay. Finally he proposed that the northwest flowing ice eventually became a separate lobe confined to the Richardson River basin. Craig also provided the first ^{14}C date for the area, marine shells from 74 m a.s.l. yielded an age of 10 530 ± 160 BP [I GSC-25] (Table 1).

The 1968 version of the Glacial Map of Canada (Prest et al., 1968) incorporates Craig's (1960) data on the Bluenose Lake Moraine, as well as the morainic ridges north of Rae River. Drumlin trends are well represented, as is the approximate marine limit.

Based on a major study of the Quaternary geology of Banks Island, Vincent (1983, p. 86) believed that, at its maximum, the Early Wisconsinan ice was considerably more extensive than it was during the Late Wisconsinan. The distal margin of the Bluenose Lake Moraine was taken as the limit of Early Wisconsinan ice whereas the proximal side of the moraine was taken as the limit of Late Wisconsinan ice.

Mapping in the Richardson River basin, as part of a Master's thesis study, led Mercier (1984) to conclude that a glacial lake, which he named glacial Lake Richardson, had occupied the Richardson River and Rae River basins. He further suggested that the eastward retreating ice led to four distinct glacial lake phases, each controlled by a well-defined outlet and marked by numerous deltas. Deglaciation of the Richardson River and Rae River basins occurred prior to 10 300 ± 240 BP (GSC-3663), the age of the oldest shells collected in the area (Mercier, 1984).

A later study by Kerr (1987a) on lithofacies assemblages refined our understanding of the evolution of glacial Lake Richardson and led to a re-evaluation of ice frontal positions. Kerr (1987a, p. 2139) concluded that a sequence of overlapping glacial lakes "remained generally confined to the western end of the ice lobe occupying the basin". For each glacial lake phase, an outlet through the present study area is suggested. Clear evidence for several of these has not been found.

St-Onge and McMartin (1987) defined six morphosedimentary zones in the Bluenose Lake region and argued that the major moraine draped to the east and north of the lake, which they named the Bluenose Lake Moraine, was an ice thrust feature of Late Wisconsinan age. In a later study (McMartin and St-Onge, 1990), they suggested that the ice core of the moraine ridges is glacier ice buried by westward thrusting at the late Wisconsinan ice limit. In the same paper they divided deglaciation of the region into four distinct phases. A ^{14}C date on the Portlandia arctica in marine rhythmites establish that deglaciation was well advanced by 11 170 ± 80 BP [TO-1231] (Table 1).

Topography and physiography

The Inman River area is part of the Horton Plains, the northern subdivision of the Interior Plains region (Bostock, 1970). Small areas in the south and southeast along Rae River and including Cape Kendall, which are underlain by Shield rocks, are part of the Coronation Hills.

The Horton Plains east of 120°W is composed to two major elements: a bedrock-controlled, flat to gently rolling plateau of thick dolomite gently dipping northwest, and the rugged relief of morainic ridges, hills, and kames of the Bluenose Lake Moraine.

The dolomite plateau gently rises from Coronation Gulf at a rate of 1-3 m/km to an undulating upland at 150-200 m a.s.l. which occupies the area east of Harding River valley. Westward from the rectilinear valley occupied by the east arm of Harding River, the upland rises up to 300-350 m a.s.l. to the foot of the Bluenose Lake Moraine. This vast plateau is thinly mantled by glacial material including till (mostly drumlinized), moraine ridges, outwash, and marine deposits. Bedrock outcrops are common and many display the effects of periglacial processes such as pingos, rock mounds, felsenmeer, and other effects of frost shattering. The plateau is bordered by generally gentle slopes rising northward from Rae River valley, westward from Coronation Gulf and southward from Dolphin and Union Strait. Along the coastline of Dolphin and Union Strait subvertical bedrock escarpment up to 60 m high convey a degree of ruggedness to an otherwise gently undulating landscape. The southeast trend of major valleys reflects jointing or faults in the bedrock, the most conspicuous example being the Harding River valley (Map 1846A, in pocket). The innumerable lakes, on the other hand, tend to closely reflect the extensively drumlinized topography with one major exception being a large, unnamed lake at the headwaters of Harding River.

Twenty to twenty-five kilometres west of the Inman River, the land rises abruptly from an extensively drumlinized surface at 250-300 m a.s.l. to a hilly region with summits averaging 650-700 m a.s.l. and reaching as high as 741 m a.s.l. This area of hills, ridges, and lakes is the Bluenose Lake Moraine (St-Onge and McMartin, 1987) and is identified as the Melville Hills on the 1:50 000 topographic maps. On smaller scale maps, Melville Hills seem to refer to a portion of the moraine belt between Bluenose Lake and Mount Davy (Fig. 1). On the 1:1 000 000 scale map, the name identifies the upland area west of Bluenose Lake; it is in this last sense that the name is used in this report.

Bluenose Lake is a large body of water extending 55 km north-south with an irregular eastern coastline so that its width varies from a few tens of metres in some narrows to over 11 km at its widest, averaging 5-7 km. Strangely enough, it was only discovered as a result of tri-metragon photography taken by the RCAF on July 14, 1948. Arthur J. Shana of the Geodetic Survey established five position fixes at the northern end of the lake in July 1949. Dr. J.P. Kelsall of the Canadian Wildlife Service and an assistant circumnavigated the lake in 1953 to study birds and mammals. As a Nova Scotian he proposed the name Bluenose for this "last major discovery in the geography of mainland Canada" (Sebert, 1974). Bluenose Lake drains northward into Amundsen Gulf through the narrow, 80 m deep, canyon cut by the Crocker River. Smaller lakes nestled between steep-sided hills do not display preferred orientations, they are often silt-laden as a result of mudflows from bordering hills. West of Bluenose

Table 1. Radiocarbon age determinations.

	Lab. Number Field Number	Latitude Longitude Elevation	Uncorrected Age Corrected Age[1]	Material Taxa	Locality
1.	GSC-4374 SV(K) 86-56	69°20'07" 119°58'52" 45 m	570 ± 60 570 ± 60	wood *Picea*	9 km west of Tinney Point
2.	GSC-4368 SV(A)-86-27	68°38'04" 119°23'12" 600 m	4760 ± 90	peat	northeast of Bluenose Lake
3.	GSC-4390 SV-86-7	69°08'23" 118°41'34" 61 m	10 700 ± 100 10 700 ± 100	marine shells *Hiatella arctica*	8.5 km south-southwest of Clifton Point
4.	GSC-4425 SV(A)-86-11	69°11'32" 118°38'49" 12-14 m	8890 ± 110 8890 ± 110	marine shells *Clinocardium ciliatum*	3 km south-southwest of Clifton Point
5.	GSC-4424 SV-86-5	69°09'32" 118°38'17" 25 m	10 300 ± 100 10 400 ± 100	marine shells *Hiatella arctica*	6 km southeast of Clifton Point
6.	GSC-4402 SV(M)-86-20	69°08'05" 118°35'45" 47 m	10 300 ± 100 10 300 ± 100	marine shells *Hiatella arctica*	9 km south-southeast of Clifton Point
7.	AECV-473Cc SV(K)-86-4	69°07' 118°31' 45 m	9600 ± 140	marine shells *Hiatella arctica*	Dolphin and Union Strait coast
8.	AECV-712Cc SV(K)-86-3	69°07'31" 118°30'55" 30 m	9300 ± 540	marine shells *Hiatella arctica,* *Macoma balthica*	Dolphin and Union Strait coast
9.	TO-536 SV(K)-86-19	68°44'21" 116°57'14" 95 m	 9610 ± 110	marine shell *Cyrtodaria kurriana*	23 km south of Cape Young
10.	AECV-713Cc SV(K)-86-22	68°44' 116°56'17" 70 m	10 420 ± 540	marine shells *Mya, Hiatella arctica*	Dolphin and Union Strait coast
11.	AECV-474Cc SV(K)-86-20	68°44' 116°56' 80 m	10 040 ± 240	marine shells *Macoma*	Dolphin and Union Strait coast
12.	I(GSC)-25	68°47' 116°56' about 74 m	10 530 ± 260	marine shells	Harding River area
13.	AECV-403Cc SV(K)-84-42A	67°48' 116°40' 73 m	9800 ± 140	marine shells *Hiatella arctica,* *Mya truncata*	Dolphin and Union Strait coast
14.	GSC-4917 SV(M)-89-4	68°39'21" 115°46'40" 99 m	9820 ± 80 9850 ± 80	marine shells *Hiatella arctica*	19 km south of Stapylton Bay

[1]Ages corrected to 0.0‰ for shells and -25.0‰ for organic samples.

Table 1. (cont.)

	Lab. Number Field Number	Latitude Longitude Elevation	Uncorrected Age Corrected Age[1]	Material Taxa	Locality
15.	GSC-4926 SV(M)-89-26	68°48'15" 115°39'12" 85 m	10 100 ± 90 10 100 ± 90	marine shells *Hiatella arctica*	12 km south of Stapylton Bay
16.	AECV-717Cc 88K-28a	68°45' 115°38' 15 m	7250 ± 180	marine shells *Clinocardium ciliatum*	Dolphin and Union Strait coast
17.	AECV-646Cc 88K-28	68°44' 115°37' 18 m	5660 ± 110	marine shells *Mytilus edulis*	Dolphin and Union Strait coast
18.	GSC-4916 SV(M)-89-10	68°22'41" 115°35'17" 100 m	10 800 ± 100 10 700 ± 100	marine shells	64 km north-northwest of Coppermine
19.	GSC-4727 SV-88-7	68°27'58" 115°34'34" 100 m	9930 ± 100 9920 ± 100	marine shells *Hiatella arctica*	74 km north-northwest of Coppermine
20.	TO-1231 SV(M)-88-11	68°14'17" 115°28'59" 125 m	11 170 ± 80	marine shells *Portlandia arctica*	46 km north-northwest of Coppermine
21.	GSC-4849 SV-88-29	68°39'31" 115°26'46" 80 m	9670 ± 150 9690 ± 150	marine shells *Hiatella arctica*	25 km southeast of Stapylton Bay
22.	GSC-4709 SV-88-15	68°19'03" 115°22'28" 100 m	9160 ± 160 9190 ± 160	marine shells	55 km north-northwest of Coppermine
23.	GSC-4848 SV-88-35	68°27'26" 115°21'23" 95 m	9680 ± 80 9700 ± 80	marine shells *Hiatella arctica*	41 km southwest of Bernard Harbour
24.	GSC-4915 SV(M)-89-6	68°14'42" 115°20'06" 115 m	9510 ± 80 9520 ± 80	marine shells *Hiatella arctica*	7 km northwest of Klengenberg Bay
25.	GSC-4747 SV(M)-88-67	68°11'30" 115°18'47" 105 m	9620 ± 170 9640 ± 170	marine shells *Hiatella arctica*, *Mya arenaria*	44 km north-northwest of Coppermine
26.	GSC-4930 SV(M)-89-45	68°24'42" 115°15'52" 105 m	9450 ± 120 9480 ± 120	marine shells	17 km northwest of Basil Bay
27.	GSC-4823 SV(M)-88-68	68°33'25" 115°13'32" 148 m	6980 ± 90 7040 ± 90	lake sediment	76 km north of Coppermine
28.	GSC-4842 SV(M)-88-69	68°32'32" 115°12'48" 105 m	7030 ± 110 7010 ± 110	lake sediment	77 km north of Coppermine

	Lab. Number Field Number	Latitude Longitude Elevation	Uncorrected Age Corrected Age[1]	Material Taxa	Locality
29.	GSC-4846 SV(P)-88-12	68°30'33" 115°10'53" 100 m	9610 ± 80 9610 ± 80	marine shells *Hiatella arctica*	34 km southwest of Bernard Harbour
30.	GSC-4696 SV(M)-88-61	68°27'33" 115°05'54" 111 m	9520 ± 80 9540 ± 80	marine shells *Hiatella arctica*	70 km north of Coppermine
31.	GSC-4749 SV-88-25	68°32'32" 115°05'40" 105 m	9610 ± 90 9620 ± 90	marine shells *Hiatella arctica*	76 km north of Coppermine
32.	GSC-4845 SV(A)-88-13	68°18'29" 114°59'49" 100 m	9110 ± 120 9120 ± 120	marine shells not identifiable	2.5 km north of Basil Bay
33.	GSC-4847 SV(M)-88-58	68°29'12" 114°59'32" 97 m	9500 ± 80 9530 ± 80	marine shells *Hiatella arctica*	33.5 km southwest of Bernard Harbour
34.	AECV-472Cc SV(K)-84-11	67°40' 114°33' 120 m	9560 ± 130	marine shells *Mya truncata*	Coronation Gulf coast
35.	AECV-404Cc SV(K)-86-62	67°40'30" 114°16'30" 120 m	9080 ± 150	marine shells *Hiatella arctica*, *Mya truncata*	Coronation Gulf coast

Lake the land rises gently from the shore line at 561 m a.s.l. to a high plateau of Cambrian and Precambrian rocks west of the map area (Jones et al., 1992).

The general configuration of plateaus and uplands is certainly pre-Quaternary. Results of our study do not provide any significant additional information to that included in the detailed and cogent discussion on the origin of pre-Quaternary physiographic evolution of the Canadian Arctic by Dyke et al. (1992, p. 16-19).

Bedrock geology

The Bernard Harbour/Bluenose Lake region is underlain by a thick sequence of subhorizontal sedimentary rocks, mostly carbonates (dolomites) of Paleozoic age which unconformably overlie Shield rocks (Fig. 2; Fraser et al., 1960). Precambrian rocks which outcrop along the Rae River southwest of Libby Lake and in the Cape Kendall area are part of the Bear Province of the Canadian Shield. They consist of northward dipping (2°-5°) sedimentary rocks, mostly sandstones, of the Rae Group which are "profusely intruded by Coronation Sills...assumed to be of Hadrynian in age" (Baragar and Donaldson, 1973, p. 3). The gabbro rocks of the Coronation Sills and dykes are responsible for the spectacular cuesta of Cape Kendall and of most of the islands in Coronation Gulf.

The Rae Group is unconformably overlain by essentially flat-lying carbonates of Paleozoic age. "Typical exposures are platy and flaggy bedded...with...fresh exposures ranging from greyish buff to pinkish buff...weathered surfaces are buff to yellowish buff" (Baragar and Donaldson, 1973, p. 12). Steep-sided valleys of modern rivers and/or meltwater channels show the best outcrops of this rather monotonous bedrock (Fig. 3).

Climate and soils

The entire Bernard Harbour area lies north of the 10°C July isotherm and therefore belongs to the polar tundra zone which is underlain by continuous permafrost. According to the Köppen climatic classification system it has a "Dfc" climate and in the Troll system a "II,2" one.

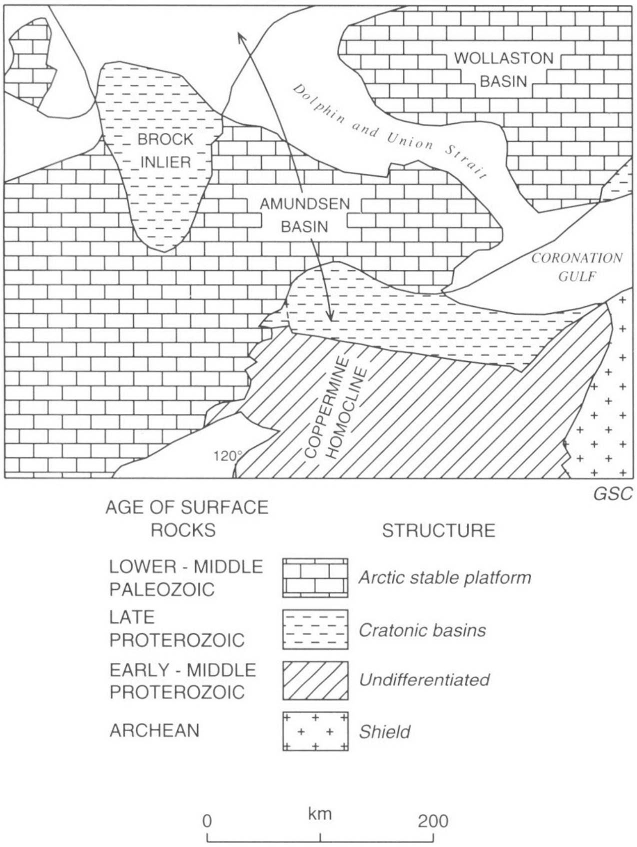

Figure 2. *Regional lithostratigraphic elements of Dolphin and Union Strait area (after Jones et al., 1992, p. 8).*

Weather records are available from Coppermine (67°50′N, 115°07′W, 9 m a.s.l.) to the south, and at Cape Young (68°56′N, 116°55′W, elevation 18 m), a DEW-line station west of the study area (Atmospheric Environment Service, 1982).

The two stations experience similar weather patterns (Fig. 4); both have long, cold, dry winters and brief, cool, damp summers. Freezing temperatures can occur in any month. On average, the frost-free period ranges from 50 to 60 days. In the summer of 1989 temperatures as high as 32°C were recorded.

Snow covers the ground for at least eight months of the year. By the end of June in 1986, 1988, and 1989 snow had stopped falling but could be expected to begin again by September. The greatest snowfall occurs in October and November, 68.4 cm (Cape Young) and 100.7 cm (Coppermine) annually. The greatest amount of precipitation falls in the summer and early autumn. On average, there are only 75.3 mm (Cape Young) and 89.4 mm (Coppermine) of rainfall during the months of June, July, and August, the growing season. The Bernard Harbour area may be classified as a cold desert.

In terms of soils the whole study area belongs to the so-called "Tundra zone" (Charlier, 1969, p. 1992) or "Cryiosolic Order soils...formed in either mineral or organic materials that have permafrost either within 1 m of the surface or within 2 m if more than one-third of the pedon has been strongly cryoturbated..." (Agriculture Canada, 1978, p. 63). Due to the persistent cold climate, pedogenesis is very limited. The slow rate of plant decomposition is reflected in the very thin humus layer (organic horizon – H). Tundra soils, in general, do not have a continuous profile, most are a combination of thin-layered sandy clay and humus. The surface is sparsely covered by lichens and mosses. According to Tedrow and Cantlon (1958, p. 178) and Charlier (1969, p. 1992) the soil in the study area can be described as "Arctic Brown".

Acknowledgments

We are indebted to student assistants, who have contributed significantly to the observations and discussions which led to the concepts contained in this report: R. Avery, H. Beaudet, D. Kerr. M. Potschin, and J. Samson. Their meticulous field observations and keen interest were essential stimulants. Helicopter support from the Polar Continental Shelf Project ensured efficient field logistics. Helicopter and fixed-wing pilots improved the quality of life in the "luxury" accommodations of YSO and Drumlin City. To Rod Stone, the competent and patient radio expeditor, a special thanks for keeping a sense of humour while catering to the often strange requests from numerous field camps. The GSC Dating Laboratory, directed by Dr. R.N. McNeely contributed essential age analyses displayed in elegant computer generated tables and maps. Suggestions by R.G. Blackadar polished the language, and comments by the critical reader, Dr. A.S. Dyke, forced us to clarify our interpretations thus significantly improving the original draft of this paper. Ms. T. Barry transformed rough sketches into artistic computer figures. The word

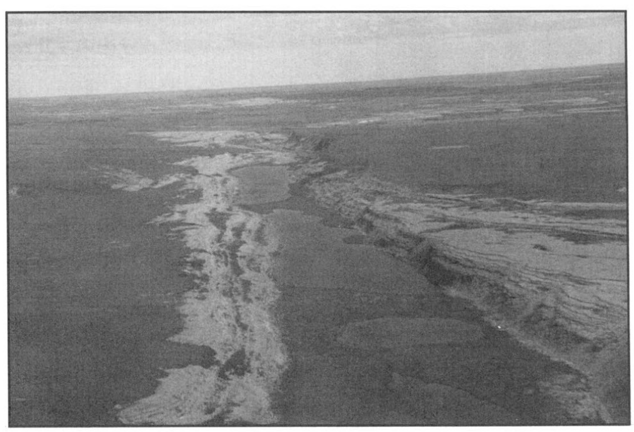

Figure 3. *Low level oblique air photograph of fluvioglacial channel carved in flat-lying dolomite bedrock, 15 km northwest of Cape Kendall; average channel depth is 50 m and width is 350 m. View is downstream towards northwest. GSC 1993-158J*

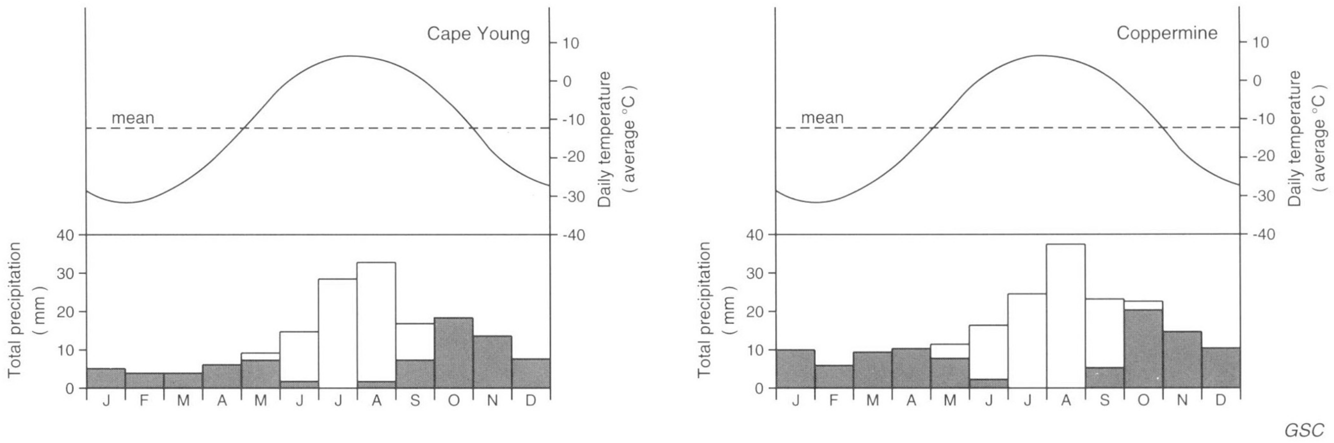

Figure 4. Climatic data for Cape Young and Coppermine, Northwest Territories, (Atmospheric Environment Service, 1982).

processing expertise of Marg Herzog and of Chantal Bélanger, both of Polar Continental Shelf Project is greatfully acknowledged.

MAP UNITS

In this section of the report the bedrock lithologies and the Quaternary sediments and landforms in the Inman River area are briefly described. Each corresponds to a map unit shown on the Surficial Geology map. This "extended legend" is intended to help the reader gain a better understanding of what is included within each unit. Since rock stratigraphic units are seldom homogeneous, it is important to understand the degree of inhomogeneity or, to put it another way, the amount of grouping that is included in each unit. In a later section the emphasis will be on how the authors have interpreted the genesis of surficial materials and landforms.

Although the legend is based on stratigraphic principles it must be realized that most events are time transgressive (i.e., the marine off-lap sequence) or that different processes can be active at the same time (i.e., ice marginal outwash in one area and marine delta construction in another). Thus there is considerable overlap from a time sequence point of view between many of the map units.

Bedrock (R)

Bedrock outcrops are common and extensive in the map area. Along Rae River the Proterozoic sandstone and carbonate with occasional gabbro sills have been eroded into spectacular cuestas with a gentle slope dipping northward and a steep south-facing exposure of dark columnar gabbro over light-coloured sandstone and/or dolostone. The rest of the map area is underlain by flat-lying, heavily jointed carbonate rocks, mostly dolomite, of Paleozoic age. Outcrops occur in upland areas which have been scraped clean by flowing glacier ice or along major valleys of which the Crocker River canyon is the most spectacular. On the more resistant dolomite beds, grooves and striae are very well preserved resulting in strikingly moulded surfaces (Fig. 5).

There is no clear indication of the importance of glacial erosion on the bedrock surface. The lack of lithologic variety precludes the use of differential erosion to estimate values. Pre-Wisconsinan steep sided bedrock valleys at right angle to glacial flow and bedrock scarps with sharp angles suggest that erosion in bedrock was minimal and that glaciation mostly entrained pre-existing debris.

Morainal deposits

Unsorted or poorly sorted glacial debris mantles extensive parts of the map area. Till is usually the base unit of the Quaternary stratigraphic sequence. In spite of its extent, good exposures are rare because erosion by fluvial or marine processes tends to concentrate the coarser elements at the surface resulting in open textured bouldery lags instead of clean sections.

Figure 5. *Striated groove and crosscutting striae on glacially moulded dolomite bedrock, 9 km southwest from the head of Stapylton Bay; striated groove towards 305° crosscut by younger striae towards 275°. GSC 204982-F*

Because of the uniformity of the underlying bedrock, till is fairly uniform in texture and composition albeit somewhat sandier with a greater frequency of Precambrian gabbro boulders in the southern part of the map area. Composition of clasts ranges from greater than 45% dolomites (up to 97%), 10-20% metamorphics, with gabbro and sandstone providing the balance (McMartin, 1990). The matrix of the till, generally, is a silty sand supporting clasts ranging in size from gravel to boulders, one metre or more in diameter.

Till, washed (gT)

Till washed refers to surfaces covered with coarse gravel and boulders. Commonly this open textured material is a lag resting directly on bedrock as a result of the washing out of fines principally by meltwater flow and by glacial lake and current activity. Outcrops are common in this unit. Occasionally, particularly when associated with drumlins, the bouldery lag mantles till (Fig. 6).

Till veneer (Tv)

Areas where the till cover is too thin to mask the underlying small scale bedrock topography are mapped as till veneer, a discontinuous cover of diamictic material seldom reaching 2 m in thickness. Small bedrocks scarps and other outcrops are common, as are boulders. Unit Tv tends to occur on bedrock highs or on bedrock slopes. Drumlins are understandably rare in this unit but large grooves 1 m or more in diameter and hundreds of metres in length carved in the thin till cover or on the underlying bedrock are not uncommon.

Till blanket (Tb)

Areas mantled by a continuous cover of till and associated drift to a thickness that completely obscures underlying bedrock structures are mapped as Tb. In the Inman River region, areas of till blanket are generally extensively drumlinized (Fig. 7). Relative to R and Tv units, Tb occurs in low areas. Wherever it was studied, either in natural or man-made exposures, this material is a light brown to buff, stony diamicton of cobbles and boulders supported in a silty sand matrix. The consistency of the fabric of the matrix-supported clasts, the texture, and fissility suggest that it was emplaced as a lodgement till while the sparse lenses of well sorted sand indicates that some meltwater flow was occurring at the base of the ice mass. Although occasionally present between drumlins, bedrock outcrops are rare in these areas and, where they occur, striae are parallel to the long axes of adjacent drumlins.

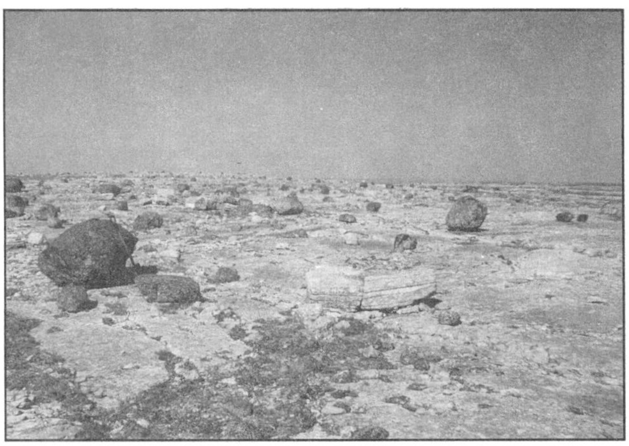

Figure 6. Erratics, mostly dolomite and gabbro, of varied size strewn on glacially polished dolomite bedrock; typical washed till (gT) map unit; 40 km south of Stapylton Bay and above marine limit. GSC 1993-158F

Till, hummocky (Th)

The westernmost part of the map area contains a thick cover of till, outwash, and associated drift with measured thicknesses of over a 100 m. The unconsolidated surficial material in the ridges, hills, and hummocks of this region is a diamicton of gravel and boulders in a silty sand matrix. Many of the hills have a steep-sided cap of sand and gravel outwash, too limited in extent to be mapped; outwash material occurs in small patches throughout the hummocky till area (Fig. 8).

Glaciofluvial and morainal deposits (Gx)

Discontinuous patches of outwash, sand and gravel, small ridges of boulders, and bare bedrock characterize this unit which tends to occur in corridors commonly, but not necessarily, along esker ridges. It is distinct from washed till because of the more frequent occurrence of patches of well sorted sand and gravel and, particularly, of ridges of boulders. Although the washing away of fines from till by meltwater is an important process, sedimentation of outwash sand and gravel in ice crevasses and in other ice proximal situations played an important role in the emplacement of this unit.

Glaciofluvial deposits (Gh)

Glaciofluvial deposits consist of well sorted sand to coarse gravel deposited by meltwater in ice-proximal situations (Fig. 9). These often massive deposits can result from sedimentation within ice caverns and result in eskers up to 50 m high and several tens of kilometres long. Eskers tend to be rare within Tb and Th units and more frequent in Tv and gT. As will be discussed, this merely reflects the nature of glacier ice in its waning stages; debris laden stagnant ice will produce eskers, relatively clean ice will not.

At the ice-land contact, outwash streams deposited broad sheets, 2 to 10 m thick, of boulders, cobbles, gravel, and sand. The material of these kames, even the boulders, is generally well rounded and becomes finer with distance from the ice contact face which is generally steep and often affected by collapse structures. Outwash sheets occur extensively within the Bluenose Lake Moraine, particularly in the northwest part of the map area.

Mount Davy is a unique and spectacular case of a mega-kame. With a summit at 531 m a.s.l., it dominates the surrounding plain by over 100 m. On clear days it is visible from the coast at Clifton Point just over 30 km to the north-northeast. Exposures along gulleys show tilted blocks of well stratified gravel, sand, and silty sand. The presence of silty clay beds implies ponding in the sedimentary sequence. Surprisingly, there are no large eskers leading to or from this massive kame which argues in favour of sediments being carried by supraglacial streams to a very large moulin in the ice. Mount Davy is certainly a candidate for the world's biggest kame (Fig. 10).

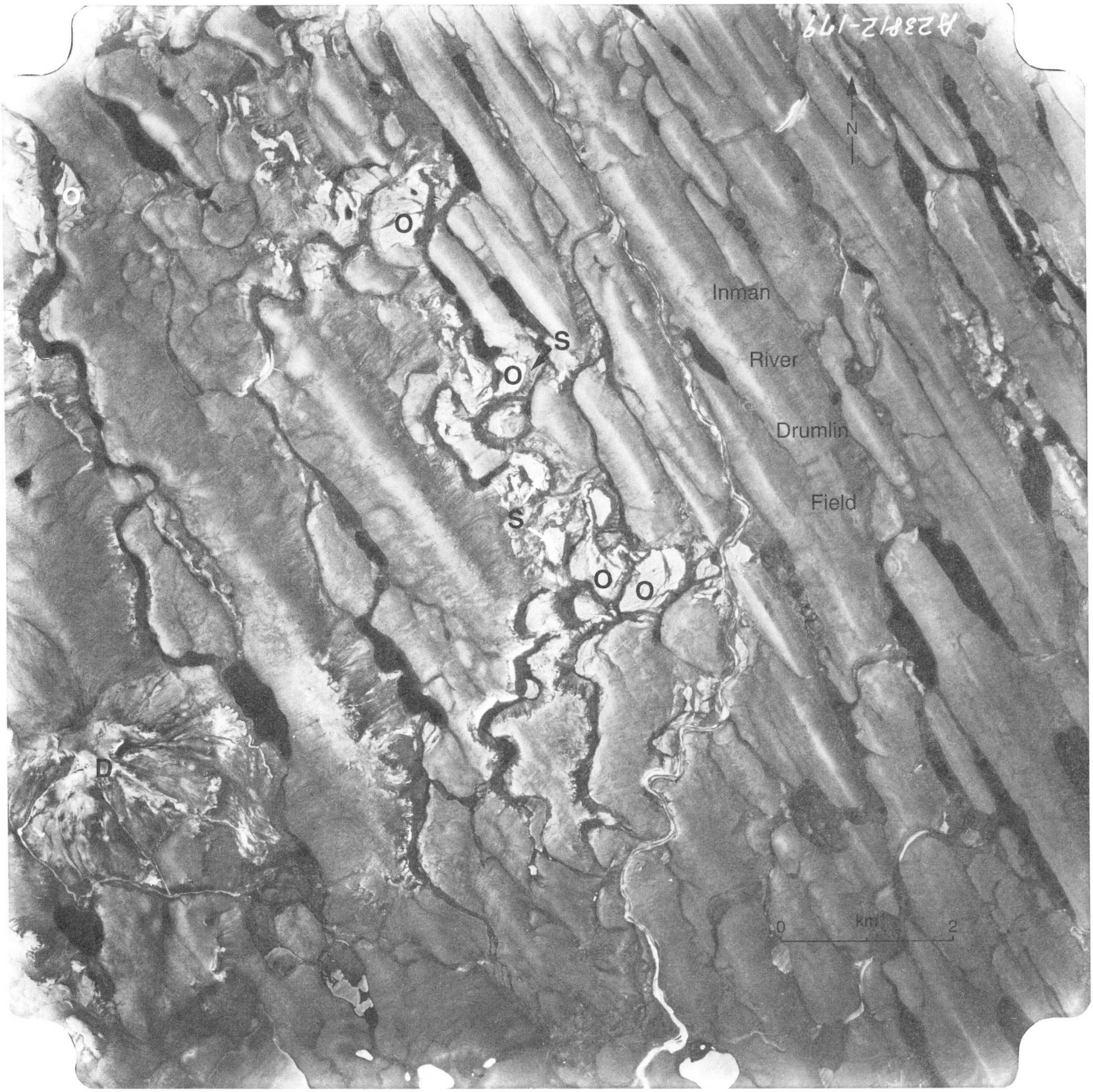

Figure 7. *Part of morphosedimentary zone C (Fig. 1) south of Clifton Point; Mount Davy (D) in southwest corner (Fig. 10); southeast-northwest drumlins are typical of the large, whaleback-shaped forms of the Inman River field; note that the steep northeast facing slope of the numerous outwash fans (O) at the mouth of meltwater channels are often scarred by slumps (S) (NAPL A23812-179)*

Glaciolacustrine deposits

During phases of ice retreat from the Bluenose Lake Moraine region, ponding occurred between the ice front and high ground to the west. As a result, glaciolacustrine sediments in the form of rhythmites, which, in some cases, coarsen upward into deltaic sands, are found in the western half of the map area and in the Rae River basin to the south.

Typically, glaciolacustrine sediments tend to be fine silty sand for the "summer" layer and silty clay for the "winter" layer. However, interlayered with these rhythmites, beds of diamicton 20 to 50 cm thick are common in some of the glaciolacustrine basins (Fig. 11).

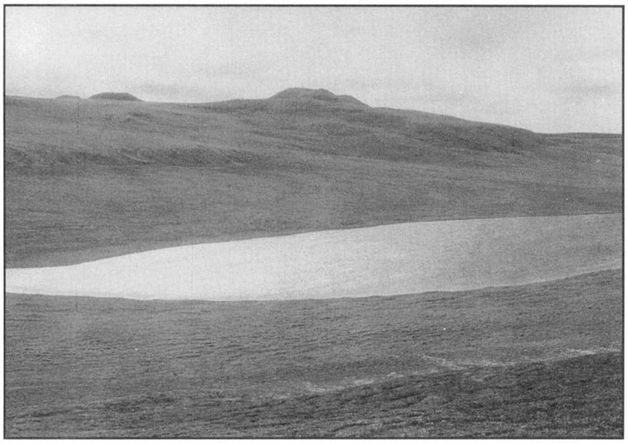

Figure 8. *Massive hummocks, part of Bluenose Lake Moraine, capped with steep sided, commonly collapsed kames of bouldery gravel and sand; view west, 20 km east of Bluenose Lake and 9 km northwest of figures 23 and 24; elevation from lake to highest hill is 20 m. GSC 1993-158I*

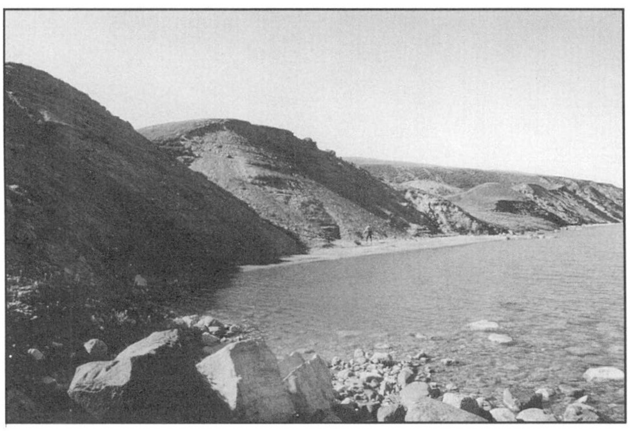

Figure 9. *Ridge of outwash sand and gravel at the south end of Libby Lake; this kame complex was deposited at the northern margin of the ice lobe occupying the Rae/Richardson rivers basin during deglaciation phase III (Fig. 54). GSC 1993-121D*

Quiet water sediments (Lp)

Areas where glaciolacustrine sediments, i.e., rhythmites and associated sediments, are thick enough to mask topographic details of the underlying material, usually till, are shown as Lp on the map. This blanket of sediments, usually whitish grey on airphotos (Fig. 12), is greater than 1 m thick and reaches 10 m in measured sections. It represents sediments deposited in ice contact or ice dammed lakes which lasted tens of years based on rhythmite counts.

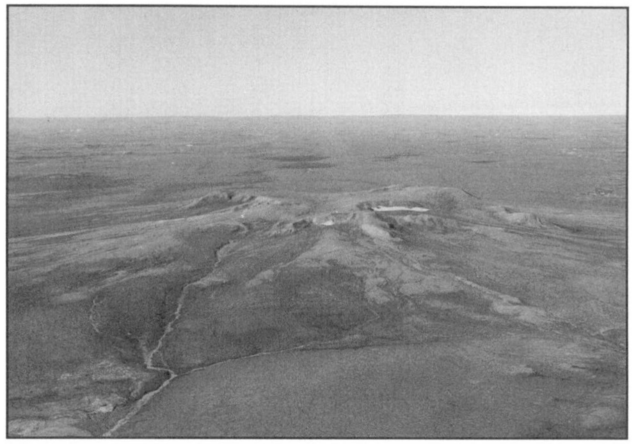

Figure 10. *Low level oblique airphoto of Mount Davy (Fig. 7); solifluction in fine sand and silt has eroded numerous amphitheatre-shaped scars in the rim of the flat top of this very large kame; view northwest. GSC 1993-158H*

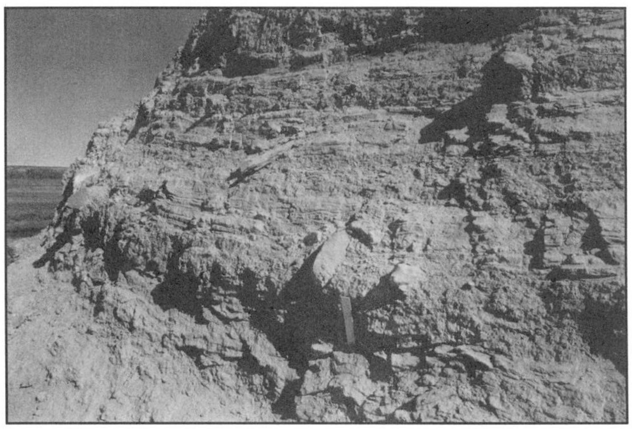

Figure 11. *Debris flow diamicton interbedded with silty rhythmites, typical sedimentary sequence in ice contact lakes within the Bluenose Lake Moraine complex.*

Figure 12. *Upstream section of Croker River: hummocky terrain and kames in western part of airphoto, note poorly defined northwest-southeast ridges and lineations; light grey areas with thermokarst ponds west of Croker River are sandy-silt rhythmites (R) of glacial Lake Bluenose; extensively channelled deltas (D) on either side of valley were emplaced in the very early phases of the draining of the glacial lake prior to the incision of the modern valley; (S) locates section in Figure 27. (NAPL A23812-154)*

N
Croker
River
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2

Nearshore sediments (Lv)

In the shallow nearshore of large glacial lakes or in the basins of ephemeral lakes, sandy silt forms a discontinuous, thin, usually less than 1 m thick, mantle over underlying material. Glaciolacustrine nearshore sediments do not display well defined rhythmites; although bedding is present, it is often lenticular and/or contorted.

Littoral sediments (Lr)

Abandoned spits and beaches of glacial lakes only occur around Bluenose Lake where terraces of crossbedded sand and gravel identify higher lake levels. Maximum thickness could not be determined but does exceed 2m. This unit is limited in extent.

Deltaic sediments (Ld)

Channelled deltas of sand and gravel mark the former level of glacial lakes on the east side of Bluenose Lake, in glaciolacustrine basins within the Bluenose Lake Moraine, and along the western and northern limit of the glacial lake sequence which occupied the Rae River basin.

The large and extensively pitted deltas at the south end of Bluenose Lake testify to sedimentation at the ice front over and around masses of inactive ice which resulted in the collapse of delta material around kettle holes following ice melt (Fig. 13). These delta sediments exceed 30 m in thickness; at 615 m a.s.l. their surface is 55 m above present lake level. At the northeast end of Bluenose Lake there is a large sand delta, also at 615 m a.s.l., with abundant evidence of collapse at its northern, southern, and eastern margins. Feeding into it from the east is a complex of pitted and collapse sand ridges, the remnant of an intricate system of channels weaving within masses of inactive ice which carried large amounts of sediments to construct a delta in glacial Lake Bluenose.

Within the Bluenose Lake Moraine, several basins opening easterly were occupied by glacial lakes ponded by ice in the Rae River and Inman River basins. At the head of these relatively small lakes (10 to 180 km^2), channelled and terraced deltas of gravel and coarse sand mark former lake levels.

In the southwestern part of the map area in the Rae River basin, numerous deltas of bouldery gravel are arranged in tiers lowering southward. Evidence of collapse towards the south and east as well as the paucity of fine sediments at elevations below the deltas, indicate that the gravel was deposited at the mouths of meltwater streams debouching into relatively fast flowing water bodies at the margin of the ice lobe still occupying the basin of the Rae and Richardson rivers (Fig. 14).

Marine deposits

Glaciomarine deposits cover a broad range of sediments: bouldery gravel to clay. As the encroaching sea penetrated the area along narrow embayments wedged between ice lobes occupying the lowlands and the uplands still covered by stagnant ice, initial sedimentation was ice contact bouldery gravel. With receding ice, arms of the sea became larger and deeper so that finer sediments could be deposited in the increasingly deep and quiet marine basin. Within the map area, the marine limit is below 70 m a.s.l. at the western margin and at approximately 160 m near Cape Kendall.

Wave-washed material (gM)

A lag of coarse sand, gravel, and boulders resulting from the winnowing of fines by marine waves and currents on material of diverse origin including till, glaciofluvial sediments, and shattered bedrock, is an extensive unit particularly in the eastern part of the map area where large drumlin fields have been extensively subdued, and in part remoulded, by this process. The gravelly lag, although extensive, is seldom over 1 m thick, and is commonly fossiliferous with paired bivalves found in the coarse sand/fine gravel fraction; in coarser material only shell fragments are found. Raised beaches occur in this unit, most commonly on north- or west-facing slopes.

Linear grooves lacking systematic orientations are a common feature in the wave washed material. They are interpreted as the markings left by iceberg keels dragging in the ocean bottom during early deglaciation time. Grooves are 150 to 500 m in length, 1 to 2 m wide and 10 to 50 cm deep. They are a common occurrence below marine limit in Arctic areas (Hélie, 1983; Dyke and Dredge, 1989).

Quiet water sediments (Mp)

There are two main facies (McMartin, 1990) in this unit: 1) an interlaminated sand, silt, and clay and, 2) a massive silty clay which commonly overlies facies 1. Unit Mp is fossiliferous and contains several species of marine ostracods, foraminifera, and bivalves. Dropstones occur but are not abundant. Maximum thickness is up to 40 m. Extensive gullying resulting in badland topography is common in this unit (Fig. 15). Flat areas away from river valleys are generally covered with peat and marshy vegetation. This unit occurs in bedrock basins which are close to the marine limit; similar bedrock basins over 20 km away from the marine limit generally do not contain any appreciable amount of fine grained marine sediments.

Nearshore sediments (Mm)

The interbedded sand, silt, and clay of this map unit occasionally contain organic-rich strata of twigs, leaves, and rootlets. Ice rafted boulders and patches of gravel are common on surfaces which are frequently strewn with marine shells. Thickness is variable but averages about 5 m.

Littoral sediments (Mr)

Subangular to well-rounded cobbles and shingles in openwork deposits or in a matrix of medium to coarse sand form spectacular flights of raised beaches in more exposed areas such as points and capes (Fig. 16). In more sheltered parts such as bays, this unit may contain silty sand and horizons of

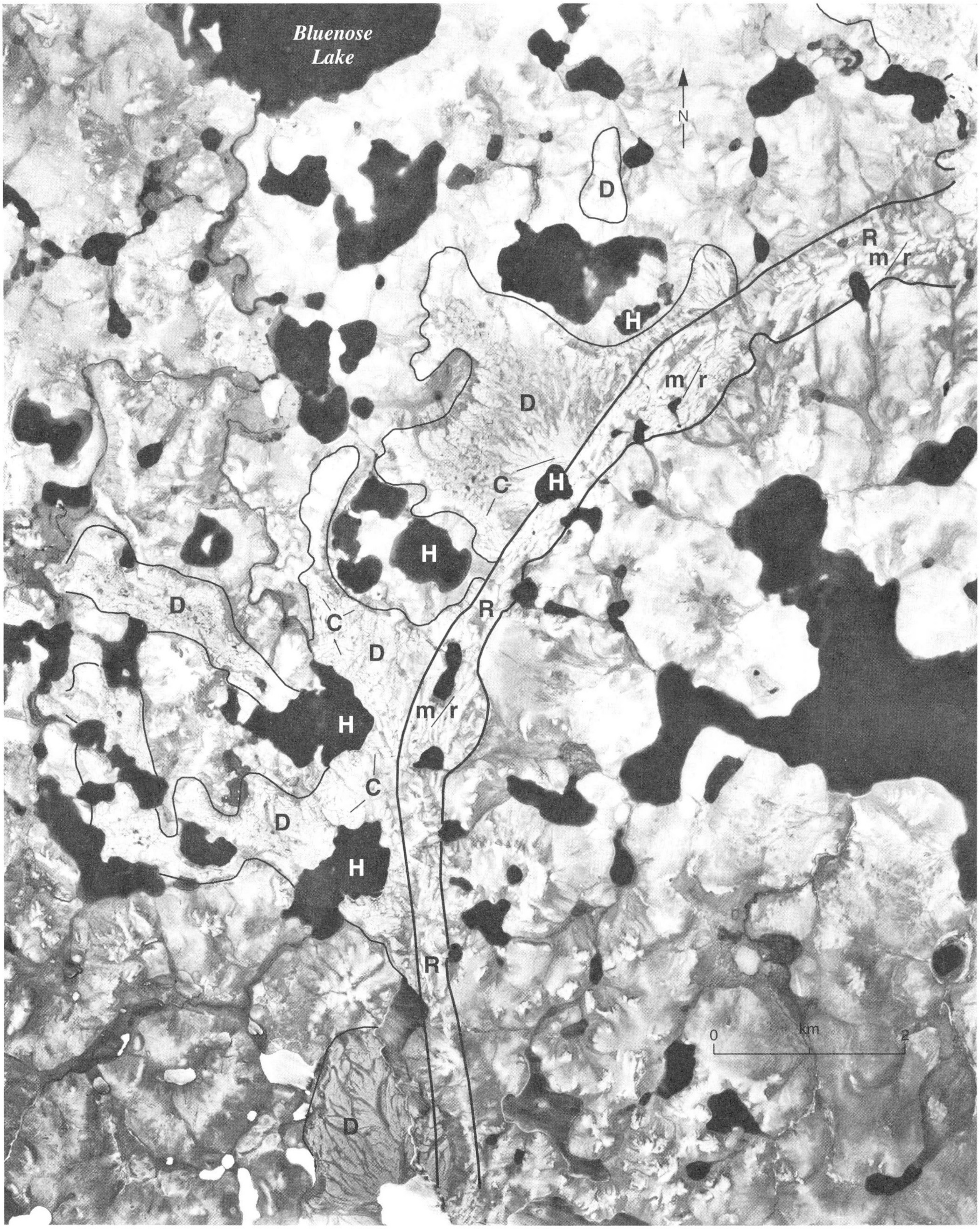


Figure 13. Part of morphosedimentary zone B (Fig. 1) south of Bluenose Lake; west of a large moraine ridge (R), extensively channelled deltas (D) are pitted by numerous kettle holes (H) rimmed by collapse ridges (C) and associated fracture zones; numerous minor ridges (mr) on main moraine are thought to reflect thrust planes (Fig. 21B). (NAPL A24079-156)

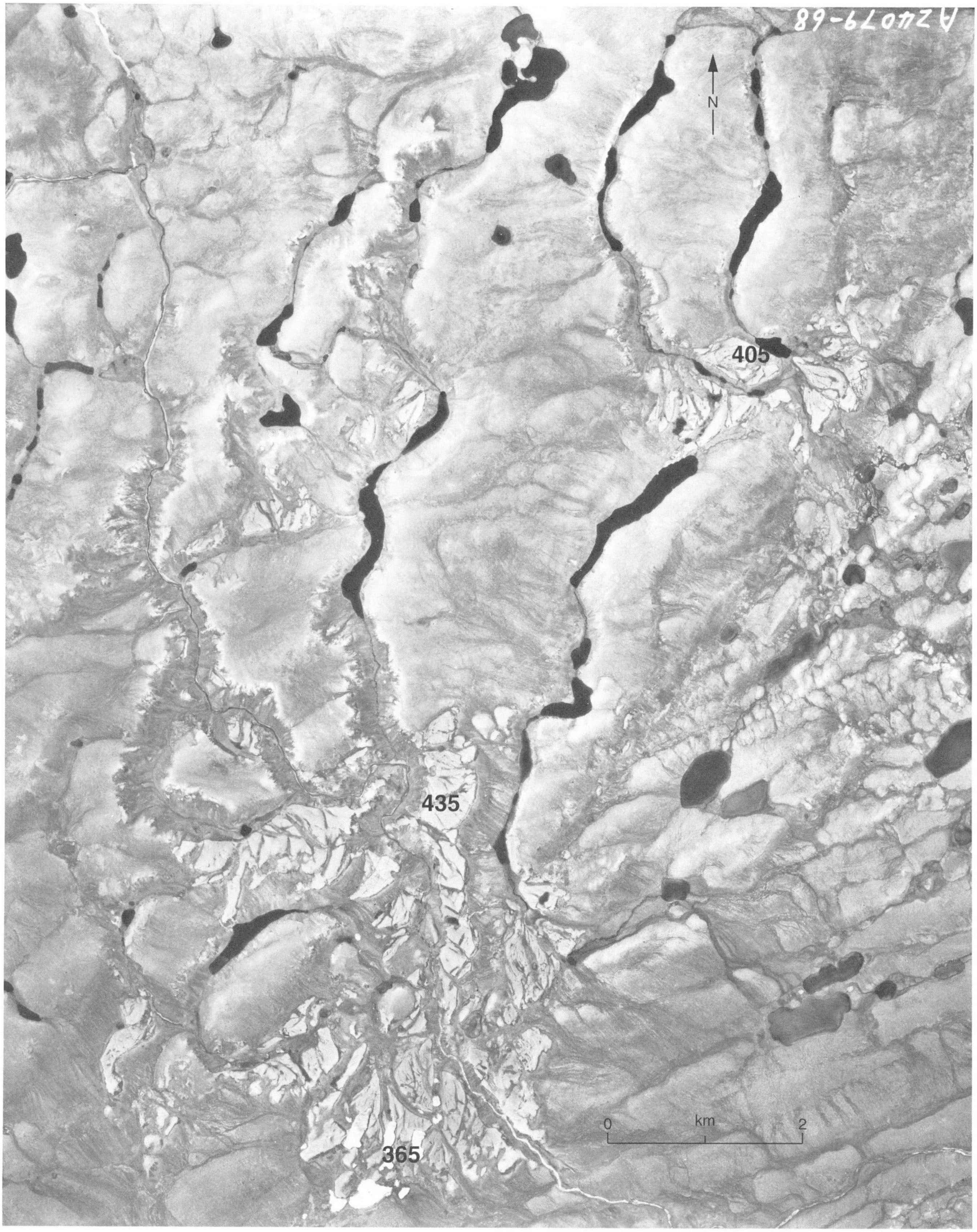


Figure 14. *Series of north-south meltwater channels now occupied by narrow lakes and south-flowing tributary creeks of Rae River. The sequence of terraced and extensively channelled bouldery deltas, ranging in elevation from 435 m to 365 m a.s.l., was built into a narrow body of water marginal to the ice lobe occupying the Rae/Richardson rivers basin. Note that there are no glaciolacustrine sediments on the drumlinized plain below and to the southeast of the deltas. (NAPL A24079-68)*

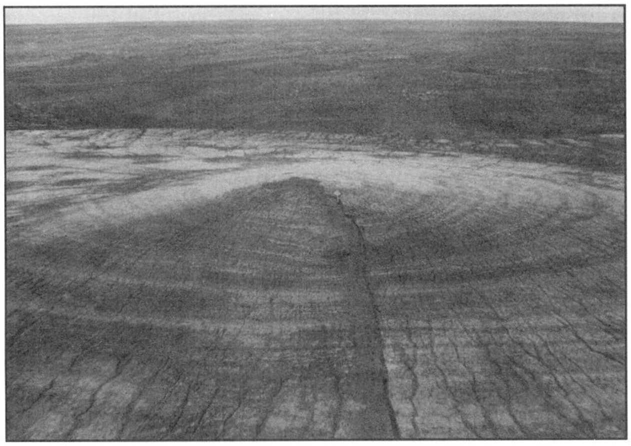


Figure 15. Low level oblique aerial photograph of a 9 m high hill of rhythmically bedded, fine grained, fossiliferous marine sediments. Rill and gullies flow from just below the vegetated apex at 61 m a.s.l. View south, 15 km northwest of Basil Bay. GSC 1993-1580

crudely bedded gravel. Thickness is highly variable and difficult to measure because of paucity of good exposures but, probably seldom exceeds 5 m

Deltaic sediments (Md)

At the upper limit of post-glacial marine incursion, delta complexes were constructed by water flowing from melting stagnant ice further inland. These sediment-laden streams deposited, at their mouths, large quantities of sand, gravel, and boulders. Paleochannels are well preserved on delta surfaces (Fig. 17). Material fines from bouldery gravel at the apex to coarse sand at the delta front. Maximum thickness of over 30 m has been documented (Kerr, 1994, p. 25).

Alluvial deposits

Stream deposited sediments are not extensive relative to other units within the map area. Seasonally fast-flowing rivers are mostly confined in narrow bedrock valleys or have boulder beds rather then extensive alluvial reaches.

Terraced sediments (At)

Terraces of cross-stratified sand and rounded gravel, which, in some instances grade downstream to deltaic sediments, occur along the lower reaches of some streams, notably the Harding River and the river draining at the head of Stapylton Bay. In rare cases, within the Inman River Drumlin Field in particular, terraced sediments may be a distal facies of outwash. Typical thickness is between 3 to 5 m but it can be as much as 10 m.

Floodplain sediments (Ap)

Gravel to gravelly sand occurs as an occasional deposit in floodplains along major rivers such as the Inman and the Hoppner. Thicknesses greater than 1 m can be determined along cutbanks of modern streams but maximum thickness is not known. Fine to medium sand at the surface is often removed by strong winds, generally from the northwest, leaving a surface covered with a gravelly lag.

Felsenmeer (F)

Felsenmeer (Fig. 18), the angular platey debris resulting from in situ disagregation of bedrock is difficult to map because, on air photographs, it is often indistinguishable from unmodified bedrock (R) or from thin till (Tv). Although it is common as patches too small to map, large areas only occur on bare upland where glacial sedimentation was limited or absent. As a unit, felsenmeer is closely associated with bedrock and thin till; the limits between these units tend to be gradational. The angular rock debris in felsenmeer ranges in size from gravel to boulders; rock mounds which will be discussed later, although not limited to this unit, are a common feature, as are isolated boulder-size erratics.

Eolian deposits (E)

Sand-blown features including blowouts and a variety of dunes are extensive and spectacular in the coastal region south of Dolphin and Union Strait (Fig. 19). Ventifacts are not common. There are two major features resulting from eolian activity in this region. One, south of Clifton Point, extends 5.5 km in a northwest-southeast direction and has a maximum width of 1.5 km. Strong winds from the northwest mobilize sand and silt from fossiliferous marine nearshore sediments and redeposit it behind tufts of vegetation, partly stabilized small dunes, and at the edge of the blowout area.

Figure 16. Beach ridges of angular dolomite shingles on the south side of the triangular-shaped bedrock hill called Qikiarrarnaq Bluff (Fig. 1); the photograph was taken at 68 m a.s.l. looking east-southeast from point A in Figure 39. GSC 1993-121A

Figure 17. Marine offlap sequence south of Clifton Point; bedrock (R) and till veneer (Tv) and drumlinized till (Tb) occur southwest of the channelled gravel and sand marine limit delta (Md), to the northeast is a classic sequence of: washed drumlinized till (gT), sandy silt/clay (Mp), sandy nearshore sediments (Mm), and gravelly beach ridges (Mr); northwesterly winds are reworking silty sand into an arrow head shaped blowout and small dune complex (Ae) also shown in Figure 19. (NAPL A19428-83)

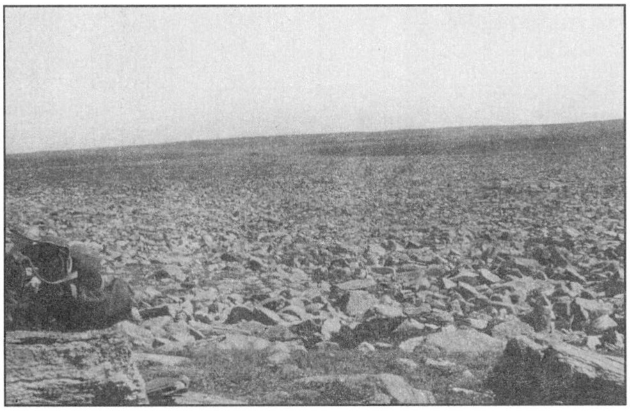


Figure 18. *Field of angular dolomite boulders and rare rounded erratics of gabbro; view northwest, 12 km west of Klegenberg Bay; typical of Felsenmeer (F) map unit. GSC 1993-158G*

The downwind leading edge of dunes, up to 5 m high, is encroaching on a flight of gravelly beach ridges (Avery, 1988).

A similar, but larger feature, is found south of Cape Young. From the modern delta of Harding River it extends southeastward for 25 km over a width of 2 to 3 km. This blowout/dune complex is easily visible on satellite imagery (see Fig. 31). A succession of small blowouts and parabolic dunes, 2 to 5 m in height, result in a hummocky complex bordered by linear, partly vegetated dunes at the margin and at the leading edge. Although indicated by a stiple pattern on Craig's (1960) map, it is not discussed in the accompanying text.

There are several smaller eolian features in the region, mostly in coastal areas, where strong winds winnow out the fines from modern beach sediments. Cliff top dunes too small to map are found on top of sections undergoing active erosion by modern streams.

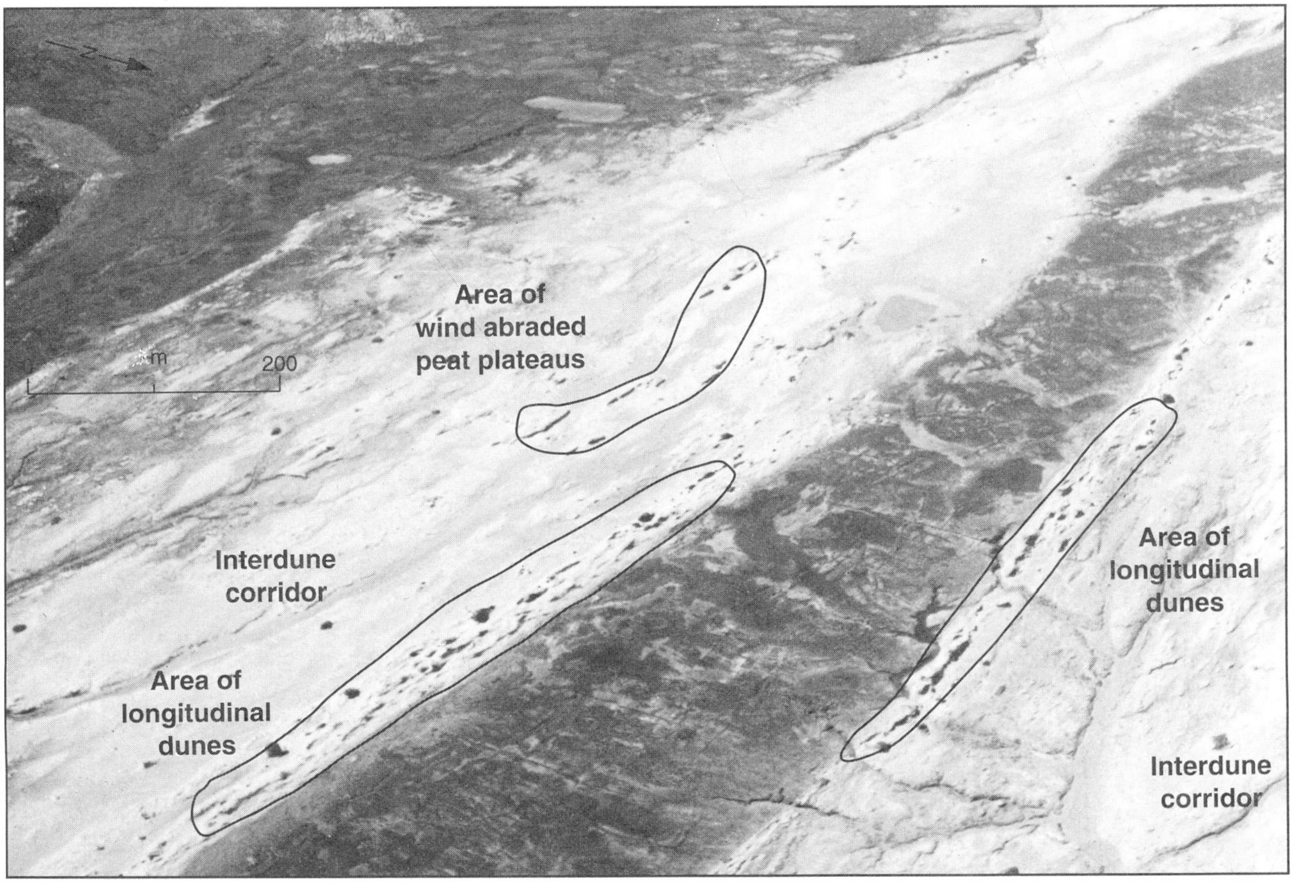


Figure 19. *Low level oblique airphoto view of part of the eolian complex south of Clifton Point (Fig. 17). View west. GSC 1993-121F*

Organic deposits (O)

Peat covers extensive areas, in particular low-lying areas, within the Inman River region. Accumulation of sphagnum with disseminated twigs reach thickness of 2 m but usually occur as a veneer 50 cm thick or less. Only the thicker deposits, obliterating the nature of the underlying material, is shown on the map.

MORPHOSEDIMENTARY ZONES

In order to discuss the genesis and interrelationships of major landforms and sediment complexes, the Inman River region has been divided into major morphosedimentary zones (Fig. 1). These morphostratigraphic units are arranged from oldest to youngest i.e., from west to east, however, it is not possible to adhere to a strict chronostratigraphic sequence because events overlap in time and space.

Felsenmeer, kames, and silt mounds complex, (zone A)

The southwest corner of the map area (Fig. 1) is part of the southeastern flank of a broad bedrock plateau to the west of Bluenose Lake, which rises to over 700 m a.s.l. and is referred to as the Brock Inlier. It is underlain mostly by carbonate rocks ranging in age from Proterozoic for the highest part of the plateau to Lower Palaeozoic on the flanks (Jones et al., 1992).

Within the map area, the generally flat to gently rolling surface is mantled by boulders of dolomite and sparse granite and gabbro erratics. Speckled on this surface are numerous steep-sided conical hills of bouldery gravel, 5-15 m high, interpreted as moulin kames (Fig. 20). Low mounds of sandy silt 3-5 m high form an apron around the coarser material. A 1 to 3 m thick deposit of outwash sand and gravel cover a small part of this area; the steep slopes to the southeast and to the east are scarred by collapse features which indicate sedimentation in contact with the western front of an ice mass.

The presence of a few erratics, kames, and associated glaciolacustrine sediments and outwash leaves no doubt that zone A was covered by glacial ice. Sediments and associated landforms strongly suggest that the last ice was either inactive for the most part, or was cold based. The Brock Inlier upland, west of Bluenose Lake, rises above 780 m a.s.l. In all current models of the Laurentide Ice Sheet this upland is thought to have been above the level of active ice flow during the Wisconsinan (Vincent and Prest, 1987; Dyke and Prest, 1987c). However it must be noted that current knowledge of the Quaternary geology of the area is based on limited field data (Craig, 1960; Klassen, 1971).

The association of mounds of sandy silt with kames is interpreted as resulting from sedimentation of coarse material within moulins in glacier ice too thin to imprint signs of active movement (Fig. 21A). As the moulin became clogged with bouldery gravel water, flow was severely restricted so that only fine sediments continued to be deposited in the englacial water body. The present subdued relief associated with these fine grained sediments is the result of collapse and solifluction following ice melt rather than a reflection of original sedimentation surface.

Bluenose Lake Moraine complex (zone B)

Within morphosedimentary zone B, the Croker River which drains Bluenose Lake has, for the first 30 km of its course, re-excavated a bedrock valley that predates the last glaciation. As a result sections up to 48 m high are exposed which show a complete glacial cycle starting with the pre-last glaciation river alluvium at the base (see Fig. 27). The alluvial deposits are overlain by sand grading to silt to sandy silt rhythmites deposited in a proglacial lake formed when the advancing ice blocked drainage to the north. The rhythmites are in turn overlain by a massive diamicton containing large boulders of dolomite, gabbro, quartzitic red sandstone, and various types of gneiss. This is believed to be a lodgment till deposited at the base of the last ice sheet when it overrode the glaciolacustrine sediments. Thin silty clay and clay rhythmites and 20-30 cm thick gravelly diamicton beds rest on the till and represent the early phase of a glacial lake. Massive sedimentation in this basin is represented by sandy foreset beds of an advancing delta which coarsen upward to top set beds of gravel. Although bedrock was not seen at the base of the section illustrated in Figure 27, it is visible in several other exposures so that there is no doubt that the coarse alluvial gravel at the base of the section is in fact the lowermost unit of the sequence and that it rests on bedrock. So far the lack of datable material does not make it possible to estimate when the ice reached this position or how long it stayed there.

In the southwestern part of the map area, the limit between zones A and B (Fig. 1) is marked by a steep sided ridge of bouldery gravel and sand 20-30 m high with an east-facing slope that is systematically steeper than the west (M in Fig. 20). The interpretation is that the ridge was deposited by meltwaters and debris flows which dumped englacial and supraglacial debris in a trench that developed within the ice. The trench is thought to correspond to zones of weakness along thrust planes which marked the margin of active ice (Fig. 21B).

All of zone B is underlain by thick layers, in some places over 100 m, of glacial, glaciofluvial, and glaciolacustrine debris. In a well exposed section the till, a bouldery diamicton is more than 3 m thick, has a sandy to sandy silt matrix, columnar jointing and prismatic fissility, and is often prone to mudflows (Fig. 22 and 23).

The most common landforms are boulder-covered hills, up to 60 m high, and hummocks interspersed with numerous lakes. This massive 20-30 km wide arcuate belt of "hummocky moraine" (Prest, 1967, p. 11), marks that part of the ice margin which got buried and preserved as an ice-cored moraine that became detached from the wasting ice sheet.

At the western edge of morphosedimentary zone B, massive ridges, up to 100 m high, of bouldery till delimit an ice frontal position which skirts Bluenose Lake to the east and north. Major ridges are either massive or a composite of numerous smaller ridges which show as linear longitudinal

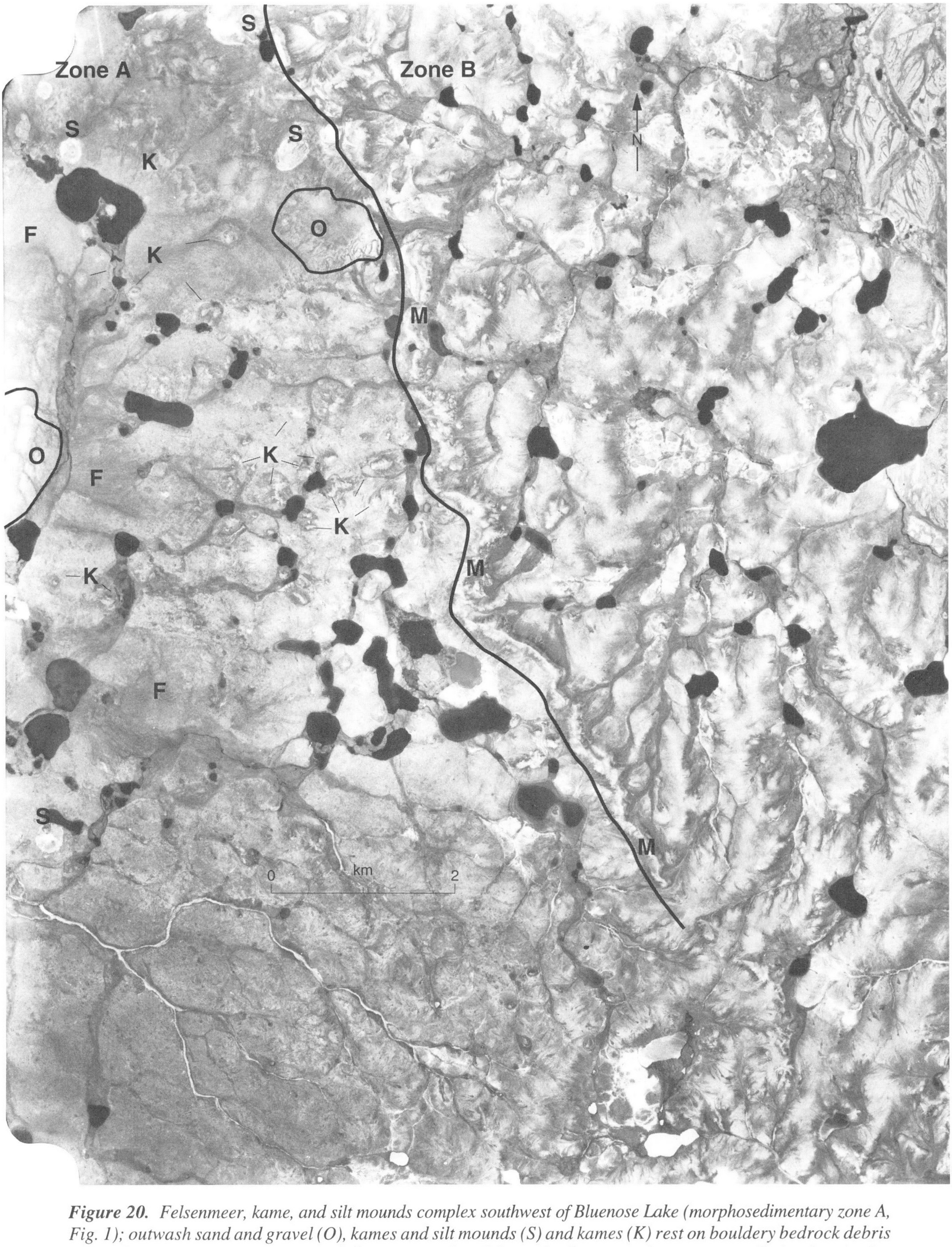


Figure 20. *Felsenmeer, kame, and silt mounds complex southwest of Bluenose Lake (morphosedimentary zone A, Fig. 1); outwash sand and gravel (O), kames and silt mounds (S) and kames (K) rest on bouldery bedrock debris (F) and sparse erratics; a steep sided moraine ridge (M) marks the approximate western limit of active ice during the last glacial maximum. (NAPL A24079-73)*

pattern on their crest (Fig. 13). This is thought to be the surface expression of steeply dipping, stacked layers of coarse bouldery diamicton emplaced by extensive thrusting up of basal material along the climbing flowlines at the margin of the ice sheet. During stagnation the debris was redistributed to the point that linear patterns reflecting thrusting planes were nearly obliterated, except in the case of the outermost ridges.

One additional feature of the Bluenose Lake Moraine deserves reporting in more detail. Within this arcuate belt of drift, commonly thicker than 30 m, several slumps have recently exposed ice-rich sediments overlain by massive diamicton. This was first mentioned by St-Onge and McMartin (1987, p. 95) but, at that time, the underlying material was not sufficiently exposed to allow speculation on the origin of the icy sediments. The slump, first observed in 1986 (and shown in Fig. 10.4 in St-Onge and McMartin, 1987), was re-examined in 1988 and again in 1989. Between 1986 and 1988 the slump had enlarged substantially (Fig. 23). It now exposed a 25 m high section comprising 15 to 20 m of icy sediments overlain by 2 to 5 m of silty sand and bouldery diamicton (Fig. 24). The icy sediments exhibit folding and

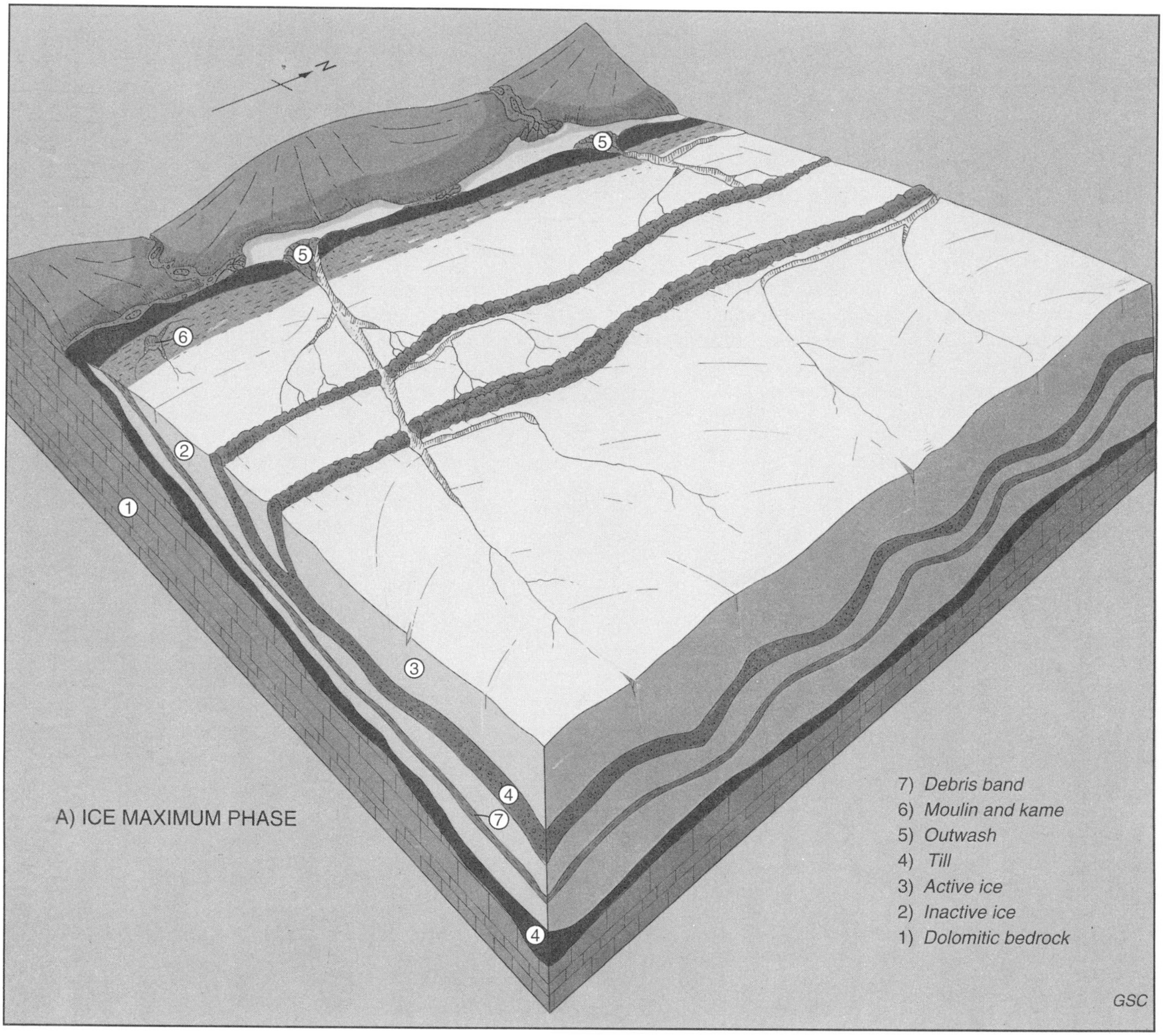


Figure 21. Block diagrams suggesting genesis of landforms in morphosedimentary zone A (felsenmeer, kames, and silt mounds complex) and zone B (Bluenose Lake Moraine) southwest of Bluenose Lake (Fig. 1). ***(A)*** *At the Late Wisconsinan glacial maximum, an apron of inactive ice rests on felsenmeer-covered bedrock. In the early stages of ice retreat the ice front is affected by thrusting which builds a moraine ridge (M in Fig. 20); outwash is deposited at the ice margin and within moulins; small ponds occupying ice collapse depressions around moulin kames fill with silty sand.*

complex deformations, and numerous large boulders, cobbles, and pebbles occur in the silty, icy sediment. Some layers also display small subhorizontal sheared ice lenses. The upper contact of the icy sediments and the bouldery diamicton is sharp, planar, and subhorizontal. Reticulate ice, developed in the matrix of the diamicton overlying the unconformity, was observed; the maximum thickness of the active layer is less than 2 m within the diamicton and is well above the unconformity between the diamicton and the underlying icy sediment. Climatic conditions since deglaciation cannot explain the sharp planar contact shown in Figure 24 as a thaw unconformity. A depth of 5 m below the surface precludes such an explanation. In the general area where the slump occurs, there are many old and recent scars which display icy sediments and overlying diamicton. These massive icy sediments are interpreted as basal glacier ice buried by glaciogenic debris carried along the hanging wall of a thrust-plane in an area of compressive flow in the ice frontal zone of an active Late Wisconsinan ice mass (St-Onge and McMartin, 1989).

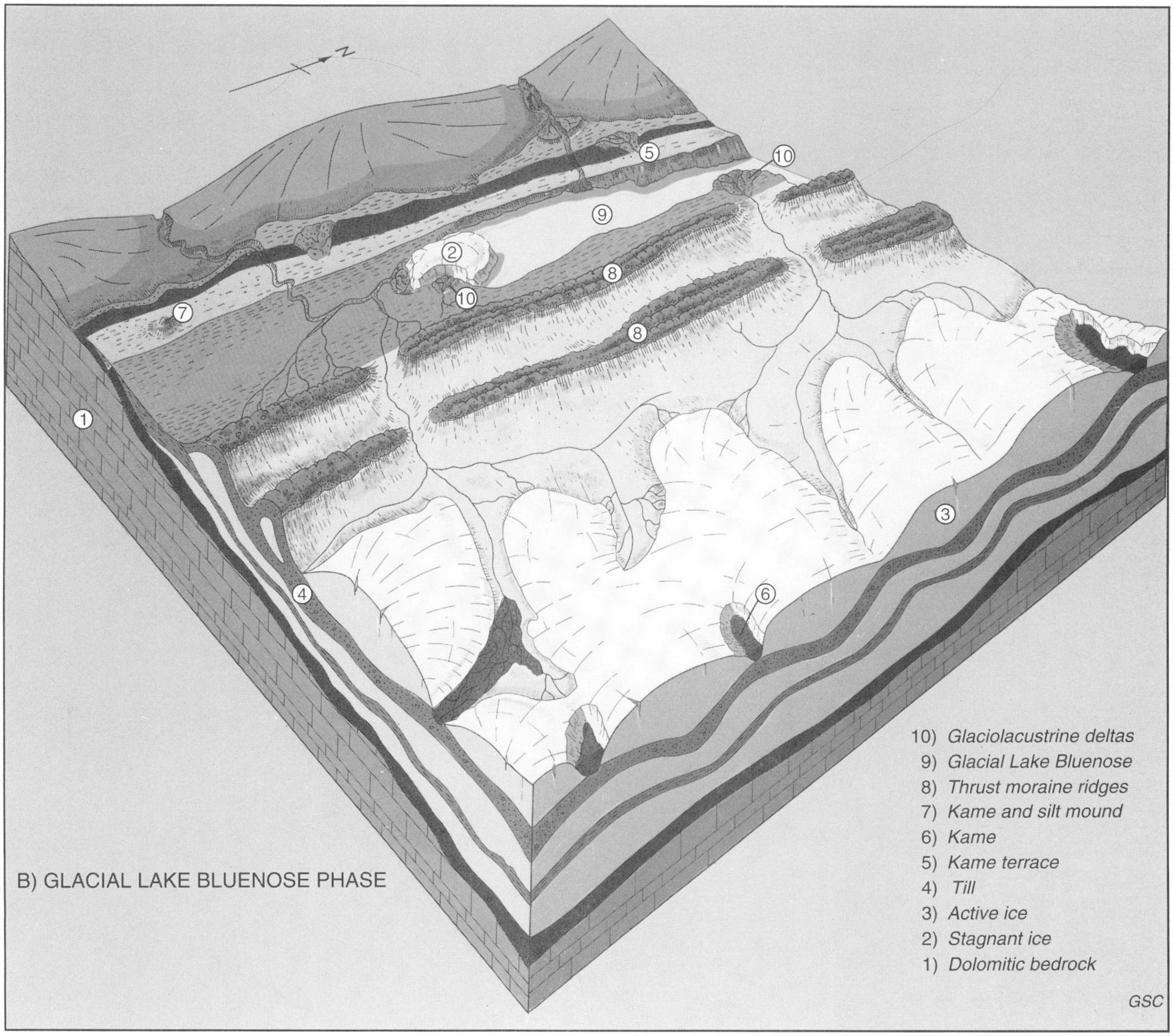


Figure 21. ***(B)*** *As a result of increased melting, the ice front is now located east of the zone of thrusting which constructed the ice cored ridges of the Bluenose Lake Moraine. Ice contact features emplaced during phase A are now exposed. Meltwater streams are rapidly building bouldery gravel and sand deltas in glacial Lake Bluenose which is still partly clogged with masses of stagnant ice.*

The observed isotopic composition is consistent with this hypothesis. The exposed icy sediments were sampled for $\delta^{18}O$ and δD determination. Ionic composition (Ca, K, Na, Mg), pH, and conductivity were also determined. The isotopic values obtained from the icy sediments vary from $\delta^{18}O$ = -28.2 to -26.9‰ ± 0.1‰ and from δD = -244 to -217‰ ± 3‰ (Fig. 25). By comparison, values obtained from the segregated ice samples (reticulated ice and ice wedge) are slightly higher ($\delta^{18}O$ = -26.1 to -24.0‰ and δD = -202 to -190‰). When these values are aligned on a δD - $\delta^{18}O$ diagram, the result is a straight line relationship with a regression slope of approximately 7.6 (Fig. 25). According to Lorrain and Demeur (1985), this slope value is expected for meteoric water from which buried glacier ice is derived (i.e., snow). Furthermore, an abrupt discontinuity is also observed between the isotopic signature of the icy sediments and that of the segregated ice bodies. This abrupt change is also observed between the hydrochemistry of the different ice masses (Fig. 26), another criteria which distinguishes buried glacier ice from segregated ice (French and Harry, 1988).

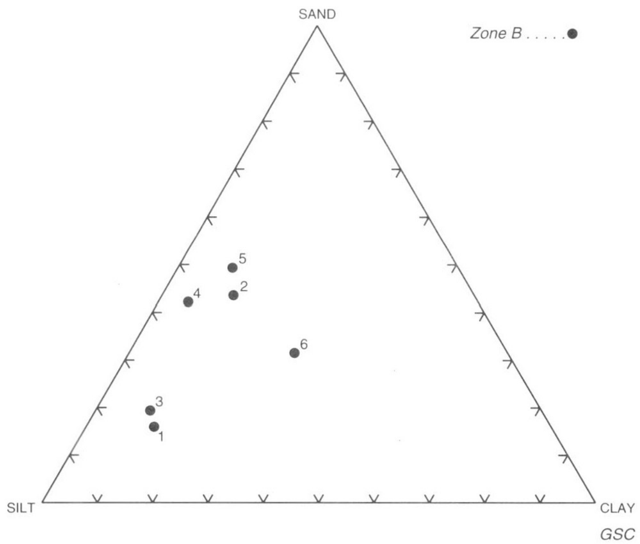


Figure 22. Ternary diagram comparing matrix in till samples from Bluenose Lake Moraine (morphosedimentary zone B, Fig. 1).

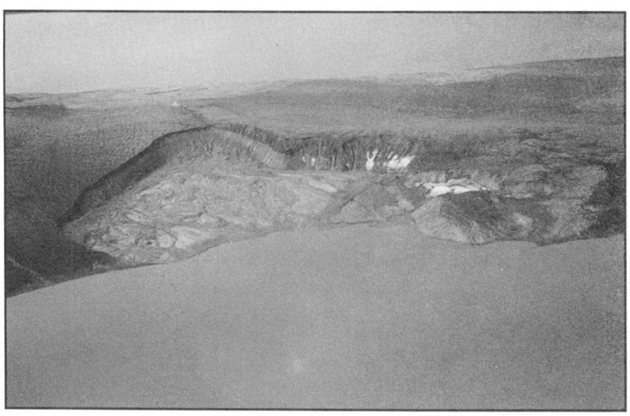


Figure 23. Low level oblique aerial photograph of the large slump exposing icy sediments beneath till sheet within the Bluenose Lake Moraine. View towards the northwest, July 23, 1988. GSC 204982-E

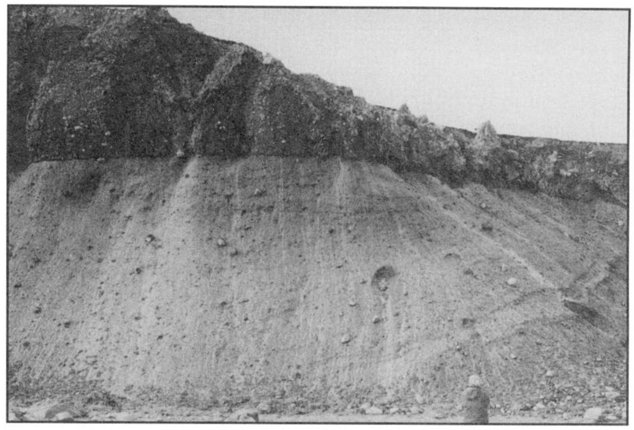


Figure 24. Ground view of the northern part of the slump scar shown in Figure 23; note sharp planar contact between the icy sediments and the overlying massive columnar till; numerous clasts can be seen in the gently folded icy sediments. GSC 204641-B

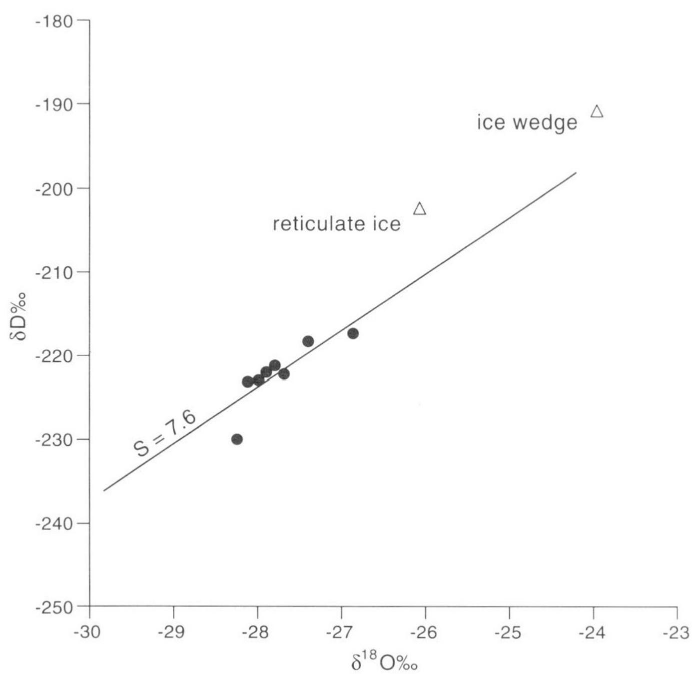


Figure 25. δD-$\delta^{18}O$ composition diagram of the icy sediment samples and of the segregated ice samples. "S" is the slope of the linear regression of the icy sediments.

When the ridges and hummocks of the Bluenose Lake Moraine were being constructed by thrusting, meltwater flowing to the west from the active ice mass deposited large loads of sediments in bodies of water which were dammed between masses of stagnant ice located in the depression now occupied by Bluenose Lake. Large "marginal deltas" (Lundqvist, 1979) were thus built on the western slope of the moraine ridges. All are pitted with numerous kettles which are bordered by collapsed slopes. The outer limits of the deltas show ubiquitous evidence of postdepositional collapse (Fig. 13). The most northerly of these deltas is on the east side of a large unnamed lake near the northeast end of Bluenose Lake. In this case the material (sandy gravel) is finer than that of the bouldery gravel deltas farther south. It is bounded on the north by a moraine ridge and kettles with associated collapse features. Its southern margin is also marked by large kettles and collapse features.

These deltas, constructed by westerly flowing waters, occur over a 70 km distance east and southeast of Bluenose Lake. Their surface is always between 660 and 670 m a.s.l. Rhythmically bedded sandy silt and clay deposits, 2-4 m thick, are found at elevations below the delta surfaces; the deposits, which blanket irregular surfaces around Bluenose Lake, are commonly pitted by kettles and distorted where exposed in sections.

All the above evidence supports the hypothesis of rapid sedimentation from an ice front into a water body occupying a depression which was extensively clogged with dead ice masses. The water plane of this lake was 40-50 m above present day Bluenose Lake and has been named glacial Lake Bluenose (St-Onge and McMartin, 1987). The areal extent of fine grained sediment is limited because of the amount of stagnant ice that was present in the basin. Lakes, kettle holes, and collapse structures (St-Onge, 1984, p. 276) below the level of the deltas all support this hypothesis. Glacial Lake Bluenose drained northwesterly, initially along poorly defined channels now occupied by a string of small lakes and marshy areas with sill elevations near 600 m a.s.l.

At lower elevation outlets become more defined, reflecting more sustained flow in the absence of masses of stagnant ice. The most spectacular trends northwesterly, passing from the map area just south of 69°15"N. Its present sill elevation is at 475 m a.s.l. The southeastern segment of the channel is, in part, in a side hill position with collapsed outwash on its eastern side. Northwestward, it flows into step terraces of bouldery outwash. Both of these features indicate that the channel initially functioned both as a glacial Lake Bluenose outlet and as an ice marginal channel carrying large quantities of outwash material.

Within the Bluenose Lake Moraine complex there are several small glacial lake basins, varying in surface area from 10 to 180 km^2, in which glacial meltwaters were held by the easterly retreating ice front. In these ice-dammed lakes sandy silt and clay rhythmites accumulated to measured thicknesses of 20 m. Beds of bouldery diamicton, 20-80 cm thick are commonly interbedded with the silty clay rhythmites (Fig. 11). These are interpreted as debris flow diamictons.

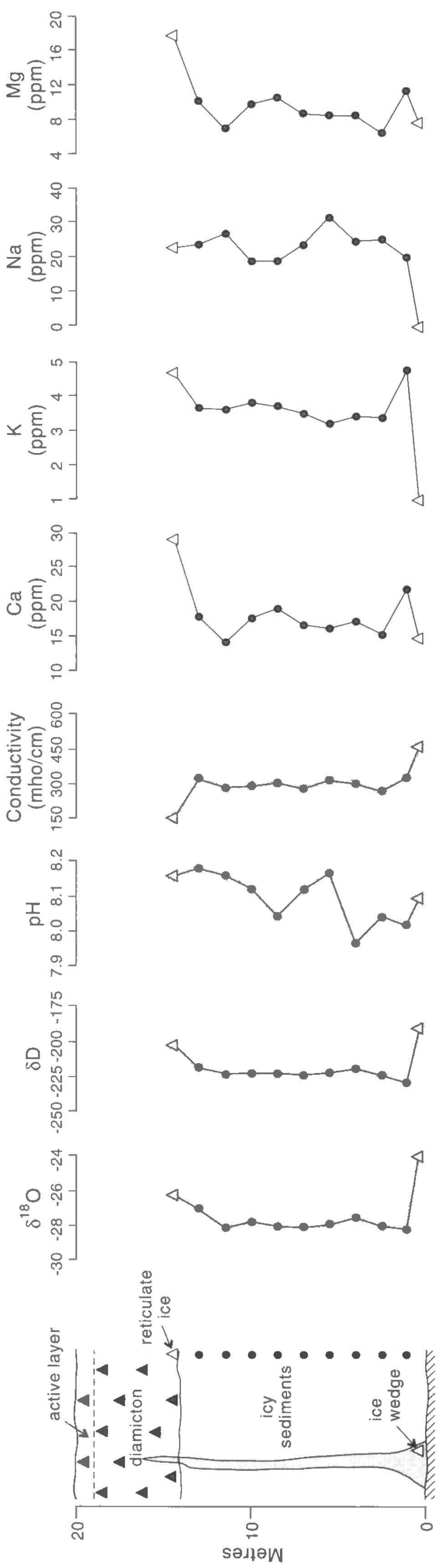


Figure 26. *Isotopic composition, pH, conductivity, and ionic composition of the icy sediment samples and of the segregated (reticulate and wedge) ice samples.*

Because clasts provide high resistance to wind or sheet wash erosion, they commonly form the capping beds of this sequence.

The Bluenose Lake Moraine complex with its system of north-trending ridges and associated high level deltas, along with numerous ice dammed lakes on its eastern flank, represents a major ice frontal position where a westerly-trending ice mass piled large amounts of glaciogenic materials. Spatial and stratigraphic relationships suggest that this is a Wisconsinan terminal moraine. Vincent (1984, p. 94, 96, 1992, Fig. 2 and 16) suggested that it is part of an Early Wisconsinan glacial limit, with Late Wisconsinan ice being limited to elevations below 300 m a.s.l. As will be discussed later, we have discovered no evidence to support this hypothesis which was based on airphoto interpretation.

Inman River drumlin field, (zone C)

The Inman River drumlin field forms a belt 5 to 45 km wide which starts just north of Rae River in the south and then forms a broad, northwesterly trending arc between the Bluenose Lake moraine complex (zone B) and the marine offlap area (zone I) fringing Amundsen Gulf and Dolphin and Union Strait (Fig. 1). The field is composed of long linear ridges up to 8 km long and 10-15 m high with the intervening swales occupied by narrow lakes or streams (Fig. 7 and 28). Dolomitic bedrock commonly outcrops between drumlins. Although no sections were found exposing the core of a drumlin, surface exposures and shallow sections show a diamicton with subrounded, mostly dolomitic clasts in a silty sand matrix (Fig. 29). Larger erratics are dominantly of gabbro and granites. The whole field appears to be composed of till resting on an essentially flat dolomite rock platform.

Outwash sand and gravel bodies are numerous and occur as dissected terraces irregularly spaced along the course of all major rivers. They commonly form broad fans with a steep east or northeast facing slopes scarred by slump features (Fig. 28). When the lobe responsible for moulding the drumlin field was undergoing regional stagnation the ice was melting more rapidly in the higher western region. The front of the dead ice was retreating towards the east and northeast allowing streams flowing from the Bluenose Lake moraine complex to deposit, at progressively lower elevations, coarse sediments in contact with the ice margin.

In the headwaters region of Inman River, drumlins parallel to the main north-northwest flow direction have been partly remoulded by a more westerly flow (see Fig. 36). The implication of this late ice movement will be discussed in a later section.

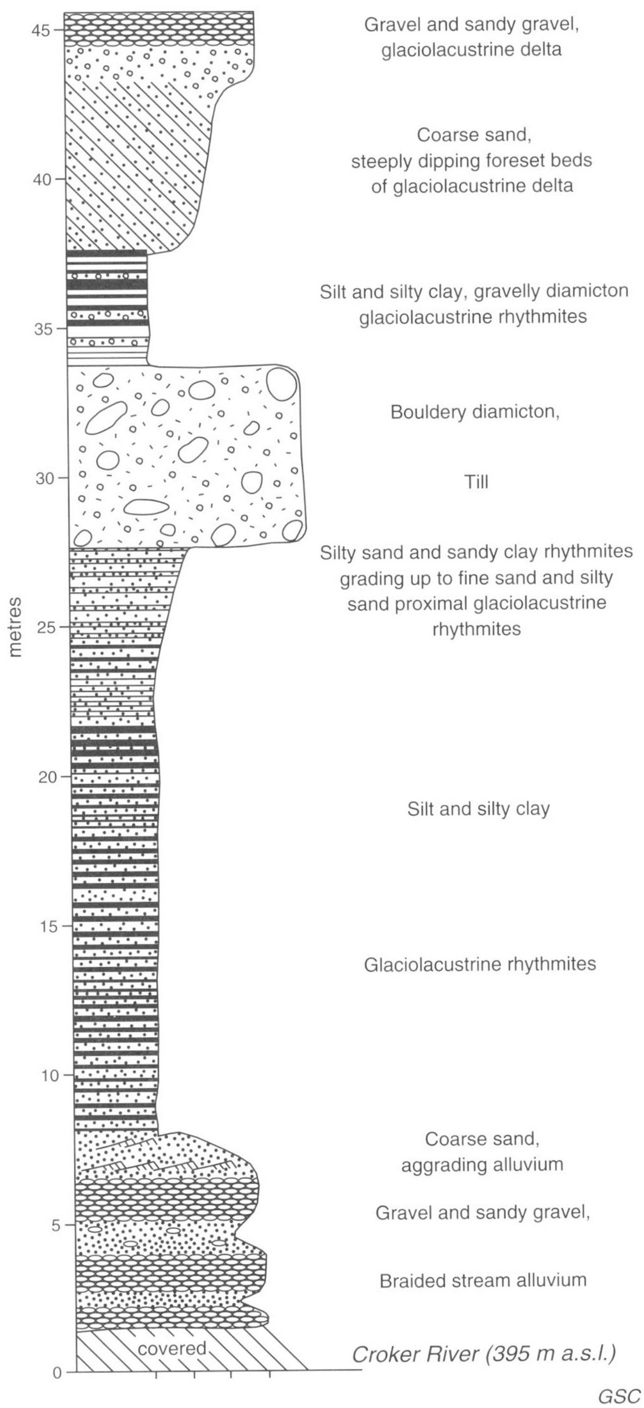


Figure 27. *Section exposed along the east side of Croker River, 68°56′58″N, 119°37′36″W (Fig. 12). The sediments exposed in this section result from a complete glacial cycle: gravel and coarse sand alluvium at the base emplaced in an environment similar to the present when river flow was unimpeded; coarse sand with climbing ripples mark the beginning of aggradation as a result of ice blocking the lower course of the river; silt and silty clay rhythmites deposited in a glacial lake which occupied the ice-dammed valley; sand and silty sand proximal rhythmites reflect encroaching ice mass; till emplaced by overriding glacier; silty rhythmites and debris flow diamicton deposited in a glacial lake during ice retreat; and at the top, sand and gravel of a glaciolacustrine delta.*

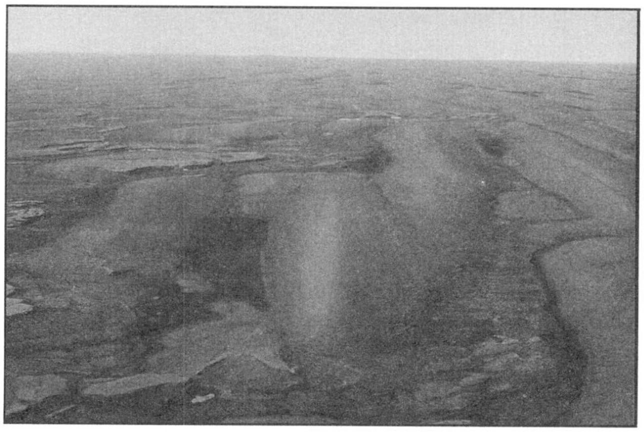


Figure 28. Part of the Inman River drumlin field (Fig. 7); note the long, whaleback-shaped drumlins and the outwash fans of gravel and sand with steep, often collapsed northeast facing slope. View northwest (down ice). GSC 1993-167D

Rae River glacial lake sequence, (zone D)

During the early phases of deglaciation, small glacial lakes were formed within the Bluenose Lake Moraine complex, particularly in small basins in the headwater regions of Rae River. Associated with these small glacial lake basins are numerous strandlines southeast of Bluenose Lake. The highest of these is at 550 m a.s.l.; others are at 520, 500, 470, 450, 420, 400, 380, 350, and 220 m. South of Libby Lake the lowest strandline occurs at 200 m a.s.l.

Deltas of bouldery gravel are associated with the strandlines and the steep outer slope of all of them shows evidence of collapse. Above 200 m, glaciolacustrine sediments are generally thin (1-3 m) with abundant diamicton beds; however, as mentioned above, in some small basins sediments are up to 20 m thick. All this evidence suggests that these features are related to narrow ice marginal lakes along a large lobe occupying the basin of Rae and Richardson rivers. This interpretation differs from Mercier (1984) who proposed various levels of a large glacial lake maintained by an ice front to the east. Mercier's hypothesis does not take into account the ubiquitous collapse structures on all glaciolacustrine deltas or the absence of fine sediments downstream from coarse bouldery gravel deltas. Both can be readily explained by deltaic sedimentation in a narrow ice marginal water body where the outer part of the delta front is in contact with, and partly on top of, ice which then collapses upon melting of the ice. In narrow water bodies at the margin of the ice, water flow velocities would be sufficient to remove sediments finer than coarse sands. This readily explains the apparent contradiction presented by large gravel deltas standing above till bodies or bare bedrock and the near total absence of distal fine grained sediments which should occur if the deltas had been constructed in a large body of quiet water.

Below 200 m a.s.l., in the lowest part of the basin which was covered by glacial lake waters for a longer period, rhythmites of silt and clay mantle the surface to a thickness of 2 m or more. South of Libby Lake, however, this mantle of glaciolacustrine sediments is found only below 140 m a.s.l., again suggesting a lowering sequence of lakes, marginal to a shrinking ice lobe. This time transgressive sequence lasted from early deglaciation until marine incursion following collapse of the ice lobe.

Upper Harding River moraine system, (zone E)

Harding River originates in a large unnamed lake in the south-central part of the map area (Fig. 1). To the north, west, and south of the lake, several sinuous morainic ridges, averaging 20 m in height but, in places, rising to 50 m above the surrounding surface, mark the frontal position of small ice lobes generated by a glacier to the east. These lobes advanced after the ice, responsible for the formation of the Inman River drumlin field, had become stagnant.

Drumlin orientations show that the northernmost ridge was built at the front of a lobe which moved westerly and then southwesterly across an older set of drumlins trending north-northwest. The middle set of ridges, east of Libby Lake, are associated with a lobe advancing northwest.

One outstanding feature in this morphosedimentary zone is a rock pingo located west of the Harding River just south of the northern limit of this zone (St-Onge and Pissart, 1990). This unique pingo, originally identified by Craig (1960), is discussed later (see Harding River rock pingo, below).

Harding River drumlin field, (zone F)

This zone comprises a field of relatively short (average length 1 km), in places crosscutting, drumlins in the northern part of the field. The trend of the drumlins changes from west-northwest in the southern part of the field to northwest in the

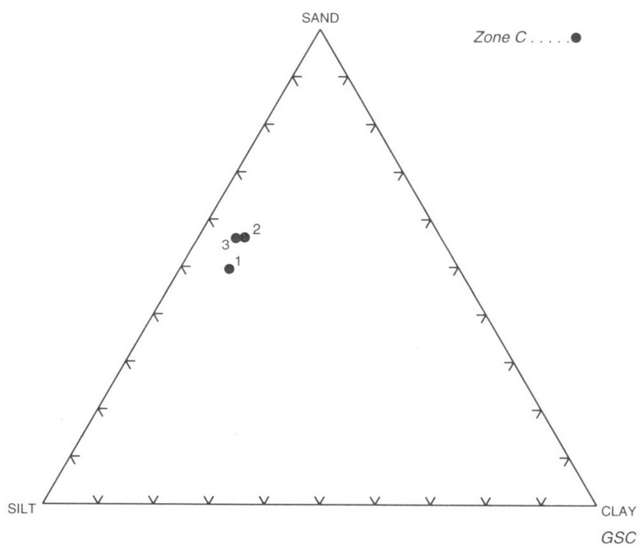


Figure 29. Ternary diagram comparing matrix in till samples from Inman River drumlin field (morphosedimentary zone C, Fig. 1).

northern part. The diamicton of the drumlins is a silty sand matrix (Fig. 30) with clasts of up to 98% dolomite; sandstone, gabbro, and granites are minor constituants.

The central part of this zone is mostly bare bedrock or till veneer with very few drumlins (see Map 1846A, in pocket). Except for the small drumlin field in the northern part of this zone, the relief strongly reflects bedrock topography; stepped scarps at the edge of small plateaus are the dominant landform. Steep-sided valleys cut in bedrock by meltwaters are also part of this rock-controlled landscape, in particular, the strikingly rectilinear valley of the lower Harding River and the upper segment of a river which flows into Dolphin and Union Strait approximately 12 km south-east of Cape Young. These two valley segments align to form a single, bedrock controlled, 55 km linear feature clearly visible on the LANDSAT MSS image (Fig. 31). Flat bedrock surfaces often display large grooves which trend parallel to the drumlins. In this upland area, where the headwaters of Hoppner and Harding rivers as well as several other unnamed rivers originate, glacial erosion predominated over sedimentation.

Although they occur in the marine offlap morphosedimentary zone (I) it is appropriate that the Harding River Falls be discussed here. The Harding River, which originates in the large lake in zone E, flows in a shallow postglacial valley through moraine ridges and then through the drumlin field in the northern part of zone G. It has carved a channel in its marine limit delta and in silty marine sediments before cascading down into the fault controlled bedrock valley which is certainly pre-last glaciation and very likely preglacial. This spectacular series of rapids and falls (Fig. 32) is 22 km upstream from the mouth of Harding River. The cascades over ledges of resistant, flat-lying dolomite beds represent a total drop of 25 to 30 m. These falls, which mark the point where a river flows from an undulating plateau surface into a bedrock valley, underscore the fact that relatively narrow bedrock valleys oriented at generally right angle to glacier flow are commonly not filled with glacial debris. This had been postulated by St-Onge and Lajoie (1986, p. 1706) for segments of the Lower Coppermine valley. Another valley, 27 km southwest of Bernard Harbour which is in part occupied by a narrow lake, illustrates this phenomena even more convincingly since a drumlin field occupies the upland on either side of the valley, yet the valley was not filled with glacial drift. It must be concluded that advancing ice will occupy a pre-existing bedrock valley and continued flow will result in sheering within the ice from valley wall to valley wall. This explains how the olisthostrome of the Sleigh Creek formation (St-Onge and Lajoie, 1986) could be deposited in bedrock valleys which had been submerged by the postglacial marine incursion.

Rae River moraine belt, (zone G)

Southeast of Libby Lake and north of Rae River in the south-central part of the map area, several east-trending massive morainic ridges of silty diamicton (Fig. 33) were formed at the northern margin of a lobe which occupied the basin of Rae River and Richardson River, immediately south of the study area (St-Onge, 1987, 1988). They rise to 270 m a.s.l. and show no sign of modification by wave action which indicates that this area was still under ice, when farther west, ice marginal lakes formed strandlines at elevations up to 550 m a.s.l. (see zone D, above).

Five sets of discontinuous ridges mark the shrinkage of the lobe from 270 m a.s.l. to 240 m a.s.l. The absence of ridges at lower elevations suggests that the ice mass occupying the Rae and Richardson rivers basin had become thin and stagnant. It soon desintegrated allowing sea water to replace the ice-marginal lake.

Marine limit kame complex, (zone H)

East of zone F in the headwaters of the river which flows into Coronation Gulf west of Cape Hearne is a narrow belt of bouldery gravel, coarse to medium sand, and patches of sandy silt rhythmites up to 25 m thick. The generally horizontal and channelled surface is pitted by numerous steep-sided kettles occupied by lakes, ponds, or bogs. Present day lakes within the kame complex are typically fringed by well developed beaches of gravel and sand. Beach material tends to be size-sorted, becoming finer towards the east (Fig. 34).

Upper surfaces of steep slopes are scarred by slumps (Fig. 35) indicating that the gravel and sand outwash was deposited between the bedrock upland of zone F and glacier ice to the northeast. Within this wedge-shaped corridor opening to the northwest meltwater was rushing into the encroaching sea. Turbulent, sediment-rich waters flowing along the ice front and between blocks of glacier ice deposited coarse bouldery gravel which now form narrow ridges, i.e., crevasse fillings between more extensive kame terraces. These are between 140 and 150 m a.s.l., just above marine limit in the

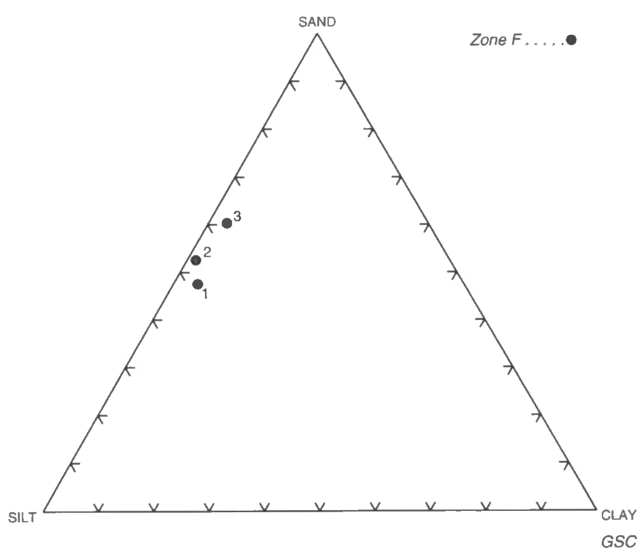


Figure 30. Ternary diagram comparing matrix in till samples from Harding River drumlin field (morphosedimentary zone F, Fig. 1).

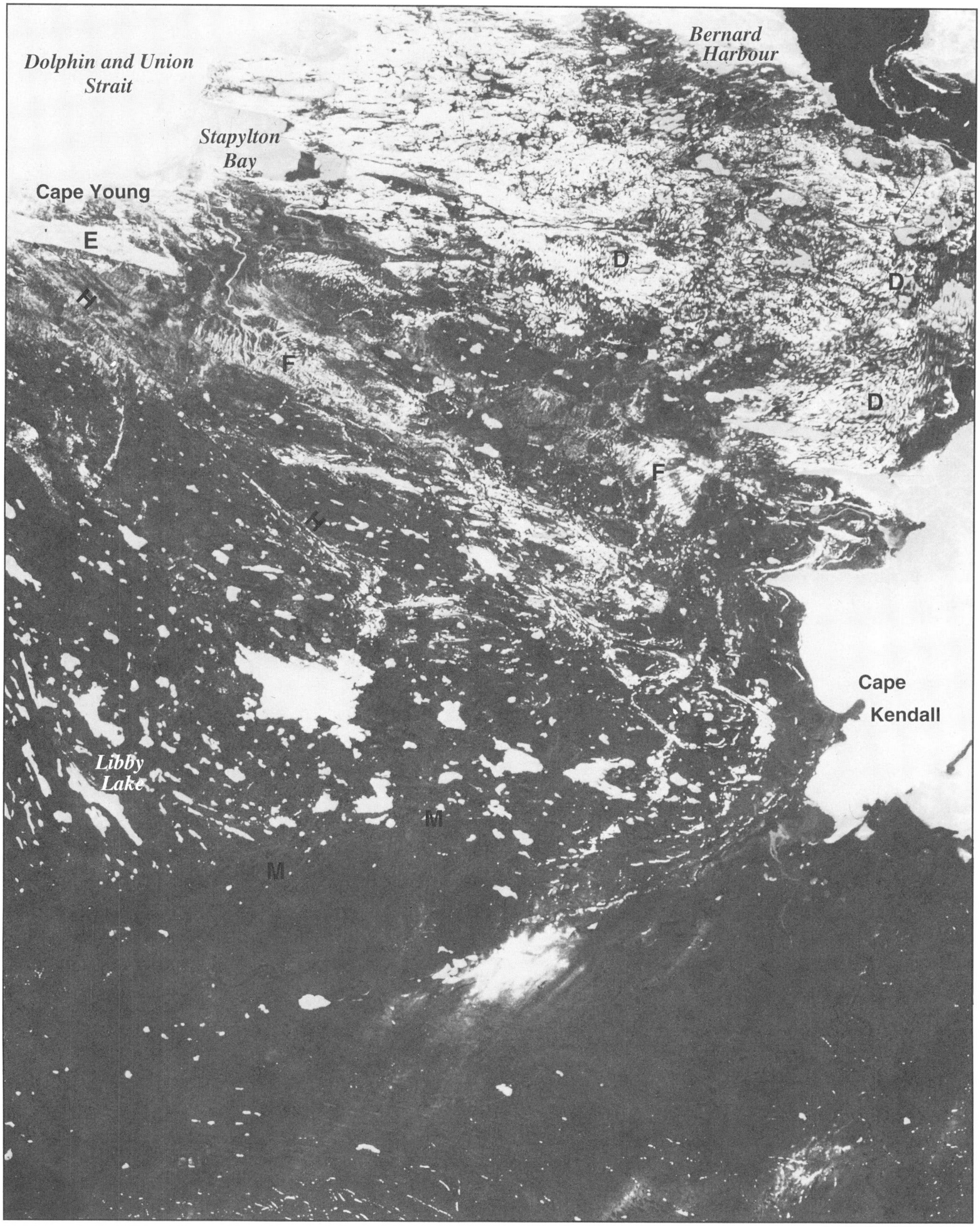


__Figure 31.__ LANDSAT image of Bernard Harbour area east of Cape Young shows pattern related to vegetation, sediment type, drainage and bedrock structure (LANDSAT-4 MSS, Band 24, July 1, 1985); morphosedimentary zones C to J are represented in this view; Harding River bedrock lineament (H), Rae River Moraine ridges (M), Bernard Harbour drumlin fields (D), fine grained marine sediments (F), Cape Young eolian feature (E).

area; the base level for meltwater responsible for the emplacement of this kame complex was the sea encroaching from the northwest.

The sedimentary facies, from coarse gravel to sandy silt rhythmites, reflect the varied sedimentary dynamics within the glaciomarine environment. In the coarse sand and bouldery gravel facies only two exposures were found, both in the eastern part of the deposit. They display beds of medium to coarse sand and lenses of silty clay. Type A and type B climbing ripples (Jopling and Walker, 1968) are overlain by draped undulating laminae which grade into, or are truncated by, horizontally bedded sands of offlap nearshore environment. Paleocurrent directions are variables reflecting local, rather then regional, trends (Fig. 36).

Within zone H several sections expose a facies of stratified fine to medium sand with minor amounts of silt and clay. Measured thicknesses range up to 12 m (Fig. 36). This silty sand is commonly underlain by a silty till (Fig. 37) and overlain by marine rhythmites; it occurs mostly at elevations at least 30 m below marine limit. Kerr (1994, p. 16) stated that: Beds are generally from 0.1 to 1.5 metres thick, exhibiting both sharp and gradational upper and lower boundaries. Paleocurrent directions are variable, but generally indicate deposition from currents flowing oceanward in a westerly to northerly direction.

"This facies is interpreted to be subaqueous outwash (Rust and Romanelli, 1975). This interpretation is supported in part by the position of [this facies] within a stratigraphic sequence which could only have formed in a proglacial environment at an ice-sea interface. These sediments consist of stratified sand removed from a glacier front by subglacial meltwater streams and deposited in a tunnel, or in front of and beyond the glacier margin into a standing body of water."

Figure 32. Segment of the falls which marks the location where Harding River cascades down, in a series of steps, from the upland where it originates into the rectilinear, fault controlled, pre-last glaciation (or older) paleo-bedrock valley that will carry it to the ocean; total drop is 35 m. GSC 1993-167A

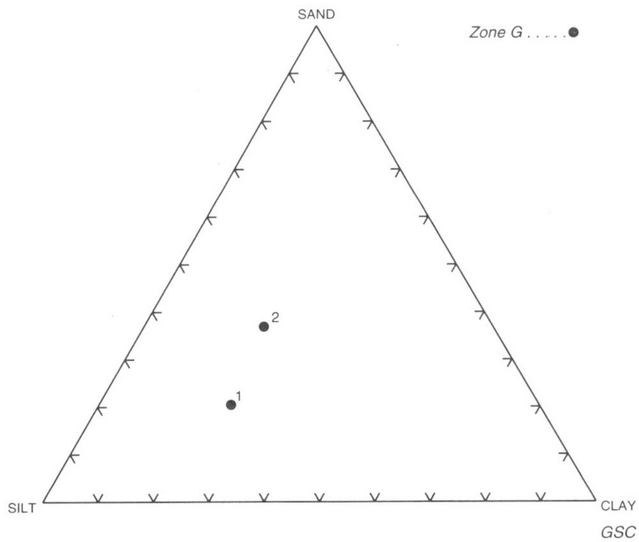


Figure 33. *Ternary diagram comparing matrix in till samples from Rae River moraine belt (morphosedimentary zone G, Fig. 1).*

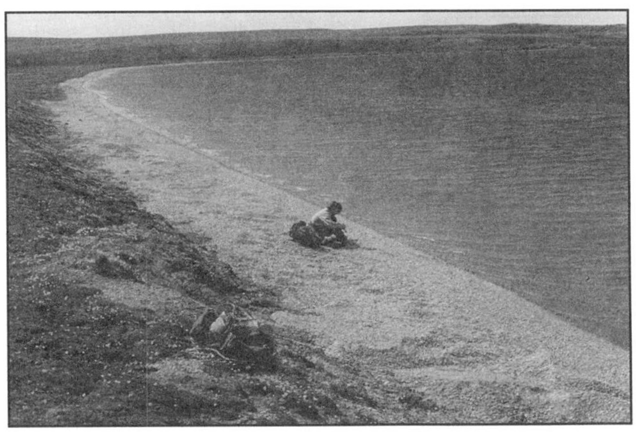


Figure 34. *Beach of pea gravel and sand at the southeast end of a 6 km long lake within morphosedimentary zone H (Fig. 1); view is to the southwest; a drumlin in zone G forms the horizon. GSC 1993-121E*

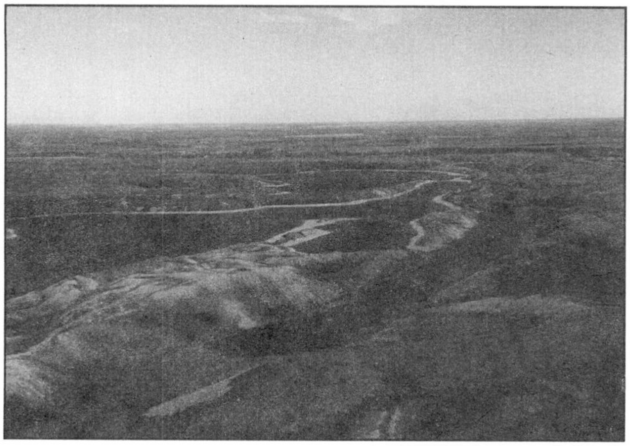


Figure 35. The extensive kame complex of morphosedimentary zone H (Fig. 1) is pitted by numerous kettles occupied by lakes; view is east, upstream; note slump scars on slopes in lower left hand corner of photograph; marine delta in centre left (north) of photograph; 8 km northwest of Klengenberg Bay. GSC 1993-158N

The fine grained fossiliferous marine sediments associated with the transitional glaciomarine complex of zone H will be dicussed in the following section.

Marine offlap area, (zones I and J)

Along the south coast of Amundsen Gulf and of Dolphin and Union Strait the postglacial marine offlap sequence widens and increases in elevation towards the east (Kerr, 1989). Approximately 1 km wide and reaching elevations of 45 m a.s.l. at the margin of the map area west of Tinney Point, it widens to 30 km and reaches elevations of 160 m west of Cape Kendall.

A typical profile from marine limit to the present shoreline (Fig. 17) crosses the following features:

1. A delta deposited at marine limit with material rapidly becoming finer downstream, grading from bouldery gravel near the apex of the delta to sand and silty sand on the delta front. Numerous well preserved abandoned channels mark its surface. The marine delta is commonly cut into an excessively bouldery kame deposit 10-15 m higher with its seaward side scarred by numerous collapse features.

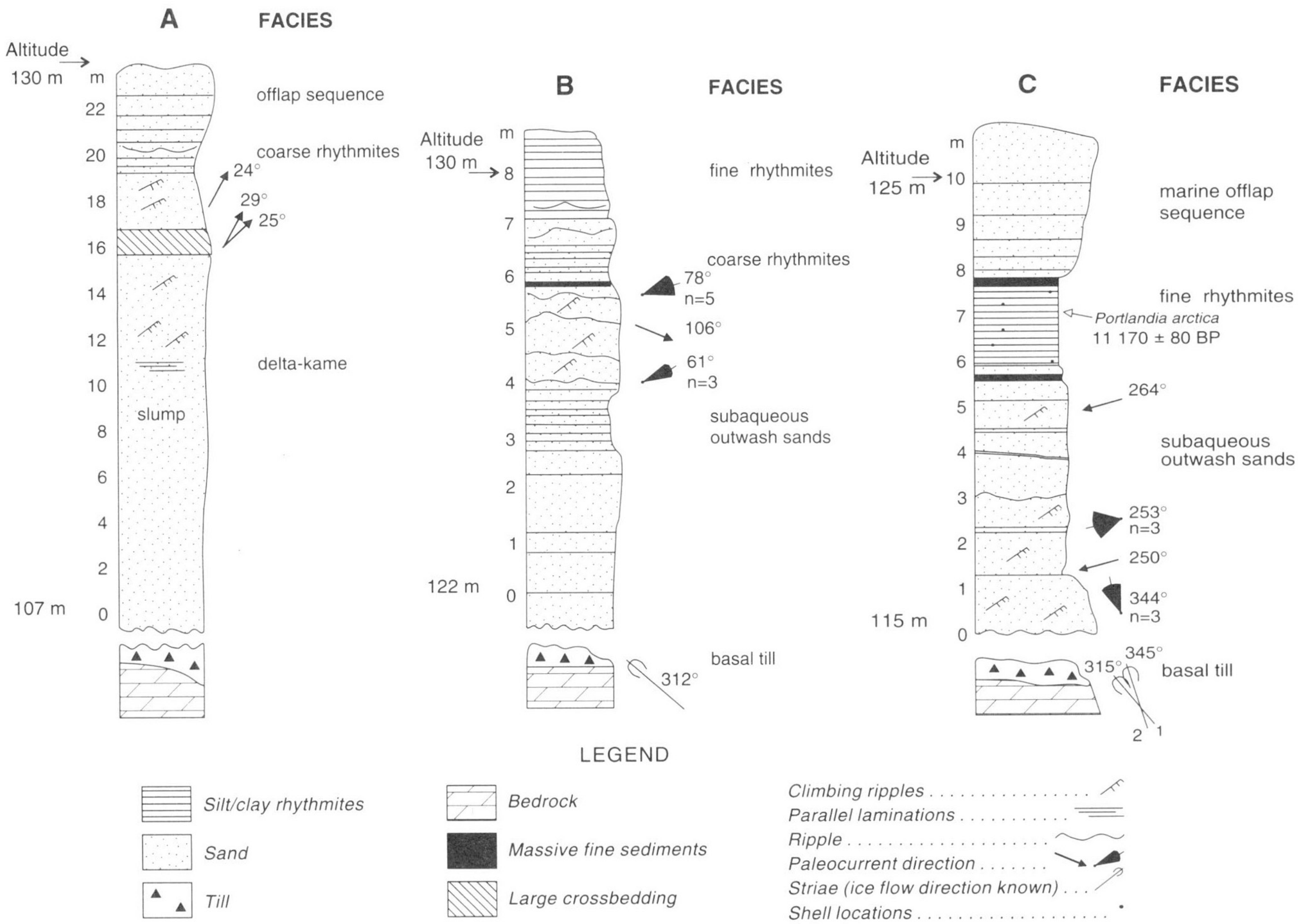


Figure 36. Lithostratigraphic sequences exposed within morphosedimentary zone H (Fig. 1) showing bedrock/till contact and outwash/subaqueous outwash transition to deep water marine silty clay coarsening upwards to nearshore sands (after McMartin, 1990).

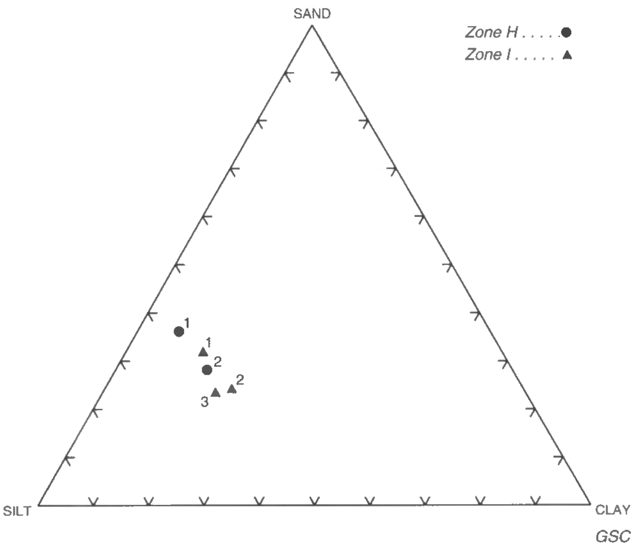


Figure 37. *Ternary diagram comparing matrix in till samples from A (•), marine limit kame complex (morphosedimentary zone H, Fig. 1), and from B (▲), till exposures in adjoining marine offlap area (zone I).*

2. Farther out in the basin a commonly rhythmically bedded and fossiliferous deposit is composed of buff, sandy silt beds containing brown sticky clay laminae. Except in bedrock valleys, the thickness of this deposit seldom exceeds 2 m. Numerous inliers of bedrock and till (drumlins) protrude through the marine material. Some sandier areas have been extensively modified by eolian activity.

3. Between the fine materials accumulated in marine basins and the flights of beaches that fringe the coastal areas, a lag of boulders, gravel, and sand mantles the surface. This lag-covered zone, which varies in width from a few tens of metres to over 40 km, is the result of winnowing of fines by wave and current action from diverse underlying materials. Marine fossils do occur in these sediments.

4. Flights of beaches of pebbly gravel lead to the modern shoreline. These raised beaches form an irregular belt of spectacular scrolls, a few tens of metres to over 3 km wide.

5. The modern beach, strewn with driftwood which becomes progressively less abundant eastward, is a 30-50 m wide strip of sandy pebbly gravel.

Although the relative extent of any given material may vary greatly, the sedimentary succession described above can be found along a transect across any part of the marine offlap sequence. For the sake of clarity, the features within the marine offlap sequence are divided into four groups: I-dominated by deposition; II-dominated by erosion processes; III-glacial landforms modified by marine processes; IV-the modern beach.

Features dominated by deposition

Marine deltas

One of the striking aspects of this sequence is that the deltas (Fig. 17 and 38) occur all along the marine limit but seldom at lower elevations. The implication is that the processes responsible for constructing these imposing sedimentary structures were short-lived and, as will be described later, confined to early postglacial time. Following the initial flush of sediments carried to the sea by powerful meltwater streams, inland ice had disappeared to the point where rivers were no longer sufficiently fed by melting ice to enable them the transport coarse sediment. As is the case today, they transported material in the range of clay to coarse sand which was redistributed by wave action and shore currents.

That most of these deltas are, in part, carved into bodies of kame gravel deposited at the margin of an ice mass, indicates that the sedimentation of coarse clastics was continuous from ice marginal to marine environments along a corridor at the margin of a shrinking ice lobe occupying Amundsen Gulf and Dolphin and Union Strait.

Marine deltas in the Bernard Harbour/Bluenose Lake area are described by Kerr (1994, p. 25) as consisting

> "...primarily of coarse sand and granules, as well as pebbles and cobbles, although the latter may entirely dominate certain deltas near marine limit. Large-scale planar crossbedding is the dominant structure in which internal bedding varies from 10 cm to 30 cm in thickness and sometimes shows fining upward cycles."

These deltas vary in thickness from 4 to 37 m. Most are fan shaped, giving rise to a lobate form with an arcuate front. They are typical of coarse Gilbert type deltas (Shaw, 1977). The channelled delta surface represents braided stream deposits (Rust, 1978). The contact between topset and foreset beds is often erosional and probably represents sea level at time of construction (Gustavson et al., 1975). Marine bivalves occur in some sections, occasionally in life position, but more commonly as single valves.

Fine grained marine sediments (zone I)

Away from the delta fronts, sediments rapidly become finer and range from fine sand to silty clay (Fig. 38). On the Surficial Geology map (in pocket) two units are distinguished: Mm, a silty sand and Mp, silt/clay. In his detailed study of sedimentary facies in the area Kerr (1994, p. 18-24) defined the following facies: 1) an interlaminated sand, silt, and clay; 2) a massive silty clay, and 3) a thinly bedded sand, silt, and clay. Surface characteristics are not sufficiently pronounced to distinguish these units on air photographs and to determine areal extent on the map. As a result, sandier units were mapped as Mm and silty clay ones as Mp. In reality, sedimentation in the marine environment tends to form a continuum from the coarse deltaic sediments to the massive muds of the deeper water environments.

Kerr (1993, p. 16-18) described these sediments as follows:

1. The interlaminated sand, silt and clay (Kerr, 1994, p. 18): "...thinly bedded to laminated fine-grained rhythmites which are found throughout the study area occur most commonly above... (subaqueous outwash deposits or more rarely till). Fine sand, silt, and clay horizons, 0.1 to 1.5 cm thick, have either sharp or gradational upper and lower contacts. Each rhythmite is defined by cyclic changes in colour and texture giving rise to striking, repetitive interbeds. Where rhythmites gradationally overlie the coarser sand deposits, beds may attain thicknesses of 10 to 50 cm and consist of graded fine sand to silty clay with internal sand or clay laminae, 0.1 to 1.5 cm thick."

This rhythmically bedded fine sand, silt, and clay facies is interpreted as a glaciomarine deposit resulting from meltwater streams discharging in subglacial and ice-marginal environments forming underflows, interflows and overflows. The thickness of this unit is quite variable, ranging from as little as 0.5 to 24.5 m.

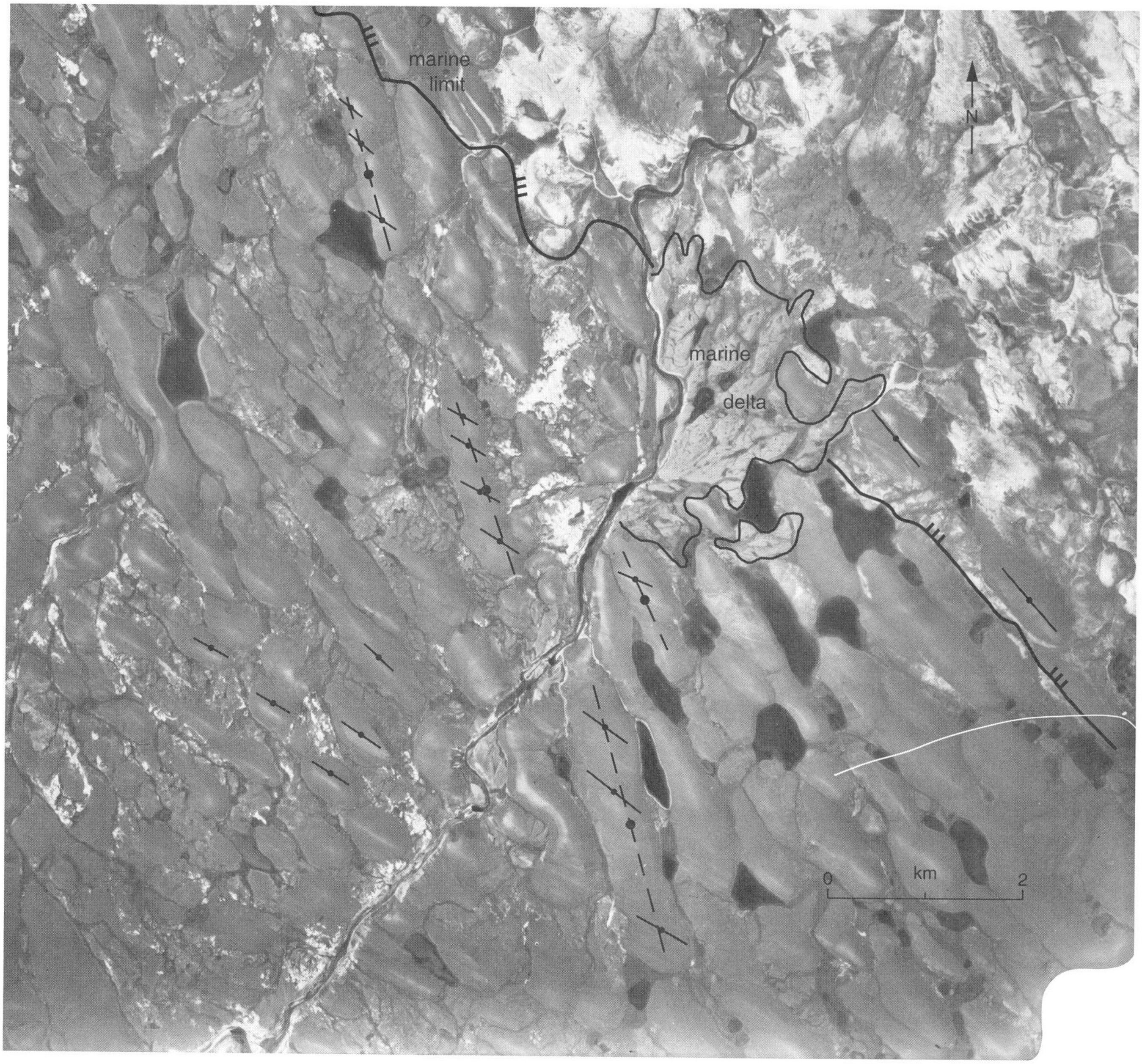


Figure 38. Glaciomarine delta at 120 m a.s.l. and marine limit along the east branch of the Harding River south of Cape Young. Note partial remoulding of northwest-trending drumlins by a west-northwest ice flow; light coloured area in northwest corner of the figure corresponds to fine grained marine sediments. (NAPL A21509-41)

2. The massive silt and clay facies (Kerr, 1993, p. 21) is composed "of mottled pinkish grey silty clay which is for the most part massive and structureless; rare thin silt or fine sand stringers and small gravelly lenses are restricted to the lower part of this facies. By volume it is the most significant sedimentary unit within the study area. It is found all along the coast, varying in thickness from 1.5 m to commonly 20-25 m.... This unit is fossiliferous and contains several species of marine ostracodes, foraminifera and bivalves. ...(This unit) is interpreted as a marine deposit resulting from suspension in quiet water environments of the most distal glaciomarine settings, as well as in the relatively deep waters of the open marine conditions which existed following deglaciation."

3. Thinly bedded sand, silt and clay facies: strata composed of sand, silt, and clay mixtures and which range from 1 mm to 15 cm in thickness form a third, fine grained marine sediment facies which Kerr (1993, p. 22-23) described as a deposit in which:

> ..."The cyclic variation in grain size and colour of each horizon defines the rhythmic nature of this deposit. Strata have gradational and sharp upper and lower contacts and may be either massive or show internal stratification....The sediments of this facies are particularly rich in organic plant debris (twigs, leaves, rootlets), as well as in situ marine macro and micro-fauna. Bioturbation, in the form of burrows, and soft-sediment deformation structures are also generally common throughout the facies. It varies in thickness from 0.5 m to 20 m, although average thicknesses are in the range of 5-10 m. The lower boundary of this facies is gradational with the massive mud as is the upper contact with (beach sediments).
>
> "The rhythmic deposition of (this facies), its stratigraphic position and fossil content, all suggest a shallow water marine environment in which bedding resulted in part from variations in sedimentation due to variable cycles of fluvial discharge in response to short-term climatic changes. Rivers draining the coastal regions are responsible for the introduction of terrestrial plant fragments into the marine environment. Deposits similar to these described by Hillaire-Marcel (1979) are characteristic of shallow, isolated marginal marine basins in which they often complete a regressive marine sequence....In the present study, it is difficult to determine the degree tidal influence has had on marine sedimentation because the tidal range in the Coppermine area is only 0.17 m (tidal Institute of Ocean Sciences, 1983)."

Features dominated by marine erosion processes

Wave washed materials (zone J)

In the Inman River area, terrain below marine limit comprises an irregular zone of generally coarse materials. These materials drape a variety of substrate including moraines of generally very coarse ice contact fluvioglacial bouldery gravel and till, drumlins of bouldery glacial diamicton (till), eskers and kames of gravel and sand, felsenmer and other bedrock debris, and to a lesser extent, areas of fine marine sediments.

The dominant feature of the marine washed area is a nearly continuous mantle of gravel and sand with some boulders. This lag, which seldom exceeds 1 m in thickness, except in swales between drumlins or similar depressions, is often fossiliferous and is a distinct sedimentary facies not linked to a particular landform. Although the material is coarse, with gravel being the predominant grain size, texture varies with the nature of the parent material; open textured bouldery gravel on top of drumlins varies to gravelly sand and occasional boulders on eskers and kames. Depressions, particularly between drumlins, are covered with coarse sand and minor gravel washed from adjoining slopes.

Lag material tends to form an armour or pavement 15-25 cm thick of subangular clasts mantling the underlying material. Wave action removed fines until the concentration of coarse material impeded or nullified any further erosion by wave action during the short summer period when the sea was sufficiently ice-free to permit wind to generated waves. The combination of short, ice free period and rapid initial uplift, discussed later, prevented much rounding of clasts. Even at lower elevations, although corners have been blunted, beach shingles tend to remain angular.

The marine lag deposits which cover large areas of Bernard Harbour region, particularly northeast of a line joining Basil Bay and Stapylton Bay, represent an important change in processes from heavy sedimentation near the marine limit to erosion by wave and marine currents during a later phase of the marine offlap event. The paleogeographic implications are important: massive sedimentation in the encroaching sea during early postglacial time, followed by a dramatic shutoff of sediment supply and erosion of emerging landforms by waves during most of the marine offlap period. The short-lived burst of sediment supplied to the sea, as pointed out earlier, is also marked by the single level of marine deltas at or near marine limit. If sediment supply had continued for a significant period during uplift, deltas would have been constructed at lower elevations as they were along the Coppermine River to the southeast (St-Onge, 1987, 1988).

Rectilinear furrows or scour, are a fairly common feature in the wave washed zone. Most are between 40 and 100 m a.s.l. and are 150 to 500 m long, 1 to 2 m wide, and 20 to 50 cm deep. They have no preferred orientation, and are interpreted as grooves formed by iceberg keels or by large sea ice pressure ridge keels (McMartin, 1990). Scours were not found within the elevation zone of well developed raised marine beaches. Either icebergs were no longer present, or scours were destroyed by beach forming processes.

Raised marine beach ridges

Along the coast and in places inland, delicate and complex scrolls of raised beaches mark the gradual fall of sea level (Fig. 39). These flights of beaches are constructed of angular

Figure 39. *Raised marine beaches along ice-strewn Coronation Gulf north of Locker Point; note intricate scrolls resulting from wave deflected by bedrock obstacle particularly around Qikiarrarnaq Bluff; wave-washed drumlins in northwest corner of the figure; A is location of Figure 16. (NAPL A18910-24)*

shingles of dolomite. They are best developed on east- or northeast-facing rock slopes or on drumlins well below marine limit, suggesting that during the early postglacial either the fetch was insufficient or the ice cover was too extensive to permit effective wave action. To a large extent they are the result of the bulldozing actions of pans of sea ice pushed on shore by strong winds, a process extensively studied by Dionne (1988; Dionne et Brodeur, 1988). Waves smooth out the jagged scars left after the ice raft has melted and washes out some of the sand and finer particles which will be redeposited as bars on the lee side of obstacles such as protruding drumlins or rock knobs. Sand beaches are confined to bays where the finer material is reworked alluvial material transported by rivers rather than sand deposited by longshore currents. Tides of less than 1 m amplitude and seas covered by ice for more than 6 months of the year preclude strong wave action, particularly when uplift imposes a further limiting factor.

No sections were found in shingle beaches. Finer beaches are constructed of a brown silty sand which is structureless to crudely bedded. Locally, it may contain horizons of coarse sand to gravel that here and there form planar crossbedded strata. This facies consistently overlies the massive silty clay and the marine rhythmite facies. These deposits are fossiliferous, containing both marine bivalves and foraminifera and may exhibit convolute bedding where stratification is evident.

Glacial landforms modified by marine processes

Although, as described above, landforms were modified by marine wave and current action, some elements such as moraines, drumlins, and eskers are very well preserved within the offlap zone as shown on the Surficial Geology map (in pocket) and are, in part, shown on the map by Craig (1960).

Moraine ridges (zone J)

South of Stapylton Bay and just north of the kame terraces, zone H, are arcuate moraine ridges convex south, southwest, or west (Fig. 40).

There are two quite distinct ridge patterns: the most northerly is oriented northwest-southeast and then west-east with small but pronounced lobations convex south; the southerly moraine forms a broad arc convex to the west and then extends in a southeasterly direction. The two systems meet at a sharp angle (Fig. 40) in a lake-pitted complex node pointing east. Both moraine systems occur within the marine offlap zone and both are flanked by fine marine sediments. Ridge crests are generally at or below local marine limit of 155 m a.s.l., although in the ridges and mounds at the node of the two systems, an elevation of 159 m a.s.l. is reached, and this small part of the moraine system may have extended above sea level during formation.

The two moraines differ markedly in several important characteristics. The ridges of the northern system are asymmetrical, with the steeper slope facing south or southwest; terraces of sandy gravel have been built at the end of gaps cut in the ridges by south flowing streams. Elevations are approximately 130 m a.s.l. on the top of these terraces. The surface material of the main ridge is a very bouldery diamicton with the larger clasts being dominated by dolomite (90% dolomite, 6-7% gabbro, and 3-4% sandstone and granite).

A section at point S (Fig. 40) shows the internal structure of the main ridge of the northerly system (Fig. 41). The lower 2.5 m of the section comprises horizontal beds of sand, gravelly sand, and minor lenses of sandy-silt. These are overlain by just over 2 m of coarse poorly sorted gravel; beds, although poorly defined, dip by as much as 30° towards 105°. The overlying 5 m of gravelly sand occasionally display beds dipping by 10° towards 110°. One metre of structureless diamicton caps the section.

North of the main ridge, and parallel to it, are a series of minor ridges and/or bands of differently coloured material. These are moulded by discreet fluting in a direction perpendicular to the ridge. The en echelon disposition of these minor ridges and bands is reminiscent of that of a tilted deck of cards and probably represent the surface outcropping of thin thrust sheets separated by low angle faults. These features become progressively diffused along the northwest arm of this moraine and are replaced by a field of small drumlins which abut against the eastern slope of the ridge.

The north-south ridge (due east of J on Fig. 1) is more massive and of much finer material in which the tan colour of the sandy silt is similar to the marine deposits in the area. For most of its length this part of the moraine is a double ridge, the eastern slope being in places slightly steeper. The western ridge displays evidence of thrusting in the form of numerous minor ridges and bands of differently coloured materials. East of the moraine, vaguely streamlined forms grade rapidly into drumlins.

The ridges are the result of rapid and intense sedimentation at or near the front of a thrusting ice sheet. The important question which remains is whether or not construction followed a readvance in a marine environment.

The very coarse, crudely stratified material of the west-east moraine does not suggest the remobilization of marine sediment by a thrusting glacier, but rather the rapid sedimentation of subaqueous outwash at the ice front (Rust, 1977) along with minor thrusting. The sand and gravel terraces on the south side of some segments of the ridge indicate that meltwater streams flowed across the moraine ridges into the arm of the sea in which was deposited the extensive fine grained sediments which cover the low-lying areas within zone J (Fig. 40).

The north-south ridges are very different. The silty matrix cannot be differentiated by grain size, colour, or carbonate content, from the silty marine sediments (Mm) immediately west of the ridges (McMartin, 1990). The implication certainly is that a glacier lobe surged from the east into an arm of the ice-marginal sea to the west thus reworking nearshore sediments which became a major component of the matrix of the ice contact material. Crosscutting striae lend further support to a minor westerly re-advance by a lobe in this area (McMartin, 1990). The absence of glacier surge in the ocean would be surprising given the destabilizing effect of rapid

Figure 40. *Stapylton Bay morainic ridges (phase IV); the east-west ridges are composed of south dipping beds and lenses of poorly sorted bouldery gravel (Fig. 41). The diamicton of the north-south arcuate ridge includes a high percentage of reddish silt, similar to the marine sediments to the east; S shows location of Figure 41. (NAPL A21509-68)*

calving at the ice front of glacier with terminus in an ocean (Dyke and Dredge, 1989, p. 208). In the absence of an exposure of the internal structure and composition of the ridges, it is not possible to determine the relative importance of subaqueous sedimentation at the ice front, and of stacking of material by thrusting.

Bernard Harbour drumlin fields (zone J")

The area east of a line joining Basil Bay and Stapylton Bay is covered by fluted forms which are collectively referred to as the Bernard Harbour Drumlin Field (Potschin, 1989a, b).

They include some of the most remarkable swarms of drumlins in Arctic Canada. Although they have been modified, some extensively, by wave action, they maintain the typical "baskets of eggs topography" (Menzies, 1984) with smooth flowing lines of "an inverted spoon elongated in a direction of ice movement...." (Shepps and Fairbridge, 1968, p. 293).

In some areas such as in the vicinity of Bernard Harbour the wave modification may be slight, due to the excessively stony materials which quickly "armoured" the drumlins. In other cases, modifications are more extensive (Fig. 42) although no forms are known to have been nearly destroyed.

In Figure 43, drumlin trend and mean direction are indicated, along with data on striae and grooves. Not surprisingly, the orientation of drumlins tends to parallel that of the striae left by the latest ice movement in the area.

The detailed work by Potschin (1989a, b) summarized in Figure 43 and Table 2 indicates that drumlins vary significantly between fields and also within a given field. Average length varies from 575 m near Bernard Harbour (field B), to 1200 m in the Cape Krusenstern (field C) and west of Cape Kendall (field E) areas. Average width ranges from 210 m (field B) to 325 m (field F). Individual drumlins can exceed 3000 m in length which, although impressive, is considerably less than values found in the Harding River drumlin field (see above section). Length- to-width ratios range from 1 to 17.1 with averages from 2.74 to 5.14 which, based on Shaw's (1983) classification, would place most of these drumlins in the "parabolic" class with a smaller number in the "spindle" class. Average heights range from 3.6 m (field B) to 12.2 m (field E) with a minimum of 1.7 m and a maximum of 27 m.

With the use of a water pump, trenches were cut into six drumlins. The sections were made at a right angle to the long axes of the drumlins to depths ranging from 65 cm to 130 cm. All sections exposed a stony diamicton with rare contorted sand lenses (Fig. 44). The diamicton is interpreted as a glacial till, an interpretation supported by the abundance of striations on most of the angular to subrounded clasts. Potschin (1989b, p. 90) concluded: "...that the material in all sections investigated indicate lodgment till with a top layer of till which was

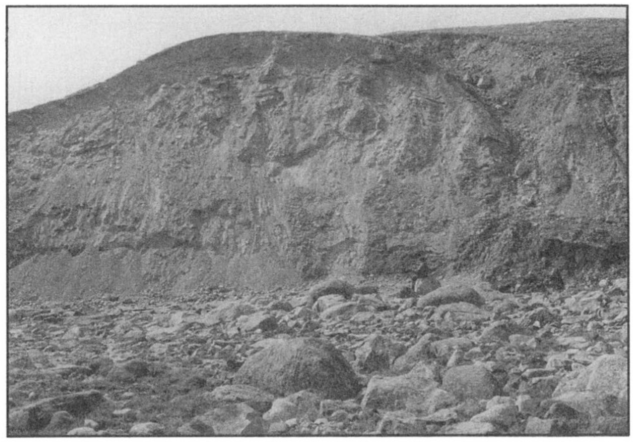


Figure 41. *A stream cutting across the east-west Stapylton Bay morainic ridges (S in Fig. 40) has exposed the coarse, poorly sorted and poorly bedded material forming the ridge. The material and its structure suggests that it is subaqueous outwash deposited at the front of an ice lobe. GSC 1993-158A*

Table 2. Morphometric data on drumlins in the Bernard Harbour region (morphosedimentary zone J', Fig. 1 and 43); from Potschin (1989b).

Zone	n	Orientation (°)	Length,L (m)	Width,W (m)	L,W ratio :1	Height (m)	Slope (°)
A	247	262	950	233	4.07	9.0	6.4
		242-315	243-3406	104-556	1.3-16.8	3.4-27	1.5-23.5
B	464	238	575	210	2.74	3.6	1.3
		193-271	209-1703	104-487	1.0-12.3	1.7-5.0	1.0-1.5
C	174	216.5	1200	233	5.14	8.5	2.3
		200-241	350-3200	104-420	1.5-17.8	3.4-13.5	0.5-4.5
D	722	266	731	214	3.42	6.0	2.8
		212-310	243-1772	104-521	1.0-7.5	3.4-10.0	1.0-6.5
E	197	307	1200	325	3.68	12.2	4.9
		253-341	313-2641	70-835	1.3-16.3	4.8-17.8	2.0-9.0
F	341	227	751	213	2.52	12.0	4.3
		195-272	272-2502	104-382	1.4-14.4	5.0-20.0	2.5-6.5

* The first line represents mean values, the second line represents the range of measured values.

reworked by the sea". Till fabrics reported by Potschin indicate that the generally weak preferred orientation deviated from the long axis of the drumlin by 8° to 65°.

Because drumlins are such ubiquitous landforms in the Inman River area it is important to discuss the probable origin of this contentious glacial landform.

The origin of drumlins has been extensively discussed in the literature particularly in the last decade. Boulton (1987, p. 31) summarized the current hypotheses as follows:

"Several major sets of processes have been invoked to account for the origin of drumlins and related forms, of which the following are current:

1. Drumlins as the product of subglacial deformation (Smalley and Unwin 1968, Evanson 1971, Boulton 1982).

Figure 42. This drumlin, 6 km west of Bernard Harbour (zone A, inset, Fig. 43), has been intensely modified by wave action which has concentrated coarse material into flights of storm beaches; top of photograph is towards the east, elevation of drumlin is approximately 20 m, note fine grained marine sediments (silty sand) in upper right hand corner. GSC 204720A

2. Drumlins as the product of subglacial lodgement (Fairchild 1907, Boulton 1982).

3. Drumlins as the product of melt-out of basal debris-rich ice (Shaw 1980).

4. Drumlins as fluviatile infillings of subglacial cavities (Shaw 1983, Dardis and McCabe 1983)."

To this should be added:

"5. Drumlins might form as remnant ridges resulting from erosion of the glacial substrate by subglacial meltwater sheets" (Shaw and Sharpe 1987, p. 2316).

Boulton (1987, p. 74) further succinctly summarized the different views as follows:

"There are two principal groups of drumlin formation theories; those which assert that drumlins are produced by glacial transport of bed material or basal debris, and those which assert that they are intimately related to subglacial fluvial activity."

In her study of Bernard Harbour drumlin field Potschin (1989b) noted that:

a) throughout the area drumlins and striae on bedrock have the same orientation;

b) that the downstream end of numerous drumlins in the area north of Basil Bay appear to be truncated (Fig. 45) resulting in concave down ice shores of small lakes;

c) that in the same area herringbone or en echelon patterns are particularly prominent (Fig. 45 and 46).

These observations led her to conclude that the erosion of a till sheet by a strong subglacial turbulent meltwater flow is responsible for the sculpting of the drumlins. This hypothesis best explains the "morphology of the drumlins, the herringbone pattern...the truncation of many drumlins and the 'negative forms between drumlins'." (Potschin, 1989b, p. 159).

There is no doubt that the "meltwater hypothesis" has several attractive features, as argued by Sharpe, for the Victoria Island drumlin field on the northside of Dolphin and Union Strait (Sharpe, 1988, 1992); however, as Sharpe (1992, p. 41) judiciously points out "Drumlin cores may not allow direct understanding of drumlin formation if they are erosional forms".

Difficulties with the erosion by meltwater hypotheses are:

a. The presence of striae parallel to the long axis of drumlins on bedrock outcrops between fluted forms is hardly compatible with erosion by powerful turbulent debris-laden flow. Such delicate features on relatively soft rock should have been obliterated and replaced by "S" forms and other water erosion features.

b. The amount of sediment that would have been removed by the assumed catastrophic flood is enormous yet there is no clear evidence of it in the marine basin exposed by postglacial uplift. As pointed out earlier, (see Marine offlap area (zones I and J) above), the sediments fine seaward from the marine limit deltas. Except for the morainic ridges in Zone J there is no significant accumulation of coarse material in the narrow corridor which existed between the ice front and the uplands to the west. Zone J, between Cape Young and Klengenberg Bay, would have been an ideal sediment trap for subglacial flood deposits from ice to the east. No evidence has been found except for Kerr's (1994, p. 15) "Facies B: cross-stratified sand and gravel". This facies, part of Zone H is discussed (in section Marine limit kame complex (zone H)). It is consistent with subglacial meltwater streams debouching from ice tunnels into the sea. It cannot account for the nature and quantity of sediments assumed by the meltwater erosion hypothesis.

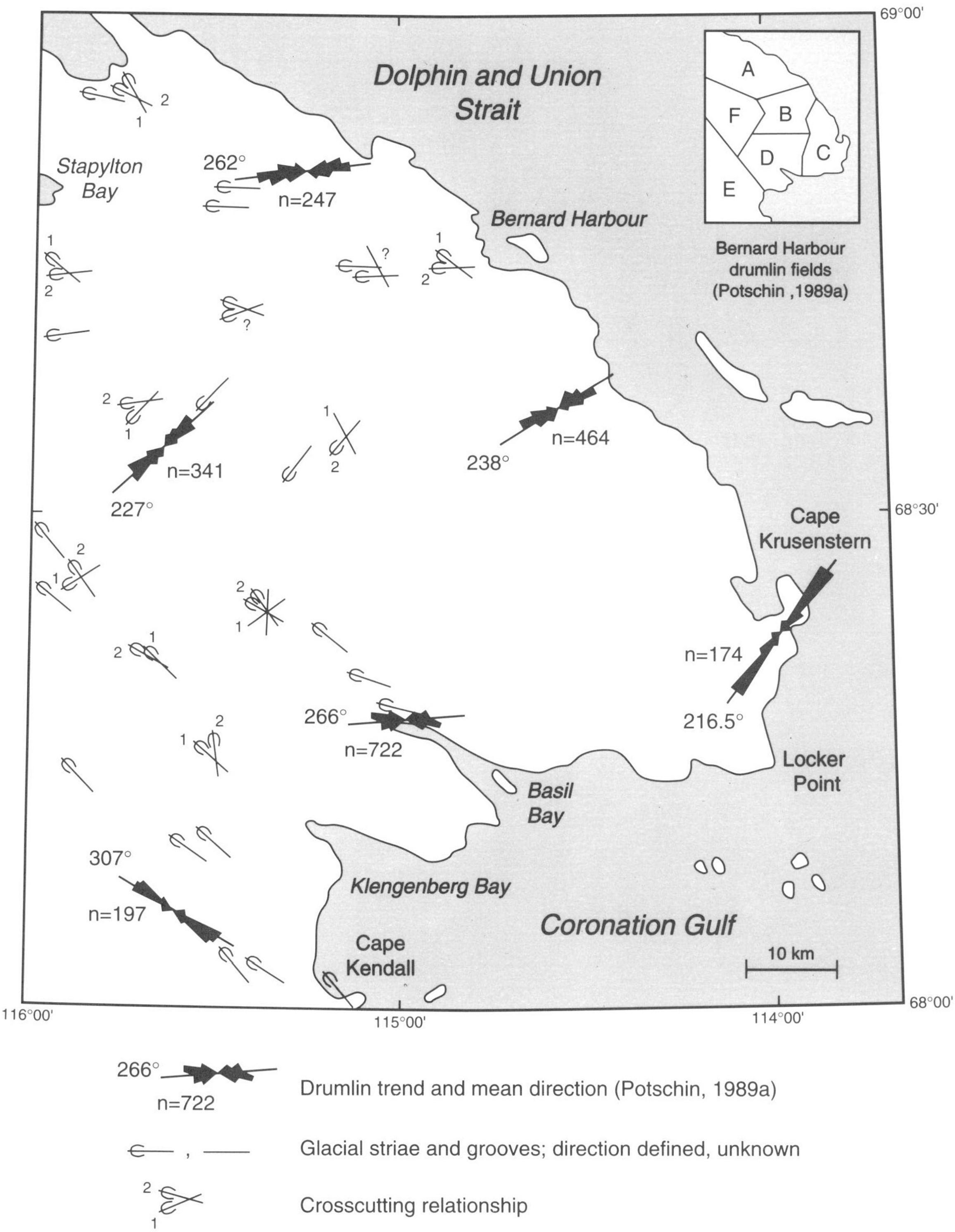


Figure 43. Drumlin trends and ice direction indicators for the Bernard Harbour region.

c. The herringbone and en echelon patterns do not require the meltwater hypothesis. The original, largely compressive flow of the ice mass responsible for the glaciation of the Inman River area flowed from east to west. During deglaciation the encroaching sea along Admundsen Gulf and Dolphin and Union Strait triggered extensive flow, generally to the northwest, but with many local variations which are reflected in crosscutting striae and, particularly near the ice front, in crosscutting drumlins as, for instance, in the northwestern part of zone H. Rapid disintegration of the ice front by calving into the encroaching sea would have created instability in ice flow dynamics with many local surges at oblique angles to the main drumlin-forming flow.

d. Potschin (1989b, p. 163) points out that "...the preferred orientation of pebbles indicates lodgment till rather than melt out till [in the drumlins]. However, [the pebble orientation] deviates slightly from the direction of the investigated drumlin crests...". In this case the erosion by subglacial meltwater hypothesis is certainly intriguing but not necessary. The pebble orientation which is consistently west to northwest (Potschin 1989b, Table II, p. 165) reflects ice movement at the time of till deposition. The orientation of the crests of the various drumlin fields vary significantly and is related to ice surges during late phases of deglaciation. Unless the till was completely remobilized during drumlin formation, Figure 45 shows that the older forms are not completely remoulded, there is no particular reason why these two orientations should coincide.

In summary, except for the truncation of the down-in tail/up-in ice nose of some drumlins, which may be more easily explained by water erosion, it is not necessary to invoke the hypothesis which proposes that drumlins in the Bernard Harbour area are the result of erosion by powerful subglacial meltwater flow. Indeed this hypothesis raises insurmountable problems in view of the facts as presently known based on the detailed mapping of the Quaternary geology of the region.

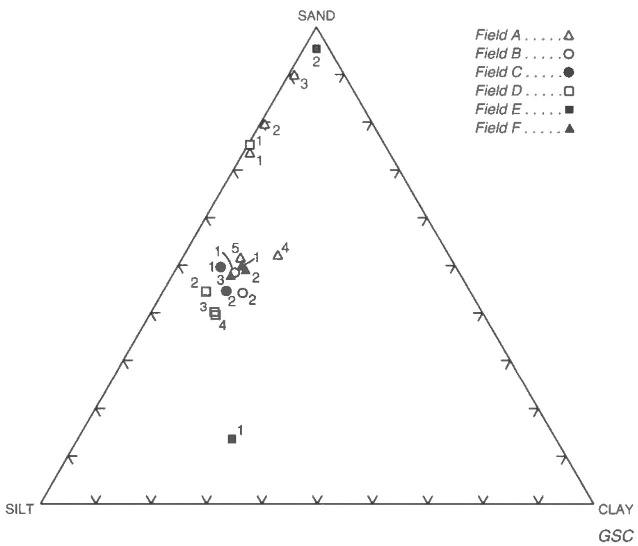


Figure 44. Ternary diagram comparing matrix in till samples from drumlin fields in the Bernard Harbour region (morphosedimentary zone J', Fig. 1 and Fig. 43).

Modern beaches

Along the present shoreline a fringe, seldom more then 20 m wide, marks the zone affected by weak tidal currents, storm waves, and more importantly, by the ploughing action of pans of sea ice (Fig. 47). Except at the mouth of rivers where the material is dominated by sand, modern beaches are composed of pebbly gravel in a coarse sand matrix. Two zones can be recognized: the lower is between 0 and 2 m above low tide line and, the upper between 2 and 5 m above low tide line.

During the spring and early summer, pans of ice pushed by wind plough through the shore zone scouring the pebbly sand and bulldozing it into ridges several tens of centimetres high (Allard and Tremblay, 1983; Dionne, 1988). These ridges are temporary, as waves will smooth them out later in the summer. This is why these forms are seldom found on raised beach ridges. Withought ice-push, as pointed out by Dyke et al. (1992, p. 51), beach ridges would not form in an environment where marine wave and current energy are dampened by ice cover and limited fetch.

The gravely sand of the 2 to 5 m above low tide zone is reworked by waves generated by late summer/early fall storms. The surface is littered with driftwood which diminishes rapidly eastward. Other debris include sea mammal carcasses and, unfortunately, human litter such as oil drums, plastic, and an extensive disposable diaper deposit.

MAJOR PERIGLACIAL LANDFORMS

Harding River rock pingo

A spectacular pingo, first reported by Craig (1960), is in the middle of a marshy area adjoining a small lake on the west side of Harding River, approximately 50 km from its mouth. The 22 m high pingo stands above the irregular surface of an upland averaging 230-240 m a.s.l. The jagged hill is composed of eight pie-shaped slabs of dolomite that have been uplifted and now dip between 35° and 48° (Fig. 48). Glacial striae and grooves are well preserved on some slabs, particularly on those facing south-southeast. At the top of the pingo is a circular grass surface surrounded by the jagged tips of the bedrock slabs that form this tight dome (Fig. 49). At the base of the bedrock chevrons a talus of debris forms an apron around the pingo.

The Harding River Pingo is the only closed system rock pingo reported in the literature so far. St-Onge and Pissart (1990) suggested that the following circumstances made the growth of the pingo possible: 1) the upper few metres of the dolomite bedrock is intensely fractured; this results in a porosity of at least 10%; 2) during deglaciation the area was covered by a glacial lake; this created favourable conditions for the establishment of a talik beneath the lake and the eventual growth of a pingo upon the partial drainage of the lake as the ice front receded. These events led to circumstances similar to those which gave birth to the numerous

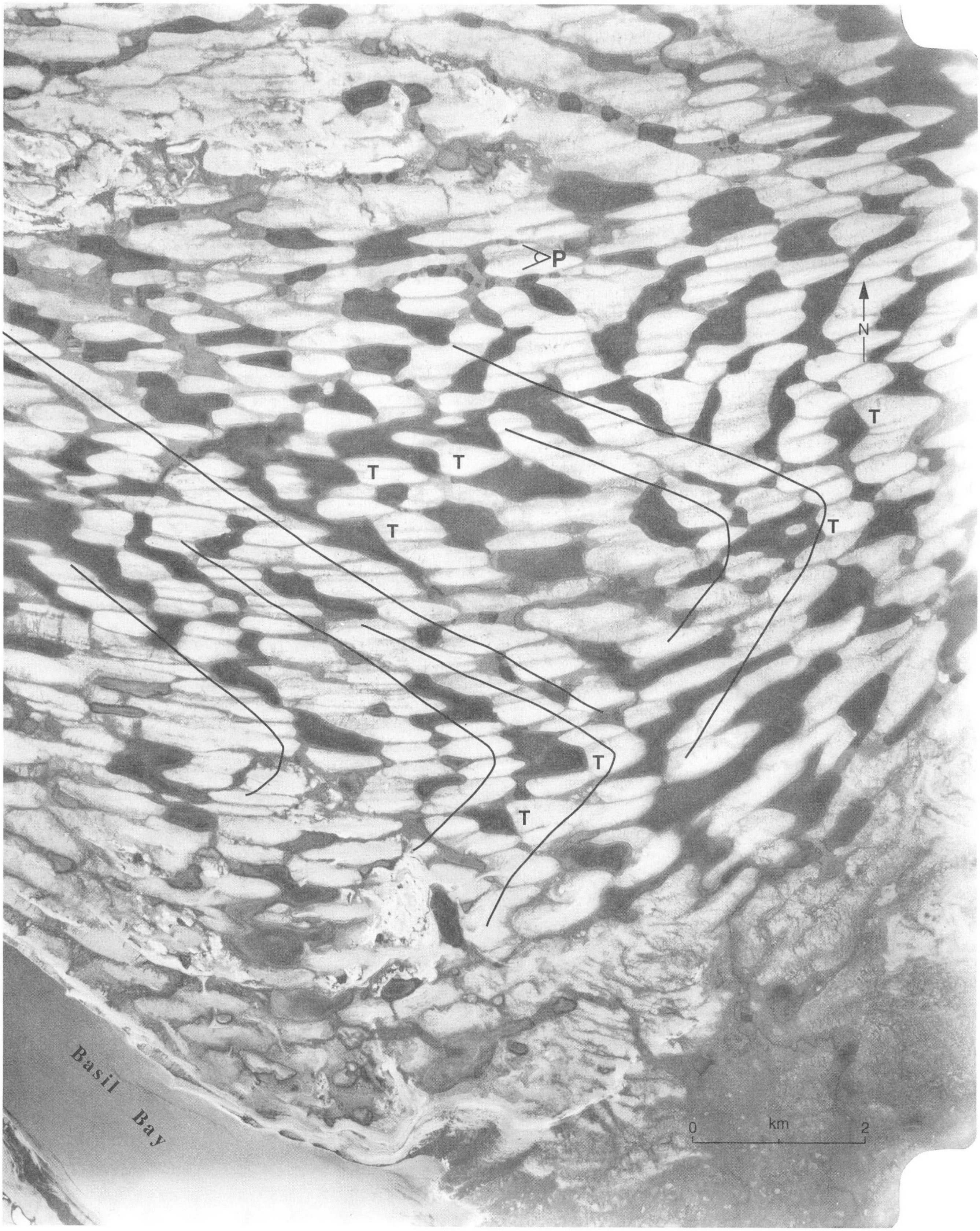


Figure 45. *Drumlin field north of Basil Bay; flow direction changes from southwest to east-west and then to west-northwest; note that drumlins align northwest-southeast then southwest to form a herringbone pattern; the down-ice portion of some drumlins appears truncated (T); P locates viewpoint for Figure 46. (NAPL A21510-230)*

Figure 46. Low level oblique airphoto looking west in the northern part of drumlin field north of Basil Bay; short stubby drumlins have been extensively modified by wave washing. GSC 1993-121H

pingos of the Mackenzie River Delta (Mackay, 1979). The occurrence of intensely fractured rock and of a glacial lake deep enough to maintain a talik that will allow the growth of an ice core is a combination of factors that is probably seldom achieved. This is the probable explanation why open system rock pingos are so uncommon.

There are several other pingos mostly in the central part of the Inman River area as shown on the Surficial Geology Map 1864A. All occur in shallow depressions, usually drained lakes or ponds, and they resemble the classic closed system pingos of the Mackenzie River Delta (Mackay, 1979).

Rock blisters

Dolomite beds, varying in thickness from 5 to 150 cm, are cut by fractures which delineate angular blocks with sides ranging in length from 10 cm to 1 m or more. Where recently exposed, the bedrock surface is smoothly grooved and striated, but where long exposed, striations and other flow indicators have been removed, and in many places the surface has been differentially dissolved into kerren. The surfaces of these platforms are commonly deformed by mounds and small rounded hills which result from the updoming of the beds of dolomite (Fig. 50; 51A, B).

Although rock blisters display great variations in detail, they have the following characteristics in common:

1. They occur on presently well drained surfaces of dolomite benches (Fig. 50).

2. They are 1 to 3 m high and 3 to 5 m in diameter (Fig. 51).

3. Mounds less than 1 m high do not display a collapsed centre.

4. A small crater is commonly present at the top of higher mounds with the collapsed blocks occupying the base of the depression (Fig. 51).

5. They are composed of imbricated angular blocks of dolomite which retain the glacially smoothed or the kerren surface characteristic of the bench surface (Fig. 51A).

6. The tilted blocks, which may comprise several beds, are between 80 to 100 cm thick (Fig. 51B).

7. At present there is no ice in the core of the mounds.

The general appearance is that of a jigsaw puzzle pushed up from underneath; when the upward pressure was removed, the jagged angularity of the blocks prevented their falling back to their original position.

There is no doubt that the rock blisters result from the uplifting of blocks which were part of a glacially smoothed surface of horizontally bedded dolomite. The shape and size of the blocks are closely related to two dominant subvertical sets of fractures trending 340° and 260°, but these fractures are not the controlling factor because, in many places, the mounds are absent from surfaces that are intensely dissected by these near perpendicular dominant sets of fractures.

The Bernard Harbour area, with a mean annual air temperature of -12.5°C, lies well within the continuous permafrost zone (French, 1976). Several excavations in glacial drift and in coarse raised marine beach sediments indicate that the active layer was up to 1.5 m thick in July 1988 and was probably thicker in bedrock (L.D. Dyke, 1984). Thus, the conditions certainly exist for the formation of frost heave features.

Rock pingos do exist in the area as discussed above. Rock blisters in the Bernard Harbour area have no ice core and, furthermore, lichen cover and weathering on block faces clearly indicate that no disturbance has taken place for a long time.

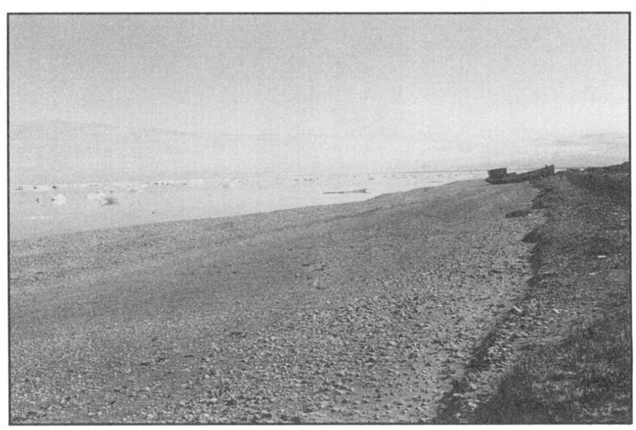


Figure 47. Modern beach near Clifton Point; the upper limit of the active beach is marked by a 20 to 50 cm scarp below which the wave washed, gently inclined surface is covered with sandy gravel and coarse sand; between this gentle slope and the shoreline several ridges have been built by ice pans pushed on shore; the abandoned 15 m wooden freighter is the "Gertrude B". GSC 1993-121B

Similar features to the rock blisters discussed here have been reported from numerous parts of Arctic Canada: the east coast of Hudson Bay (Dionne, 1981), northern Quebec (Payette, 1978), eastern District of Mackenzie (DiLabio, 1978), and Baker Lake area, District of Keewatin (A.S. Dyke, 1984). In all cases, domed structures occur in jointed or sheeted rock and in discharge areas where water moved in joints and bedding planes.

In his study of frost heaving of bedrock in permafrost regions, L.D. Dyke (1984) described "dome-shaped rock heave features" which, although in quartzite, appear identical to the rock blisters discussed here. Dyke proposed that features formed by the expulsion of excess water pressures in the saturated active layer reach an equilibrium height and show little further movement. Domes are similar to the closed-system type of pingo. Water saturated fractured dolomite and a progressing freezing plane result in heaving forces from high

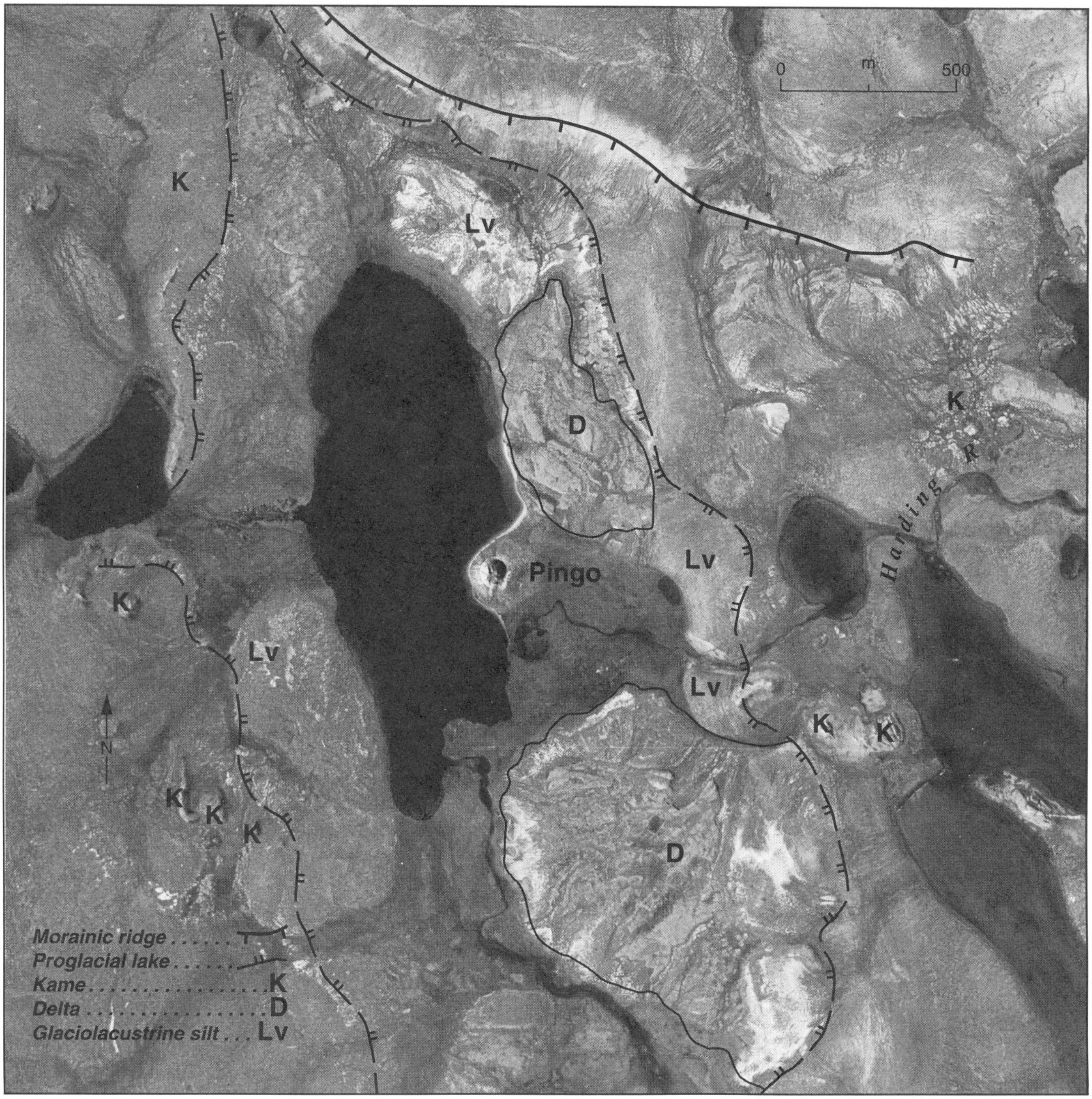


Figure 48. In this enlargement of airphoto NAPL-A21509-94 (Fig. 56) the Harding River rock pingo is shown in the centre of a glacial lake which existed during deglaciation Phase III (Fig. 54).

Figure 49. This low level oblique airphoto of the Harding River rock pingo shows the uplifted pie-shaped dolomite slabs, the mantle of debris surrounding the base and the jagged nature of the summit rim. GSC 205295A

pore water pressures. In the dolomites of the Bernard Harbour area the near vertical fracture sets and the subhorizontal bedding acted as the controlling discontinuities that favoured the development of ice-cored domes.

Doming of bedrock layers usually produces a dilation and when the ice lens melts, partial collapse or total disruption occurs depending on the relative thicknesses of the ice body and rock layer. Dyke was successful in predicting the maximum size of dome-type features in central District of Keewatin from relationships between the total dilation and the height of the dome. In the Bernard Harbour area, because rock blisters have partly collapsed, maximum original size can only be estimated – approximately 5 m in diameter by 3 m in height. The dilation for that diameter and height may have been so large that the ice core was exposed, leading to eventual melting and collapse during the annual thaw of the active layer. Once the miniature pingo had collapsed into a mound of blocks, conditions no longer existed for the formation of the high pore water pressures that would allow the process to start again.

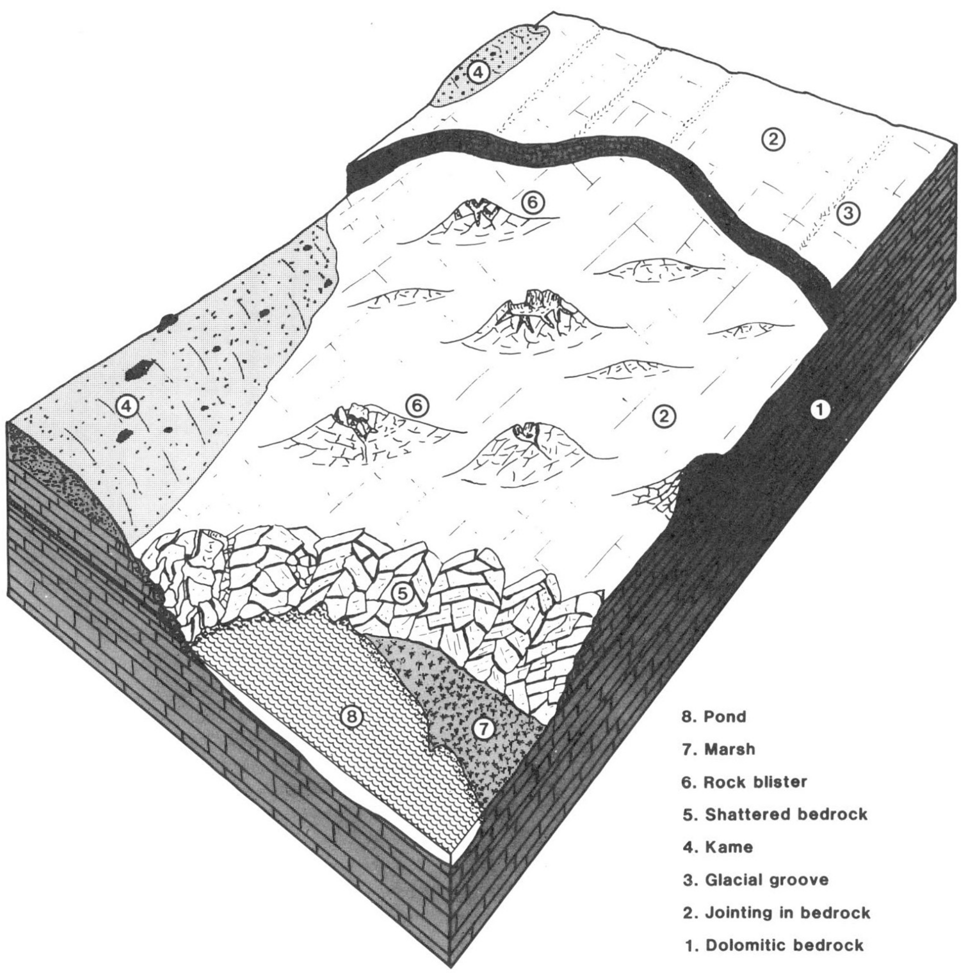


Figure 50. Block diagram illustrating typical location where rock blisters occur. From St-Onge et al. (1989).

Rock blisters are built of angular weathered blocks. Given that the only reasonable explanation for the formation of rock blisters is the growth of ice in a confined space and since no ice is present today, when were conditions most favourable for their formation?

The mounds occur on bedrock surfaces that have been smoothed by glacier ice or by meltwater abrasion, therefore they postdate the last glaciation. No permafrost existed under the surface which was uncovered as the glacier retreated (Washburn, 1979, p. 61). The growth of permafrost would have started soon after the area became ice-free or when the sea retreated from the coastal regions. The melting of snow supplied all the water required to saturate the bedrock surface which had become intensely fractured as a result of decompression by ice removal if not earlier.

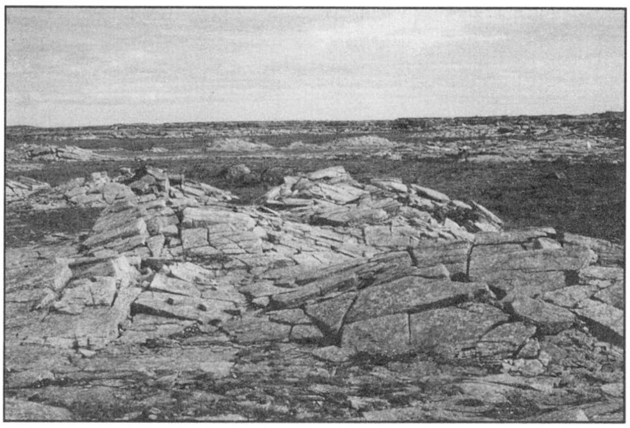


Figure 51A. Group of rock blisters in similar conditions to those illustrated in Figure 50; note the generally even size, dominantly rectangular shape of the blocks, and open central crater; the glacially smoothed surface is clearly visible both on the undisturbed bench surface and on the uplifted blocks. GSC 204662C

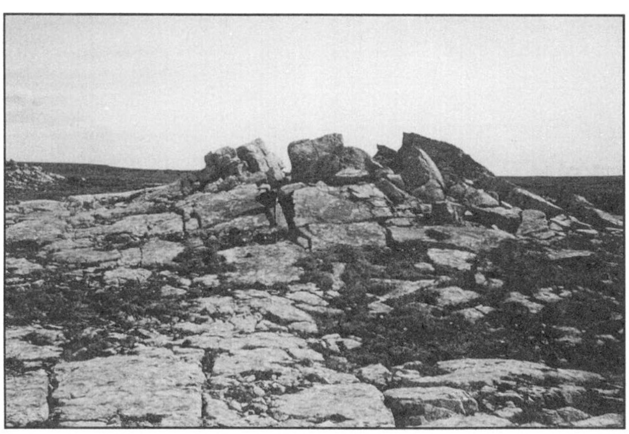


Figure 51B. Detail of a rock blister showing the uplifted beds, rectangular shaped blocks, and collapsed centre; shovel handle, centre photo, is 50 cm long. GSC 204662B

Permafrost established to a depth of a few metres provided the impermeable lower barrier required to generate high pressures in the active layer during the annual freeze-back. The updoming of the rock blisters, resulting from the enormous pressures generated by ice growth in the saturated active layer, shattered the bedrock surface. The uplifted and tilted blocks did not fall back into place but came to rest as an open work fabric which precluded any further development of rock-breaking pressures in the active layer. Each rock blister thus acted as a pressure release valve for the immediate area surrounding it.

Rectangular grid in bedrock

Rock blisters represent a particular response to high pressure generated during the annual freeze-back of the active layer of newly established permafrost in this area underlain by beds of dolomite. Similar conditions in other lithological units produced different results. In shaly sandstone in the Locker Point area (Fig. 1) frost heaving has produced rectangular grids up to 50 by 120 m, bounded by 2 to 3 m deep trenches (Fig. 52 and 53) as a result of uplifting and tilting of beds on either side of presumed vertical fractures. Similar features are found on Wollaston Peninsula (Sharpe, 1992, p. 18) and in the eastern Arctic (Dredge, 1992, p. 316) and at many other sites in the central Arctic (L.D. Dyke, 1984, p. 5-6; Dyke et al., 1992, p. 94).

LATE WISCONSINAN DEGLACIATION PHASES

The ice frontal positions identified in Figure 54 are inferred from the distribution of major continuous morainic ridges, which commonly display thrust features, ice contact meltwater channels, extensively washed surfaces, glaciolacustrine sediments and lake strandlines, and major lithofacies assemblages (McMartin and St-Onge, 1990). No attempt has been made to assign specific ages to any of the ice frontal positions.

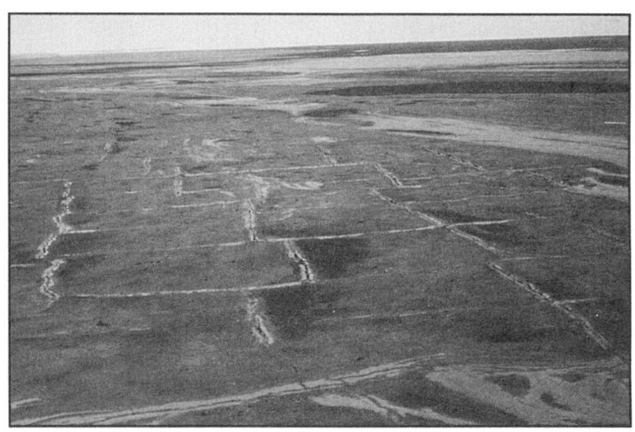


Figure 52. Rectangular grid, up to 50 by 120 m, delimited by uplifted shaly sandstone beds; note trenches extending across raised marine beaches in the lower right of the low level oblique airphoto. GSC 204662A

Figure 53. Uplifted shaly sandstone beds and trenches forming margins of 50 m grid pattern. GSC 204662D

The whole sequence is unarguably Late Wisconsinan with age determinations on marine shells providing minimum ages for the deglaciation of areas covered by the postglacial sea. The trend of intermediate ice frontal positions is also indicated where the presence of morainic or other features permit.

Bluenose Lake phase

To the south and west of Bluenose Lake, the limit of Late Wisconsinan active ice is marked by a complex morainic ridge with a steep easterly slope (Fig. 13) The ridge is interpreted as glaciogenic debris carried along the hanging wall of a thrust-plane at the limit of the actively flowing ice (Fig. 21; St-Onge and McMartin, 1987). West of this ridge the drift is thin and is restricted mostly to moulin kames, thin glaciofluvial deposits with collapsed structures to the east, and thin, usually contorted rhythmites also collapsed to the east (Fig. 1, zone A).

South of the Bluenose Lake region, in the Bebensee Lake area, the early westward-trending ice was also recognized by Prest (1985) who proposed a series of deglacial events with glacier recession generally towards the east.

North of Bluenose Lake, the ice flowed to the northwest into Amundsen Gulf, while south of the lake it skirted the southern limit of the Melville Hills to flow west and eventually to the northwest into Hornaday River basin or to the southwest into Great Bear Lake basin (Klassen, 1971). Following this maximum advance to the Late Wisconsinan ice margin, thinning occurred and drawdown within Amundsen Gulf and Great Bear Lake resulted in numerous surges and redirection of ice flows as explained in the following events.

The ice thrust moraine lies at approximately 600 m a.s.l. and marks the western limit of Late Wisconsinan ice advance on the eastern slope of the Melville Hills (Fig. 54). Several lines of evidence support this hypothesis. Sediments exposed along Croker River (Fig. 27) record a complete glacial cycle. Pre-last glaciation fluvial gravel and sand at the base, which are overlain by a fining upward sequence of coarse to fine sand, are interpreted as resulting from the blocking of the lower part of the pre-last glaciation Croker River. The glacial lake rhythmites of silt and silty clay are overlain by a massive bouldery diamicton interpreted as a basal till. The till is in turn overlain by rhythmites formed in an early postglacial lake, which are capped by coarse, steeply dipping foreset beds of coarse sand and gravel, which on the surface display numerous collapsed structures on the proximal eastern side (St-Onge and McMartin, 1987, p. 98, Fig. 10.7). These delta kame deposits cannot be distinguished in nature, in freshness, or in any other way from a whole flight of similar features which extends from the area just north of Bluenose Lake to the delta kames which mark the marine limit.

The age of this ice limit is contentious and may not be resolved until datable material is found in sections along the Croker River or in stratigraphically equivalent sites. Based on air photo interpretation Vincent (1984) proposed that this ice limit was probably Early Wisconsinan and that the Inman River drumlin field was the result of Late Wisconsinan glaciation; however, our work clearly demonstrates that the Bluenose Lake Moraine and the Inman River drumlin field are results of distinct phases of the same glacial event.

The Bluenose Lake Moraine contains numerous small glacial lake basins within major east-facing re-entrants in the hummocky region (Fig. 1 and 11). Generally speaking, these lacustrine deposits end abruptly to the east where they terminate in channels trending to the north-northwest and ending in the Inman River drumlin field. There is, therefore, a continuity of landforms and lithofacies assemblages between the Bluenose Lake Moraine and the drumlin field. It can be further argued that if the limit between the Inman River drumlin field and the Bluenose Lake Moraine marked a significant age break, as suggested by Vincent (1984, p. 94, 96), a major northwest-trending meltwater channel should be found north of Bluenose Lake. This channel would have allowed drainage during the late Wisconsinan phase when ice would have occupied only the Inman drumlin field and adjacent lowlands. Such is not the case. Indeed, Croker River is superimposed on the drumlin field across which it has carved its valley. This is only possible if it is assumed that the northward flowing river draining the Bluenose Lake basin flowed on top of the ice across part of the Bluenose Lake Moraine and across the drumlin field. All the evidence strongly supports the hypothesis that the Bluenose Lake Moraine and the Inman River drumlin field are part of the same glacial events, although they represent distinct ice dynamics within this complex.

Active ice flowing upslope along the east side of the Melville Hills and attaining an elevation of 600 m implies great thickness of ice to the east including southern Victoria Island. Sharpe (1992, p. 60) reached the conclusion that "...Late Wisconsinan ice covered all of Wollaston Peninsula to ice limits lying well beyond southwestern Victoria Island (Sharpe, 1984), perhaps at the west end of Amundsen Gulf as shown on the Glacial Map of Canada (Prest et al., 1968)".

To resolve these "apparent incongruities in interpretation of the Late Pleistocene deposits and ice limits" Vincent (1992, p. 233) proposed that "...extensive Keewatin Sector Ice of the Laurentide Ice Sheet may have first advanced in northwestern Canada during the Sangamonian (broad sense)/early Wisconsinan and remained there until it finally disappeared in the Late Wisconsinan". This view is not incompatible with our data which strongly support a Late Wisconsinan ice retreat from a front resting on the eastern slope of Melville Hills just east of Bluenose Lake. However, we have no indication of when the ice build up occurred nor do we know in which previous glaciation did glacier ice override Melville Hills (Jones et al., 1992, p. 31). As pointed out earlier, evidence for active ice is lacking in morphosedimentary zone A. This suggests that if such a glaciation ever occurred, it must be quite old, i.e., Mid-Pleistocene or older. Our information is obviously sparse and there is a need for detailed mapping west of Bluenose Lake, but our results to date support the model proposed by Dyke and Prest (1987b).

Inman River phase

When the westward flowing ice margin receded to the Inman River plateau, the compressive flow, typical of the Bluenose Lake phase, was replaced by extensive flow which was redirected towards the calving bay in Amundsen Gulf to the northwest. This high velocity flow moulded drift into large drumlins named the Inman River drumlin field by St-Onge and McMartin (1987). The northward trending ice flow observed in the Bebensee Lake area (Prest, 1985) is presumed to be associated with this phase, indicating that the Amundsen Gulf drawdown captured flow up to 250 km inland.

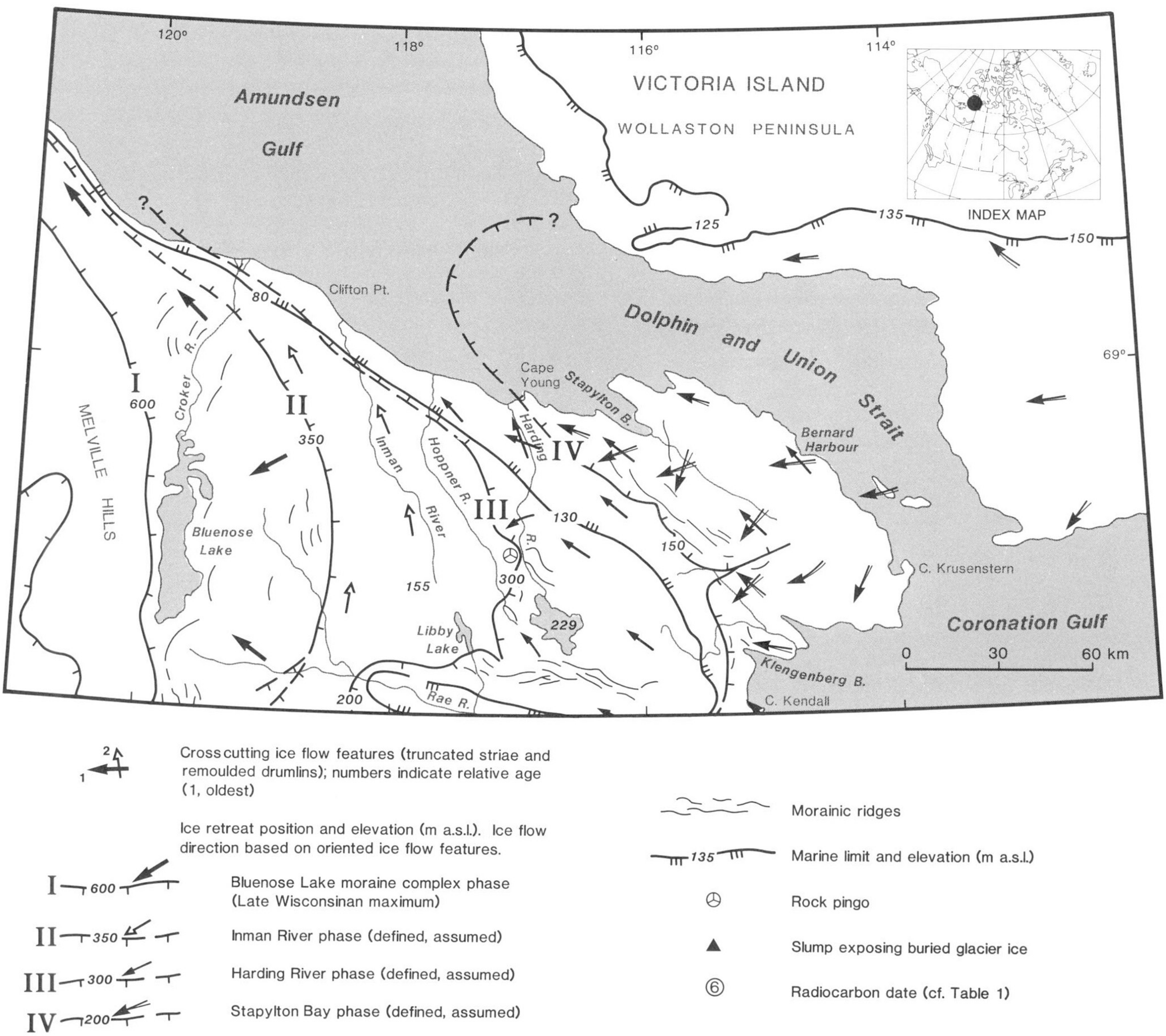


Figure 54. Late Wisconsinan ice retreat phases south of Dolphin and Union Strait.

Final ablation of the ice was by massive downwasting as demonstrated by the absence of morainic ridges of any sort. Several esker ridges, which wind between the drumlins, also suggest downwasting over extensive areas. However, outwash fans with ice contact features to the east and northeast, suggest that the stagnant ice mass was shrinking towards the lowlands in an east or northeast direction (St-Onge and McMartin, 1987).

In the Coppermine River valley region, St-Onge (1987) defined the Forcier Lake Moraine Phase which is the most westerly ice frontal position recognized in that area; with an elevation between 325 and 400 m a.s.l., it is possibly part of the 350 m ice frontal position of the Inman River phase. This was suggested by Kerr (1987a, p. 2139).

Harding River phase

As a result of rapid thinning, the northwestward flowing ice was divided into two distinct lobes centred in major basins. It is probable that the evidence for westerly- and southwesterly-flowing ice north of Bebensee Lake (Prest, 1985) represents the early inception of the lobe which was to occupy the broad basin of the Richardson River and Rae River. The second lobe that resulted from the shrinking of the main ice mass to lower ground occupied the channels of Amundsen Gulf and Dolphin and Union Strait.

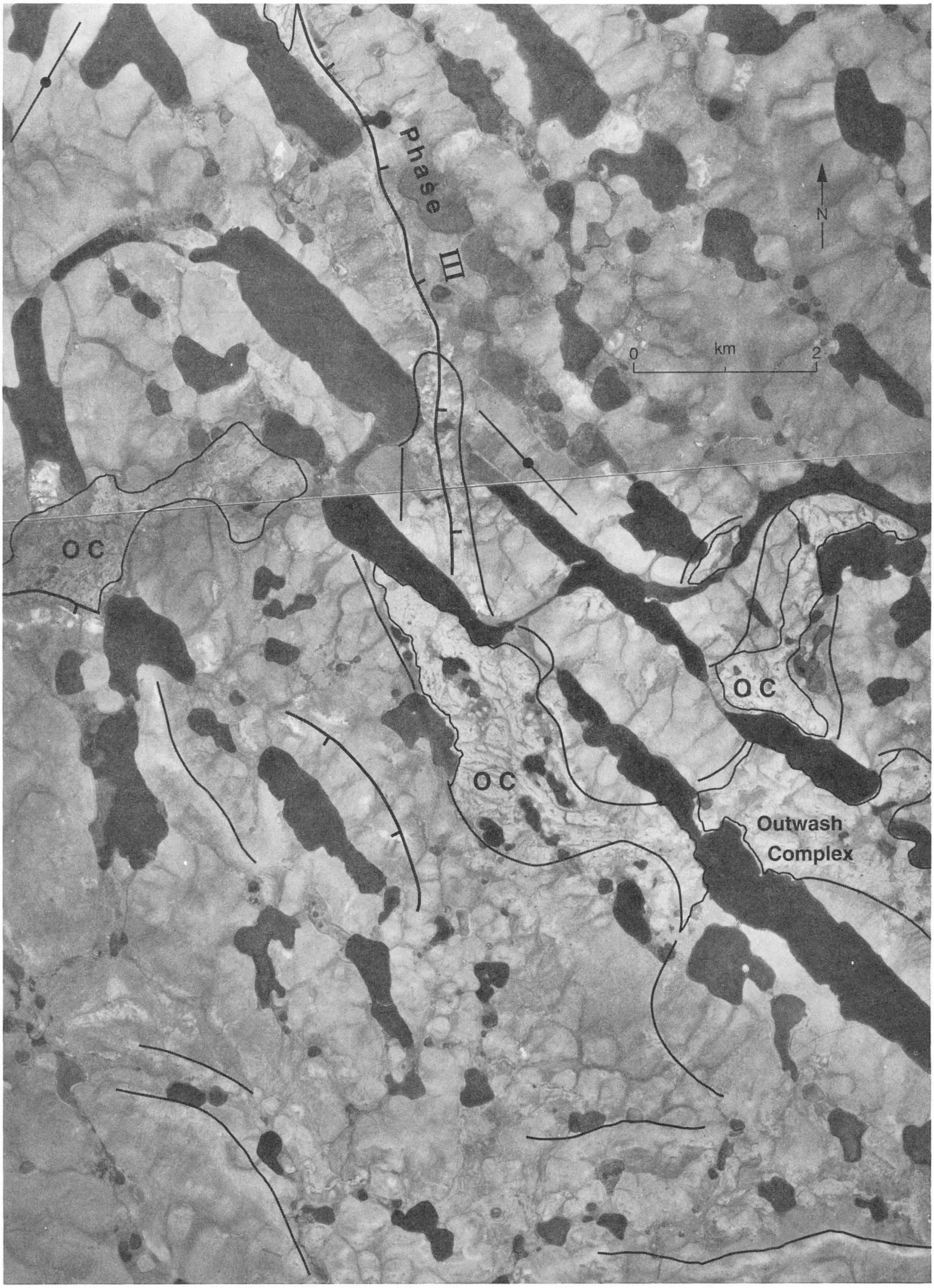
West of the large unnamed lake from which Harding River originates (Fig. 54, Lake "229"), a major, north-south morainic ridge, averaging 20 m high, occurs at 300 m a.s.l. (Fig. 55). This nearly continuous ridge displays a distinctly steeper east-facing slope; outwash fans and deltas rest on its western slope. It represents the frontal position of the ice mass occupying the upland area between the two lobes centred in the depressions to the north and south. Two major breaches in the morainic ridge are filled with large outwash terraces of gravel and sand which are bounded by several minor bouldery gravel ridges oriented northwest-southeast (Fig. 55, 56). Drumlins associated with these minor morainic ridges indicate that the major ice front was affected by small crosscutting surges. North of Harding River, the ridge strikes towards the northwest and becomes a gravelly kame occupying an extensively washed glaciofluvial corridor which ends abruptly at marine limit at 115 m a.s.l. (Map 1846A, in pocket). The drumlins, along the coast, from Hoppner to Croker rivers, are oriented towards the northwest. The ice flow in that area was identical during both Inman River and Harding River phases.

Between Libby Lake and Lake "229", the ridge extends to the southeast into a pitted outwash complex bordered by collapsed structures (Fig. 55). This results from intense sedimentation between the ice resting on the upland and the Richardson River and Rae River lobe to the south. Numerous ice marginal lakes were formed during this phase; one example is ancestral Lake Libby in which fine sediments were deposited up to 20 m above the present lake level. Surging by the lobe occupying Richardson and Rae river basins into ice marginal lakes at its western extremity, led to the formation of westward- and northwestward-trending drumlins (St-Onge, 1988). Sedimentation of till and outwash, along the steep northern slopes of the basin, produced a complex of large morainic ridges trending east. These ridges rise from 240 to 280 m a.s.l. and are concentrated in a corridor 70 km long and 10 km wide (Fig. 1, zone E). East of Libby Lake, the ridges have not been modified by wave action nor do they comprise glaciolacustrine sediments. Therefore, they represent ice marginal positions prior to the existence of ice marginal water bodies and to the incursion of marine waters. At this stage, ice marginal lakes existed only at the western extremity of the lobe (St-Onge and McMartin, 1987). Shrinking of this lobe below elevations of 240 m permitted the coalescing and extension of the ice-marginal lakes to a point where they fringed its northern margin, thereby accelerating the disintegration of the ice by extensive calving. Glacial lake outlets were identified by Kerr (1987a) at elevations of 240, 210, 180, and 165 m a.s.l. west of Klengenberg Bay. However the extensive stripping of large bedrock surfaces by meltwaters and the large deposits of bouldery gravel at the downstream end of these channels suggest that these corridors, which may have served as glacial lake outlets, were carved mostly by high energy meltwaters flowing from glacier ice to the east resting along the steep slopes west of Cape Kendall.

South of Cape Young, the marine incursion from the northwest led to a minor redirection of the ice flow towards the west-northwest. Numerous northwest-striking drumlins have been modified by a west-northwest ice flow (Fig. 38). Intersecting striae show a similar relationship. Craig (1960) originally proposed that a lobe coming from Victoria Island had been overridden by another lobe moving to the northwest. However our evidence suggests that these features are minor changes in the direction of flow near an unstable ice front.

As the lobe occupying Richardson and Rae river basins was shrinking, so was the ice lobe occupying the channel of Amundsen Gulf and Dolphin and Union Strait. Eventually this allowed the sea to invade the isostatically depressed lowlands along a corridor between the land and the ice lobe (St-Onge and McMartin, 1987). The deglaciation of the coast and the synchronous marine invasions occurred prior to 11 170 ± 80 BP (TO-1231), a radiocarbon date obtained from *Portlandia arctica* shells collected at 125 m a.s.l. in glaciomarine rhythmites overlying subaqueous outwash gravels and sands (Table 1, Fig. 55, 57). The corridor between sea and ice made it also possible for marine waters to invade the Richardson and Rae rivers basin from the northeast. The oldest radiocarbon age on marine shells in the basin is 10 300 ± 240 BP (GSC-3663, Table 1) obtained from shells collected at 90 m a.s.l. This elevation is well below the marine limit of approximately 150 m a.s.l. in the west at the head of the basins but which rises to 170 m along Coppermine River to the east (Mercier, 1984). The difference in age is probably due to the absence of an early marine fauna in a marine arm in which large quantities of meltwaters flowed.

During the final stages of the Harding River phase, the ice front was in contact with the sea. Near marine limit large areas were intensively washed and modified by wave and current actions. Sedimentation from meltwaters debouching into the ocean was intensive, resulting in the accumulation of a thick sequence of marine rhythmites (Kerr, 1987b, 1993).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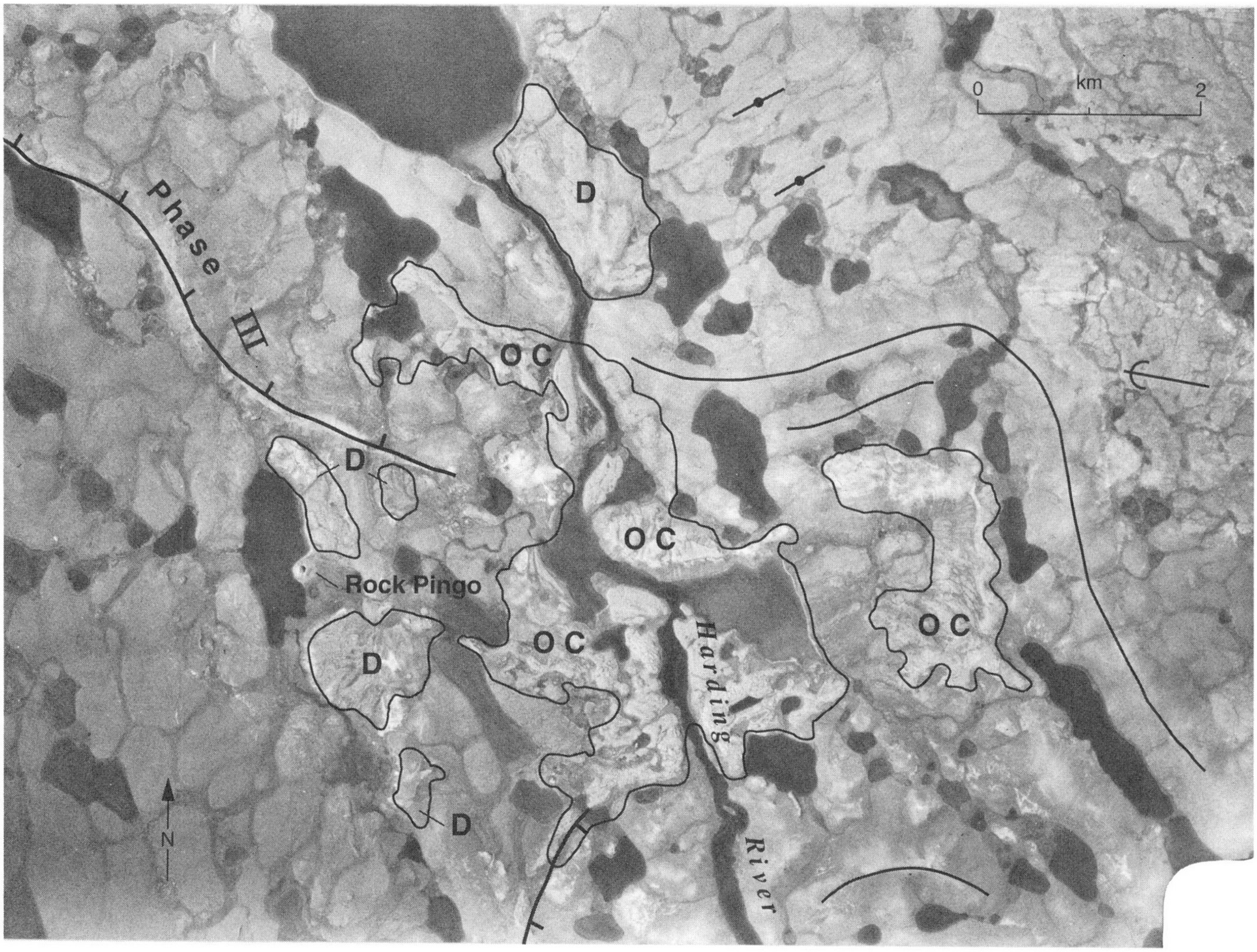

Figure 56. Morainic ridges of deglaciation Phase III (Fig. 54) interrupted by an outwash complex (OC) with collapse structures to the east and glaciofluvial deltas and fans (D); Harding River rock pingo is located in the centre left of the airphoto. (NAPL A21509-94)

Figure 55. Southern portion of the morainic ridge and outwash complex of deglaciation Phase III (Fig. 54); note contrasting texture between Inman River drumlin field west of the ridge and minor ridges (solid lines, stipple indicate steeper slope) to the east; drumlin orientation also differs on either side of the main ridge; outwash complex (OC) mark suture zones within the ice mass eroded early during deglaciation hence the jumbled nature of the deposit resulting from collapse; the morainic ridges at the bottom of the airphoto are part of morphosedimentary zone E (Fig. 1). (parts of NAPL A21510-147 and -197)

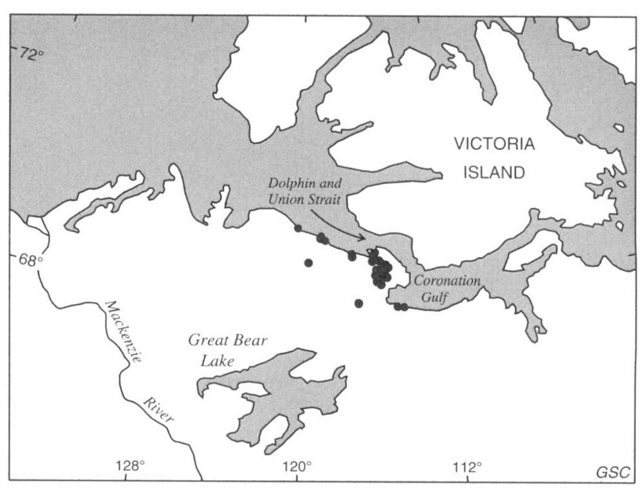

Figure 57. Sites of radiocarbon dated material south of Dolphin and Union Strait.

Figure 58. De Geer moraines draped over southwest-trending drumlins northeast of the arcuate moraine ridge shown in Figure 40; these features are generally less than 10 m wide, 40 m long and not higher than 1 m. GSC 1993-158U

Stapylton Bay phase

The western limit of the Stapylton Bay phase is marked by an important end moraine complex. This phase marks the beginning of a complete re-orientation of the retreating ice lobe centred in Dolphin and Union Strait. A southwestward flow from Victoria Island to the northeast crosscuts the north-westward flow of Harding River phase (Fig. 54). The entire coastal area in the Dolphin and Union Strait region was deglaciated by the end of this phase which occurred prior to 10 215 BP, as indicated by an age determination obtained from *Macoma calcarea* shells collected at 85 m a.s.l. near the mouth of Tree River, south of Coronation Gulf (Craig, 1960, p. 6).

This ice front, which terminated into an arm of the sea, was very unstable. Minor surges flowed in various directions as documented by intersecting striae and by the presence of push moraines composed, in part, of reworked glaciomarine sediments. Numerous minor moraines draped over drumlins are likely DeGeer moraines formed at the grounded margin of the ice on top of the drumlins (Fig. 58).

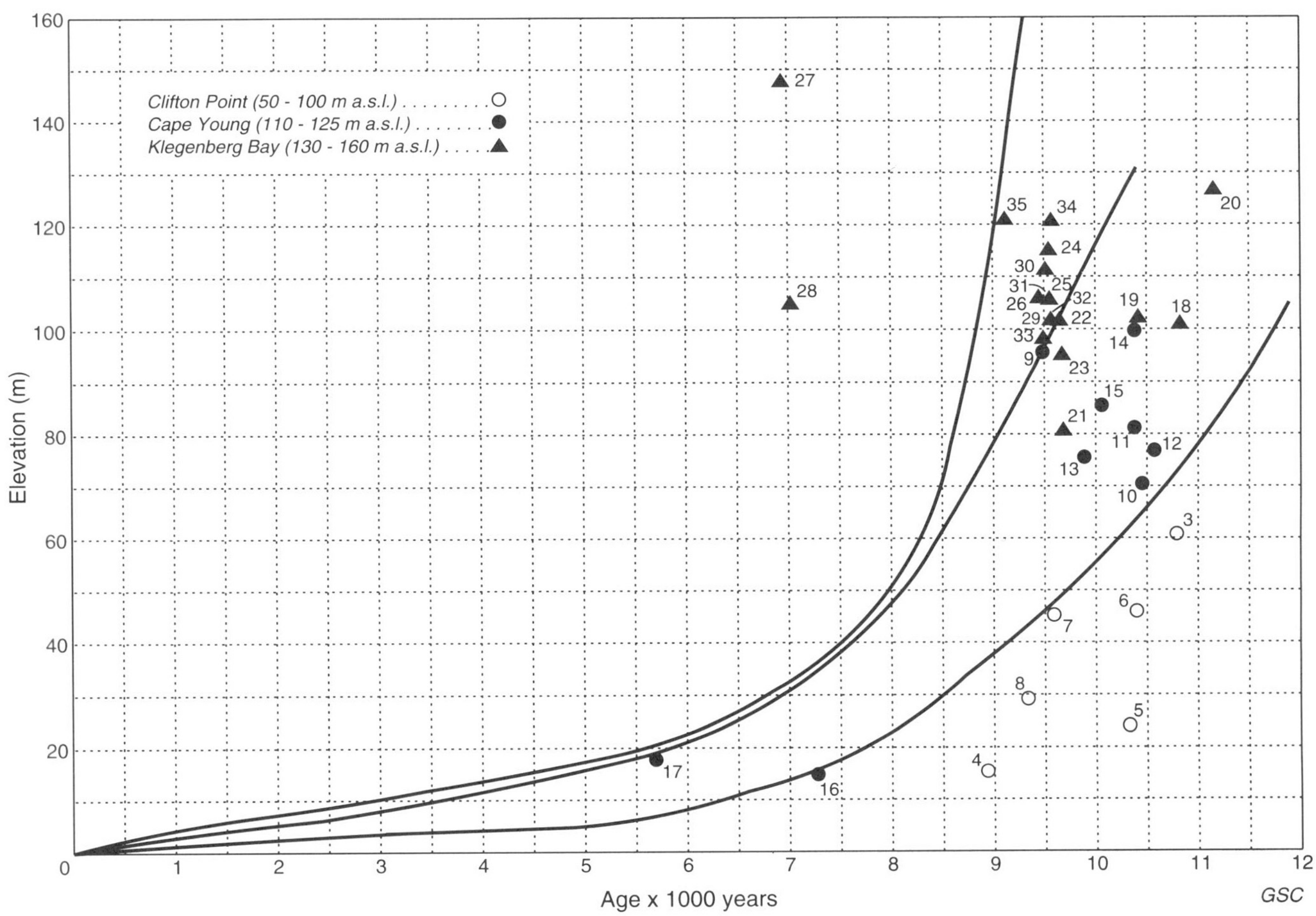

Figure 59. Emergence curves for Clifton Point, Cape Young, and Klegenberg Bay showing increased postglacial emergence west to east and corresponding increased initial uplift. Numbers beside each symbol correspond to numbers on Table 1.

On Wollaston Peninsula of Victoria Island, a major lateral moraine complex – the Colville Moraine – was constructed, along with large outwash terraces, deltas, and fans before 10 710 ± 100 BP (GSC-3566; Sharpe, 1988; 1992, p. 67). This would suggest that the Colville Moraine was constructed on the north side of the same ice lobe responsible for the construction of end moraines between Klengenberg and Cape Young on its south side.

The invasion of the Richardson and Rae river basins by the sea led not only to the rapid disintegration of the Richardson-Rae lobe but also to the collapse of the ice mass occupying Coronation Gulf.

This allowed the postglacial sea to encroach on the south shore of Coronation Gulf where, along the present Coppermine River valley, it reached an elevation of 170 m a.s.l. (St-Onge and Bruneau, 1982). Following the collapse of the ice mass, the keel of icebergs produced innumerable grooves over the areas that were still covered by the postglacial sea.

Marine limit

South of Dolphin and Union Strait, the marine limit rises from west to east and is time transgressive as it is on Wollaston Peninsula north of the strait (Sharp, 1992, p. 67). Isostatic rebound was greater over Coronation Gulf where the thick glacier ice was part of the Inlandsis (Dyke and Prest, 1987c). Farther west the Wisconsin ice was reduced to a lobe occupying the Amundsen Gulf trough leaving the Melville Hills to the south unglaciated. As a consequence isostatic loading and rebound were comparatively small. In early postglacial time ice disappeared from the uplands and was concentrated in shrinking lobes occupying the Dolphin and Union Strait/Amundsen Gulf channel and the Richardson and Rae rivers basin. Marine incursion marked by deltas, washed zones, and beach ridges entered the area from the west as a narrow inlet al.ong the ice in Amundsen Gulf and higher ground to the south as initially suggested by Craig (1960, p. 3). At the western margin of the map area, west of Tinney Point, the marine limit is approximately 55 m a.s.l. (Kerr, 1987b, 1994) rising to 160 m a.s.l. near Cape Kendall (McMartin, 1990, p. 128). On Wollaston Peninsula, Sharp (1992, p. 68) found that the "sea appears to have entered the area in an erratic fashion..." resulting in restrained rebound in some areas. We have found no evidence to that effect. Our results and those of Kerr (1987b, in press) suggest that the marine incursion proceeded in an orderly fashion, west to east, along an embayment at the margin of ice lobes wrapped around the Bluenose Lake/Bernard Harbour upland.

Postglacial emergence

Radiocarbon age determinations, mostly on marine organisms (Table 1), constrain the postglacial marine incursion and uplift (Fig. 59). We have no date at the marine limit thus the timing of ice retreat, marine incursion, and uplift is a matter of extrapolation. The oldest date (TO-1231, 11 170 ± 80) is on *Portlandia arctica* fragments at 125 m a.s.l., 10 km west of Klegenberg Bay (Table 1, #20). The implication is that, further west, in the Tinney Point area, marine incursion probably occurred as early as 12 000 BP as suggested by Dyke and Prest (1987b). The curves are drawn on the hypothesis that the western area, where uplift was 50 to 100 m, did not experience a very dramatic initial uplift because the downwarp was so small. This is strongly suggested if we accept that the initial marine incursion took place around 12 000 BP. Towards the east, with increasing amplitude, the rate of initial uplift more closely resembles emergence curves from heavily glaciated regions (Dyke and Dredge, 1989, p. 211). For the Coronation Gulf area the current rate of uplift is approximately 0.5 m per 100 years. This figure is close to the 0.46 m per 100 years reported by Blake (1963) and Kerr (1994) for areas further east.

SUMMARY AND CONCLUSIONS

Glacial landforms and lithofacies assemblages provide unambiguous evidence that ice reached the eastern slopes of the Melville Hills in the Bluenose Lake region south of Amundsen Gulf during the Late Wisconsinan. There, westerly flowing ice under compressive flow built a large thrust moraine system within which glacier ice has been preserved. Whether glacial ice covered the area during all or most of Wisconsinan time cannot be argued on the evidence we have gathered. The only clear stratigraphic evidence we have is for one nonglacial-glacial-postglacial cycle with retreat from a maximum on the east slope of Melville Hills occurring in Late Wisconsinan. We have no evidence to indicate when englaciation occurred.

Following the Late Wisconsinan maximum ice advance, three stages in ice retreat can be defined. Each of these is the result of shrinking ice masses which gradually confined ice lobes to major depressions such as Dolphin and Union Strait to the north and the basin of the Richardson River and Rae River to the south. Flow dynamics were primarily controlled by rapid drawdowns generated by extensive calving in Amundsen Gulf. The result was a readjustment of the glacier profile by extending flow conditions which led to the moulding of spectacular drumlin fields such as the one in the Inman River region, and the complex, commonly curvilinear swarms of the Bernard Harbour region.

One major conclusion of this study is that early during the Harding River phase of deglaciation, the sea occupied a corridor between the dolomite upland and the ice lobe in the lowlands. It is in this arm of the sea that most of the postglacial marine sedimentation occurred. Once the area was completely ice free, sedimentation had all but ceased and marine action was confined to wave washing of progressively emerging landforms such as occurs along the present-day coastline. One major difference, however, is that the rapid disintegration of the ice lobes calving into the sea resulted in pervasive iceberg scouring.

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APPENDIX 1

GRAIN SIZE ANALYSES
(see Fig. 1 for location of zones)

No.	Sample#	Material	Lat.	Long.	*Sand	Silt	Clay
Bluenose Lake Moraine, zone B							
1	SV86-29	Till	68°39'41"	119°24'11"	15.82	70.92	13.26
2	SV86-30	Till	69°05'14"	119°28'44"	43.83	42.39	13.28
3	SV86-31(b)	Till	68°14'02"	119°33'13"	19.70	70.31	9.99
4	SV86-34	Till	69°01'08"	120°32'21"	42.37	52.24	5.39
5	SV86-37	Till	68°19'29"	120°01'40"	49.35	40.64	10.01
6	SV89-17	Till	68°35'00"	118°59'20"	31.54	38.24	30.22
Inman River Drumlin Field, zone C							
1	SV(A)86-28	Till	68°53'36"	118°27'36"	49.61	41.49	8.90
2	SV(A)86-29	Till	68°55'40"	118°23'02"	56.68	35.16	8.16
3	SV(A)86-30	Till	68°56'13"	118°21'11"	55.95	37.49	6.56
Rae River glacial Lake sequence, zone D							
1	SV86-43	gT	68°02'18"	119°03'05"	85.03	6.01	8.96
Harding River Drumlin Field, zone F							
1	SV(M)86-6	Till	68°14'10"	115°37'25"	47.7	48.3	4.0
2	SV(M)86-64	Till	68°18'36"	115°41'07"	52.5	46.0	1.5
3	SV(M)86-65	Till	68°18'40"	115°42'26"	60.2	36.4	3.4
Rae River Moraine Belt, zone G							
1	SV88-10	Till	68°03'15"	116°04'32"	20.13	55.81	24.06
2	SV88-11	Till	68°04'06"	116°04'36"	37.78	41.23	20.99
Marine limit kame complex, zone H							
1	SV(M)88-10	Till	68°14'20"	115°29'17"	36.2	56.3	7.5
2	SV(M)88-14	Till	68°23'50"	115°49'45"	28.2	55.2	16.6
Marine offlap (Klengenberg Bay area), zone I							
1	SV(M)89-31	Till	68°10'33"	115°30'24"	32.4	54.6	12.9
2	SV(M)89-28	Till	68°08'49"	115°20'41"	23.8	52.8	23.4
3	SV(M)89-7	Till	68°14'42"	115°25'49"	23.4	56.2	20.4
Marine offlap (Bernard Harbour Drumlin Field), zone J' (see also Fig. 43)							
Field A							
1	SV(P)88-1	Till	68°45'36"	114°59'50"	71.60	24.92	3.48
2	SV(P)88-2	Till	68°45'45"	115°02'14"	79.82	20.18	0.00
3	SV88(P)-4	Till	68°43'57"	114°59'53"	90.16	8.98	0.86
4	SV88(P)-20	Till	68°47'00"	114°50'30"	51.96	31.11	16.93
5	SV88(M)-63	Till	68°46'03"	114°46'51"	51.1	38.3	10.6
Field B							
1	SV88(P)-13	Till	68°30'02"	115°05'30"	48.60	40.13	11.27
2	SV88(M)-56	Till	68°30'02"	115°05'30"	44.4	40.6	15.0
Field C							
1	SV88(P)-17	Till	68°23'10"	114°13'55"	44.37	44.34	11.29
2	SV88(M)-66	Till	68°23'10"	114°13'55"	43.44	49.8	6.9
Field D							
1	SV88(P)-6	Till	68°17'34"	114°52'52"	76.44	23.25	0.31
2	SV88(P)-9	Till	68°21'28"	114°38'37"	44.55	47.37	8.08
3	SV88(P)-11	Till	68°20'36"	114°39'23"	40.04	48.50	11.46
4	SV88(M)-32	Till	68°20'36"	114°39'21"	40.4	48.1	11.5
Field E							
1	SV(M)89-5	Till	68°33'13"	115°55'26"	13.2	58.2	28.6
2	SV(A)88-6	gT(?)	68°36'36"	115°50'02"	95.65	2.40	1.95
Field F							
1	SV88(P)-15	Till	68°39'08"	115°35'22"	49.05	39.61	11.34
2	SV88(P)-18	Till	68°35'24"	115°17'26"	49.60	38.18	12.22
3	SV88(M)-57	Till	68°39'08"	115°35'22"	47.7	42.2	10.1

In Zone J', "Fields" refer to Potschin's drumlin field as shown in Figure 43.