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GLACIAL SEDIMENTS AND LANDFORMS,
SOUTHERN VICTORIA ISLAND, N.W.T., CANADA

By

David Robert Sharpe

B.Sc., University of Toronto, Toronto

M.Sc., University of Colorado, Boulder, Colorado

This thesis was written under the auspices of
the Ottawa-Carleton Geoscience Centre

A thesis submitted to the School of Graduate Studies
University of Ottawa in partial fulfillment
of the degree of
Doctor of Philosophy
Department of Geology



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This thesis for the Doctor of Philosophy degree by
David Robert Sharpe
has been approved for the
Department of Geology

by

Dr. J. Shaw and Dr. H.M. French, Co-Supervisors
for
Dr. B.R. Rust

Chairman of Examining Committee

Dr. T. Fowler, Chairman of the Department

To my parents,
Bob and Peggy Sharpe
and
Gordon, Amanda and Daniel

ABSTRACT

GLACIAL SEDIMENTS AND LANDFORMS,
SOUTHERN VICTORIA ISLAND, N.W.T., CANADA

Wollaston Peninsula and most of southern Victoria Island comprise Palaeozoic carbonate lowlands, scarps, and tableland situated between rises and arches of underlying Precambrian sedimentary and igneous rocks. Quaternary sediments are hummocky, thick and ice-cored near escarpments where ice flow resistance, thrusting and meltwater concentrated glacial debris; thinner, streamlined drift occurs in lowlands. Quaternary sediments are mainly Late Wisconsinan in age. Glacial sediments predominate but surficial raised marine and periglacial sediments are noted. Fluvial modification of the landscape is minor.

Many of the spectacular glacial landforms on Wollaston Peninsula are streamlined and indicate formation under thick, warm-based (i.e. free subglacial water) glaciers. A set of distinctive landforms, including ground moraine (with small moraines and marginal channels), hummocky moraine, lateral moraines, and streamlined forms, relates to varying flow conditions within one major glacial advance. Stratified drift within moraines indicates the importance of glaciofluvial processes in addition to ice action. Collectively, these landforms record ice-marginal retreat, marginal stagnation following compressional flow, surging, flooding and regional stagnation during deglaciation. Freeze-on and ice stagnation trapped extensive bodies of drift-rich ice in zones of hummocky moraine. Arborescent networks of narrow eskers record subglacial meltwater drainage beneath major ice lobes and long broad eskers record subaerial deposition by meltwater of mainly supraglacial

derivation.

Late glacial events are dated relative to the incursion of the sea during deglaciation of northwest areas of Wollaston Peninsula by about 12 000 BP. Active ice-marginal conditions existed just before 10 000 BP, during formation of the large Colville moraines. Ice downwasted in the area causing glacier thinning.

Prominent ground-ice features include pingos, thermokarst scars, and debris-flow lobes. Ground ice occurs as massive icy bodies, ice-wedge ice, and buried pingo ice. Based on its setting in hummocky moraine, its stratigraphy, debris content, and isotopic composition, the massive ice is likely buried glacial ice. Landscape modification by thermokarst erosion has produced ubiquitous diamictons similar to till, and landforms similar to glacial forms. Thick ground ice bodies exist only above marine limit and the limits of streamlined landforms because permafrost degradation occurred below these limits. Widespread thaw slumps and large-scale thermal contraction cracks also indicate ice-cored terrain.

RÉSUMÉ

FORMES DE TERRAIN ET SÉDIMENTS GLACIAIRES DU SUD DE L'ÎLE VICTORIA, TERRITOIRES DU NORD-OUEST, CANADA

La péninsule de Wollaston et de la majeure partie du sud de l'île Victoria comprennent des basses terres, des escarpements et des plateaux composés de carbonates paléozoïques tous situés entre des crêtes et des arcs de roches sédimentaires et ignées sous-jacentes, d'âge précambrien. Les sédiments du Quaternaire sont bosselés et épais et contiennent des noyaux de glace près des escarpements, là où l'écoulement compressif de la glace, les charriages et les eaux de fonte ont concentré les débris glaciaires. Les formes profilées se limitent aux dépôts plus minces des basses terres. Les sédiments quaternaires datent principalement du Wisconsinien supérieur. Les sédiments de contact glaciaire abondent mais on observe aussi des sédiments marins. Les formes périglaciaires sont aussi répandues. L'influence des cours d'eau sur la topographie est négligeable.

Un grand nombre de formes de terrain sur la péninsule de Wollaston sont profilées, ce qui indique qu'elles se sont formées sous un glacier épais dont la base n'était pas gelée. Un ensemble de formes à savoir des moraines de fond (accompagnées de petites moraines et de cheneaux latéraux), des moraines bosselées, des moraines latérales et des structures profilées, reflètent les diverses conditions d'écoulement des glaces survenues lors d'un épisode majeur d'avancée glaciaire. Des dépôts glaciaires stratifiés présents dans la plupart de ces dépôts morainiques indiquent l'importance des processus fluvioglaciaires dans leur mise en place. L'ensemble de ces formes de terrain marquent le retrait de la marge glaciaire, de la stagnation du front

glaciaire suite à l'écoulement compressif, la présence de crues, d'inondations, et d'une stagnation régionale. Le regel et la stagnation des glaces ont piégé de grands volumes de glace chargés de débris glaciaires dans des zones de moraines bosselées. Des réseaux dendritiques d'eskers étroits témoignent de l'écoulement des eaux de fonte sous-glaciaires sous de grands lobes glaciaires, et des eskers longs et larges témoignent d'un dépôt subaérien par des eaux de fonte d'origine principalement supraglaciaire.

La transgression marine survenue durant la déglaciation des régions nord-ouest de la péninsule de Wollaston fournit un âge d'environ 12 000 ans pour les formes de retrait tardiglaciaire. La marge glaciaire était encore active un peu avant 10 000 ans, au cours de la formation des grandes moraines de Colville. Les glaces ont fondu dans ce secteur, ce qui a entraîné l'amincissement du glacier.

Les pingos, les cicatrices de thermokarst et les lobes de solifluxion constituent les principaux indices de glace enfouie. La glace enfouie se présente sous forme de glace massive, de coins de glace et à l'intérieur des pingos. Sa position dans les moraines bosselées, sa stratigraphie, son contenu en débris et sa composition isotopique indiquent que la glace massive est probablement de la glace de glacier enfouie. La modification de la topographie par l'érosion thermokarstique a produit des diamictons semblables au till et un relief semblable à celui des formes de terrain d'origine glaciaire. Les épaisses couches de glace dans le sol ne peuvent exister qu'au-dessus de la limite marine, celles-ci ayant été détruites sous l'ancien niveau de la mer. Les nombreux affaissements de terrain ainsi que les fentes de contraction thermique indiquent un relief à noyaux de glace enfouie étendu.

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1.0 INTRODUCTION

1.1 Scope and purpose

This thesis addresses the genesis of complex and enigmatic glacial landforms, such as hummocky moraine, streamlined terrain, and eskers on southern Victoria Island. Several fundamental questions are posed: are present models of their origin adequate and how can they be applied in high latitudes where Pleistocene ice may not have completely melted away? Is analysis of landforms as separate entities sufficient or are more encompassing landform/sediment associations required to improve understanding?

The study area is situated near the former northwestern margin of the Laurentide Ice Sheet. The northern margins of the great continental ice sheets are little studied when compared to their southern margins. Yet, marginal areas are sensitive regions of environmental change. Not only do they record the waxing and waning of glacial and periglacial conditions, but they also contain major depositional sequences within glaciated landscapes, particularly where underlain by erodible sedimentary rocks. Such "soft" substrates and tidewater glacier conditions (common to Victoria Island) may have affected the behaviour of the Quaternary ice sheets. The wasting of the latest ice sheet is particularly well recorded by the spectacular pattern of glacial landforms found across much of southern Victoria Island. Following deglaciation permafrost conditions have prevailed, with fluvial processes having only minor impact on the glaciated landscape. Buried ground ice of either segregation or glacial origin records a complex Quaternary history.

This thesis describes the Quaternary geology of southern Victoria Island, particularly Wollaston Peninsula, south of Kagloryuak River and north of Coronation Gulf (Fig. 1.1a). Descriptions of the landforms and surficial materials of this area provide information for two basic endeavours. First, an understanding of Quaternary sediments provides baseline data for future economic activity; Victoria Island is an area of economic interest as gas pipelines are proposed to cross western Victoria Island. Second, study of the well preserved glacial record of Victoria Island provides an unparalleled opportunity to assess the impact of the last glaciation on the Canadian landscape. Thus, the spectacular array of glacial, periglacial and postglacial landforms provided a broad rationale for research.

A purpose of the thesis is to increase understanding of glacial sedimentology. The thesis also examines the relationship between frozen soil, ground ice, glacial sediments and arctic landforms. The relationships between landforms and sedimentary sequences is especially stressed. Landform models based on the study of enclosed sediments and ground ice provide powerful, predictive, geologic tools that are of vital economic and environmental interest. Such models allow better evaluation of arctic terrain.

The general goals of this thesis are:

1. To describe the landforms and sediments of southern Victoria Island with respect to their composition, sedimentology and origin.
2. To outline the Quaternary history by recording and dating Pleistocene and Holocene events.

The specific objectives and major hypotheses of the thesis concern the origins of several major glacial

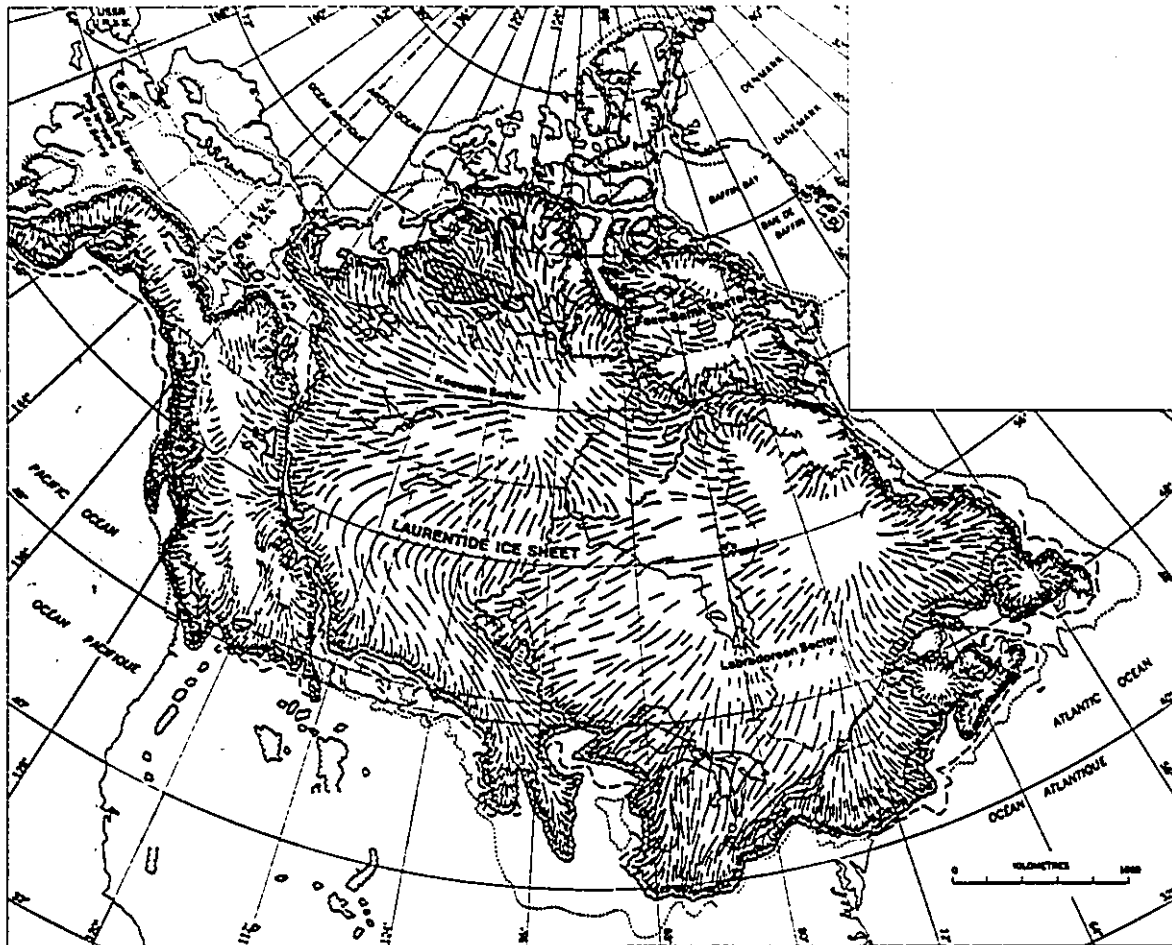


Figure 1.1. a. Location map of southern (highlighted) Victoria Island (outlined) and extent of Laurentide Ice Sheet (from Prest, 1984).

landforms:

1. Is hummocky moraine a product of supraglacial processes, subglacial processes or a product of preservation of ground ice, either of segregation or glacial origin?
2. Are streamlined landforms of erosional, depositional or of combined erosional/depositional origin? What features and processes best explain the form and origin of streamlined forms on southern Victoria Island?
3. Are eskers readily understood in terms of form and sedimentology? Are eskers an aid to reconstructions of the continental ice sheets during deglaciation?

As an integral part of this thesis, a landform/sediment map has been produced (GSC Map 1650A, scale 1:250,000; Sharpe and Nixon, 1989). This map and adjacent unpublished maps by the author, covering most of southern Victoria Island, are summarized in a series of page figures to guide analysis.

1.2 Previous Quaternary studies

The earliest reports of Quaternary geology on Victoria Island are by Collinson (1889) and Jenness (1923) who observed glacial erratics. O'Neill (1924) reported morainic material on Wollaston Peninsula, while Johansen (1924) observed glaciated bedrock on Richardson Islands, off the south coast.

Early maps depicting the glaciated areas of North America show only the southeast coast of Victoria Island had been ice covered (Chamberlain, 1907). Ice flow over southern Victoria Island was indicated by Chamberlain (1913) and his

view prevailed until recently (Wickenden, 1947). An early map by Tarr and Martin (1914) showed an independent ice cap on Victoria Island and other Arctic islands. More recent maps, however, clearly show ice flowing northwestward from an accumulation centre in Keewatin onto Victoria Island (cf. Alden, 1924; Prest, 1984; Fig. 1.1a).

Later reconstructions of ice flow, including the concept of a single source area over Hudson Bay, first proposed by Flint (1943) and revived by Hughes and Denton (1981), set the stage for ongoing discussion of ice-flow history. Presently held ice-flow reconstructions rely on the premise that marine-based ice sheets contained ice streams which flowed outward along inter-island channels, dominating the discharge of the Laurentide Ice Sheet, and overtopping most of Victoria Island. The ice stream concept was extended to include major east to west flow (across inter-island channels) from M'Clintock Channel across Victoria Island (Dyke, 1984; Dyke and Prest, 1987).

The earliest comprehensive study of Quaternary geology on Victoria Island was by Washburn (1947), who reported on the glacial sediments, land emergence, periglacial landforms and geomorphic processes. He argued that striae, eskers, other glacial landforms and their deposits and evidence for emergence indicate that all of Victoria Island was probably glaciated. At the same time, Washburn cautioned that sea ice may have produced striae around coastal areas and that solifluction deposits are easily confused with till. He suggested that the striking glacial topography indicates a glacial cover over all of Wollaston Peninsula, and that the glacier subsequently stagnated. Washburn mapped eskers and crevasse fillings on southern Victoria Island and noted the glaciofluvial nature of Mount Pelly, a prominent broad ridge

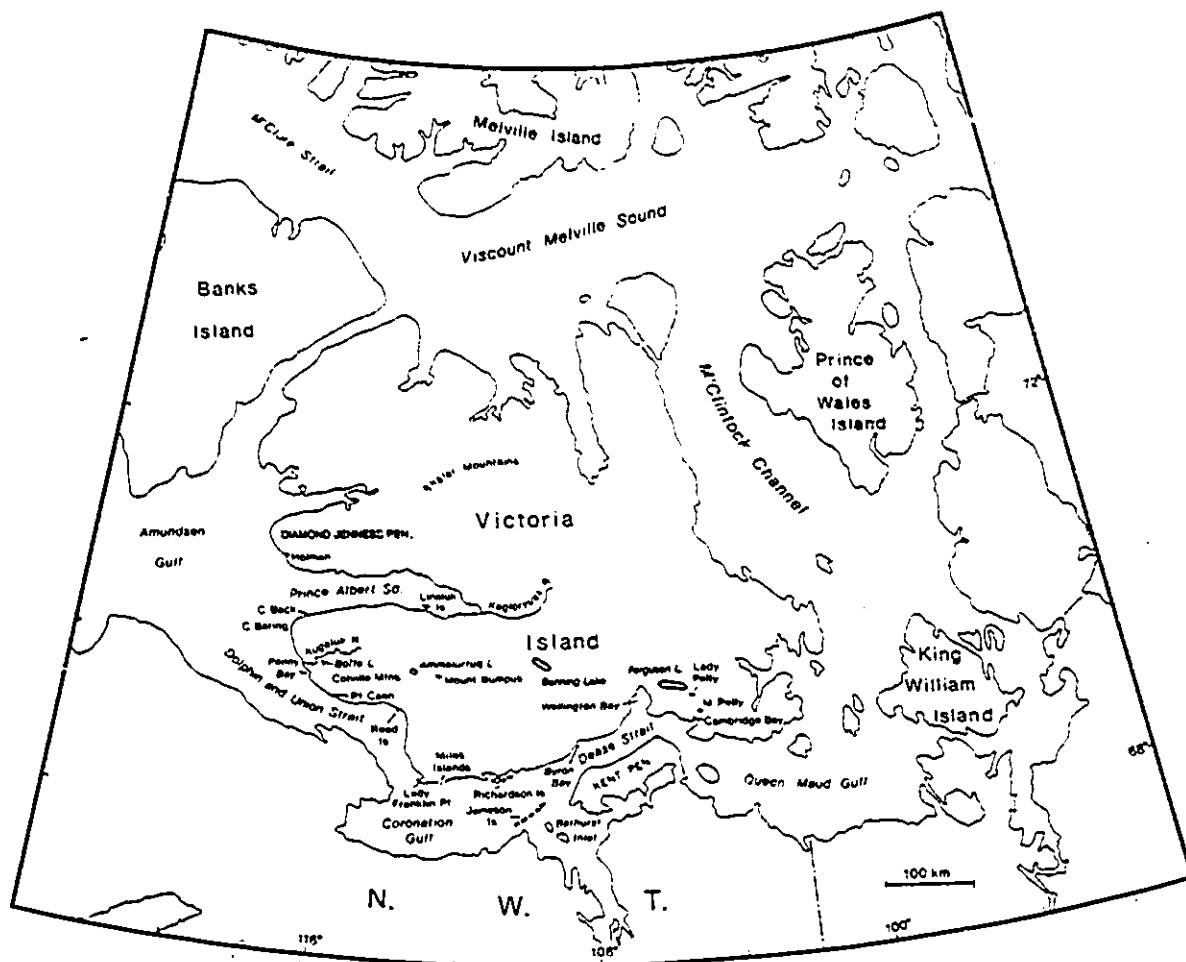


Figure 1.1. b. Location of places mentioned in text.

(Fig. 1.1b). He suggested an ice divide east of Victoria Island and north of Queen Maud Gulf. Of special interest is his observation that mass wasting, especially solifluction, far outweighed fluvial activity in terms of landscape modification. His study of patterned ground and other periglacial forms was a benchmark in the study of permafrost terrain in Arctic Canada.

Subsequently, Fyles (1963) systematically mapped selected surficial landforms of Victoria and Steffansson islands. Of particular importance is his conclusion that the latest ice occupied the major sounds and straits adjacent to Victoria Island. This conclusion makes it difficult to distinguish between marine and lacustrine limits. Fyles noted undeformed stratified sediments lying below a thin diamicton in many parts of the Island. These and other diamicton masses are difficult to separate from non-glacial diamictons such as colluvium or solifluction debris. He also noted eskers, up to 160 km in length, connected to meltwater channels, on the order of 100 km in length, and attributed complex cross-cutting of streamlined forms to control by an ice-sheet topography characterized by thin ice and distant accumulation areas.

Recent studies by Vincent (1982) and Hodgson (1987) cover areas north of the study area. In addition, overviews by Vincent (1989), Dyke and Dredge (1989) provide a regional context for this study. There are also several preliminary accounts on the area examined in this thesis (Sharpe, 1984; Sharpe, 1985; Sharpe, 1987; 1988a; Borowiecki, 1984).

1.3 Bedrock geology

The bedrock geology of Victoria Island was mapped by Thorsteinsson and Tozer (1962) and more recently, for

Precambrian areas, by Christie (1964), Young (1981), and Campbell (1978; 1983). A generalized bedrock geology map (Fig.1.2) shows thin, flat-lying, dominantly lower Palaeozoic carbonate rocks, underlying southern Victoria Island. These rocks lie unconformably on Upper Precambrian sediments and older deformed rocks extensively exposed north of Wollaston Peninsula, and to the south on the mainland and adjacent islands. Wollaston Peninsula is underlain by Ordovician and Silurian grey and yellow-brown weathering dolomite, minor chert, shale and sandstone. Underlying sandstone, minor shale, siltstone and dolomite of Cambrian or early Ordovician age occur along the south coast of Wollaston Peninsula.

The major Precambrian formations comprise terrestrial and marine sediments of the Rae and Shaler Groups (Fig. 1.2). The Upper Precambrian, Glenelg Formation (lowest Rae Group) found on the Richardson Islands consists of sandstone, limestone, shale, siltstone, dolomite and conglomerate (Campbell, 1985). Younger Rae Group and the Glenelg Formation rocks outcrop in the Miles Islands off the south coast.

Bedrock to the south of Queen Maud Gulf, Dease Strait and Coronation Gulf, is of interest because it is the source area for Precambrian erratics (Fig. 1.3 a,b) carried onto Victoria Island during regional glaciation. Campbell (1983) describes the Precambrian Rae Group rocks of the Coronation Gulf, including the Richardson Islands area and the Jameson Islands, extending south of Victoria Island to Bathurst Inlet. The Rae Group rocks comprise red to white, sandstones, siltstones, dolomite and shale, and a diagnostic rock type, an orange weathering biohermal carbonate with stromatolites. These rocks are identified in the drift of southern Victoria Island (Fig. 1.3b). Two formations

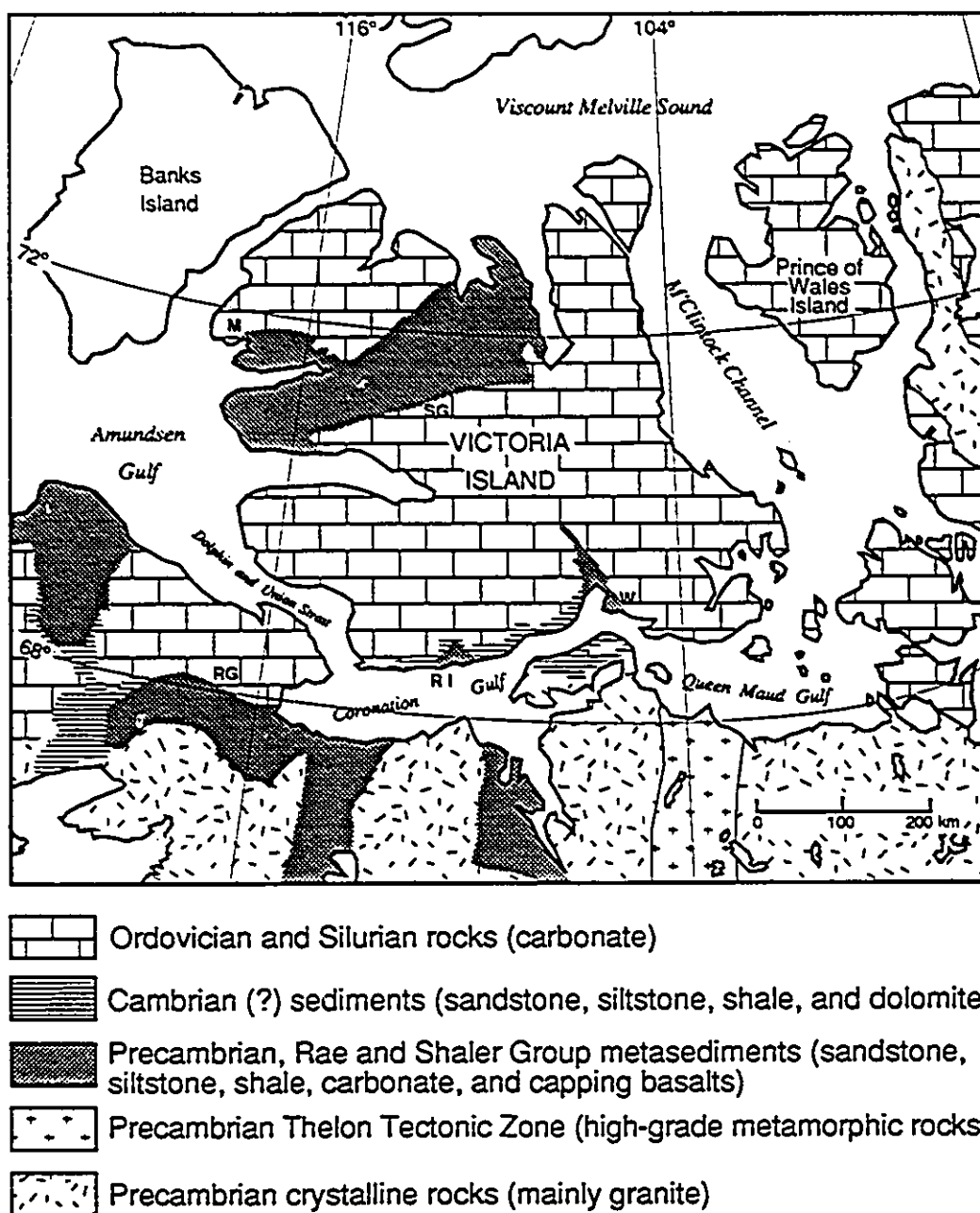


Figure 1.2. Generalized bedrock geology of Victoria Island and adjacent southern mainland. SG, Shaler Group; RG, Rae Group; RI, Richardson Islands; W, Wellington inlier; M, Minto arch. After Thorsteinsson and Tozer (1962) and Campbell (1985).

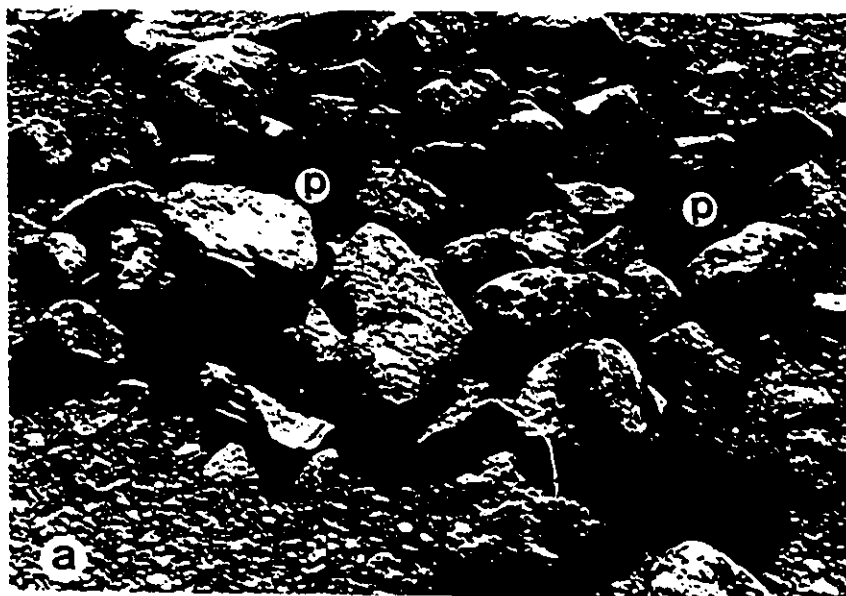


Figure 1.3. a. Precambrian erratics (p) derived from areas southeast of study area. Glove is 20 cm long. GSC 204442-B



Figure 1.3b. Rusty-weathering biohermal carbonate with stromatolite. Scale is 8 cm long. GSC 1991-165

originally assigned to upper Rae Group were shown to be Palaeozoic (Campbell, 1983). Rae Group rocks are correlatives of the Shaler Group of west central Victoria Island (Fig. 1.2), which consist of sandstone, siltstone, dolostone and limestone, including distinctive orange weathering stromatolites and evaporite units (Young, 1981). Dixon (1979) showed that rocks of the Wellington inlier of southeastern Victoria Island are distinct from those of the Duke of York inlier north of Richardson Islands. The rocks near Wellington Bay are red to pink quartzitic sandstone, conglomerate or siltstone; cherty ironstone occurs near Washburn Lake. These rocks are considered to be of fluvial origin rather than marine, as are the Rae and Shaler Groups. Rocks similar to those of the Wellington Bay area occur to the south on Kent Peninsula (e.g. the Burnside River Formation (Dixon, 1979; Campbell and Cecile, 1981)).

Thompson et al. (1985) described the Thelon Tectonic Zone (Fig. 1.2) which forms the boundary between the Slave and Churchill structural provinces. This zone marks the western extent of high grade metamorphic terrane comprising pink gneiss, migmatite and granitoid rocks derived from the unmetamorphosed Goulburn Group clastic and carbonate rocks (Campbell and Cecile, 1981; Hoffman et al., 1984). These metamorphic and granitic rocks are particularly good indicators of ice movement from the mainland onto Victoria Island.

1.4 Structural setting

Victoria Island is situated on the southern margin of a stable Palaeozoic carbonate platform attached to the North American craton. The Arctic Islands comprise three major tectonic or structural elements; 1) the Canadian Shield, 2)



the Arctic Platform and 3) the Innuitian Tectonic Province (Trettin, in press). The first two elements are found on Victoria Island while the Innuitian Province includes younger terrane on islands to the north. The Arctic Platform borders and covers the Canadian Shield to the south, except for rises and inliers on Victoria Island, in particular, Minto Arch and the Wellington and Duke of York highs (Fig. 1.2). This platform comprises undeformed strata, of Lower-Middle Palaeozoic age, which are preserved in post-depositional structural basins bounded by Precambrian topographic highs (Minto Arch and Wellington high). Movement associated with possible structural basins or along fault-bounded blocks has affected the Arctic since Precambrian times (Kerr, 1980) and may have controlled part of the structural arrangement of Victoria Island.

Broad channels and sounds dissect Victoria Island and the study area (e.g. Coronation Gulf, Dolphin and Union Strait and Prince Albert Sound; Fig. 1.1b). These channels form part of the structural setting of Victoria Island and their origin and that of other arctic seaways is in debate. Subaerial Tertiary drainage systems (Fortier and Morley, 1956; Bornhold et al., 1976), deep glacial scour (White, 1988), and tectonic or structural control (Kerr, 1980; Sanford et al., 1985; England, 1987) have all been advanced as primary mechanisms to explain channel formation.

A major portion of the Arctic has experienced a complex sedimentary and tectonic history since Cambrian time (Kerr, 1980), but little of this record is present on the stable platform forming Victoria Island (Fig. 1.4). The inter-island seaways may record some of this history and a structural control for their origin was proposed by Sanford et al., (1985, their figure 4). In their synthesis, adjacent

islands and Wollaston Peninsula are pictured as possible large, fault-bounded blocks that may have been affected over time by periods of quiescence and reactivation. The fracture patterns (Fig. 1.5) which parallel some of the major structural lineaments identified from the Shield (Sanford et al., 1985), lend support to the idea of fault-block movement. These fractures and blocks may reflect continuing tectonic activity along such ancient orogenic belts as that of the East Greenland orogen. There is however, no evidence on southern Victoria Island comparable to that for a fault and graben origin of Lancaster Sound (Kerr, 1980). More recent fault-block movement, associated with the Boothia uplift on adjacent Prince of Wales Island, may have occurred in Holocene time due to glacial unloading (Dyke et al., 1991).

Recent theories on the distribution of major ice streams (along lines of topographic control or first order geology; Denton and Hughes, 1981) and the effectiveness of continental ice sheets in producing deep erosion (White, 1988) have favoured a glacial and glaciofluvial origin for the seaways.

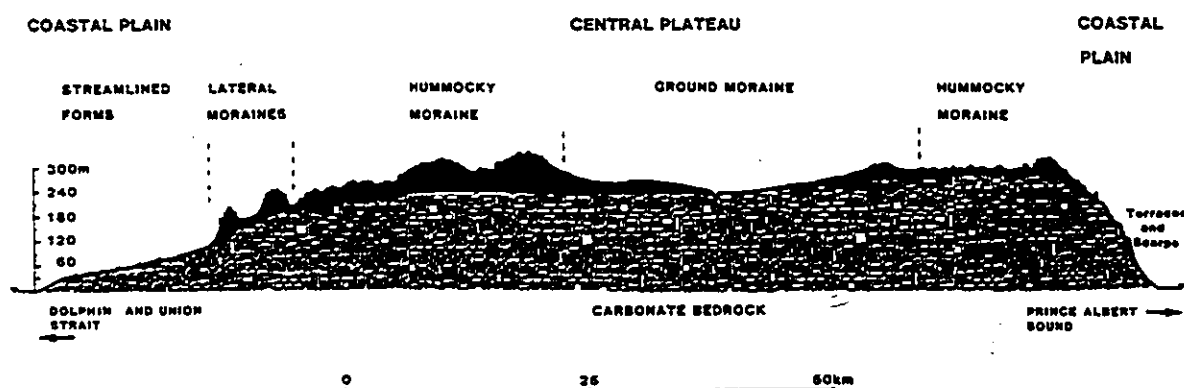


Figure 1.4. Cross-section along longitude 114° W showing Palaeozoic carbonate plateau and escarpment that is the core of Wollaston Peninsula; major landform belts superimposed.

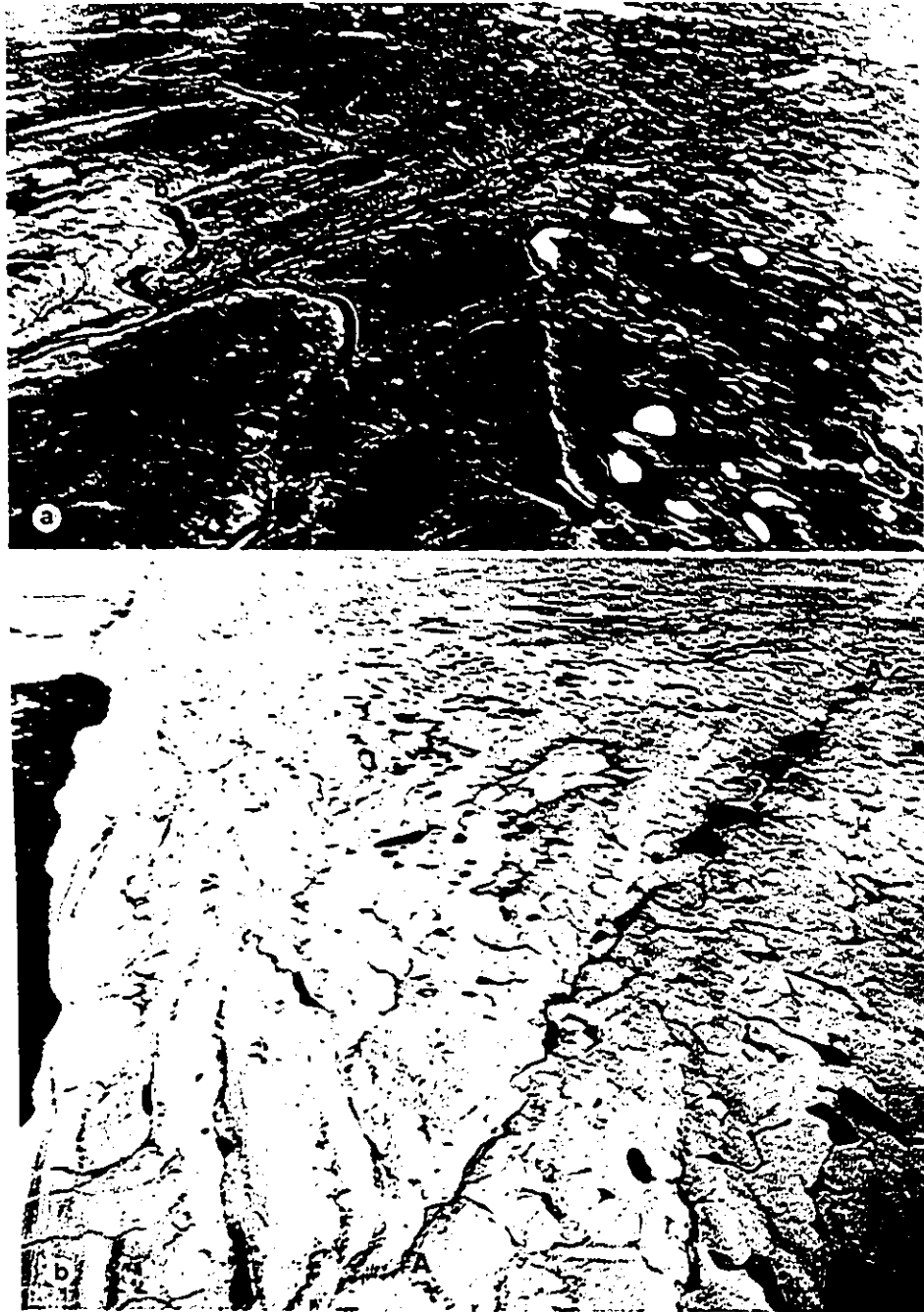


Figure 1.5. Bedrock lineaments, Wollaston Peninsula: a. Fractures control drainage along Kugaluk River. Lineament a-a parallels lineament A-A (Fig. 1.5b). Lineament b-b is orthorhombic pair to lineament a-a . NAPL T480R-17 b. view to east along north coast. Lineament A-A occurs on Figure 1.6. NAPL T345L-70.

1.5 Topography and physiography

The physiography of southern Victoria Island reflects bedrock lithology and structure, glacial activity and post-glacial emergence. The relief of Wollaston Peninsula is controlled by Palaeozoic carbonate strata eroded to a major escarpment or series of escarpments which encircle the peninsula (Fig. 1.4 and Map 1650A). These divide the peninsula into an interior plateau and a coastal plain.

The escarpments are drift-mantled exposures of Palaeozoic carbonate rock that form 10-50 m high re-entrant scarps along coasts and rivers. Many of these are fracture- or joint-controlled (Fig. 1.5) and can be mapped from LANDSAT images (Fig. 1.6). These fracture patterns may represent major northwest lineaments related to large fault-bounded blocks identified on a continental scale (Sanford et al., 1985). The major northwest fracture (Fig. 1.6) parallels the trace of dyke swarms (related to Proterozoic hotspots and rifting) which have a focal point off the northwest tip of Wollaston Peninsula (LeCheminant and Heaman, 1989). Some of these fractures may be confused with periglacial patterns or glacial lineaments on aerial photographs.

Smaller scarps, cuestas and buttes form part of the coastal plain surrounding Wollaston Peninsula. Bedrock on the carbonate plateau is at about 250 to 350 m a.s.l., while overlying drift (and ice?) adds another 60 m to 100 m of relief. The highest drift reaches 500 m a.s.l.

Draping the major escarpment topography is a broad hummocky morainic complex that encircles Wollaston Peninsula. A preliminary (classical) interpretation suggests that this moraine pattern results, in part, from the carbonate plateau diverting the main glacial flow and providing flow resistance which caused thick sediment accumulation. In contrast, the



Figure 1.6. LANDSAT images of Wollaston Peninsula show tonal patterns related to vegetation, sediment type, and drainage: a. northwest Wollaston Peninsula showing bedrock lineaments and thin drift (LANDSAT B58-10-9 102-45 7, July 13, 1975);

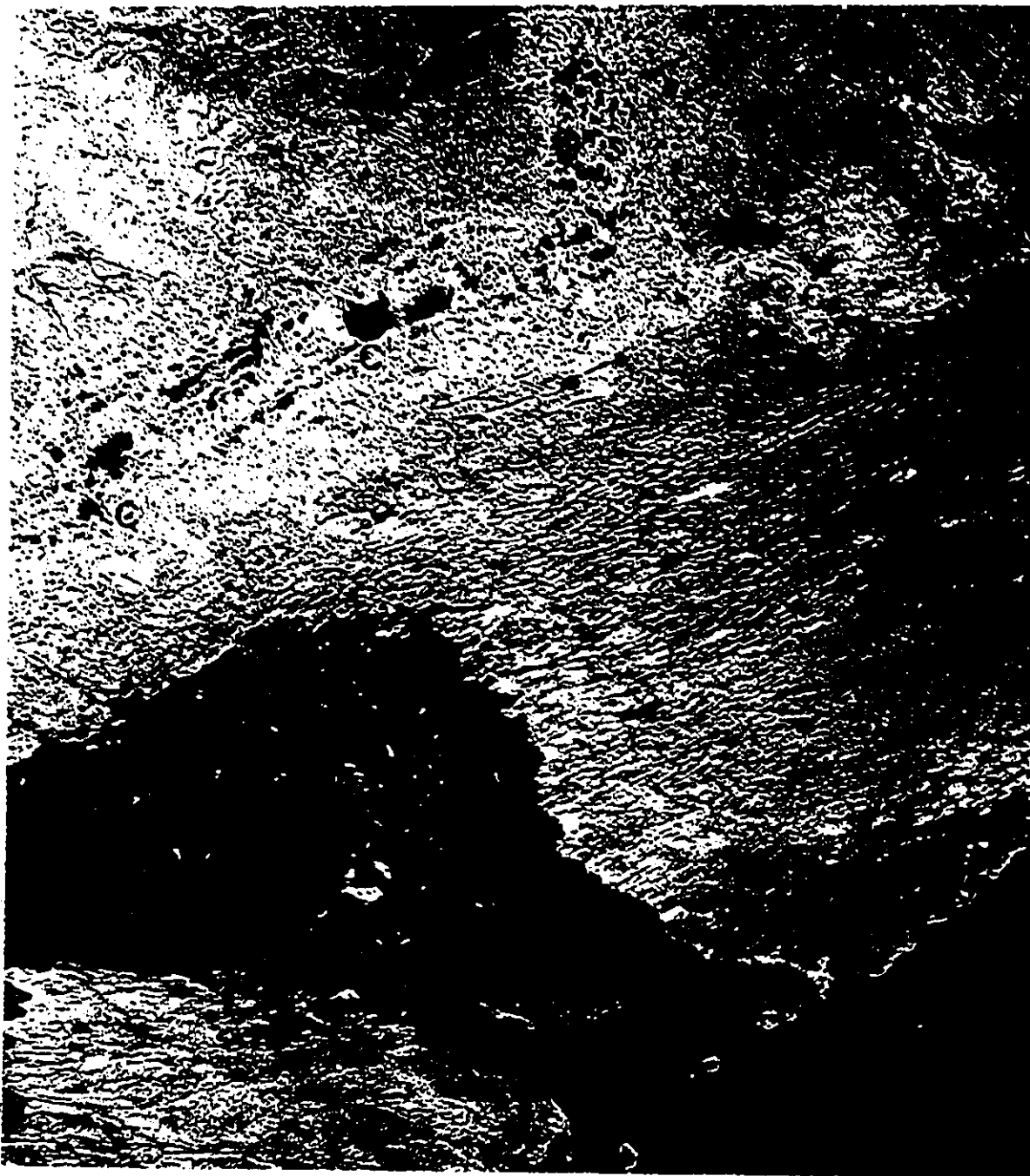


Figure 1.6. b. southern Wollaston Peninsula showing major glacial landforms (Fig. 3.1). See Figure 3.1 for labels; CC'C" is Colville moraine (LANDSAT A55-11-21 103-45 7). Arrow shows location of Fig. 4.3.

lowlands and central plateau have thinner drift. The striking, streamlined lowlands stand in contrast to the hummocky morainal areas, the Colville Mountains (Fig. 1.4).

Lakes are concentrated and unoriented in moraines, less frequent and oriented in drumlin fields. By contrast, the central plateau has few lakes. It is underlain by level or shallow dipping carbonate rock only thinly covered by unconsolidated sediment (Fig. 1.4); a poorly drained featureless plain results. This level terrain has a subdued look typical of older, weathered, drift surfaces when compared to the more youthful-looking, morainic topography.

Flights of raised beaches, bars and bluffs, along coastal areas from sea level up to 150 m a.s.l., record the rapid land emergence in the postglacial period. Permafrost and periglacial landforms mark local physiography.

1.6 Climate, vegetation and soils

Climate

Climatic stations with continuous records (20 years) are located along southern Victoria Island (Canadian Climate normals, 1982). The mean annual temperature is -12.9°C and the mean temperature of the warmest month (July) is 6.6°C with daily maximums about 10°C at Lady Franklin. Stations at Byron Bay, Holman Island and Cambridge Bay show similar temperatures. Mean annual ground temperature of about -14°C indicates permafrost.

Annual precipitation is low on southern Victoria Island (11 cm); the result is a polar desert. The maximum monthly value occurs as summer rains (2-2.5 cm/month) whereas snowfall is low. Snowfall has an early autumn peak coinciding with the period of maximum extent of open water.

Lake ice usually remains on a portion (30%) of large lakes until early August. Scattered, wind-blown, late-lying snowbanks may persist to early August during years with cool, cloudy weather.

Winter winds of over 100 km/hr are commonly recorded at Lady Franklin, Holman, and Byron Bay; these affect snow cover, vegetation, active layer thickness, and periglacial features, as well as controlling aeolian processes.

Vegetation

The relationship between surficial materials, topography, climate, and plant communities was investigated by Edlund (1983, 1989). Two arctic ecosystems are identified on western Victoria Island; the low arctic (erect shrub zone) is the most widespread.

The low arctic ecosystem has a rich flora (more than 150 vascular species) with common dwarf shrub (*Dryas* and *Salix*) and herbaceous legumes (*Oxytropis*, *Artemisia*, and *Kobreisa*). Wetlands have thicker vegetation of sedges, grasses, heaths and especially dwarf willow 25 to 50 cm in height. Sheltered sites may have dwarf birch (*Betula glandulosa*). Occasionally, in very protected sites, willows up to 5 m high have been reported north of Wollaston Peninsula (Edlund and Egginton, 1984).

A mid arctic ecosystem (dwarfed and prostrate shrub zone) in northern Wollaston Peninsula is marked by low diversity of low arctic herbs. Wetlands have prostrate shrubs of *Salix arctica* but not *S. reticulata* or *S. Polaris*. Heath is only represented by *Cassiope*.

Plant distribution is closely related to soil type, drainage and moisture regime, an important link in sediment and landform mapping from aerial photography (Appendix I).

Soils

Soils, which respond to climatic variables, parent material, relief (drainage), vegetation, and the effects of time, offer clues on past climatic conditions and the nature of climatic change. They aid in differentiating between formerly glaciated and non-glaciated areas and the length of time that regions have been ice free. In this thesis soils are used to provide relative dating information (e.g. Birkland, 1978) and to infer active layer processes.

All soils examined are cryosolic as they are underlain by continuous permafrost but only a few show extensive turbic or frost-churned development (Agriculture Canada, 1987). Many soils show horizon development (e.g. Ugolini et al., 1987) with recognizable and consistent textural and trace element trends in the soil profiles (Sharpe, 1991). Soil profiles appear to be a response to several processes: pedogenic, cryogenic (periglacial), and aeolian (Plate I). Soils on southern Victoria Island are Late Wisconsinan in age and may develop faster there than in other late Wisconsinan arctic terrains (Woo and Zoltai, 1977) due to climate and the texture of carbonate substrate (Tarnocai, personal communication, 1987).

1.7 Permafrost

Because Victoria Island lies within the zone of continuous permafrost (ACGR, 1988), a variety of periglacial landforms and ground ice features inter-related with glacial landforms (e.g. hummocky moraine) has developed. Permafrost conditions result from present day mean annual temperatures less than -10°C and relatively snow-free (windswept) winters.

1.7.1 Thermal conditions

Few data exist concerning ground temperatures on Victoria Island. At one site in silty soil, August ground

temperatures ranged from -1.2°C near surface to -16.2°C at 12.2 m (Appendix II).

The extent of permafrost is determined from indirect indicators such as periglacial landforms and estimated extent of ground ice. Ice wedge polygons are ubiquitous and form rapidly on modern floodplains and in shallow lakes. Permafrost thicknesses of >13 m have been observed (EBA, Consultants, 1980), but depths of several hundreds metres are probable (Taylor, 1988). Thicknesses range between 340 m and 660 m on Melville Island and 355 m and 600 m on Bathurst, Cornwallis, and Devon islands (Taylor, 1988), all beyond the Late Wisconsinan ice margin. On Russell Island, 200 km northeast of Victoria Island, ice-bonded permafrost extends from a surface elevation of 114 m to a depth of 300 m, based on resistivity in a borehole through Palaeozoic rocks (Hardy and Associates, 1984).

The active layer is the surface layer of ground subject to annual thawing and freezing in areas underlain by permafrost (ACGR, 1988). Test pits dug in till, July 1982 and 1983 recorded permafrost between 75 cm and 120 cm, in well drained, unvegetated, upland sites. Sites showed oxidation to the maximum depth of the active layer. Active layer depths of only 75-80 cm were measured in pits on siltier, slightly moister and better vegetated sites.

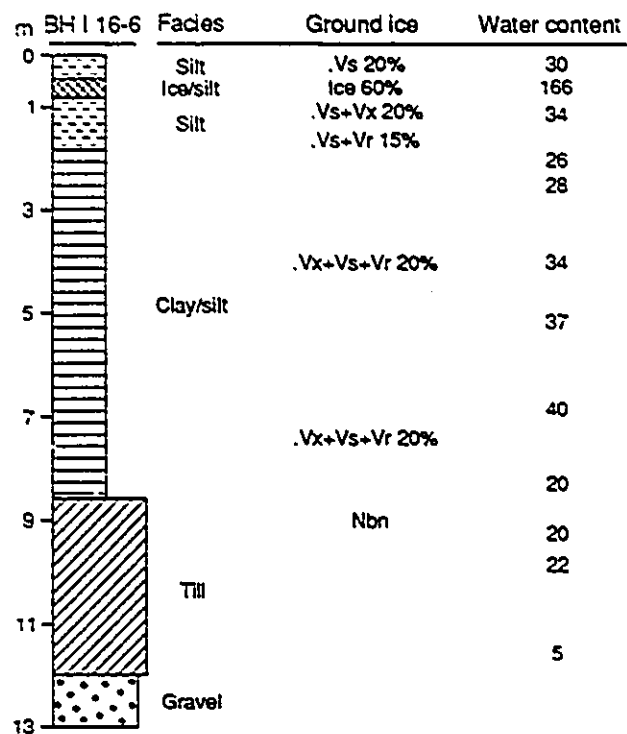
Data from boreholes drilled in August 1980 indicate active layer depths mainly between 0.6 m and 0.8 m on vegetated, silty soils, whereas depths of 1.0 m and 1.1 m were recorded on coarse sediment, till, or gravel (E.B.A. Engineering Consultants, 1980).

1.7.2 Ground Ice

Little ground ice has been seen on southern Victoria Island despite the widespread periglacial landforms

(Washburn, 1947) and the inference of extensive segregated ice in the western arctic (Mackay, 1971), although a large icing seen on LANDSAT images (Fig. 1.6b) was confirmed on the ground. Borehole logs of a transect across the southern part of the area show low ice contents and little segregated ice in till, particularly on upland sites (Fig. 1.7). Although marine silt and clay contain higher ice contents, little ice occurs as thick lensing or wedges. For example, at one site no excess ice occurred in till to depths of 13 m, whereas the overlying marine silt and clay contained up to 20% (by volume; ice crystals, coatings, or oriented ice, Fig. 1.7). Ice lenses 5 mm thick and thinner vertical sheets occur sporadically in some holes (E.B.A. Consultants, 1980). Most holes had 20-25 cm intervals showing 10-25% by volume and a few thin ice lenses. One hole revealed ice-rich till containing 90% ice by volume. At some sites, the ice-bonded sediment showed no visible ice. For example, fabric studies at Banning Lake (Fig. 1.8) show that clast fabric within the active layer was not rearranged by either ice action or frost wedging, though solifluction reoriented the fabric downslope. Few ice wedges were observed in the field but the widespread occurrence of thermal contraction crack polygons suggests that ice wedges are common. These polygons may also indicate the presence of extensive buried ground ice on Victoria Island (Sharpe, 1988a, 1989).

Massive ground ice was observed at several sites where ground ice slumping was active (Fig. 1.9). Massive ground ice may correspond with the distribution of large (30 m to 100 m), nonsorted polygons, common on hummocky and lateral moraine, but not on adjacent streamlined terrain below washed limits where sediments are thin (Fig. 1.10). Former thermokarst erosion scars, kettles, thermokarst lakes,



V_s - stratified ice

V_x - ice crystals

V_r - random oriented
ice

N_{bn} - non bonded

Figure 1.7. Borehole geotechnical log from southern Wollaston Peninsula (from Polar gas report by E.B.A. Consultants 1981).

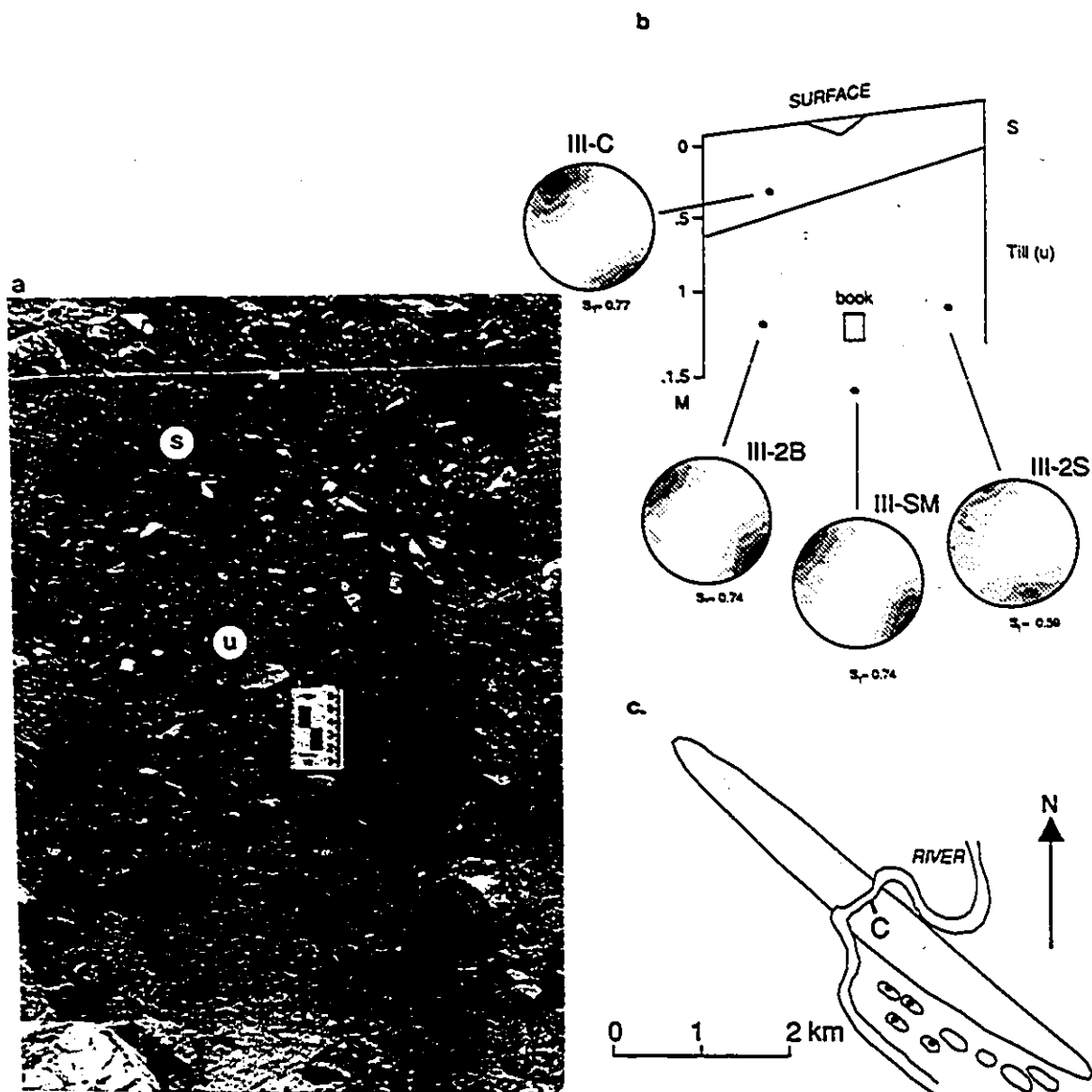


Figure 1.8. a. Drumlin sediment, 50 cm below the surface shows a primary fabric undisturbed by frost action: solifluction material (S); undisturbed till (u). GSC #1991-169A. b. Strong fabric (parallel to drumlin axis) in till (u) has been redirected downslope by solifluction (S). c. Location of section (C).

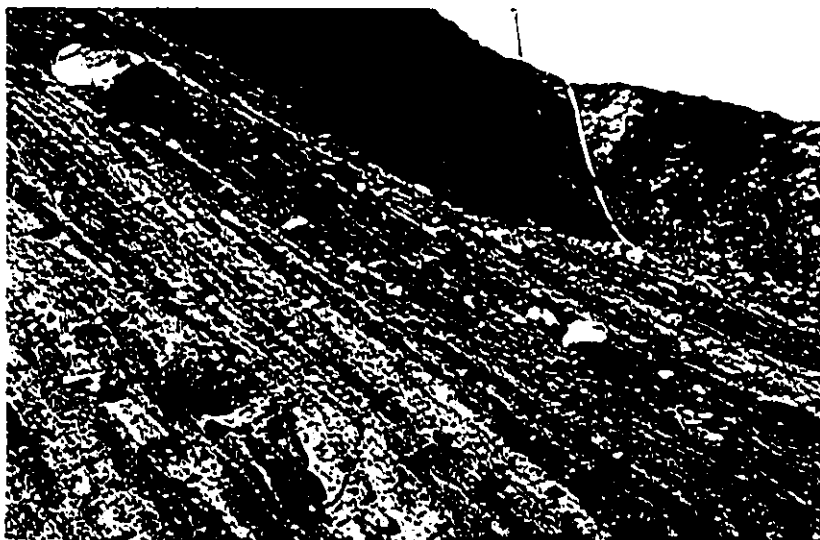


Figure 1.9. Exposure of massive ground ice Ammalurtuq Lake. GSC 204876-B.

and hummocks are indirect indicators which extend the area of probable buried ground ice (Fig. 1.6). Because massive ice may be of either glacial origin and Pleistocene age (Sharpe, 1988a) or of segregated origin and Holocene age (Sharpe and Michel, in press), and because it is intimately related to the origin of hummocky moraine, it is discussed more fully in Chapter 2.

Pingos and rock domes

Several relict pingos have been identified on Wollaston Peninsula by Washburn (1947, 1980) and Fyles (1963). They range from tens to hundreds of metres across and from 5 m to more than 25 m high. They consist of a central depression or crater-like area surrounded by a large circular or arcuate ridge or rampart (Sharpe, 1991). The ramparts are dissected in places by large gullies and by concentric cracks or fractures and, less commonly, by radial cracks. Some radial cracks may have formed as dilation cracks caused by ice-core expansion (Mackay, 1985; 1988); these appear continuous with large polygonal borders. One section through a rampart shows

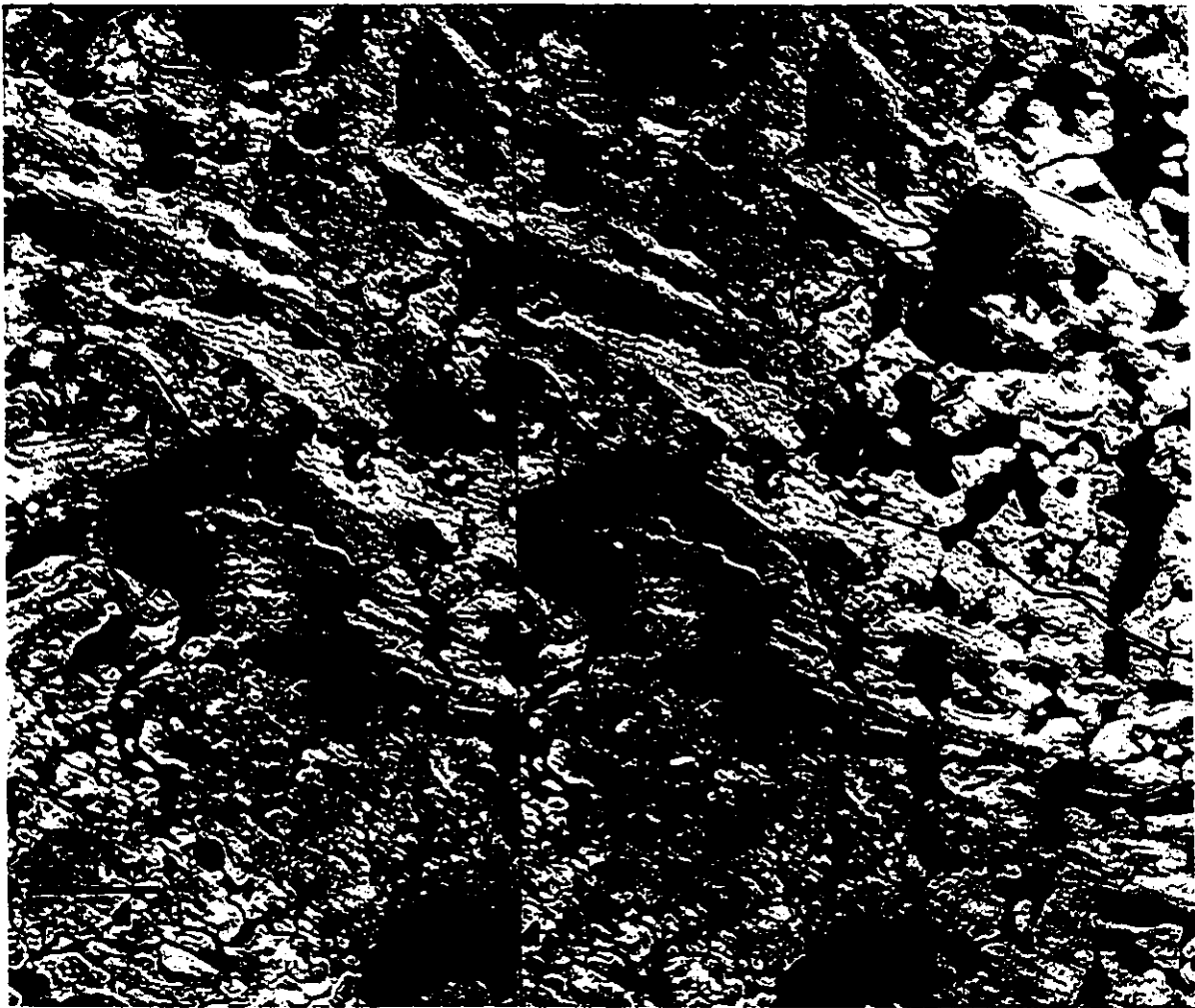


Figure 1.10. Large polygons developed on diamicton (morainic) surfaces above marine limit (highlighted). Polygons may indicate the presence of buried ground ice (Stereopair from NAPL A15787-112,113)

beds of stratified sand and silt containing moss (Fyles, 1963). These beds are the upthrust or domed sediments of glacial Lake Wollaston (Sharpe, 1984; Map 1650A).

Similar pingos, which form in drained lake beds, in the Mackenzie delta are considered to be of the closed-system type formed by permafrost aggradation and cryostatic pressure (Mackay, 1979). They provide unambiguous evidence of permafrost, with probable mean annual ground temperatures of -5°C or lower (Mackay, 1988). The Wollaston pingos appear to have formed under conditions different than present. Their growth probably began immediately after glacial Lake Wollaston drained about 10 500 years BP (Sharpe, 1984). Pingos grew between 3500 and 7000 years, BP on Banks Island (French and Dutkiewicz, 1976; Pissart and French, 1976).

Rock domes or rock "blisters" in the study area are dome-shaped rock heave features, 1-3 m high and several metres across, composed of blocks of carbonate bedrock (Fig. 1.11a). They are found in low, moist terrain of southern Victoria Island and adjacent areas (St-Onge et al., 1989). Rock domes form in the active layer by expulsion of water under pressure, analogous to closed-system pingos (L. Dyke, 1984). Similar rock heave has exploited the common joints and fractures on the carbonate lowlands (Fig. 1.11b).

1.7.3 Thermokarst activity

Many landforms and features produced by the melting of ice-rich permafrost occur across southern Victoria Island. This process is common over large areas of hummocky moraine particularly around Ammalurtuq Lake (Fig. 1.9), where thaw erosion produces retrogressive scarp headwalls and debris-flow aprons. The terrain includes broad areas of previously disturbed slopes and sediment defined by relict scarps, scars and debris-flow aprons.

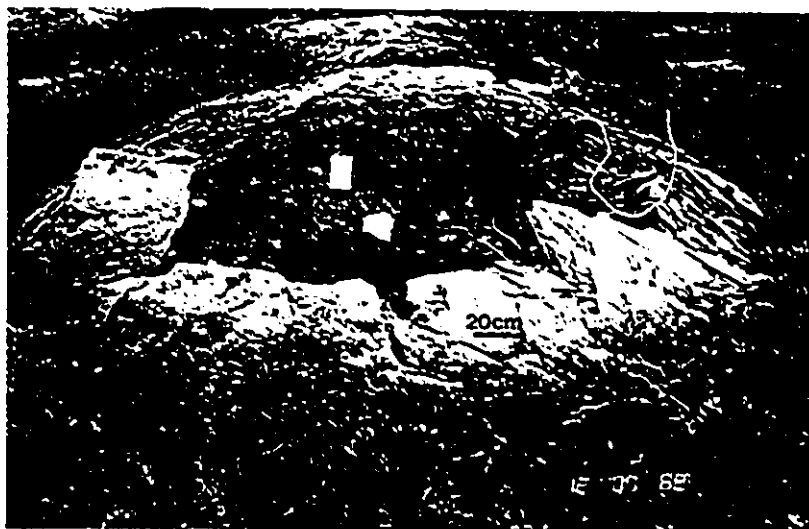


Figure 1.11a. Rock burst. GSC 204763-A



Figure 1.11.b Rock heave along fracture and joint system in carbonate bedrock. Square (S) is 80 m. GSC 204439-A.

Thermokarst lakes or thaw lakes are water bodies occupying a closed depression formed as the ground subsided with ground ice melting (ACGR, 1988). They are common on southern Victoria Island, particularly in hummocky moraine and near the Ferguson Lake esker complex (Chapter 4). They are small, rounded to elongate lakes (0.1-0.5 km) set between flutings (e.g. Rampton, 1988): larger examples (0.5-2 km) cross-cut flutings (see Fig. 4.13) or cut into drumlins. Most lakes have smooth shorelines indicating little recent expansion though some show extensions of several kilometres. Others have drained leaving a ring of slumped sediment around the lakeshore (Plate II). Thermal erosion has preferentially enlarged the east side of some thaw lakes due to degradation by waves and lake ice (e.g. Harry and French, 1983; Carter, 1987). Thermokarst gullying is prominent around large thaw lakes where thaw erosion proceeded through the Holocene.

1.7.4 Patterned ground

Common patterned ground features are nonsorted circles, polygons and nets. The most common of these are irregular circles or polygons (Washburn, 1980), about 0.5 m to 3 m in diameter, with slightly domed centres, rising several centimetres above their borders. These forms are best developed on solifluction diamicton or till, but they also occur on silty marine and sandy or gravelly sediments. Nonsorted circles and polygons, with relatively flat and often bare central areas, are called mud polygons (Washburn, 1947) or mud boils (Shilts, 1978).

The smallest nonsorted patterns are polygons, 10-25 cm in diameter, nested within larger forms. They are thought to be "desiccation" polygons caused by expansion and contraction cycles related to wetting and drying.

Large-scale nonsorted patterned ground, 30 m to 100 m

in diameter, includes polygons with high centres, low centres, and fissures. These large forms occur mainly on well-drained diamicton above marine or lacustrine limits (Fig. 1.10).

Smaller nonsorted polygons, with high or low centres and irregular margins, occur in lowlands on fine-grained soil. Low-centred polygons, 10 m to 20 m across have thin active layers, in some cases only 25 cm deep. Large associated cracks probably contain ice wedges at depth.

Coarse textured sediments on deltaic and glaciofluvial outwash are characterized by large-scale, nonsorted polygons, 10 m to 40 m in diameter (Fig. 1.12). Fissures 25-100 cm deep and 20-50 cm wide form the borders of these polygons. On delta or terrace fronts the forms become larger and elongated downslope (Fig. 1.12).

Sorted patterned ground is less common than the nonsorted variety. Sorted polygons, 0.5 m to 4.0 m in diameter, were found on moist outwash sediment and on very wet morainal slopes. Mechanisms for producing sorted circles include various types of upfreezing (Anderson, 1988) and free convection cells (Hallet and Anderson, 1988). Convection cells may be related to heat during thaw which sets up density contrasts and drives a circular motion in the active layer (Kranz, Gleason and Caine, 1988). In a similar manner, convection cells may be driven by buoyancy forces that arise seasonally in thawing ice-rich soil due to differential compaction as the thawing front moves downwards (Hallet and Waddington, 1991).

1.7.5 Mass movement

Solifluction features, already documented on eastern Victoria Island (Washburn, 1947), occur across southern

Victoria Island, particularly on morainal slopes. The most prominent solifluction sheets or lobes have overrun raised shoreline features (Fig. 1.13). Morainal areas, terrace or lobate solifluction forms allow differentiation of periglacial forms from ice-contact features, in situ till sheets, or debris-flow deposits. The presence of organic debris provides the best means of distinguishing solifluction deposits from till. Without extensive trenching it is hard to confirm the suspicion that large areas of terrain are covered by solifluction debris. Dating of incorporated organics will help determine when solifluction was most active. However, based on the relative age it appears to have been most active in immediate postglacial times.

1.8 Methods of investigation

This study is based on field work and both office and laboratory analysis. Field work was mainly carried out in 1982-1985, 1987 and 1988 by means of twelve, week-long camps. Ground surveys consisted of traverses using 3-wheel Hondas and walking. Helicopters were used for sampling traverses and detailed site investigations.

1.8.1 Field methods

Typical field surveys included determinations of landform, sediment type, surface conditions, lithology of boulders and cobbles (in places), drainage conditions, composition, fabric and genesis of the material. Searches were made for datable material, particularly marine fossils. Sedimentary work included basic measurements, lithofacies description, sampling and photography.

Detailed sediment logging was carried out at selected sites on sections exposed and cleaned by washing, using

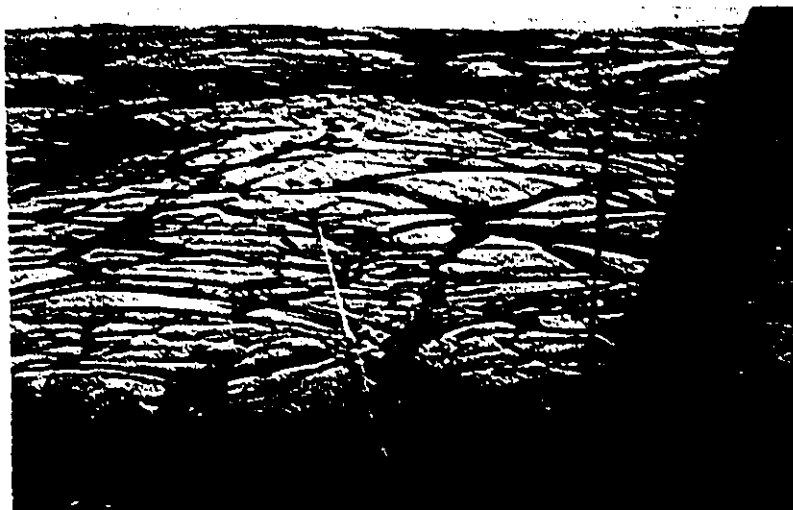


Figure 1.12 Large polygons with fissures on deltaic sediment; GSC 204766-J



Figure 1.13 Solifluction lobes overrunning raised marine terraces. Circular depressions are hunting pits
GSC 2047664-K

hand-portable water pumps. Trenches were dug when no natural exposure or slope was available. Soils pits were also dug and soil profiles were measured to investigate: 1) soil and sediment genesis; 2) the relative age of drift surfaces; 3) the nature of the active layer; and, 4) till geochemistry (Sharpe, 1991). These excavations also provided data on sediment composition, structure, fabric, depth of the active layer and presence and nature of ground ice.

1.8.2 Laboratory methods

Laboratory methods mainly involved standard techniques for analyzing samples for texture, geochemistry, and mineralogy. Laboratory analyses were conducted at the Geological Survey of Canada sedimentology laboratory, except for geochemical analyses which were carried out at Bondar-Clegg and Co. Ltd. Textural analysis consisted of disaggregation, sieving and pipette analysis on particles 2 to 0.002 mm (2 μ m). Geochemical analyses were carried out on the clay fraction (<2 μ m). The following trace elements were analyzed by atomic absorption, following hot leach treatment with nitric and hydrochloric acid: Co, Cr, Cu, Fe, Mg, Mn, Ni, Pb, and Zn. Colorimetric techniques were used for As analysis and fluorometric methods were used for uranium determinations. Accuracy and precision estimates for the trace elements are at Bondar-Clegg, Ottawa (1986).

Total and organic carbon contents were determined on the silt-plus-clay fraction using an electric induction furnace-combustion gas carbon analyzer (LECO) and acid digestion techniques, the difference being taken as percent nonorganic carbon. The value produced is equivalent to a calcium carbonate content that would be required to give the measured result, that is expressed as a percentage by weight.

The clay fraction was analyzed for clay minerals. Heavy mineral analysis was carried out on the fine sand fraction (63-250 μm). Bromoform (Sg. 2.83-2.85) was used to separate the light and heavy mineral fractions.

1.8.3 Office

Fabric data are displayed as contoured stereo-net projections and statistics on the strength and significance of the directional data are calculated (Woodcock and Naylor, 1983).

The maps central to this work were compiled mainly by mapping landforms and sediments on aerial photographs. More than 2000 photographs (scale of 1:60,000 flown in 1958) were annotated using tonal patterns and forms that could be related to ground observations (see numbers on Map 1650). Oblique aerial photographs (Trimetragon series) aided landform analysis. The Quaternary geological map presented here (Sharpe and Nixon, 1989; Map 1650A; Fig. 1.14) depicts the surficial materials and landforms, with genetic interpretation. The deposits are described in the legend of this map and three major landform-sediment associations are highlighted in Chapters 2 (hummocky moraine), 3 (streamlined landforms, and 4 (eskers). Appendix III presents a summary of the geologic units on southern Victoria Island and relates them to landforms (Fig. 1.14) and materials.

An important component of the mapping program was the use of LANDSAT images to identify regional patterns (Fig. 1.6). LANDSAT images aided the identification of several large-scale features depicted on Map 1650A: ground moraine, hummocky moraine, individual end moraines, lateral moraines, large outwash deposits, large deltaic accumulations, drumlin fields, other longitudinal landforms (e.g. fluting or

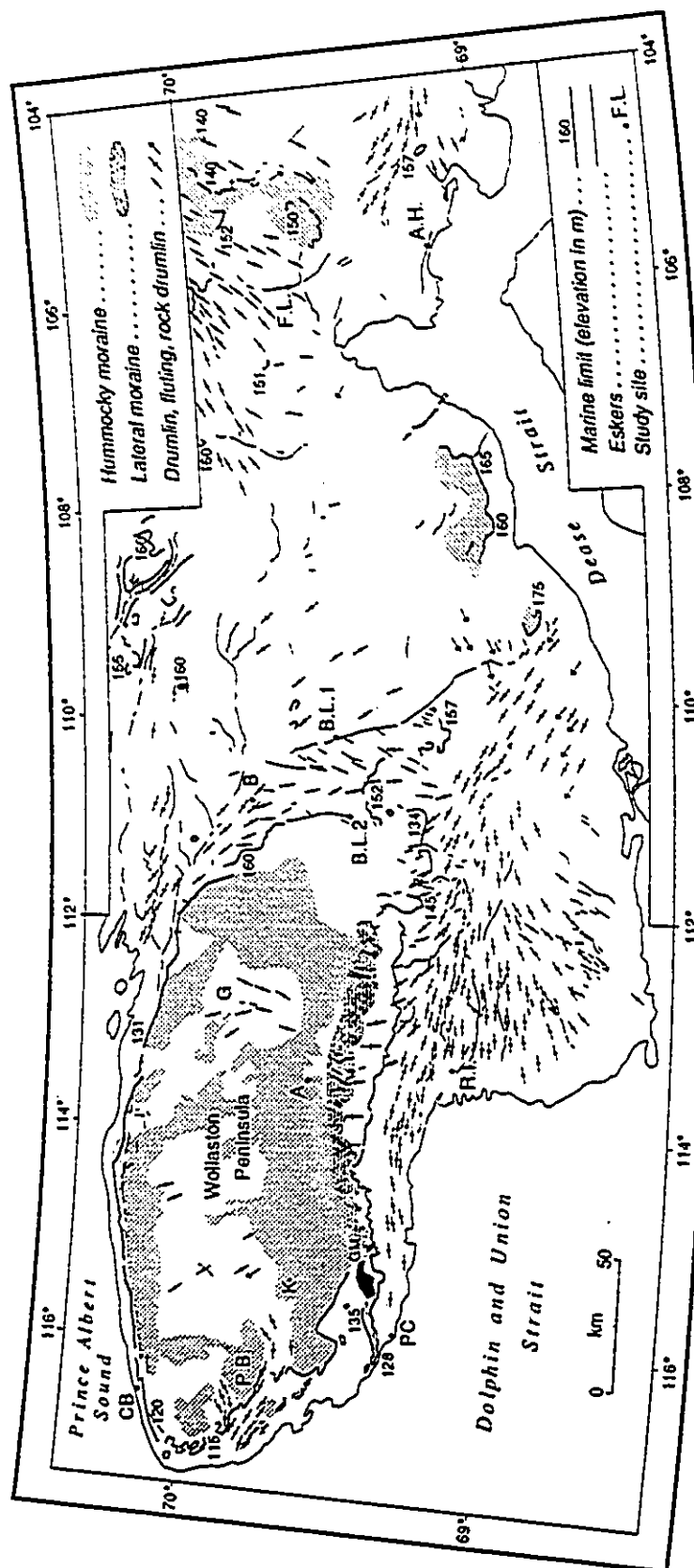


Figure 1.14. Map of glacial landforms southern Victoria Island. R.I., B.L., F.L., A.H., P.C., C.B., P.B., study sites (see text). G=glaciomarine fan.

drumlinoids), and marine limit where it is defined by vegetation contrast, were present (Sharpe, 1988a).

Regional-scale structural patterns, jointing and faulting systems in the underlying carbonate bedrock, could also be readily identified because of the large areas covered on LANDSAT images. For example, some joints are easily misidentified as linear glacial features (e.g. Clark and Boulton, 1991), while others are enhanced by linear ridges of rocks produced by postglacial frost heave (Fig. 1.12b).

Colour and tonal patterns were related to moisture regime, vegetation, and ultimately soil textural and permafrost conditions on the ground (Appendix I). For example coarse-textured, well-drained, poorly vegetated (< 25% ground cover) morainal areas appear light in colour (Fig. 1.6). White areas in topographic lows (present seashores) are mainly unvegetated marine silt deposits (possibly redistributed by wind). Other light areas indicate sparse vegetation on sandy morainic or outwash areas. Dark colours are well vegetated (<75% ground cover) areas, and coastal areas where fine marine sediments occur below marine limit.

2.0 MORAINES

This chapter discusses moraine landforms in an attempt to report on their composition and understand their genesis. The genesis of many glacial landforms is poorly understood and their sediments are important to understanding their origin. The spatial arrangement of landforms (landform associations) is also important to interpretation (Punkari, 1982) and this has been used to make inferences about glaciological conditions during their formation (Clayton and Moran, 1974; Boulton and Jones, 1976; Aario 1977). Landform analysis is especially informative in small-scale mapping where regional relationships and transitions between landforms can be observed (Aario, 1977; Shilts et al., 1987; Shaw, 1988; Bouchard, 1989). Thus, landform models rely on knowledge of landform-sediment links and landform associations.

Links between glacial landform models, glacial thermal regime, permafrost, and periglacial terrains are relevant to work on Victoria Island. Glaciers may be cold-based or have cold marginal zones (Weertman, 1961), which are important controls on sedimentation and on landform sets (Boulton, 1972b; Clayton and Moran, 1974). Glacial cover is also a major control on permafrost and periglacial conditions.

Ground moraine, hummocky moraine, end moraine, lateral moraines, and streamlined landforms are prominent on Victoria Island (Fig. 2.1a). These landforms occur in zones that may result from different flow dynamics within one ice lobe or ice stream (Sharpe, 1984; 1988a). Others discuss landform zones in adjacent areas (St-Onge and McMartin, 1987) and on a continental scale related to the inner areas of large ice

sheets (Shilts et al., 1987; Alysworth and Shilts, 1989a,b; Bouchard, 1989). The landforms are systematically arranged in belts (Fig. 2.1a,b) and are discussed in order. In addition, the sedimentary characteristics of landforms with exposed sediments are presented: 1) to describe the sediments within specific landforms; 2) to record their arrangement and structure; 3) to investigate the origin of landforms (depositional or erosional); and 4) to allow sedimentological inferences to be drawn bearing on ice dynamics, bed conditions, and ice retreat patterns.



Figure 2.1 b. Oblique aerial view of major set of landforms, Wollaston Peninsula, looking east. 1. Elongate drumlins, 2. Stubby drumlins 2a. Drumlin margin, 3. Flutings, 4. Deformed moraine, 5. Lateral moraine, 6. Hummocky moraine, 7. Ground moraine. NAPL T328L-204.

2.1 Ground moraine

Ground moraine, smooth to rolling drift composed of subglacial diamicton and subglacial outwash (Goldthwait, 1989), covers a large area of low-relief terrain, particularly on the central carbonate plateau of Wollaston Peninsula (Fig. 1.4, 2:1a). The plateau consists of featureless terrain, broken by minor moraine segments and by a network of drainage systems, which mark a series of ice-marginal positions (Sharpe, 1984). The drift is thin, sandy silt diamicton interpreted as glacial sediment, till, reworked by solifluction in places, with minor areas of glaciofluvial sand and gravel and sporadic occurrences of glaciolacustrine silt.

Drift composition

A regional till, with uniform grain size, rests on stratified sediment (Sharpe, 1988a). This till sheet has various styles of lower contact: conformable, sharp, transitional, but not deformed. Bedrock directly underlies thin till in areas showing subdued fluted forms. This till is lithologically similar to diamicton found elsewhere in end and lateral moraines and on the surface of hummocky moraine. Its characteristics, grain-size distributions, pebble lithology, till-matrix carbonates and trace element concentrations are presented in Appendix IV, which also includes information on sample locations, and the bedrock and surficial geology of southwest Victoria Island (Nixon, 1988).

A high percentage of striated carbonate clasts indicates the importance of local erosion, incorporation, and transport of basal sediment. The matrix of most tills sampled, contains 50% or more matrix carbonate and samples taken ~0.2 m above bedrock show very high carbonate concentrations in the matrix (e.g. > 80%; Sharpe, 1991).

The grain-size characteristics of till from Wollaston Peninsula reflect homogeneous carbonate bedrock (Sharpe, 1991). Till-matrix textures show little regional variation (Appendix V) and they are similar to those of tills formed on carbonate rocks in other areas, for example, southern Ontario (Appendix V). This correspondence results from mechanical abrasion during glacial transport that produces a terminal size related to rock type (Vagners and Dreimanis, 1971). Abrasion of carbonate rocks produces an abundance of silt (Vagners and Dreimanis, 1971; Halderson, 1987), although, silt is also produced from these rocks by in situ weathering (Tedrow and King, 1962).

Drift composition may be used to infer glacial movement (e.g. Shilts, 1973; DiLabio, 1981). Proportions of carbonates (Fig. 2.2a) provide the best indicators of ice movement in southern Wollaston Peninsula. Very low proportions are recorded in samples from the area northwest of the Richardson Islands, which are underlain by Precambrian gabbroic and low-grade metasedimentary rocks. This dispersal pattern produced a shadow (dilution), the trend of which contrasts with the east to west flow direction indicated by streamlined landforms. Till trace elements (copper, chromium and nickel, Appendix VI) also show slight anomalies located over and downflow from Richardson Islands. In particular, Cu shows a strong anomaly to the west of the Richardson Islands (Fig. 2.2b), which also suggests northwest dispersal.

On the basis of morphology this regional till was originally considered to be of Middle Wisconsinan age or older (Vincent, 1982), but it is now argued to be of Late Wisconsinan age (Sharpe, 1984; Vincent, 1989). The lithology and age of the ground moraine drift provide no constraints to

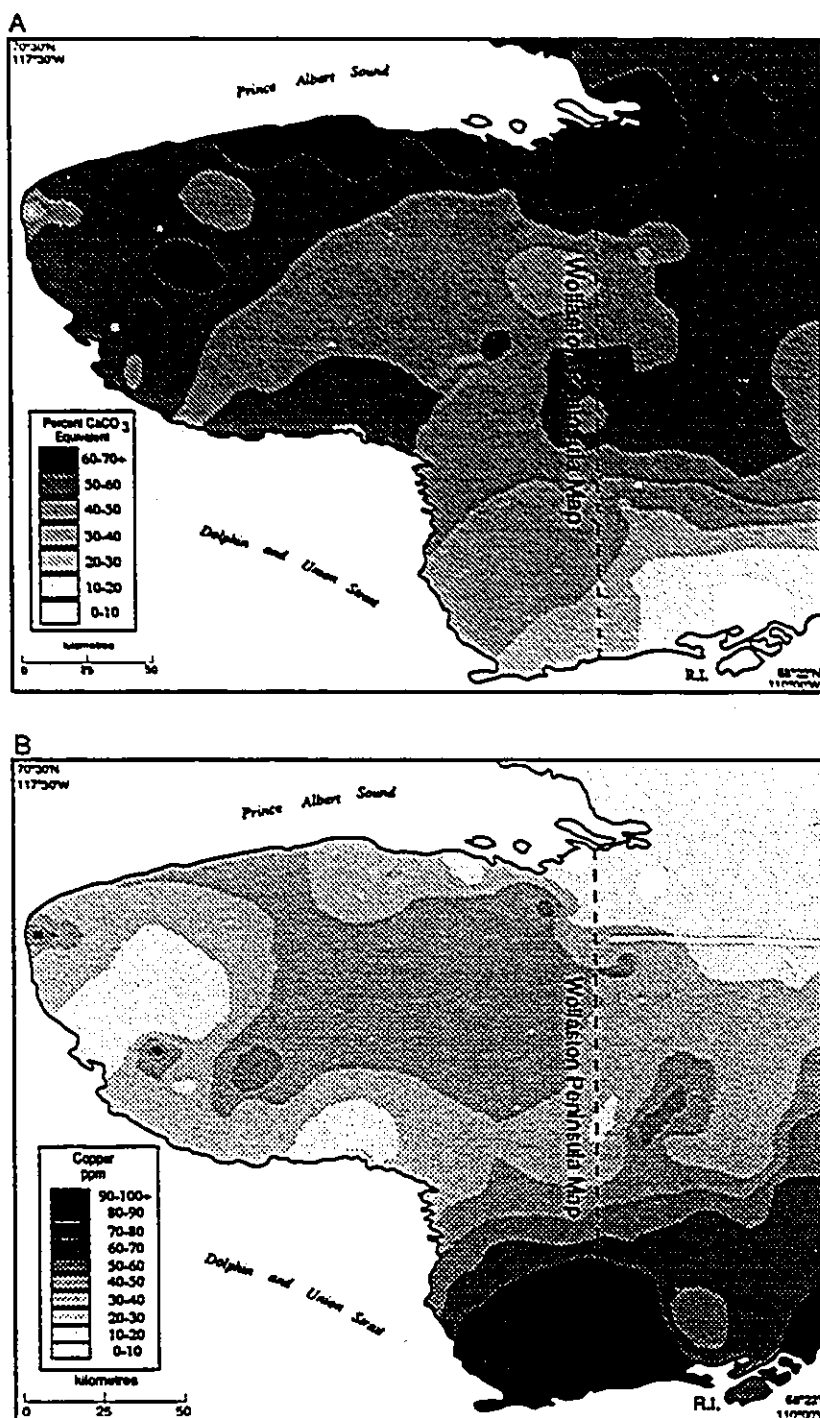


Figure 2.2. Contoured plots of till variables showing statistically significant patterns for western Victoria Island: a. carbonate content (2-81%); b. Cu (15-469 ppm).

a discussion of the nature and origin of ground moraine sediments in thick and thin drift areas.

2.1.1 Thick drift areas (Kugaluk River and Boffa Lake)

Drift thickens where ground moraine is transitional to local drumlin fields or fluted terrain. The two measured sections in thick ground moraine record stratified sediments beneath till. These lower sediments are considered part of the ground moraine deposits where there is no deformation at the contact between the units and where the contact is conformable and possibly gradational (Fig. 2.3a), but they may also be older sediments. The thick sediments at Kugaluk River consist of a lower massive, stony, sandy silt diamicton (A), overlying poorly sorted, matrix-supported gravel (B), which rests on an erosional contact. Unit B is overlain by a thick sequence of stratified sand (graded and massive) (C and E), with some disseminated organic material, and rhythmites in silt and clay (D and G). A transitional zone with interbeds of thin silty diamictons (F), grades upwards to a rhythmic unit (G; Fig. 2.3a). A sharp, conformable contact occurs at the base of a, massive, pebbly, sandy silt diamicton (H); this is interbedded at its upper contact with a weakly banded silt (I) containing marine shells (Fig. 2.3).

The origin of the upper diamicton and its relationship to underlying stratified sediments is central to interpretation of the sequence. The massive nature of the uppermost diamicton, its thickness (5-10 m), abundance of striated and faceted clasts, and, the strong preferred fabric aligned parallel to ice flow (Fig. 2.3a) all indicate subglacial or basal deposition. This diamicton rests on underlying undisturbed stratified sediments in a similar fashion to melt-out sequences (Shaw, 1982), although sandy

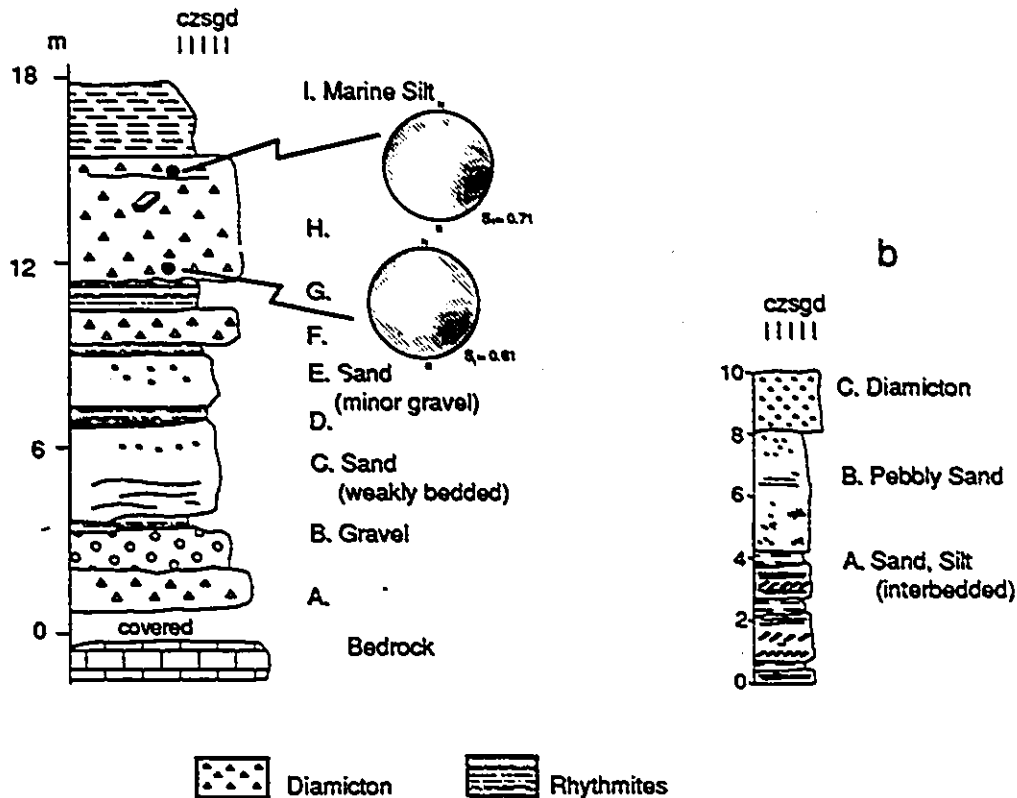


Figure 2.3. Measured sections and stratigraphy of diamicton and stratified sediments: a. Kugaluk River (Site 872); b. Boffa Lake (Site 846). Size scale: c=clay; z=silt; s=sand; g=gravel; d=diamicton.

lenses and sorting within the diamicton, usually indicative of former englacial conditions, are lacking. Sediment gravity flows are ruled out because they are generally not more than a few metres thick; they may have a random fabric but more commonly show a preferred fabric not aligned with ice flow (Lawson, 1979); and they commonly occur as multiple units. The stratified sequence beneath the diamicton unit fines upwards from gravel to rhythmically laminated silt and clay with no hiatuses. These considerations suggest that diamicton H was deposited subglacially from debris-rich ice which moved over marginal outwash without deforming it or was let-down onto subglacial beds. The interbedding at the top

of the diamicton unit indicates some sediment flows, possibly as ice was lifted from its melted bed by meltwater.

The disseminated organics within sandy beds are clearly not *in situ* and were washed into the site during sedimentation. The infinite date (Sharpe, 1991) presumably relates to an earlier, former ice-free period (section 5.1).

Most contacts are gradational involving interbedded units (I, Fig. 2.3 a). Such a sequence of beds appears to portray continuous deposition in standing water by high-energy density currents. The coarser sediments are interrupted by suspension deposits of low-energy density currents at the base of the sequence. Silt bands near the top of bed H suggest sediment flow as ice lifted from the bed. The transition to massive silt with marine fossils near the top indicates a shift to marine deposition.

The sequence fining upwards from gravel to clay is opposite to that expected for deposits associated with a terrestrial ice advance. Therefore, it may be older and unrelated to the overlying till. More likely the sequence represents glaciofluvial sedimentation in deep water (fan deposition; Rust, 1988) followed by grounding of debris-rich ice (Powell, 1990).

A similar, yet thinner depositional sequence shows diamicton over undeformed stratified sediment, pebbly sand, rippled sand and silt at Boffa Lake (Fig. 2.3b).

2.1.2 Thin drift areas

Areas of thin drift are indicated by frost-weathered fracture patterns and exposed bedrock surface structures. Sediment cover in these areas is 1-2 m thick. In places there is faint fluting, which may be transitional from hummocky moraine, (Fig. 2.4a); elsewhere the landsurface is

featureless. Eroded areas may contain rock drumlins (Fig. 2.5). One metre deep pits show a predominance of thin compact diamictons, of similar composition, but with slight colour differences or interbeds of sorted sand or loose, poorly sorted gravelly sand (Fig. 2.6). Contacts are sharp where near-surface beds of aeolian fine sand rest on diamicton; this indicates little mixing from cryoturbation (Plate II). Whereas fabrics from these diamicton beds show weak preferred orientation in featureless areas, there is consistent preferred orientation of clast fabric in fluted areas (oriented ridges, Fig. 2.4b).

2.1.3 Discussion

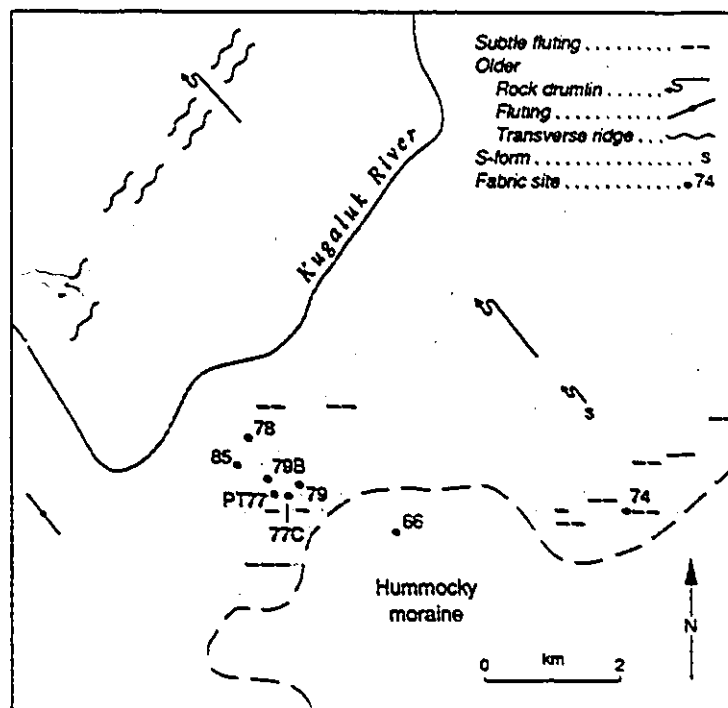
Ground moraine is usually viewed as mainly consisting of lodgement till or other basal tills e.g. a till plain (Boulton and Paul, 1976; Boulton, 1982; Goldthwait, 1989). Here, ground moraine, including thin drift with minor fluting, comprises sandy beds interbedded with diamicton (e.g. Sharpe, 1988c). This implies that the stratified deposits are not overridden older sediments (Fig. 2.3a) and that glaciofluvial sediment was delivered to the subglacial environment (Gustavson and Boothroyd, 1982; 1987). Where interbedding is equivocal, underlying undeformed beds may simply be older beds unrelated to the overlying diamicton and an integrated sediment model cannot be applied.

Level ground moraine in central Wollaston Peninsula has been interpreted as older terrain (Vincent, 1982), but because stratified ridges link this ground moraine to hummocky moraine (Fig. 2.4; 2.6) a younger, Late Wisconsinan, age is preferred.

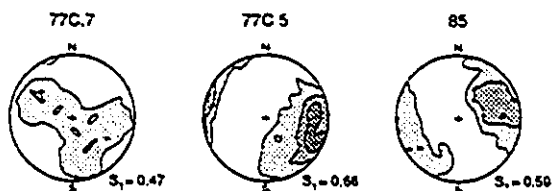
This level ground moraine was thought to have been affected by cold-based glacier conditions (Sharpe, 1984), but



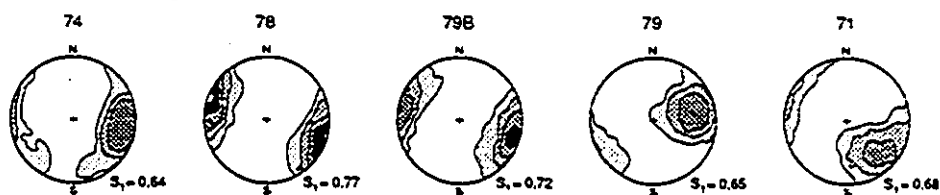
Figure 2.4. a. Aerial photograph of thin drift transitional from hummocky moraine. Note transitional ridges (t), rim ridge (rr), transverse ridges (tr) and erosional drumlin (d) with transverse ridges. NAPL A16129-113.



A. Non-oriented ridges



B. Oriented ridges



C. Hummocky moraine

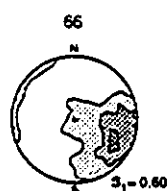


Figure 2.4b Map of (fluted?) ground moraine linked to hummocky moraine by narrow transitional ridges (subtle fluting). Fabric diagrams relate to ground moraine, flutings and hummocky moraine.



Figure 2.5 Oblique view of rock drumlin with thin sediment cover and transverse ridges. Location on Figure 3.1 and 3.3. Scale bar = 1 km. GSC-205193D.

Figure 2.6. b) Exposed stratified sediments.



Figure 2.6 a) Oblique aerial photograph of elongate or "linear" ridges. GSC 204439-C.



these frozen basal conditions could not have been continuous in time or space unless the sorted beds were pre-existing or the meltwater required for their deposition was derived from supraglacial sources or from sources further up ice.

2.2 Hummocky moraine

2.2.1 Introduction

Hummocky moraine is amongst the most complex glacial terrains and several models of formation have been proposed for this landform. Hummocky moraine is a descriptive term used to group drift accumulations comprising hummocks, mounds, ridges, drift plateau, and kame and kettle forms associated with ice stagnation. For some, it has a supraglacial connotation related to supraglacial origins (cf. Clayton, 1964; 1967; Boulton, 1972; Lawson, 1979); others consider it subglacial in origin (Hoppe, 1952, 1957; Stalker, 1960; 1973; Aario, 1977; Berendregt and Stalker, 1989 a,b). Hoppe (1952) considered moraine plateau with rim ridges (Veiki moraine) as clear evidence of formation by subglacial squeezing. He based this conclusion on reports of basal crevasse filling on modern glaciers (Gripp, 1929) and detailed fabric studies in Norbotten, Sweden. Stalker (1960) further demonstrated a subglacial (pressed) origin by recording deformed bedrock masses in hummocks on the Canadian prairies. However, Clayton (1967) argued that moraine plateau containing collapsed lake sediments provided clear evidence of a supraglacial origin for hummocky moraine in North Dakota. Subsequent thermokarst melt (Clayton, 1964) and slumping of supraglacial sediment results in topographic inversion and sediment reworking clearly displayed at glacier margins (Boulton, 1972; Lawson, 1979; 1981; 1984). Gravenor

and Kupsch (1959) considered that combined processes, ablation of supraglacial sediment and squeeze up of basal sediment, produced hummocky, ice disintegration forms on the prairies. Low-relief hummocky moraine, 5-10 m high hummocks in basal till, occurs in Keewatin (Aylsworth and Shilts, 1989) and central Québec (Bouchard, 1989).

Drift accumulations in hummocky moraine may be explained by glaciotectonic models which invoke ice-marginal compression at bed rises (Moran, 1971; Clayton and Moran, 1974; Moran et al., 1980; Mooers, 1990). Marginal compressive flow, which produces thick supraglacial drift sequences (cf. Clayton, 1964) as observed at Matanuska Glacier, Alaska (Lawson, 1979), is inferred to have had the same effects in North Dakota (Clayton, 1967). To entrain thick drift or large blocks of sediment, subglacial water, forced by a pressure head to the ice margin, is refrozen to the cold marginal zone (Weertman, 1961; Clayton and Moran, 1974). Marginal compression may also cause englacial folding and shear, producing Rogen moraine, a transverse form related to hummocky moraine (Shaw, 1979; Minell, 1979; Bouchard, 1980). A form continuum, fluted forms, to streamlined forms, to hummocky forms was used to argue for a genetic relationship between flutings, drumlins, and hummocky moraine (Hoppe, 1952; Lundqvist, 1969; Aario, 1977).

On Victoria Island, where ice-cored sediments, permafrost conditions, and complex groundwater conditions are present, the nature of hummocky moraine is further complicated (Sharpe, 1989). These landscapes are part of broader areas, termed hummocky terrain, in the western arctic (Prest et al., 1968). For example, hummocky terrain near Tuktoyaktuk is largely thermokarst (Rampton, 1974; 1988), and ice-cored moraine is common on adjacent, eastern Banks Island

(Vincent, 1982; French and Harry, 1988; French and Harry, 1990) and the mainland (St-Onge and McMartin, 1990).

Setting

A broad zone of hummocky terrain (up to 45 km wide and several hundred km long) encircles Wollaston Peninsula (Fig. 1.14; Map 1650 A; Fig. 2.1 and 2.7), except for a few locales where it is cross-cut by drumlin fields (Fig. 1.14). Streamlined forms also cut hummocky terrain, leaving streamlined residuals in places across southern Victoria Island (Fig. 1.14). The main hummocky forms comprise large irregular, diamicton-covered uplands (hummocks), depressions (mostly lake-filled), and sloping plateaus (Fig. 2.8a;) adjacent to solifluction stripes. Linear ridges (rim ridges?) are associated with large thermokarst depressions. Large, rounded hummocks commonly have isolated mounds, kames (Fig. 2.8b) or deltas perched on the surface (Fig. 2.8c). These latter features are sparsely vegetated, well-drained, and composed of sorted, stratified to massive sand and gravel. This hummocky terrain differs from smaller, uniform-relief hummocks, moraine plateau and rim ridges of the prairies (Stalker, 1960). In addition, hummocky uplands on Victoria Island commonly have retrogressive thaw slumps (e.g. Mackay, 1966; French, 1974), ranging in age from very old (possibly early postglacial) to modern where there is active, retrogressive failure associated with buried ground ice.

Ground ice

Massive ground ice, observed at several sites within hummocky moraine, was best exposed at Ammalurtuq Lake (Fig. 1.9). There, an active ground-ice slump, 8-10 m high and 50 m wide, exposes 5 m of ice below a 1-2 m cap of diamicton and above accumulating sediment. The ice is massive and relatively free of debris except for embedded, faceted

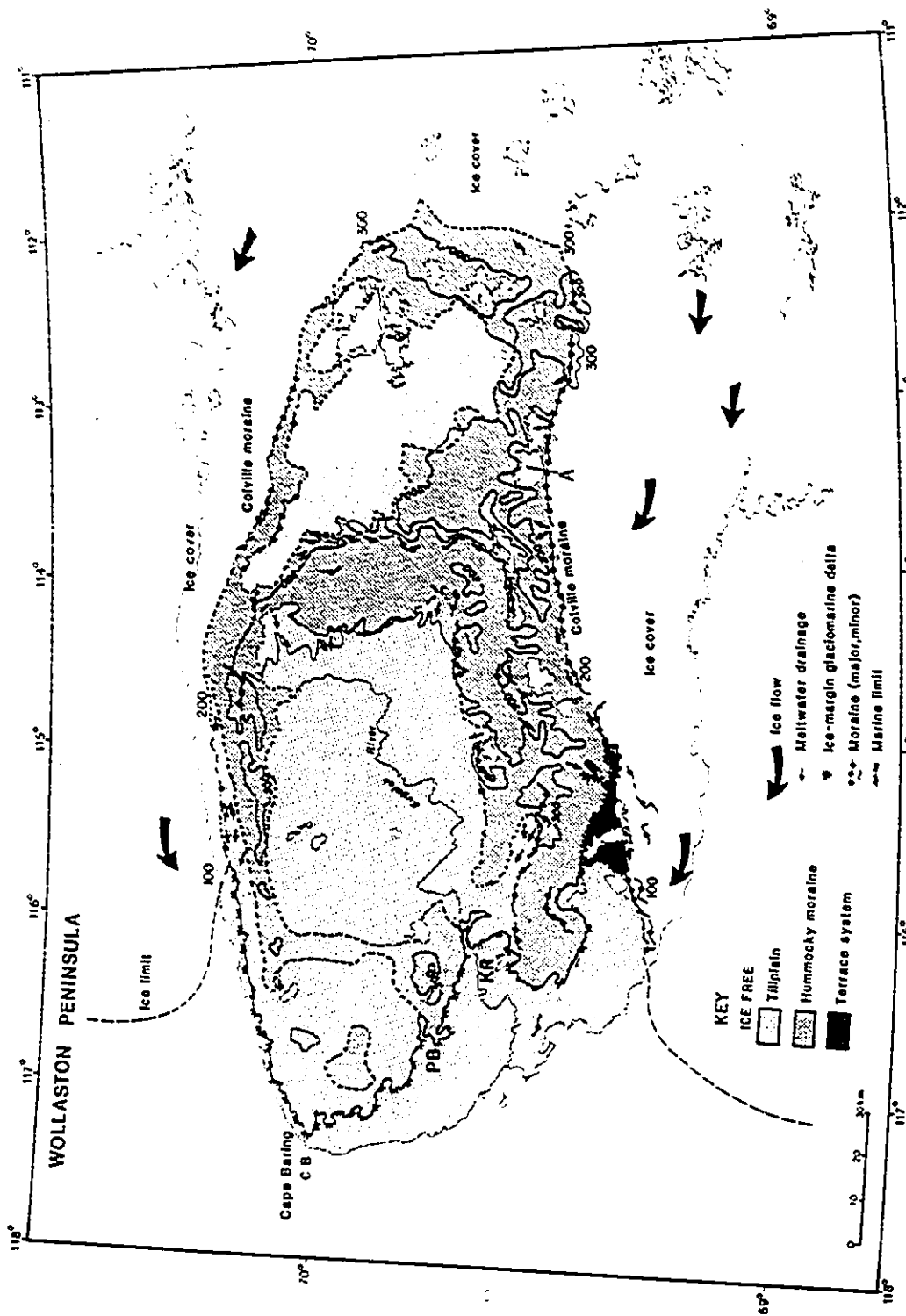


Figure 2.7. Map of hummocky moraine and configuration of active ice front. Elevations above sea level shown. A = Annalurtuq Lake, PB = Penny Bay, KR = Kugluk River.

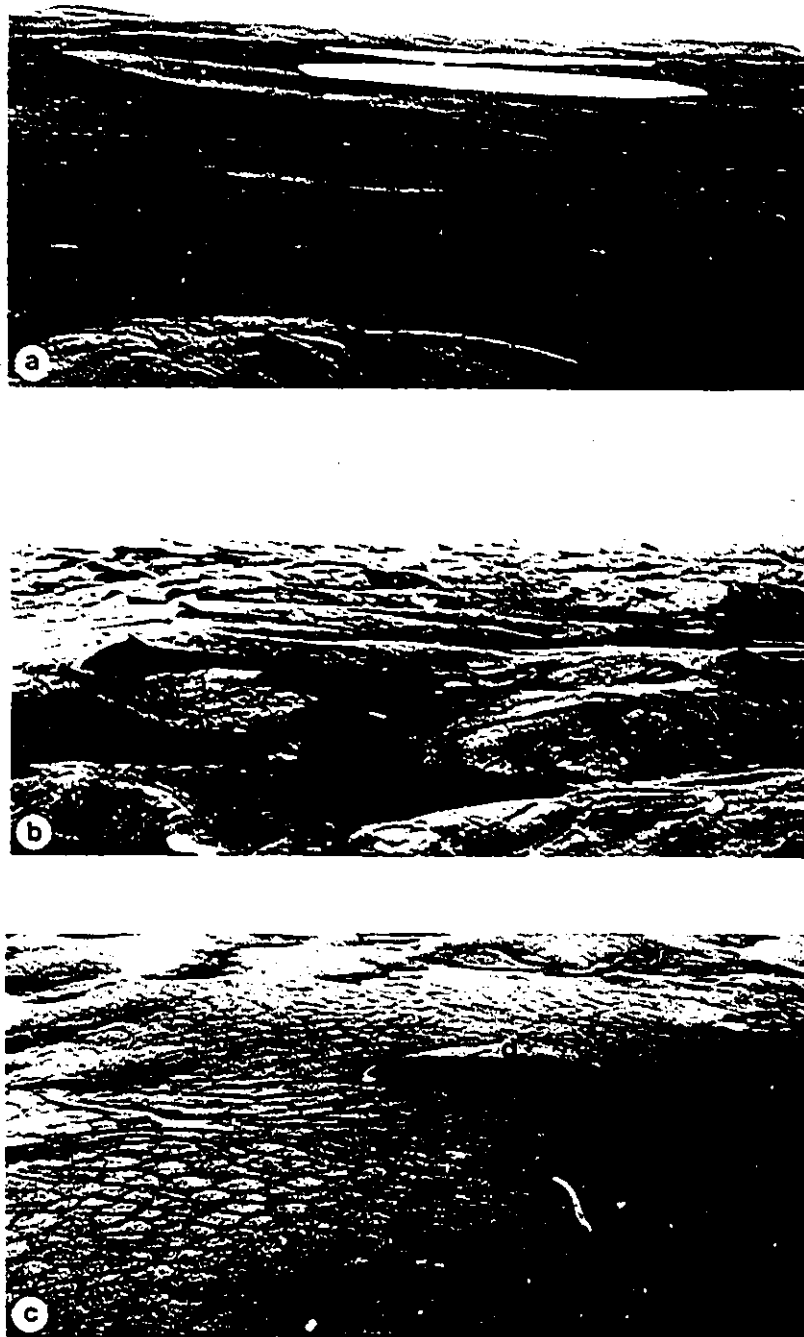


Figure 2.8. Hummocky moraine forms: a. sloping plateau (p) with solifluction stripes, GSC 204766-C; b. irregular mounds, debris cones and kames; GSC 204766-F; c. surface delta (d) on hummocky moraine uplands; GSS 112124-B.

boulders 30-40 cm in diameter. A few steeply dipping, debris-rich bands of ice were also observed. In one location, a vertically-oriented ice wedge with vertical banding was found penetrating the massive ice.

The area of inferred ground ice can be extended by mapping the scars of former ground ice slumps around Lake Ammalurtuq (Fig. 2.9) and across mapped hummocky moraine (Fig. 1.14).

The presence of massive ground ice in hummocky moraine also appears to correspond with the distribution of large (30-100 m), nonsorted polygons (Fig. 1.10). This correspondence between large polygons and ground ice distribution can be made across southern Victoria Island (Sharpe, 1989). Other surface features, active ground-ice slumps (Fig. 1.9) and former thermokarst erosion scars (Fig. 2.9), in association with large polygons, are also common on

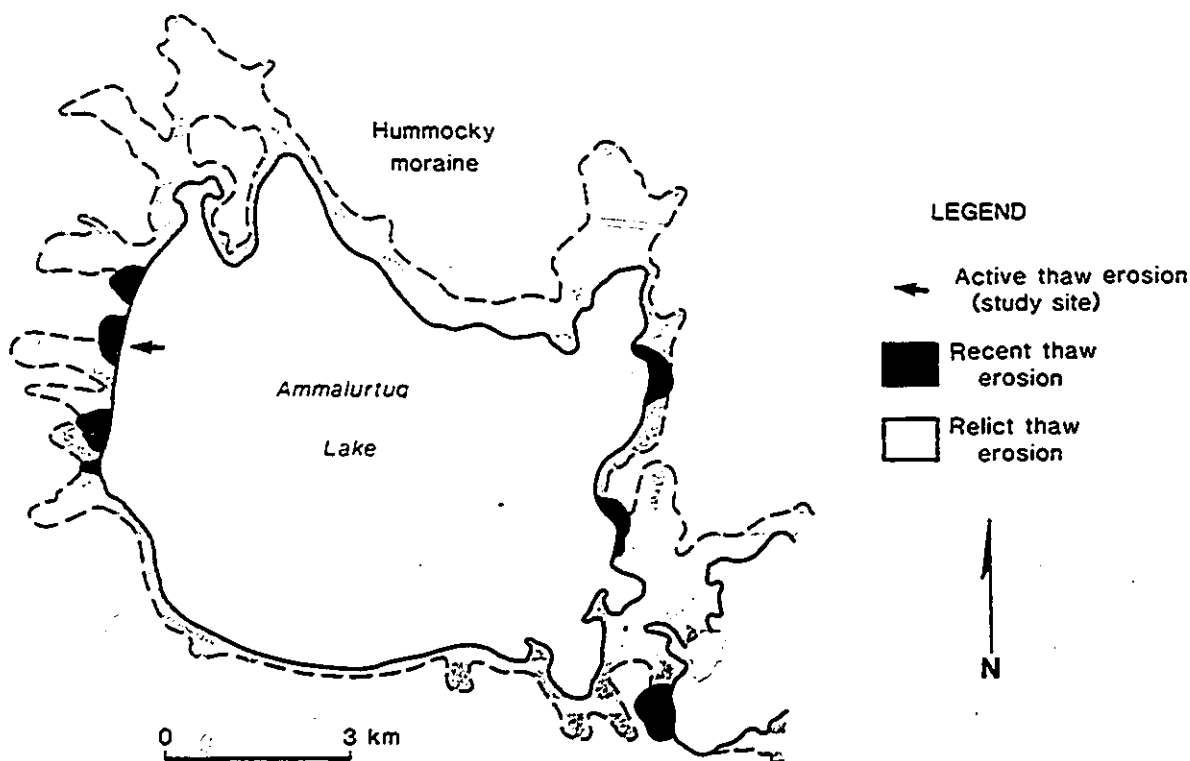


Figure 2.9. Map of thermokarst erosion scars around Ammalurtuq Lake.

hummocky moraine and on large lateral moraines. Kettles, thermokarst lakes, and hummocks are additional indirect indicators extending the area of probable buried ground ice to include most hummocky moraine (Map 1650A; Fig. 1.6). The pattern of large-scale polygonal forms, common on hummocky moraine, likely indicate thermal contraction cracking of massive buried ice (Mackay, 1984). In addition, adjacent, washed hummocky moraine (and end moraine) below marine limit, and streamlined terrain have no large polygons and only a thin (1, 2 m) sediment cover (Fig. 1.10). This difference in drift thickness likely reflects the removal of ground ice more than the removal of thick sediment during streamlining events and marine washing (Fig. 2.10). Elsewhere, a thick cover of slumped diamicton protects any buried ice from melting except at exposed lake, river or slump bluffs (Fig. 1.9).

This interpretation of ground ice extent (Fig. 2.10) suggests that the prominent hummocky and lateral moraines on Victoria Island are largely ice-cored (Sharpe, 1988a; 1989). A similar conclusion of extensive ice-cored terrain was reached to explain lateral moraines on adjacent Banks Island (French and Harry, 1988; 1990).

Most of the massive ice is interpreted to be Pleistocene in age and glacial in origin (Sharpe and Michel, in press). However at some sites, oxygen isotope analyses on the massive ice are insufficient to support either a Pleistocene glacial origin or a Holocene segregated origin independent of assessing the sediment record.

2.2.1 Measured sediments

Sediments in hummocky moraine may be useful to its

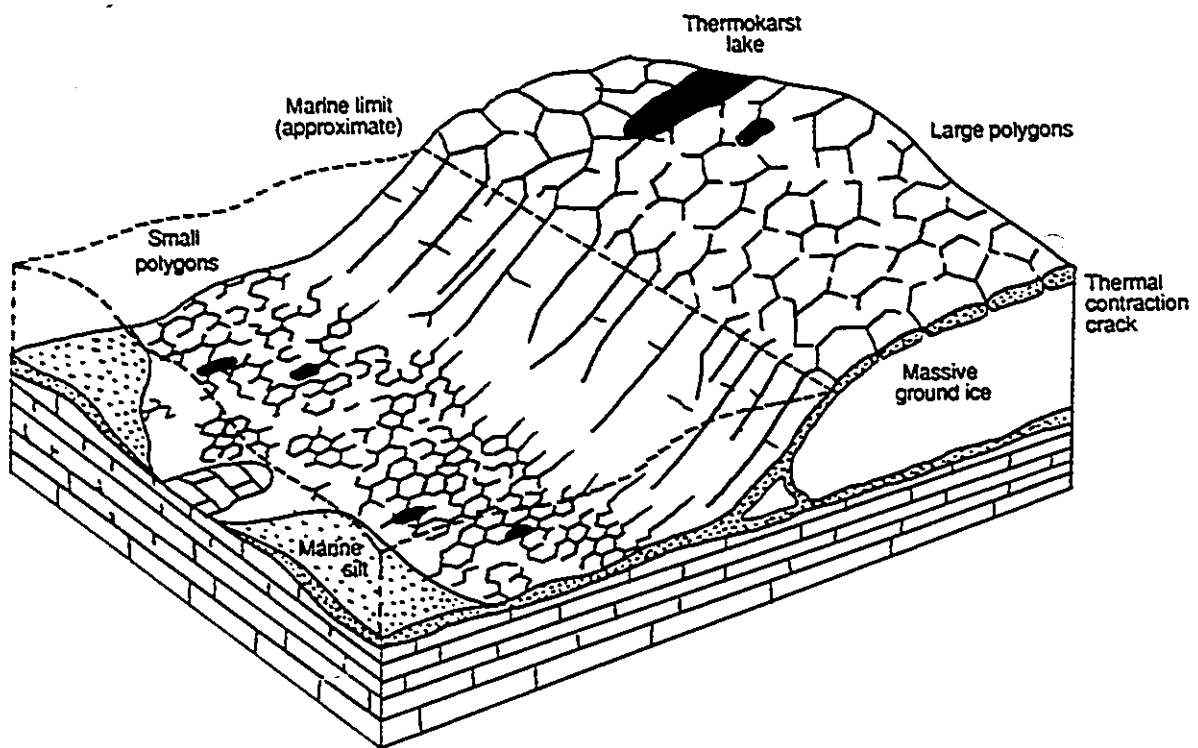


Figure 2.10. Schematic model of ground ice occurrence related to large, polygonal, thermal contraction cracks.

interpretation but few sections were studied: Ammalurtug Lake, Penny Bay and Kugaluk River (Figs. 1.14, 2.7). Surprisingly, the sections are dominated by sorted, stratified drift with only a low percentage of diamicton beds. This contrasts with interpretations in the literature that hummocky moraine forms by resedimentation of diamictic glacial debris from glacier surfaces (Lawson, 1979; 1983, 1984; Eyles, 1984).

Sediment fabric

Lawson (1979) used sediment fabric to help characterize basal ice, melt-out till, sediment-flow deposits and ice-slope colluvium in hummocky terrain along the ablating margin of the Matanuska glacier, Alaska. Melt-out till and basal ice showed high (0.75-0.95) S_1 values (the degree of scatter of (clast) axes about the mean) and a close grouping of the trend of the mean axis to regional ice flow (Lawson, 1979, fig. 46). Sediment flows and ice-slope colluvium may show strong alignment but lower (~0.5) S_1 values and a high variability in the trend related to control by local slopes (Lawson, 1979, fig. 63). He also noted an increase in S_1 values (0.5-0.7) as the water content and hence the mobility of the flows increased.

Ammalurtug Lake

The section at site 135 shows an undeformed, horizontally-bedded sequence consisting of thin diamicton beds (A), gravels and sands (B,C) capped by slope debris (Fig. 2.11a). The sand is mainly massive, the fine sand and silt normally graded. The underlying gravel is poorly sorted, weakly stratified, and interbedded with diamicton. The poorly developed structure in the sand and gravel units suggests that they are mass flow (Lowe, 1982) or hyperconcentrated flow deposits (Smith, 1986). The diamicton

beds are interpreted as debris flows because they are thin and interbedded with stratified sediment. The thin-bedded, finer sediments (fine sand and silt) indicate deposition in ponded conditions.

In the section at site 136II (Fig. 2.11b) tilted, faulted and distorted beds (Fig. 2.12) are interbedded with undisturbed horizontal beds. Finely laminated silt, fine sand and clay beds are interbedded with sand, gravel and diamicton beds. Interbedding of disturbed and undisturbed units indicates that the disturbances are syndepositional; failure occurred immediately after sedimentation of one unit and prior to deposition of the next.

The section at site 136III (Fig. 2.11c) occurs low in the hummocky moraine landscape within a large hummock with 20-40 m of relief. About 10-12 m of sediment, displaying solifluction lobes, covers the top of the section. Beds are horizontally bedded and include diamicton, gravel, sand and laminated silt and fine sand. Some beds are normally graded (Fig. 2.12) while sand, gravel and diamicton beds are massive or crudely bedded. Diamictons show diffuse to preferred fabric, unrelated to regional ice flow to the northwest.

Many beds in these three sections indicate deposition as sediment flows (Lowe, 1982). Suspension settling explains deposition of graded fine sand. Alternation of coarse and fine-grained units suggests an episodic sedimentation mechanism. Laminated silt and fine sand beds indicate quiet-water deposition. Repeated density underflow, including debris flows, may be inferred as the processes of deposition for the more massive units. Extensive local ponding can clearly be inferred from these sedimentary successions and the small surface debris cones and deltas (Fig. 2.8 b,c).

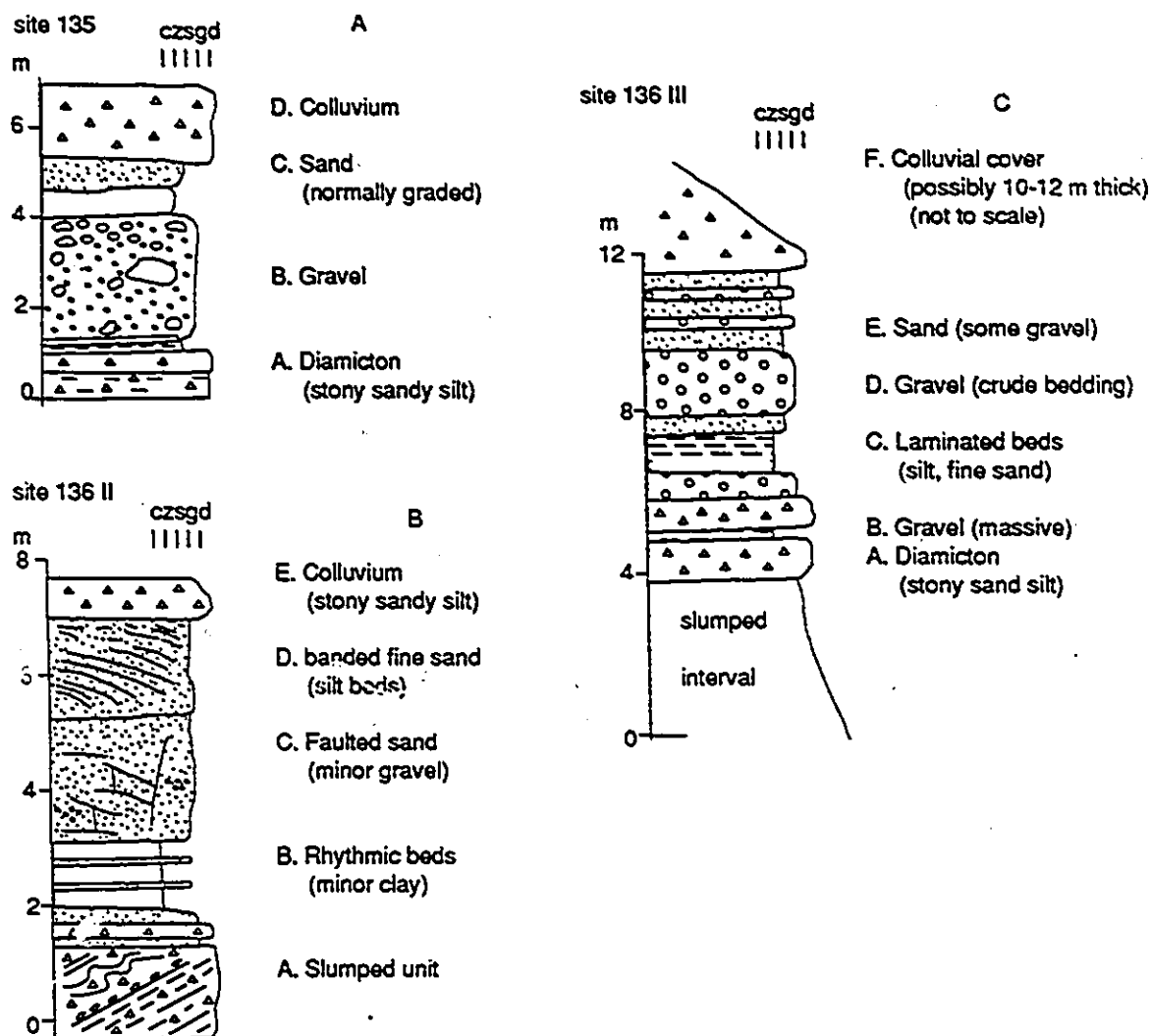


Figure 2.11 Sediment sequences in hummocky moraine from Ammalurtuq Lake; a. showing horizontal strata, site 135; b. Stratified beds with intra-bed disturbance, site 136II; c. Interbedded diamicton, sand and gravel and rhythmic silt within hummocky moraine, site 136III. Size scale: c, clay; z, silt; s, sand; g, gravel; d, diamicton.

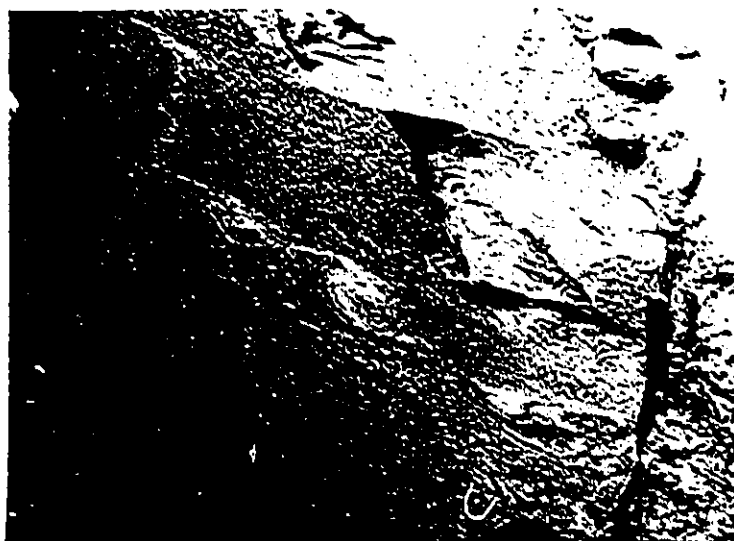


Figure 2.12. Vertical-standing, faulted, disturbed beds in hummocky moraine. Knife for scale; handle points up. GSC 1991-179.

Penny Bay

Hummocky moraine at Penny Bay comprises diamicton overlying sand and gravel. One section shows a metre of diamicton overlying 7 m of very coarse, poorly-sorted, boulder gravel containing pockets of gravel without matrix; thick diamicton occurs in an adjacent section. The diamicton is massive but fabrics reveal preferred orientations possessing an overall random pattern (Fig. 2.13). This represents debris flow or slumping processes controlled by local slopes rather than regional ice flow (Lawson, 1979). The frequent perched deltas on the surface indicate that surface run-off and ponding were common during stagnation. The perservation of these deltas indicates that melt and collapse are not pervasive.

Kugluluk River

Hummocky moraine south of Kugaluk River is transitional to small narrow elongate ridges (Figs. 2.4, 2.6) and commonly contains interbedded diamicton and sorted sand (Fig. 2.6).

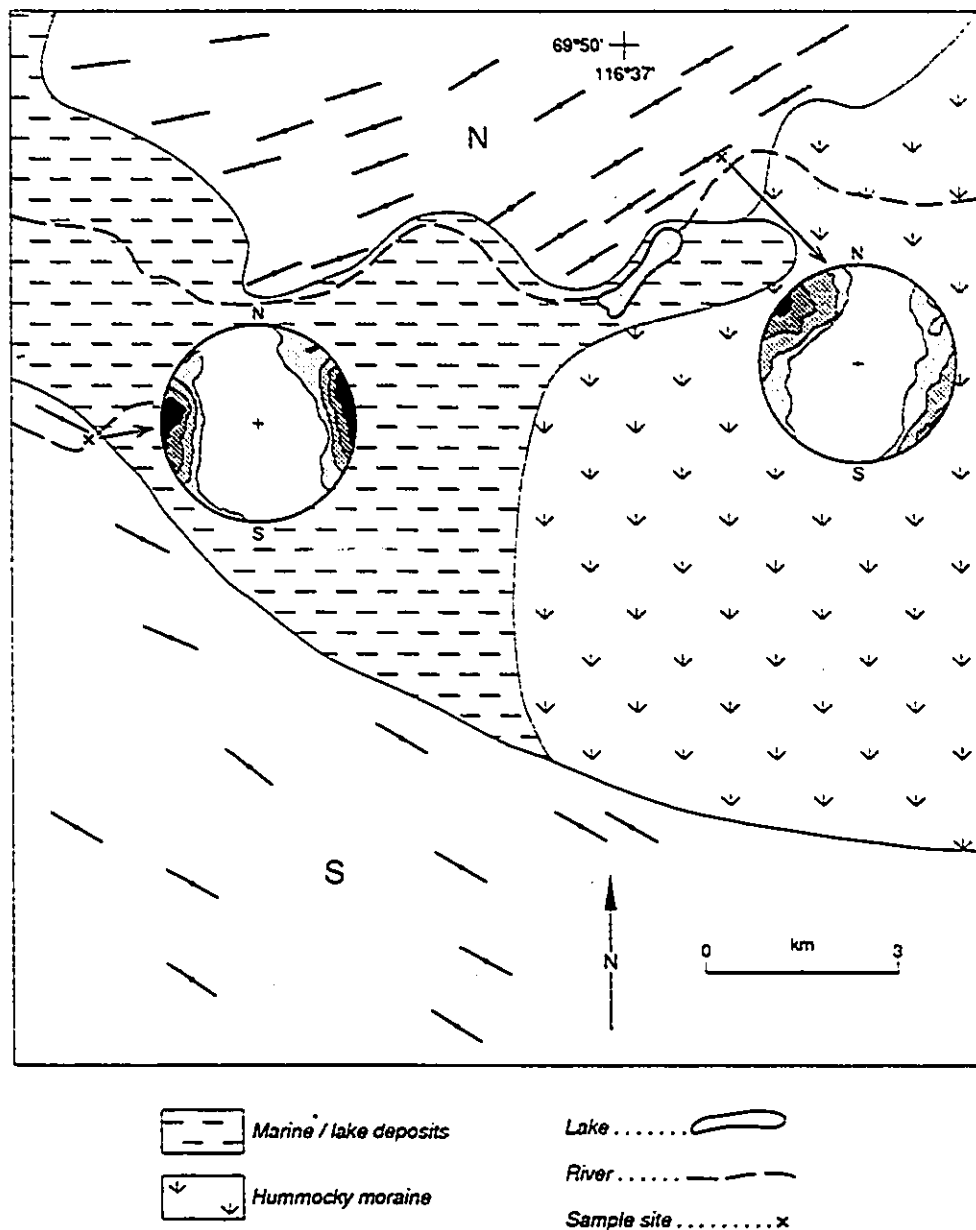


Figure 2.13. Hummocky moraine fabrics compared to fabrics in adjacent streamlined terrain, Penny Bay. Map based on NAPL photograph.

Fabric data show a preferred orientation, not in alignment with the latest east-west fluting, but with earlier flow to the northwest (Fig. 2.3). This may indicate debris-flow activity, sediment slumping, or inherited englacial fabric (Lawson, 1979). These hummocky forms comprise small hummocks and irregular ridges 2-5 m high, with no surface indication of ground ice (e.g. large polygons or thaw slumps).

Sections in thermokarst areas

Present-day thaw of massive ground ice, and slumping of accumulating sediment, is eroding and redepositing local debris (Fig. 2.14) in hummocky moraine. This produces thick massive diamictos on southern Victoria Island, but such beds are not common in the described sections (Fig. 2.11). Locally, present-day thermokarst processes also produce stratified fine sediment in small ponds in slump depressions. Supraglacial ponds were possibly larger in the past and thick stratified beds may have been deposited in them (Fig. 2.11). In addition, limited deformation of strata shows that widespread subsurface melting has not occurred because the thick sediment insulates the massive ground ice (Fig. 2.14); the sediment thickness is greater than the thickness of the active layer (~1 m), and has been since early Holocene time. A diamicton cap (~2 m) is sufficient to prevent massive ground ice melting from above (Fig. 2.11b).

Sections in sediment adjacent to an active thaw erosion site record past melting and slumping events (Fig. 2.15b). Measured sections comprise six main units: thick massive bouldery diamicton with striated clasts (units A, C, and F); an interbedded sequence of sand, silt, clay with ice and organic material, and minor gravel (units B, D and E); and interbeds of sand and clay with dated organics (unit B and



Figure 2.14. Hummocky moraine undergoing active thaw erosion. a. slump debris; GSC 204876-A. Note relict slump scars in adjacent terrain.



Figure 2.14. b. exposed massive ground ice with diamicton cap. Note large clasts embedded in ice and thin insulating cap (~ 2m). GSC 204441.

D). The whole sequence is horizontally bedded and shows little deformation. Ice lenses within clayey horizons (unit A and E) appear to be segregated ice. Organic material within the sequence has been dated at 8300 ± 90 BP (Wat-1640) and 7960 ± 120 years BP (Wat-1363) (Fig. 2.15b). These results indicate repeated melt and slumping in early post-glacial times (paraglacial sedimentation) followed by a long period of relative slope stability. The sequence may include more recent reworked slump beds but, to demonstrate this, further dating of deposits is necessary.

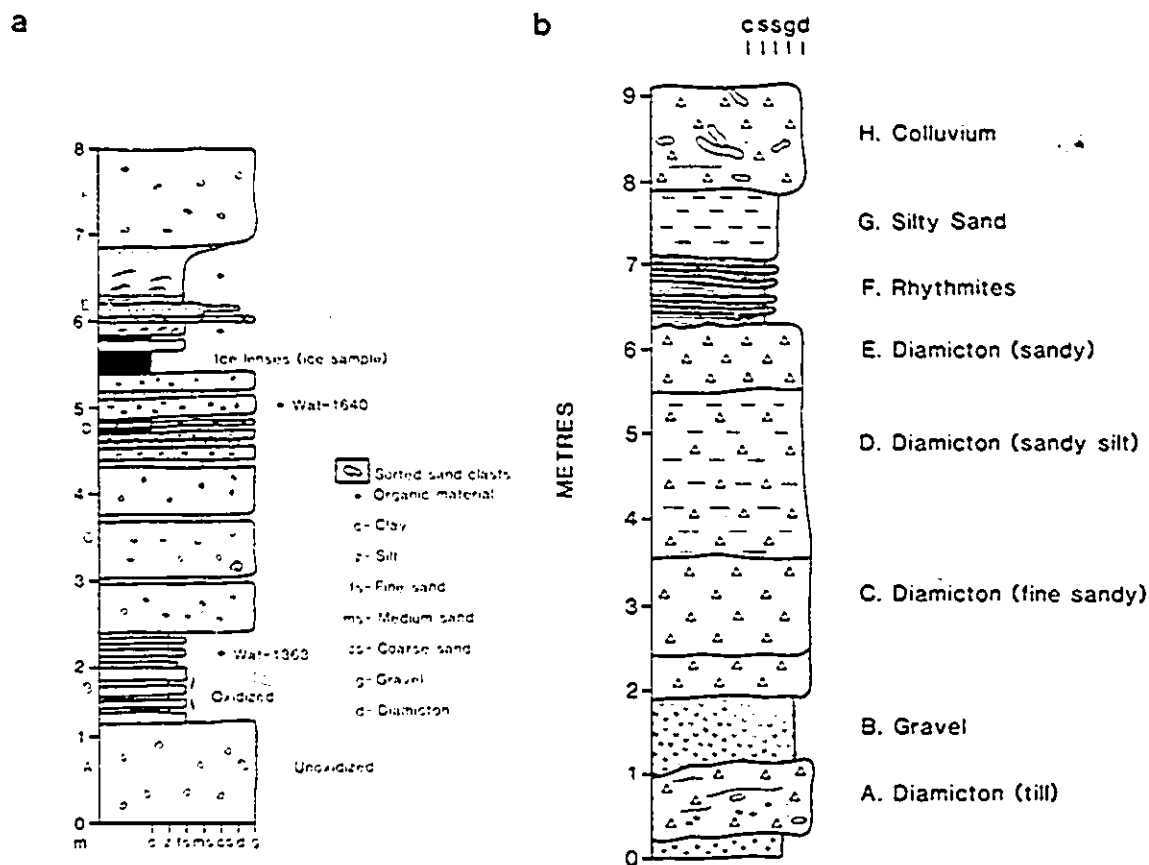


Figure 2.15. a. Sediment sequence in thermokarst erosion terrain at Ammalurtuq Lake. b. Idealized section in hummocky moraine from the Matanuska Glacier, Alaska (from Lawson, 1983).

2.2.2 Models of hummocky moraine formation

Most models of hummocky moraine involve debris-covered ice, so that means of moving debris to the glacier surface are important (e.g. compression, rising flow, shearing or thrusting at obstructions). Hummocky moraine is situated down-ice from and partly mantles a bedrock plateau forming the core of Wollaston Peninsula (Fig. 1.4). Where such scarps face up flow they cause compressive flow in overriding ice. Topographic obstructions are key elements in some models of hummocky moraine formation which use rising flow lines, or sheared flow, as an explanation for large volumes of supraglacial debris (Clayton, 1967). Hummocky moraine on southeastern Victoria Island, however, is not only tied to topographic controls.

Other models, based on study of recent glaciers (Boulton, 1967; 1968; 1972a; Lawson, 1979; 1984), concentrate more on explaining the sedimentary succession in hummocky moraine. Lawson describes sequences from Matanuska Glacier, Alaska, of mainly debris-flow diamicton. Supraglacial drift causes differential melt of an ice surface which promotes slumping, melting of freshly exposed ice, and topographic inversion. In the Alaska case, sediment is moved to a supraglacial position by marginal shearing with only minor topographic influence. This, together with subsequent slumping, produces a complex sequence of diamictons with only minor sorted sediment (Fig. 3.15b). The process of topographic inversion was applied in an explanation of hummocky moraine of the Missouri Coteau (Clayton, 1967) and the resultant landscape was referred to as thermokarst (Clayton, 1964). Similar processes inferred for permafrost regions have been termed "retarded deglaciation" where glacial ice is suspected (Astakov and Isayava, 1988).

On Victoria Island, there is little evidence of large-scale topographic inversion and sediment deformation (Fig. 2.11) as proposed in thermokarst models involving buried glacial ice (Clayton, 1967; Eyles, 1984). One reason for this is that extensive ground ice may still underlie much hummocky moraine (Sharpe, 1988a, 1989). The presence of small, undisturbed, perched deltas and kames on the hummocky moraine surface (Figs. 2.8b,c) suggests little disturbance of underlying hummocks. This implies that supraglacial debris is thick enough to prevent basal ice melting.

Evidence contradicting topographic inversion models, stratified beds, horizontal bedding, and few deformed strata, from Ammalurtuq Lake and Penny Bay requires revision of models for hummocky moraine formation (Lawson, 1979, 1984; Eyles, 1983, 1984).

Ground ice considerations

The suggestion of extensive buried ground ice (Fig. 2.10) inferred from active thaw erosion, relict thaw slumps, and the distribution of large polygons within hummocky moraine constrains potential models of hummocky moraine formation.

Widespread, massive ground ice is likely segregated ice (formed *in situ* under permafrost conditions) of Holocene age or older, glacier ice, or possibly refrozen meltwater of Pleistocene age. Segregated ice is very common in the western Arctic (e.g. Mackay and Dallimore, *in press*). However, preliminary results from southern Victoria Island favour a glacial origin or Pleistocene age for the massive ice based on debris and clast content, sharp upper contact, regional setting within hummocky moraine, dated stratigraphic position, oxygen isotope values, and the trend of co-isotopic variation (Sharpe and Michel, *in press*). A Holocene age and

segregation origin for the massive ice is unlikely because oxygen isotope values are low and there is a lack of sufficient water source given low present-day precipitation.

Neither the occurrence of modern buried glacier ice, except for remnants of alpine glaciers buried within end moraines (Østrem, 1964; 1965; Driscoll, 1980), nor relict glacier ice have been well documented. However, recent work indicates that glacial-age ice or glacier ice may exist in present permafrost terrain (Kaplyanskaya and Tarnogradskiy, 1976; Mackay et al., 1978; Astakhov and Isayava, 1988). Glacial ice was reported from northwest Victoria Island (Lorrain and Demeur, 1985), and Mackay (1983) suggested that the existence of buried glacier ice in the Mackenzie Delta region cannot be ruled out. Michel (1985; 1986a) reported probable glacier ice on Herschel Island and along the northern Yukon Coastal Plain. Dallimore and Wolfe (1988) report probable buried glacier ice from Richards Island, and French and Harry (1988) suggest that ice within the Sandhills Moraine of southern Banks Island is glacial in origin, as is ice in Bluenose Lake moraine (St-Onge and McMartin, 1989).

A relationship between ground ice and meltwater drainage may have aided formation of hummocky moraine or ice-cored terrain because extensive ground ice requires large volumes of water for its formation. Some argue that water expelled from freezing sediments (segregation) is not sufficient to explain large ice volumes in the Mackenzie delta area (Rampton, 1974), but Mackay (personal communication, 1989) indicates that segregated ice bodies 10-20 m thick by 1-2 km wide may be formed from *in situ* pore water. Alternatively, basal glacial meltwater, forced to and beyond the ice margin by ice overburden pressures, may have produced thick marginal ground ice bodies as it refroze

in marginal permafrost (Weertman, 1966; Rampton, 1974; fig. 58, 1988). Large tunnel valleys at Eskimo Lakes (Mackenzie delta area) indicate that massive outpourings of stored glacial meltwater cut through a frozen glacier fringe. In this way, post-surge meltwater floods (e.g. Kamb et al., 1985) inferred for southwestern Victoria Island (Sharpe, 1988a; Chapter 3) may have refrozen in marginal sediments of the hummocky moraine belt. Powerful meltwater flow may explain the coarse gravel and numerous sorted sediments within this hummocky moraine (Fig. 2.11). In addition, whereas the Mackenzie delta has high groundwater tables and large reserves, the raised moraines on Victoria Island have no such large porewater reservoirs and present-day precipitation is very low (~11 cm/a). Thus hummocky moraine may have developed due to burial of glacial ice or because thick ground ice formed as a result of meltwater refreezing either after normal drainage or after flooding.

2.2.3 Discussion

The abundance of ice-cored terrain on Victoria Island makes a thermokarst model for hummocky moraine formation attractive. But the exposed sediments at Ammalurtuq Lake (Fig. 2.11) are not those expected of widespread topographic inversion. Hoppe (1952) and Stalker (1960; 1973) suggested a model of hummocky moraine formation that emphasizes subglacial deposition. Some support for subglacial formation may be found in forms such as rim-ridges, but they may also be explained as ice-marginal forms. Evidence for ice press includes low diamicton ridges (Map 1650A) which form a criss-cross pattern similar to basal crevasse patterns (Hoppe, 1952). More common are large ice-stagnation features such as kettle and thermokarst lakes. Berendregt and Stalker

(1989a,b) suggested an ice-press model involving injection of sediment under glacial hydrostatic pressures somewhat analogous to those developed in pingos (Mackay, 1987). However, none of these models can be successfully applied to Wollaston Peninsula forms because of the remaining buried ice.

Topographic and glaciologic controls do, however, suggest that supraglacial drift was prominent on former glaciers covering Wollaston Peninsula. Wollaston plateau (Fig. 1.4) deflected glacial flow to form a drift-rich zone of marginal sediment. This is of importance to hummocky moraine models which depend upon different positions of deposition and the locations of zones of sediment concentration. With this in mind, I discuss first, sediment concentration on the upflow side of the Wollaston plateau and second the anomalous concentration on the downflow side of the plateau.

Thick glacial ice was present in the lowlands and major basins (Prince Albert Sound and Dolphin and Union Strait) adjacent to southwestern Victoria Island, while ice thinned at the margins over the central plateau (Fig. 1.4). This topographic arrangement suggests that thinning ice, rising flowlines, and thrusting occurred at the glacial margin. This apparently led to a large area, on the upflow side of the scarp, with intense compressional flow which moved basal debris to a supraglacial position. This topographic setting is very similar to the Missouri Coteau, a classic hummocky moraine landscape (Clayton, 1967). However, ice flow over and around the distal part of the plateau would have had no comparable compressive component (Fig 1.4). In addition, during downwasting, the deflected ice flow would tend to isolate slower-moving ice masses and induce stagnation over

the plateau (Fig. 1.4).

Marginal shear

Shear along the margins of surging ice may also concentrate debris, and cause thickening of debris-rich ice by compression as Kamb suggested for Variegated Glacier (Kamb et al., 1985). Surging and lateral compression along the plateau margins would have been followed by quiescence, thermokarst processes and redistribution of supraglacial debris (Clayton; 1967; Wright, 1980). Thus, supraglacial debris could have been carried to the downflow end of the plateau, but this explanation works only for thin ice; otherwise debris would have been carried over the plateau.

Thrusting models

Glacial thrusting at or close to the margin is another mechanism for concentrating sediment (Clayton and Moran, 1974; Moran et al., 1980). Such thrusting may be caused by the stress gradient produced by the rapid decrease in ice thickness near the margin (van der Wateren, 1985). Thrusting caused by marginal thinning may be enhanced by topographic obstructions, by a transition from a warm bed to a marginal permafrost zone, or by both of these situations. Thrust systems associated with hummocky moraine in Minnesota apparently resulted from excess subglacial meltwater flow within a subglacial aquifer (Mooers, 1990). This flow, in addition to hydraulic lifting, produced enough heat to make the frozen toe unstable and susceptible to thrusting over a 1-2 km zone. As discussed, freeze-on of meltwater may also have occurred.

Few good descriptions of the sediments from these settings are available to test models of drift accumulation. Recently, Lawson (1982, 1984) provided descriptions of sediment facies from Alaskan hummocky moraine (Fig. 2.15b).

The prominence of debris-flow sediments at Matanuska Glacier compared to high proportions of stratified sediment in permafrozen hummocky terrain on Victoria Island may reflect the very active melting and slumping in the modern, temperate, high-precipitation environment of Alaska.

Role of glacial meltwater

Glacial meltwater also influences marginal sedimentation (Gustavson and Boothroyd, 1987). The abundance of sorted and stratified beds in hummocky moraine on Victoria Island may also be linked with the concept of subglacial flooding (Dardis, 1983; Shaw, 1983; Shaw and Kvill, 1984) which is proposed to explain adjacent streamlined terrain (Sharpe, 1988a). Large amounts of sediment could have been transported to the moraine during such floods. Several metres of gravel, some with no matrix, and ice accretion may have resulted. Channels cut into the moraine, some with inset eskers, attest to subsequent meltwater flows that may have delivered sediment to the moraines, although the eskers are small and too infrequent to account for large amounts of sediment. In addition, meltwater floods may explain the lee side accumulation of drift as sedimentation in the low pressure (flow expansion) margin downflow of the Wollaston plateau (Fig. 1.4).

The high volumes of ice in the hummocky moraine explains how meltwater flows easily cut through hummocky moraine in late stages of deglaciation, but left small drumlin ridges intact (Figs. 1.14 (PB), 2.12 and Map 1650A). These inferred subglacial meltwater flows eroded more effectively through ice-rich sediment than through the drumlins that are considered to have been free of massive ground ice.

An interesting aspect of the proposed meltwater,

ground-ice and hummocky moraine association is that marginal migration and concentration of meltwater is also apparently a product of thrusting at glacier margins (Talbot and Brunn, 1989). However, marginal drainage (or lack of it), in concert with probable frozen glacier margins and the overriding of permafrost, probably enhanced marginal compression and thrusting (Clayton and Moran, 1974; Mooers, 1990). For example, large hydraulic gradients existed to produce the pattern of tunnel channels and eskers overlapping the Wollaston escarpment; a pattern indicating a connected drainage system within the Read Island lowlands (Fig. 2.1). Thin ice and low ice profiles are suggested by low gradients on lateral moraines (Sharpe, 1988a; 1991). A damming effect at the escarpment rise, coupled with a permafrost fringe, helped raise groundwater pressures in marginal subglacial flow, leading to induced thrusting.

In summary, topography, marginal thinning, thrusting, permafrost and meltwater discharge are potentially important factors in formation of hummocky moraine on southern Victoria Island. However, the strong possibility that extensive buried glacier ice or refrozen meltwater remains favours the use of the term ice-cored moraine for this hummocky terrain.

2.3 End, lateral and deformed moraine

The major moraine belt that encircles Wollaston Peninsula (e.g. Colville moraines; (Fig. 1.14; 2.1) contains a huge amount of drift and ice. Volume estimates are very high ($5 \times 10^7 \text{ m}^3/\text{km}^2$) compared to those for the ground moraine ($2-3 \times 10^6 \text{ m}^3/\text{km}^2$) which contains relatively little sediment and ice given its area. This mass of sediment is considered to represent active ice-marginal deposition in two environments: (1) *subaqueous*, where the glacier was in

contact with the sea; and (2) subaerial, where the ice was grounded above marine limit. These two environments produced distinct suites of sediments. In addition, deformed moraines are discussed relative to form and surface sediment.

2.31 End and lateral moraines

An end or lateral moraine complex (informally named the Colville moraine) comprises two or three sharply defined morainal ridges (Fig. 1.6b) and an adjacent outwash system comprising ice-marginal terraces and fans. This moraine belt is 25 km wide and individual ridges may be 3-5 km wide. Colville moraine has been traced for 200 km east-west from near sea level to 360+ m a.s.l. on central Wollaston Peninsula. This profile suggests a former minimum ice surface gradient of about 1.8 m/km or lower, possibly ~ 1.0 m/km, when uplift is taken into account. If the ice was 500 m thick, basal shear stress is estimated at about 25 kPa. This value is similar to those proposed for late glacial lobes extending into the sea (Shreve, 1985b; Clark and Josenhans, 1986). Low values (< 10 kPa) were proposed for areas underlain by "soft" impermeable sediments (Fisher et al., 1985; Beget, 1986). Colville moraine was probably constructed subaquatically in Prince Albert Sound and in Dolphin and Union Strait, west of a very large glaciomarine delta at 135 m a.s.l. (Fig. 2.1a), and subaerially for most of its eastward extent, despite this low profile. An ice shelf may have developed offshore (Hodgson and Vincent, 1984), but large shelves are subtended by polar glaciers (Meier and Post, 1987). Temperate glaciers are inferred for southern Victoria Island but evidence from thermokarst patterns means that frozen beds cannot be ruled out.

The eastern sector of the end moraine was above marine

limit and formed subaerially. It is folded and deformed into ridges that can be readily identified on LANDSAT images (Fig. 1.6b). The deformed moraines, which appear as thrust sediment in places, form a 10-12 km portion of the 25 km wide Colville moraine system. The elevation of these thrust or shear moraines indicates that ice shearing extended up to at least 225 m a.s.l. (north-south transect, Fig. 2.1a). An ice gradient as low as about 1 m/km may be deduced for the glacier during this intense shearing. Gradients of less than 2 m/km have been considered to characterize surging glaciers (Wright, 1973; Mathews, 1974; Clarke et al., 1984; Clayton et al., 1985) and ice streams (Benson, 1987). Kamb et al. (1985) observed intense lateral shearing and thrusting during a surge of Variegated Glacier, Alaska. Therefore, surging probably influenced moraine formation (Sharpe, 1988a).

Subaqueous ice-marginal deposition

Raised moraine segments (Map 1650A) found below marine limit were deposited subaqueously. Such sediments form the Colville moraine adjacent to and west of the large glaciomarine ice-marginal delta (Fig. 2.1a) on southern Wollaston Peninsula and along much of northern Wollaston Peninsula.

A 10-km-long segment of raised lateral moraine (50 m a.s.l. at the top), at Cape Back on the north side of Wollaston Peninsula, provides an exposed section (Fig. 2.16). Local marine limit is 120 m, so deposition occurred when ice was grounded in about 70 m of water. The section contains two massive diamictons and three diamicton sequences interbedded with gravel, silt and fine sand. All beds are horizontal and undeformed. Several of the beds have transitional contacts; for example the sediments (Fig. 3.14b) pass from massive diamicton (D) into stratified sediments (E)

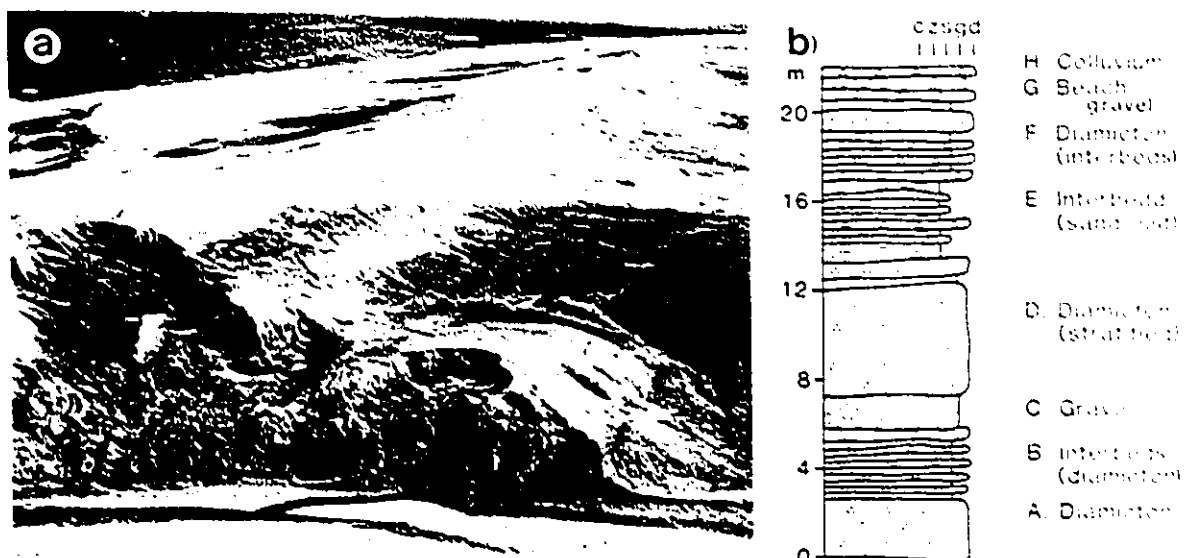


Figure 2.16 a. Lateral moraine on the north shore, Wollaston Peninsula at Cape Back; oblique aerial view of exposed sediments and Prince Albert Sound in background to northeast. GSC 204223-W. b. Section measured in lateral moraine at Cape Back, northern Wollaston Peninsula; moraine occurs at 50 m a.s.l. whereas marine limit is >100 m a.s.l. c. Thick massive diamicton with flow-nose structure adjacent to thin stratified beds in Cape Back lateral moraine. GSC 20422-T d. Graded sand bed (above knife) within morainal deposits of Cape Back moraine. GSC 204223-S



and back to a diamicton (F) interbedded with sand.

The style of deposition at Cape Back end moraine is exemplified by interbedded diamicton and sorted beds exhibiting characteristics of sediment flow deposits, particularly debris-flow sequences (Lowe, 1982). Large flow noses which appear to represent plug or very viscous flow at the margins of a large debris flow, are found downslope from the massive diamicton (Fig. 2.15c). Many of the sand beds in the section are graded (Fig. 2.15d). They are similar to graded beds deposited by turbidity underflows (McCabe et al., 1984; Powell, 1984; Dardis, 1985; Smith, 1985). The inferred tidewater setting and the sedimentary features indicate periodic deposition by turbidity currents and subaqueous debris flows from a glacial source. The landform is representative of a series of lateral or end moraines adjacent to Prince Albert Sound and another set of moraines adjacent to Coronation Gulf.

Subaqueous deposition is marked by interbedded gravity flow deposits. The distinctive facies for subaqueous moraines produced at retreating, grounded ice margins appear to be the product of meltwater sedimentation and gravity flow processes (e.g. Powell, 1984; 1990; Booth, 1986; Høltedahl, 1989). Lateral moraines produced in this style contrast with those formed by terrestrial glaciers in that they contain abundant stratified drift (Eronen and Vesajoki, 1988). Subaqueous end moraines may form as simple push moraines or by debris-flow deposition (Evenson et al., 1977); some consist of massive diamicton (Kruger, 1985; Booth, 1986). Subaqueous push moraine deposits may be deformed by shearing (Van der Meer, 1987).

Subaerial deposition

Evidence for a subaerial ice-marginal depositional

setting is provided by ice-contact glaciofluvial terraces (Fig. 2.17) along the shores of Wollaston Peninsula (Fig. 2.7). The outwash terraces are formed in coarse, cobbly, boulder gravel deposited as ice retreated eastwards into and along Prince Albert Sound (and Coronation Gulf). The length of the glaciofluvial system may be as much as 100 km, with predominant paleoflow directions towards the west and inland. Thick ice remaining offshore at a late stage, was the probable style of deglaciation in inter-island channelways. This conclusion is supported by late-stage esker development northward toward the Colville moraines.

Deformed moraine

The low glacial profiles described above suggest that ice occupying Dolphin and Union Strait may have been surging during moraine formation. Observations on the surging Variegated Glacier, Alaska, show that a major shear zone

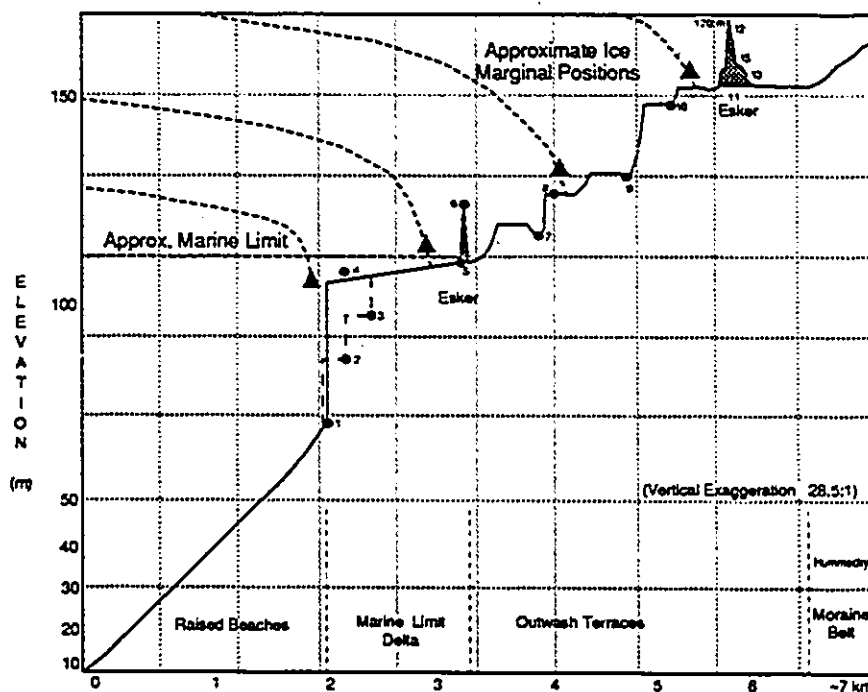


Figure 2.17. Cross-section of coast and approximate ice positions south side of Prince Albert Sound near Cape Bach. Dots are measured elevations.

developed along the glacier margin as ice flowed against a height of land (Kamb et al., 1985). Confined shear during surging apparently produced thick bodies of debris. Similar shearing may have been responsible for the deformed moraines of the Colville moraine system. The edge of the carbonate plateau (Fig. 1.4) would have acted as the confining margin. The deformed moraines are broad ridges, plate-like in places, with bouldery gravel and blocks (2-3 m in diameter) derived from the local carbonate bedrock. These blocks may have been detached by hydraulic jacking (Röthlisberger, 1981) or meltwater-induced thrust (Mooers, 1990). The moraines may include thrust sheets in some cases, although no large thrust blocks or planes (e.g. Van der Meer, 1987) were observed. The thrust and compressive flow models considered for hummocky moraine may also apply here (e.g. Clayton and Moran, 1974). A surge model of sedimentation following marginal thrusting (Sharp, 1985) suggests that individual ridges within a thrust complexes may be surface expressions of the slices of an imbricate thrust sheet (Fig. 2.18). The set of deformed ridges on southern Wollaston Peninsula may have been formed in this way by a single surge or several closely spaced surges (cf. Wright, 1980). These models also involve large blocks of ice bounded by thrust sediment that fit the evidence for ice-cored ridges cited above.

2.4 Relationship to streamlined forms

Streamlined features are the final landform type to be discussed in a landform series comprising ground moraine, hummocky moraine, end-lateral moraines, and streamlined forms. This association of landforms appears to have resulted from one major glacial advance and retreat cycle (Sharpe, 1984, 1988a). Compared to hummocky forms the

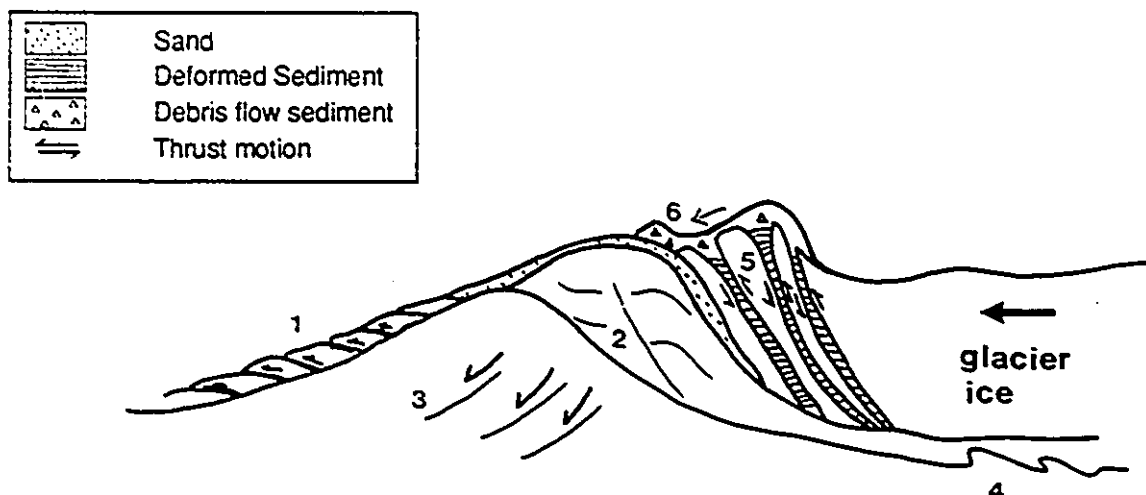


Figure 2.18. Ice-marginal thrust moraine formed by a glacier surge in Iceland (after Sharp, 1985). 1. Lobate complexes due to lateral spreading. 2. Thrust ridges. 3. Normal faults. 4. Water escape structures. 5. Englacial bands raised along thrusts. 6. Supraglacial debris ridges.

streamlined landform set (Fig. 2.1a) may have formed with less resistance to formative flows. The streamlined landforms show a gradation of forms: subtle flutings, to drumlinoid ridges and long drumlin ridges. This apparently reflects a change in flow strength.

The spectacular nature of these streamlined forms and the enigma of their origin require special attention and a separate section is devoted to the description of their pattern, form, sedimentary character, and possible origin.

2.5 Summary

Landform models and landform association may aid understanding of glacial landforms on southern Victoria Island. Zoned landforms indicate that one flow event produced landform transitions (Fig. 2.1). Landform transitions include ground moraine to hummocky moraine, which in turn is transitional to streamlined forms (Fig. 1.14). Victoria Island landform models reflect the influence of

topography and permafrost.

Ground moraine includes diamictons and interbeds of sorted sediment, rather than expected massive diamictons. Hummocky moraine contains complex deposits which may be related to a variety of factors: topography, marginal thinning and thrusting, permafrost, and meltwater migration or flooding. Whatever other factors are involved in hummocky moraine formation the escarpment topography of Wollaston Peninsula is a fundamental control as it blocked drainage along with a permafrost fringe, and induced shearing.

Hummocky moraine occurs in three main areas in North America: the central plains, Keewatin, and the western Canadian Arctic. Prairie forms are mainly subglacial in origin produced by ice press (Stalker, 1973) and marginal thrust (Moran et al., 1980). Some supraglacial debris occurs but thermokarst is minor. Keewatin forms are low-relief till hummocks likely formed as subglacial bedforms related to adjacent streamlined forms and Rogen moraine (Aylsworth and Shilts, 1989). In the western Arctic, hummocky terrain is largely ice-cored. Some hummocky terrain is simply ice-cored thermokarst derived from melt of extensive segregated ice bodies (Mackay, 1971; Rampton, 1988). Thick morainal areas, within late Wisconsinan limits (e.g. southern Victoria Island), are largely ice-cored because buried glacier ice and refrozen meltwater are insulated by thick sediment cover.

Lateral moraines are subaqueous (interbedded sediment flows) or subaerial (massive diamictons or thrust structures). Lateral shear, marginal thrust, and a frozen fringe aided the formation of deformed moraine ridges.

Thick ground ice, inferred from large polygons and ground-ice slumps controlled formation of hummocky moraine and lateral moraine.

3.0 STREAMLINED LANDFORMS

3.1 Introduction

A variety of streamlined landforms are particularly prominent on southern Victoria Island (Fyles, 1963). They include flow-parallel drumlins, drumlinoid ridges and flutings in sediment or rock. Understanding the origin of streamlined landforms, particularly drumlins, is an important step in deciphering the sequence and glaciological processes shaping the Pleistocene geology of Victoria Island. In this respect, landform associations are a key because most drumlin fields show strong links between streamlined forms, tunnel channels and eskers (Wright, 1973). The timing of these landforming processes also places constraints on models of formation and landscape evolution.

Southern Victoria Island drumlins are set in a complex of streamlined fields in surrounding areas (Fig. 3.1). More than a dozen cross-cutting relationships appear between streamlined landforms in the study area. If streamlined terrain is formed late in the glacial cycle (cf. Boulton, 1982), then the streamlined cross-cuts may represent the action of rapid subglacial processes under late-glacial ice cover (Fig. 3.1).

Streamlined landforms contain important records of their formative flows but no consensus exists on even such a basic issue as the relative importance of erosional or depositional mechanisms. This has implications for the relevance and the apparent validity of sedimentological approaches to drumlin origin because drumlin composition does not yield information on drumlin formation if they are erosional in origin. Erosional drumlins may be identified

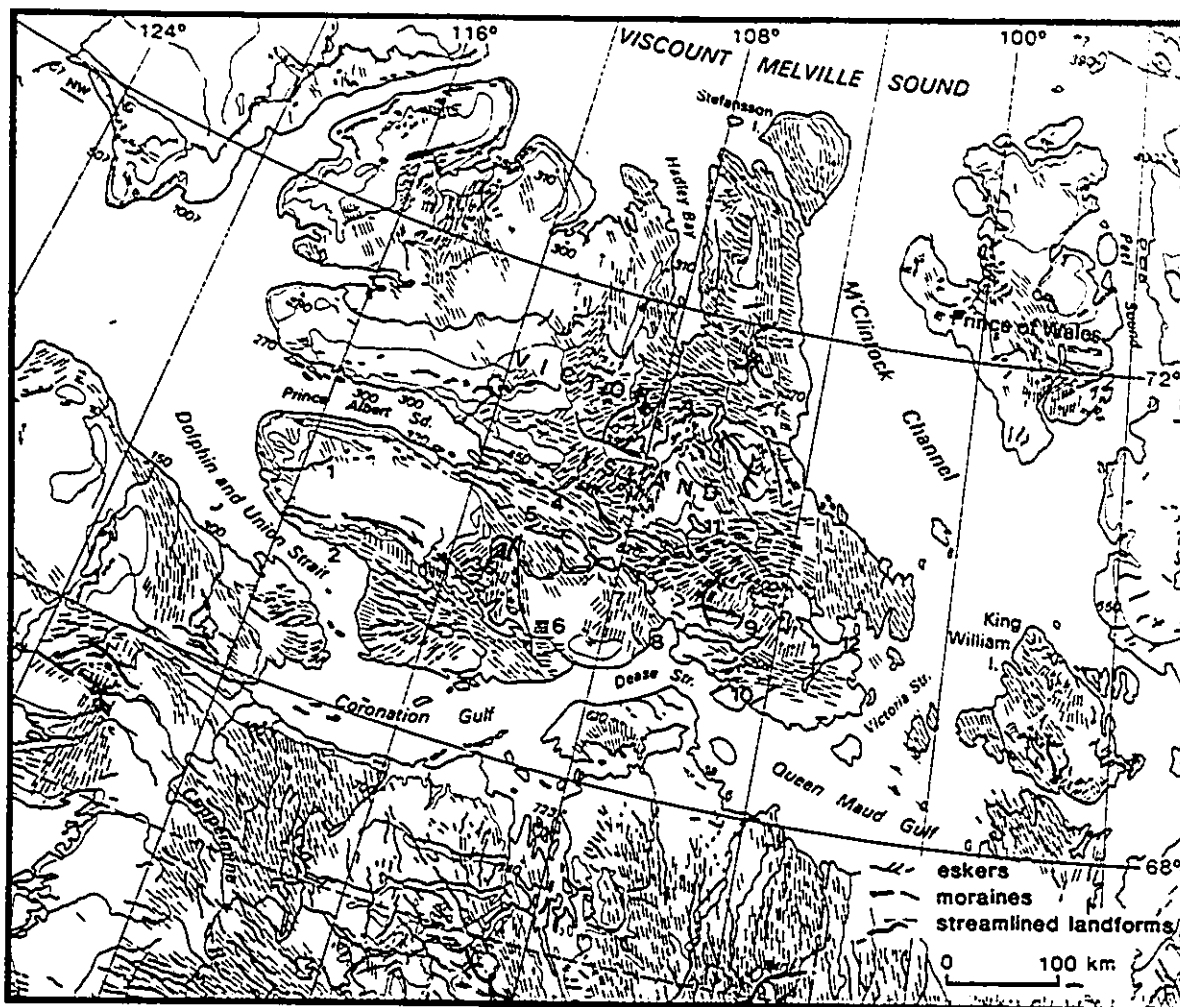


Figure 3.1. Sets of streamlined landforms on southern Victoria Island, showing cross-cut relationships numbered from oldest (1) to youngest (12) (from Prest et al., 1968).

using form, surface lags or displaced sediment as clues to their origin (Gravenor, 1953; Menzies, 1984; Shaw and Sharpe, 1987).

A depositional origin (particularly for diamicton-cored drumlins) has been favoured (Menzies, 1984; 1987). The most common view suggests that subglacial sediment has been moulded into drumlins (Smalley and Unwin, 1968). Such depositional models of drumlin formation also involve sediment accretion during the drumlin formation (Muller, 1974). Recently, links between rapid ice flow in ice streams and deformation in subglacial sediment (Alley et al., 1986; 1987) have been extended to drumlin formation (Boulton, 1986; Boulton and Hindmarsh, 1987; Boulton and Hindmarsh, 1989; Boyce and Eyles, 1991). A depositional origin, which suggests that sub-ice scours were filled with glaciofluvial sediment, (Shaw and Kvill, 1984; has been applied to some forms on Victoria Island (Sharpe, 1985; 1988a). Others proposed that lee-side cavities were filled with glaciofluvial sediment during drumlin formation (Dardis, 1983). Glaciofluvial origins have been dismissed on the basis of theoretical models (Boulton, 1987), ignored by others (Dyke and Morris, 1988; Aylsworth and Shilts, 1989a,b) but in some cases, sedimentological data favour fill by glaciofluvial and sediment flow processes (Dardis, 1987) rather than pervasive deformation (McCabe and Dardis, 1989 a,b; Clayton et al., 1989).

It follows that the hypothesis of depositional drumlins can more readily be tested if their sediments can be observed and evaluated. Erosional drumlins by contrast require an understanding of erosional processes or information on eroded sediment redeposited beyond the immediate area of study. Regional landform relationships (landform associations)

assume increased importance in the explanation of drumlins.

3.2 Setting, pattern and form

Major drumlin fields lie in low areas surrounding Wollaston Peninsula and in smaller areas of hummocky terrain across southern Victoria Island. Some fluted forms extend into higher terrain where they merge with large lateral moraines along southern Wollaston Peninsula (Fig. 2.1). The streamlined landforms rest on flat to gently stepped carbonate lowlands. Bare rock is semi-continuous and appears to define scour zones particularly on the up-flow margin of drumlin fields (Figs. 3.2 and 3.3a). The bedrock carries few striations, partly because of post-glacial weathering. Particularly common are erosional forms such as flutes, scallops, and crescentic forms, collectively termed s-forms (Kor et al., 1991). Exposed bedrock generally carries a lag of large boulders.

Several prominent drumlin fields occur on southern Victoria Island (Fig. 3.2). The most prominent field, Read Island, is significant because of its size, spectacular divergent drumlin array, and its relationship to adjacent landforms, particularly hummocky moraine (Chapter 2.0) and an associated array of eskers (Chapter 4).

This lowland drumlin field, 50 km wide where confined in the east, expands to more than 100 km wide in the west, and is 150 km long (Fig. 3.2). Streamlined forms are larger in the central and lowest parts of the field, and their size and definition decrease outwards. This coherent pattern is suggestive of formation by a single, coherent flow. For example, elongate drumlins, shorter, oval drumlins, and a field of flutings (1, 2 and 3 of Fig. 2.1a and b) occur in sequence from the centre to the perimeter. These features

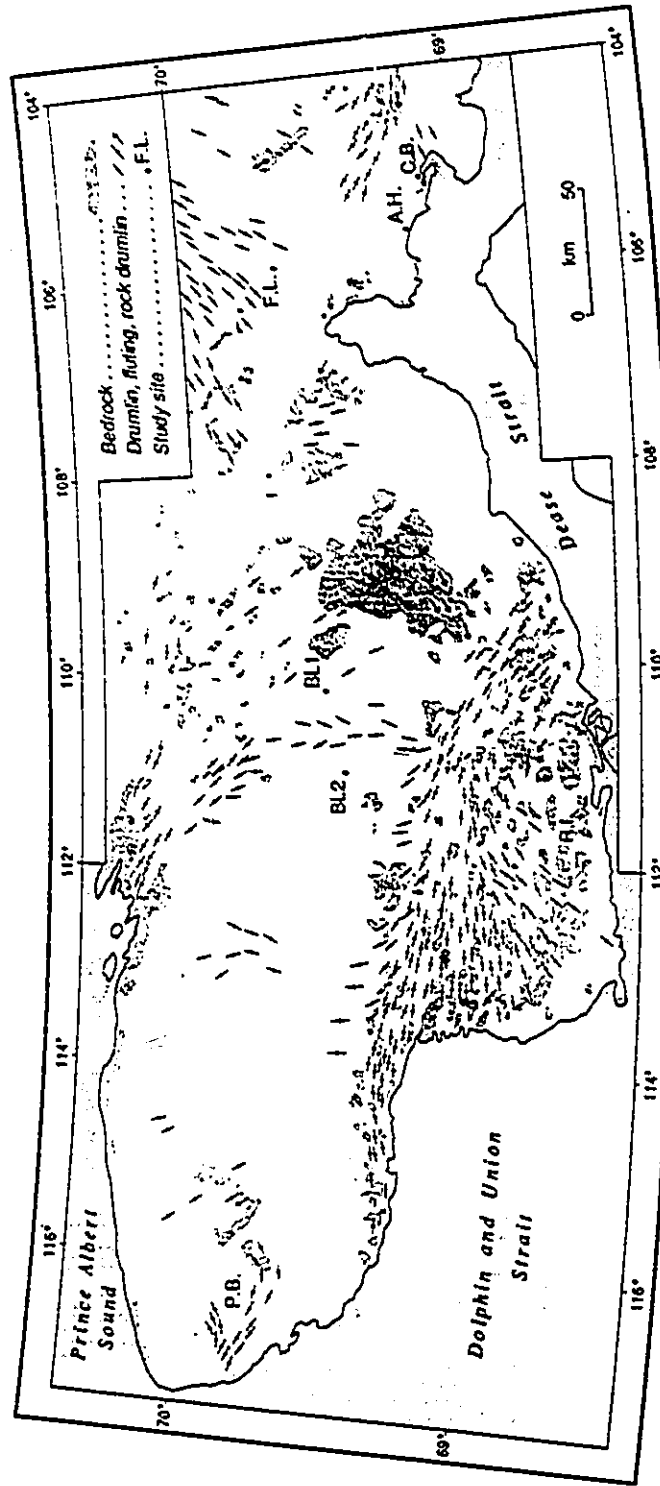


Figure 3.2. Map of streamlined land forms and exposed bedrock. Three drumlin fields (Read Island (R.I.), Penny Bay (P.B.) and Banning Lake (B.L.) studied on Victoria Island.

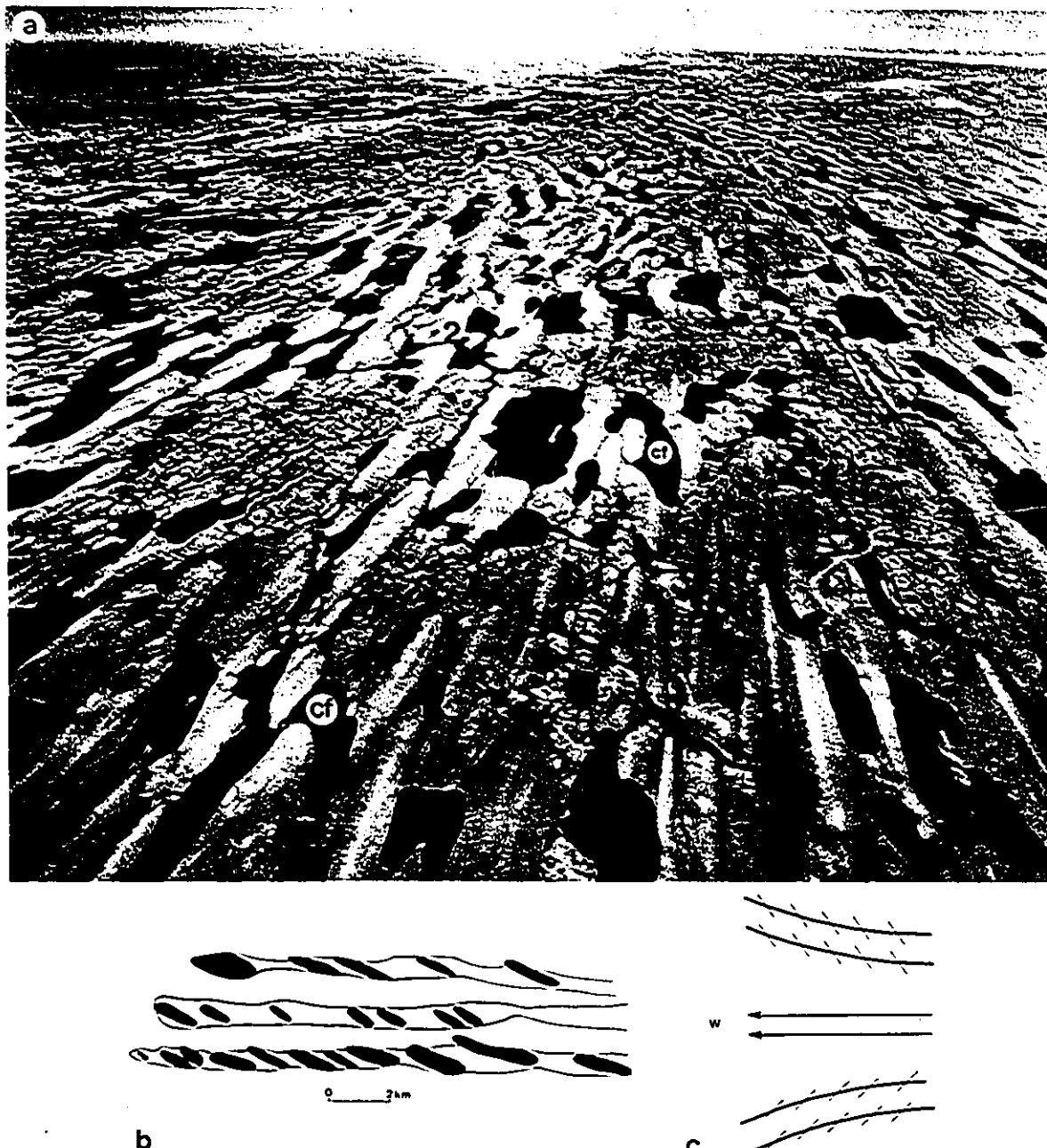


Figure 3.3. a. Aerial view of drumlin field, south coast of Wollaston Peninsula (Fig. 4.2) shows transition from elongate forms (1) to curved composite forms (2) and drumlinoid forms (3). Note that J-shaped form (J) and herringbone pattern occur only within elongate pattern near 2. Comma form (lakes) are marked CF. Flow towards viewer. See Figure 1.6A for location. NAPL T32-L-188 b. Cluster group of drumlins showing oblique orientation. c. Echelon, herringbone flow pattern cross-cuts and shows opposite direction on north and south margins of flow.

are discernible on conventional aerial photographs (Fig. 3.3a), as well as on LANDSAT images (Fig. 1.6b). The margin of the field is defined by a large moraine met by subtle flutings, cut through by channels and draped with a few eskers (Fig. 1.14).

3.2.1 Elongate and short forms

In plan view, elongate drumlins of the Read Island field are mainly symmetrical spindle forms (Fig. 3.3a); less common are parabolic and transverse asymmetrical forms (Allen, 1982, p. 185; Shaw and Kvill, 1984). These drumlins are in a field 15-20 km wide and more than 50 km long. Individual forms are about 3 km long, 100 m wide, and 10-30 m high. Some drumlin clusters are arranged diagonal to flow (Fig. 3.3). The diagonal patterns occur in offset or en echelon arrays (parabolic or comma shaped) aligned oblique to or parallel to the spindles. In places, the diagonal elements impart a herringbone pattern to the field (Fig. 3.3a). This pattern consists of a series of diagonal drumlins, in sequence, that results in longer linear elements (Fig. 3.3b). These herringbone patterns occur as dominant elements of other large drumlin fields on Victoria Island (e.g. north of Prince Albert Sound, Fig 3.1).

Short, oval drumlins occur marginal to elongate drumlins. They are mainly isolated spindle or parabolic forms generally about 100 m long, tens of metres wide, and 10-15 m high. Clusters are less common than for elongate drumlins. They are more closely spaced than elongate drumlins; groups of drumlins form a divergent pattern, rather than being oriented parallel with elongate drumlins (Fig. 3.3a). Long, en echelon clusters are parallel to the elongate forms in the larger drumlin field. The transition zone from areas with drumlins to those with no drumlins

contains poorly formed, small, oval drumlinoid features. This transition seems to be related to sediment availability (enhanced erosion); the adjacent landform suite, glacial fluting, occurs on terrain with very little drift.

3.2.2 Flutings

A prominent, 10-20 km wide band of fluted, thinly drift-covered terrain, occurs between the drumlin fields and the large lateral moraines. These flutings, which form curved bands from 1 to 5 km long, are prominent on aerial photographs yet imperceptible on the ground. Long flutings occur close to long drumlins. This fluted terrain represents a transitional zone of erosion (or minor deposition) set between large drumlin fields and ice-marginal deposition (moraines) upslope. If the drumlins are erosional, then the flutings represent a zone of enhanced erosion.

3.2.3 Sinuous forms

In several areas (e.g. Banning Lake and Ferguson Lake) sinuous drumlin ridges are contiguous and interspersed with straight-crested ridges. Ridges vary from small features, 5-10 m high and 0.5-1 km long, in a narrow (2-3 km) zone at Banning Lake (see Fig. 3.9), to 10-15 m high and 1-3 km long, in a broad (10-25 km) zone at Ferguson Lake (see Fig. 4.10). The overall effect is to give a braided appearance to the drumlin field.

3.2.4 Interpretation of pattern and form

Systematic patterns, arrangements of forms, and different forms across the drumlin fields require explanation. The gradual transition across the Read Island field suggests that varying velocities operated within the

flow. Such transitions occur in velocity gradients measured in valley glaciers and but they are not noted in ice streams (Benson, 1987). These transitions also invite comparison with other flow systems such as the thalwegs of rivers where bedforms may show transitions from large forms to smaller forms at the flow margins (e.g. dunes to ripples). The pattern of offset or *en echelon* forms in drumlins is common also to sedimentary beds displaying groups of bedforms at two scales (Allen, 1982, fig. 5-4). For example, Dzulynski and Walton (1965) found similar *en echelon* forms, and erosion marks, on turbidite beds. The herringbone drumlin patterns are analogous to erosional flow marks produced experimentally in flumes (Allen, 1969; 1971) by single flow events.

Differences in individual forms, zones of similar forms, and transition between zones within fields, indicate a coherence to the formative flow. Individual forms (spindles, parabolic and asymmetric) have close analogues with erosional and depositional features of turbulent flows (Allen, 1982; Shaw and Kvill, 1984). Streamlined forms are also explained by turbulent flow separation (Komar, 1983). Sinuous forms show a braided pattern that may also indicate turbulent flow conditions.

The pattern and form of Read Island drumlins are best related to single complex flows. Proposed glaciofluvial origins for drumlins draw analogies between drumlins and turbulent flow bedforms (cf. Shaw, 1983; Shaw and Sharpe, 1987; Shaw et al., 1989). To some however, these arguments do not exclude the direct or additional action of glacial ice (Dardis, 1985, 1987; Dardis et al., 1984; McCabe and Dardis, 1989a,b). Boulton (1987) and Boulton and Clark (1990) argue, in an alternative explanation of the herringbone relationship within drumlins on southern Victoria Island, that subglacial

sediment was moulded by two ice-flows with different flow directions. This explanation is unlikely because the crossing patterns occur near the margins of the field but not in its central part (Fig. 3.3c). The oblique elements also show opposite sense of rotation on opposite sides of the field, which implies three flow events in the Boulton and Clark explanation; this leaves a prohibitively narrow central flow zone and poses difficulties regarding flow continuity (Fig. 3.3c; Map 1650A). The pattern may be best explained as secondary flows, perhaps related to paired vortices (Allen, 1971; 1982).

Defining flow width is a major problem in the study of fields of streamlined forms. There are few ridges near the margins of the streamlined fields on Victoria Island, although marginal areas might be expected to have been zones of maximum sedimentation. Read Island drumlins are bordered by Colville moraines, yet flutings (and later-formed eskers) meet the moraines orthogonally. Where there is no topographic control, the eastern end of one drumlin field is clearly bordered by only a narrow ridge (Read Island; B, Fig. 3.1). Surface sediments of these bordering moraines are of coarse gravel. These ridges have been termed moraine traces (Fyles, 1963) on Victoria Island, shear moraines on adjacent islands (Fig. 3.1 C, Dyke, 1987), or strike-slip features (Attig and Clayton, 1990).

The extensive areas that are sediment-poor with s-forms and scattered boulder lags appear to indicate sediment stripping.

3.3 Sediments in streamlined landforms

Emphasis in this section is placed on drumlins because fluted forms are discussed in landform associations with

Ferguson Lake eskers (Chapter 4) and in another report (Sharpe, 1985). Landform studies, including those on drumlins at Read Island, Banning Lake and Penny Bay (Fig. 3.2), indicate formation beneath Late Wisconsinan ice (Sharpe, 1984; Borowiecki, 1984). This ice apparently melted in place following formation of these landforms because no ice-marginal deposits (e.g. morainal ridges or outwash systems) are found in the drumlin fields (Fig. 1.14; Map 1650A). The only glacial deposits associated with the drumlins are small eskers that wind between the larger drumlins within broad valleys or tunnel valleys (Sharpe, 1988a). Bare bedrock is common, particularly at the up-flow end of the field near Dease Strait (Fig. 3.2).

3.3.1 Read Island

The Read Island conclusions are based on a detailed examination of two drumlins (Fig. 3.4a): a large drumlin with exposures in the flank and one end, and a second, smaller drumlin with a well-exposed cross-section. In addition, fabrics were measured in two trenches and two excavated pits.

Drumlin section 1 (site 1165, Map 1650A)

Three major facies are noted in a 5 m exposure in a drumlin flank (Fig. 3.4b). The lower two are horizontal and represent intact, primary deposition. Sandy facies A comprises beds of normally graded sand 10-15 cm thick, massive medium sand, pebbly sand, and thin beds of diamicton (Fig. 3.4b). Facies B consists of two beds of diamicton, 10-15 cm and 60-70 cm thick, interbedded with well-sorted, very fine sand. Facies B is gradational from facies A in that the frequency and thickness of diamicton beds increase upwards, whereas the sand units become less frequent and are thinner.

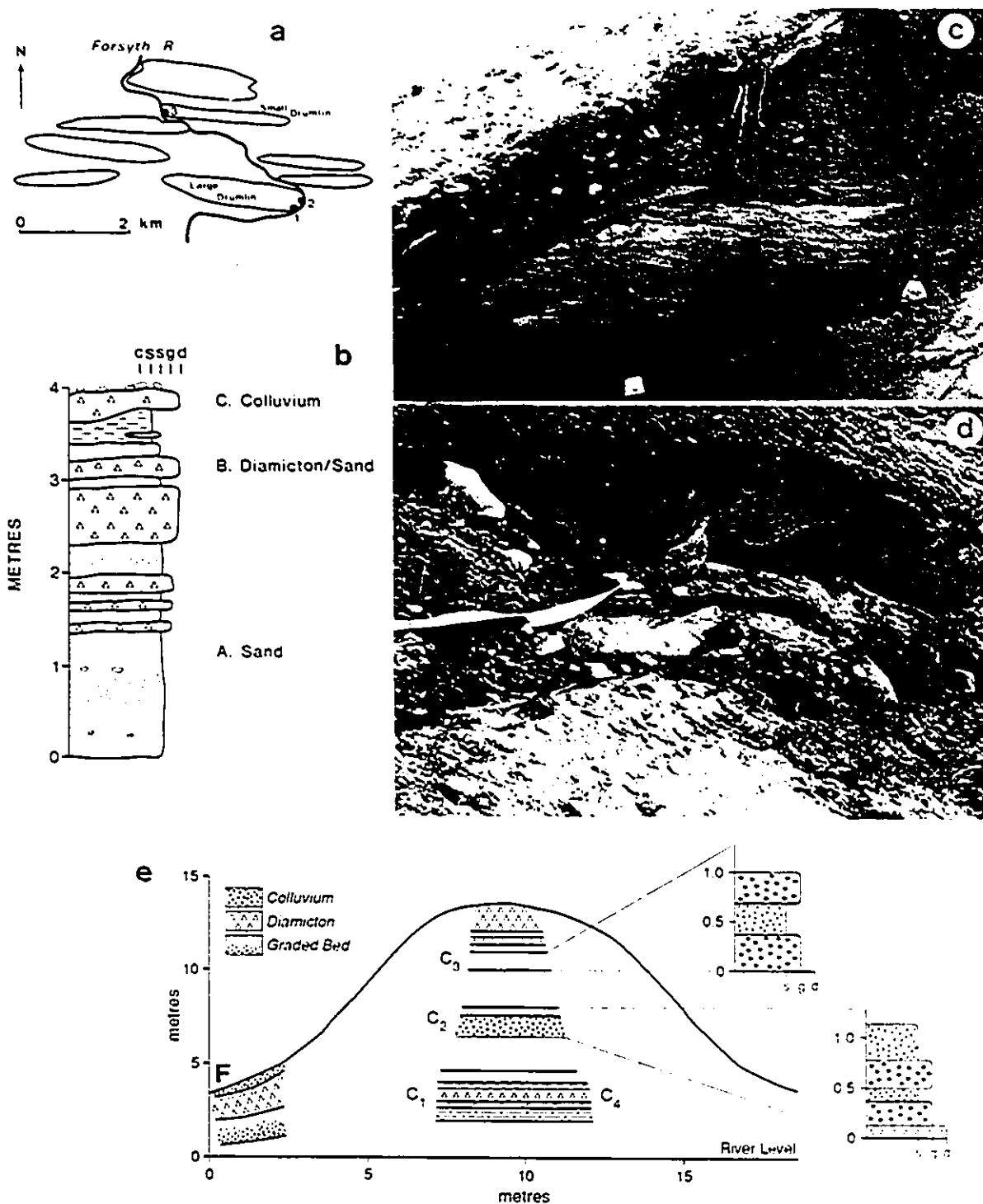


Figure 3.4. Read Island drumlin field. **a.** Location, sections 1-3. **b.** Section in side of drumlin. **c.** Sediments (a,b,c) in side of drumlin, (section 1). GSC 204439-D **d.** Sediments (a,b), section 2. GSC 204255-X **e.** Cross-section in small drumlin (section 3).

Facies C has a sharp unconformable contact with facies B (Fig. 3.4c) and consists of massive clayey silt and a lag of surface cobbles. Facies C, which dips with the slope of the drumlin and crosscuts the bedding in facies B (Fig. 3.4 B,C), comprises reworked slope material. The clayey silt contains marine shells and forms a colluvial unit. This relationship between facies C and B also clearly shows that facies B and A are *in situ* and have experienced little or no frost deformation.

Crossbeds in unit B give westerly paleocurrent directions, approximately parallel to the drumlin long axis, and in the expected direction for subglacial flow (Fig. 3.4c).

Drumlin section 2 (site 1168, Map 1650A)

An exposure in the up-flow or east end of the drumlin Section 2; Fig. 3.4a) reveals two primary facies (Fig. 3.4d). A lower apparently massive, pebbly medium sand with pebbly horizons includes fining-upward sequences of gravelly sand to sandy diamicton (a). Striated clasts in the diamicton beds indicate little secondary transport after glacial abrasion. A second facies (Facies b) consists of massive beds of silty fine sand with occasional coarse clasts. The contact between these two facies is conformable in the upper trench where the beds are parallel to one another. The contact is sharp in a lower trench where facies B shows horizontal beds. An upper facies, silty diamicton with marine shells, cross-cuts lower beds (Fig. 3.4d,C). Diamicton in Facies B has weak preferred fabrics sub-parallel to drumlin axis (Fig. 3.5). Because they were located 6-8 m below the surface of the drumlin these sediments are distinct from surficial diamictons produced by mass wasting (Fig. 3.4c).

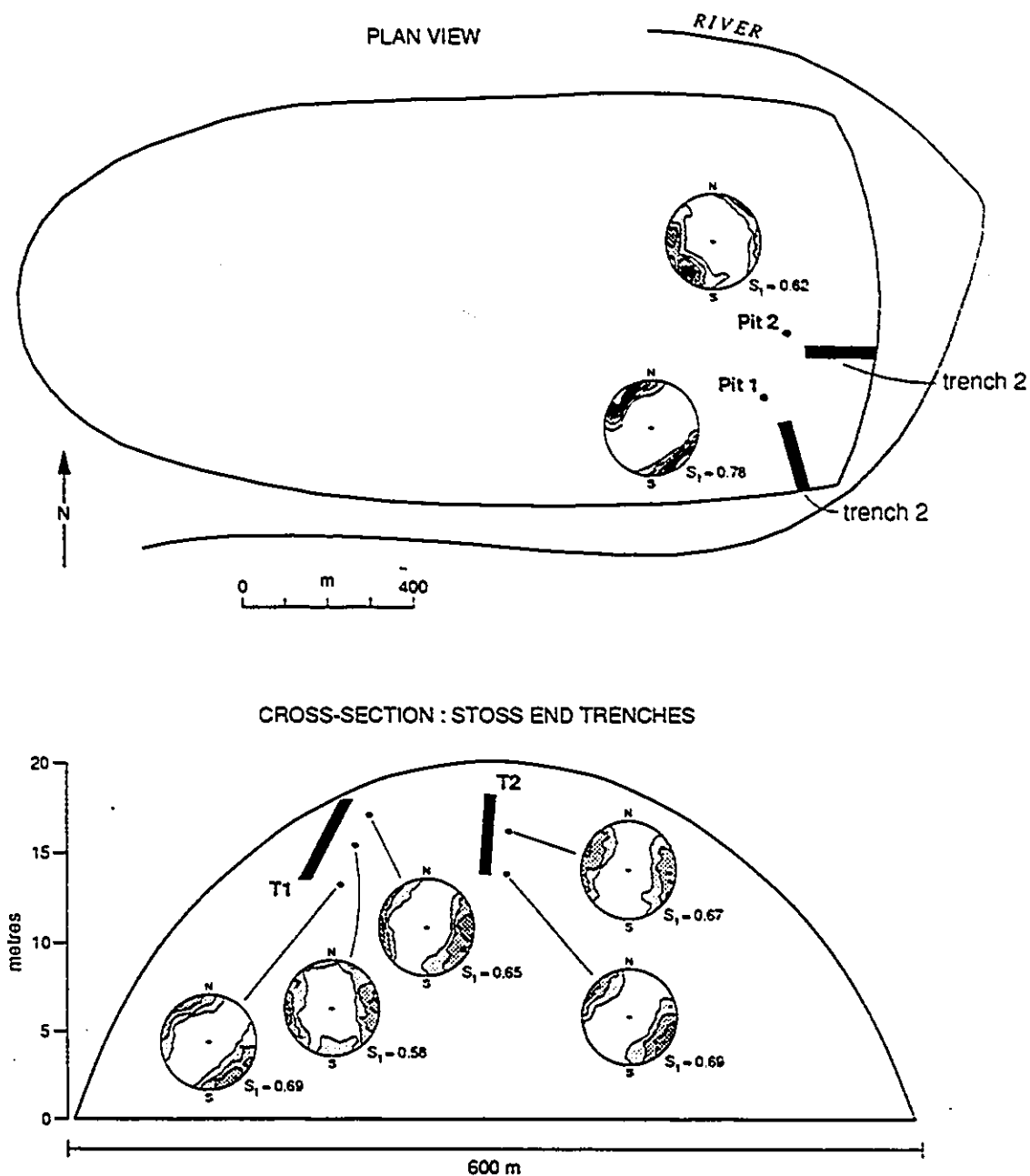


Figure 3.5. Fabric of drumlin sediments in side and end trenches and surface pits (1 and 2). From Borowiecki, 1984.

Pits dug on a level drumlin surface revealed 70 cm of massive (to finely laminated) sandy silty diamicton showing well-developed fissility. Clasts have a weak to strong fabric (Fig. 3.5, pit 1,2) oriented southeast-northwest and north-east southwest, that is subparallel compared to the westerly orientation of the drumlin field. The northwest-southwest orientation corresponds to the pattern of low carbonate dispersal from Richardson Islands (Fig. 2.2a). Weak fabrics from level sites may be frost disturbed but probably are not soliflucted. The contact with underlying stratified sediment was not observed in the pits. About 50 cm of interbedded, fossiliferous, marine gravel, pebbles and clay at the surface represent washed diamicton or till material. One large boulder at the surface showed striations parallel to the drumlin axis (Fig. 3.6).

Small Drumlin

A small drumlin, with well-exposed sediment where a river has cut a 10 m high section through its distal end, provides further sedimentary evidence on drumlin genesis (Fig. 3.4e).

Flank sediment

Here, a graded unit (1.5 m thick) changes from well-sorted, fine gravel at the base of the exposure, to a coarse pebbly sand near the top. Overlying this, with a sharp contact, is 1 m of massive, fine sandy silt diamicton containing rip-up clasts. The graded unit and the diamicton are located on the flank of the drumlin (F, Fig. 3.4e) and may represent later sedimentation.

Core Sediments

The lower 1.5 m of sediment in the centre of the drumlin (C₁) consists mainly of bedded sands (alternating 10-15 cm beds of massive very fine sand and massive fine to

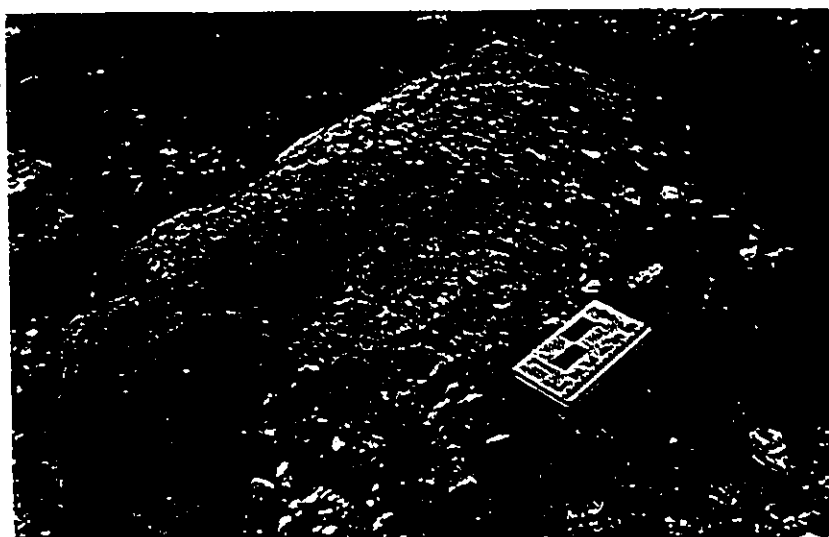


Figure 3.6. Large boulder on surface of Read Island drumlin with striation parallel to drumlin long axis. GSC 205193-B.

medium sand; C_1 Fig. 3.4e). A few reddish silty sand diamicton units occur in this sequence. In places, there are poorly defined sets of ripple drift cross-lamination (1-5 cm sets) indicating westward paleoflow.

A 1 m high covered interval separates the lower (C_1) and intermediate (C_2) measured sections (Fig. 3.4e). Section C_2 consists of several normally graded sequences of gravelly sand to well-sorted, medium fine sand, about 50 cm thick. Above another covered 1 m interval, additional graded units appear (C_3).

The section is capped by a loose, massive diamicton consisting of stones, sand, and silt, 0.5-1.5 m thick. Unfortunately, the exact nature of the contact between the

top diamicton and the underlying gravel grading to medium sand was not observed. However, marine shell fragments found in the loose surface diamicton indicates littoral reworking.

This small drumlin contains mainly horizontal beds of sorted sediment and glaciogenic diamicton. A colluvial unit covers the side slope and an underlying diamicton unit may also be colluvium.

3.3.2 Interpretation

The planar geometry of the stratified beds in the drumlin cores shows an absence of major deformation of the type expected in deformation models for drumlins (Smalley and Unwin, 1968; Boulton, 1982; 1988). The following points are conclusive on this issue:

1. Stratified deposits (some with cross-laminations) are interbedded with diamicton units.
2. Deformation is lacking in any of the stratified units including thin clay beds.
3. Graded beds, typical of sediment gravity flows, are common.

Stratified deposits and bedding appear to be horizontal and truncated by the side slope deposits. This contradicts a model of a fill conformable with drumlin form though cavity fill may have occurred.

The oriented surface boulders with striations parallel to drumlin axis and the subparallel fabric in surface diamictons indicate late but only minor surface modification of the drumlin by direct glacial shear. Of note are the clast orientations showing a northwest trend which parallels the low carbonate dispersal pattern (Fig. 2.2). This

suggests that Read Island drift was emplaced during northwest flow, and later eroded by westerly flow.

3.3.3 Discussion of Read Island sediments

Earlier workers have described sedimentary structures of drumlin sediments (e.g. Slater, 1929; Muller, 1974). They show evidence of contemporaneous and conformable deposition of bedded sand, gravel and diamicton (Menzies, 1984). Undeformed stratified beds were widely observed (e.g. Upham, 1897; Jewtuchowicz, 1956; Jaatinen, 1960; Lamporski, 1972). Sharpe (1985) made similar observations on southeastern Victoria Island.

Only a few studies include detailed sedimentological studies of sorted and stratified beds within drumlins (Dardis and McCabe, 1983; Shaw, 1983; Shaw and Kvill, 1984; Sharpe, 1987; Hanvey, 1987). DeJong et al. (1982) report regular alternation of sedimentary layers (including graded units and cross beds) and an absence of deformation structures within drumlins, but they fail to assess the significance of these beds to drumlin formation. Dardis and McCabe (1983) explains sand-cores in Irish drumlins by deposition in water-filled subglacial cavities (e.g. by cohesive sediment gravity flows, grain flows, turbidity currents and fluvial deltaic sedimentation). They conclude, however, that drumlin shape resulted from ice erosion without deformation. But, deformation of drumlin sediments has been reported elsewhere (Whitecar and Michelson, 1979).

The Read Island sediment logs show a close relationship between stratified sorted sediment and thin, massive diamicton. This indicates the same environment of deposition for both sediments. These facies are compatible with deposition in subglacial cavities by sediment gravity flow

(Dardis and McCabe, 1983). Dardis et al. (1984) favour a lee-side position of deposition. Shaw (1983 and 1985) also suggested cavity-fill mechanisms to explain the origin of some drumlins; the cavities in this case were eroded by meltwater and filled by glaciofluvial and debris flow processes.

The conventional hypothesis is that undeformed stratified drumlin cores (with interbedded diamicton) represent pre-existing proglacial sediment that was reshaped and eroded by ice (Gravenor, 1953; Smalley and Unwin, 1968; Menzies, 1979; Kruger and Thomsen, 1984; Boulton, 1987; Kruger, 1987; Boyce and Eyles, 1991). However, strong evidence in support of this view has not been established. Rather, sedimentary evidence supports the alternative view that stratified cores may represent subglacial sedimentation (Shaw, 1983; Dardis and McCabe, 1983; Shaw and Kvill, 1984; Dardis et al., 1984, and Shaw, 1985).

The following evidence places constraints on the interpretation of stratified drumlin cores:

1. Drumlins must be accretional if undisturbed sediment is conformable with drumlin shape (e.g. Sharpe, 1987; Shaw et al., 1989).
2. A depositional origin is favoured if the drumlin deposits show a systematic arrangement of facies and paleocurrents relating directly to drumlin form (no such complete observations have been made).
3. Drumlin sediment may not give direct evidence on drumlin formation if the landforms are erosional.

If drumlins are erosional forms, then such contemporary landforms as the Colville moraines and hummocky moraine which

formed downflow may record evidence of the formative flow. Fluting patterns (and later-formed eskers) run directly into the moraine belt (Fig. 2.1), possibly indicating a common genesis. In addition, hummocky moraine comprises a high percentage of stratified drift which is not easily accounted for by models of formation based on compressive flow and supraglacial sedimentation (Lawson, 1985).

The evidence described from Read Island is too scanty to clearly support a depositional or an erosional explanation of drumlin formation. However, the interbedding of sandy units and diamicton, and the undeformed, finely preserved bedding directly contradicts hypotheses requiring wholesale subglacial deformation or shearing. Meltwater deposition alternating with debris-flow sediment released from the ice explain the observed sedimentary sequences within drumlins and may have been important in drumlin formation on Read Island. Small-scale cross-bedding structures indicate flow to the southwest, parallel to drumlin orientation and the sense of regional ice movement to the southwest. Diamicton beds are interpreted as products of flows of debris melting out subglacially. Late ice action may be indicated by striations on surface boulders. The combination of these features favours sediment deposition in a subglacial environment. The low number of observed sites leaves open the possibility of lee-side meltwater and debris-flow deposition (Dardis, et al., 1984), although one site is on the stoss end of a drumlin (site 2, Fig. 3.4a). Dispersal patterns and fabric data trend to the northwest, suggesting that the westerly oriented drumlins may be erosional in origin. With this in mind, erosion by subglacial meltwater (Shaw and Sharpe, 1987; Kor et al., 1991) may account for the undeformed sediments.

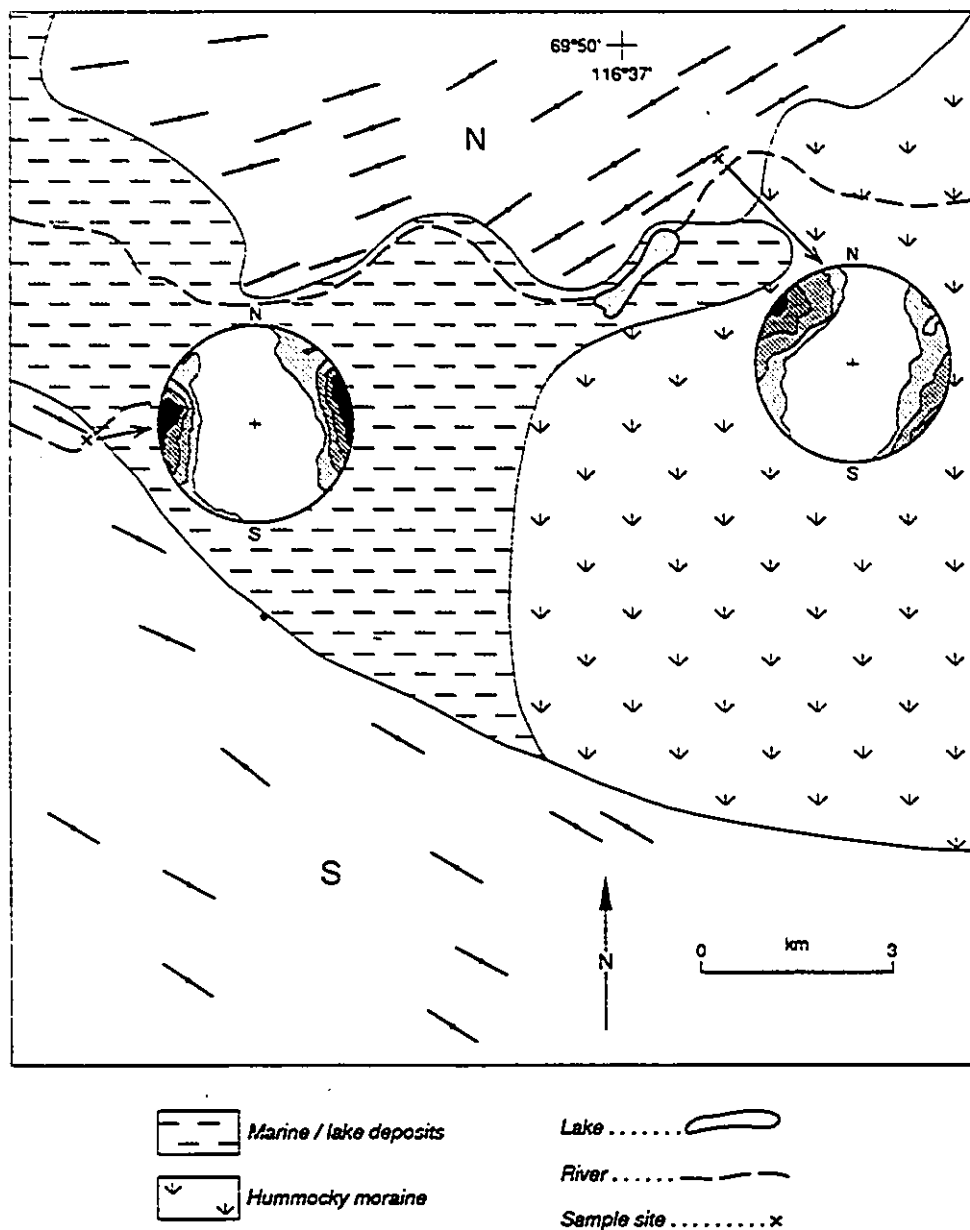


Figure 3.7. Till fabric plots from Penny Bay. N field has transverse fabric; S field has sub-parallel fabric. Map shows location (x) of fabric sites on field (N), NE-SW orientation and main field (S), oriented NW-SE.

3.3.4 Penny Bay drumlin field

Drumlins studied at Penny Bay comprise two separate drumlin fields that merge and cut through a belt of hummocky moraine (Map 1650A; Fig. 3.7). The drumlin fields consist of: 1) divergent forms (10-20 m high; ~1 km long) sweeping through hummocky moraine from the northeast to the west in a prominent field, 5-7 km wide; and, 2) subdued forms (5 m high; 1-2 km long) in a field ~10 km wide (Fig. 3.7). The second field, oriented northwest-southeast, also cuts across hummocky moraine from central Wollaston Peninsula (Fig. 3.8). Drumlins at the margin of this hummocky moraine are elongate forms and they are not lateral moraines (Fig. 3.8, s).

Both sets of drumlins provide opportunities to study internal composition although sections are only a few metres deep. The sediment encountered was compact massive diamicton, moderately stony (cobbles and pebbles) with abundant sandy-silt matrix. Clasts are strongly faceted and striated. There are occasional sand stringers in the diamicton and minor sand pockets occur under cobbles and pebbles.

Fabrics

Many authors studying glaciogenic sediment and interpreting glacial environments have successfully employed fabric analysis of diamictons in combination with morphological, sedimentological, and other directional data (cf. Shaw, 1977; Lawson, 1979; Boulton, 1982; Stanford and Michelson, 1985).

A 1-3 m high section at a stream cut in a well-formed drumlin in field N (Fig. 3.7), exposes massive, moderately stony, sandy silt diamicton with a crumbly structure below 15 cm. A high percentage of the clasts are striated, polished and faceted. Several large (15-25 cm long) clasts are



Figure 3.8. Aerial photograph of streamlined terrain at Boffa Lake (B.L.), the upper end of the Penny Bay (P.B.) drumlin field. NAPL Trimet T480L-16. Note comma form (cf) and streamlined form (s) on the margin of the field where it cuts through hummocky moraine (HM). View downflow, westward to Amundsen Gulf.

faceted and striated "bullet-shaped" forms (e.g. Boulton, 1982). The orientation of these large, abraded clasts is subparallel (230-275°) to transverse (340°) to the orientation of the drumlins (220-260°). Many fabrics show preferred orientation; some are also oriented transverse to the axis of streamlined landforms in field N (Fig. 3.7). Fabrics from drumlins in field S are subparallel to the drumlin axes, and best developed in trench exposures. Those in surface pits were weaker and thought to be reoriented by ice (Borowiecki, 1984), but their variable azimuth more likely relates to surface processes.

The preferred fabric, sorting, striated and bullet-shaped clasts indicate direct glacial transport and deposition (i.e. till), rather than debris-flow activity, as inferred for Read Island drumlin deposits. The diamictons are either lodgement or melt-out; the "bullet-shaped" clasts suggest lodgement. Due to its massive, very dense character, paucity of sand stringers or partings, and the presence of oriented (parallel to landform), polished, striated, "bullet" clasts, the till is considered to be lodgement but deformation till may not be eliminated (Alley, 1991). Deposition of drumlin sediment was by ice flowing west or northwest and not southwest. Fabrics in adjacent mainland drumlins are also oblique to drumlin axes (Potschin, 1989).

Erosional origin

On evidence from the regional landform patterns drumlins at Penny Bay may be erosional. The central Wollaston Plateau consists of low ground moraine surrounded by high hummocky ice-cored terrain (Figs. 1.14, 3.2; Map 1650A). The Penny Bay drumlin fields cut through 10 m to 50 m of hummocky moraine to the west and to the south of the study sites; this limits the size of the ice sheet that may

be reconstructed to explain this cross cutting relationship (Fig. 3.2). A small ice cap situated on Wollaston Peninsula, which subsequently flowed out onto the lowlands of western Wollaston Peninsula, would be required to explain drumlin formation by ice erosion. Small ice caps would likely have been cold-based and, thus, are unlikely to have eroded thick ice-rich sediment (Echelmeyer, 1987). Ice streaming would also have been limited by the small reservoirs of ice within the stagnating ice producing hummocky moraine (Fig. 3.2) and the limited thickness (about 10 m) of deforming substrate (Alley, 1991).

A larger, downwasting ice sheet, covering most of Wollaston Peninsula, may have trapped subglacial meltwater on the interior plateau. If this was released during breaching of the hummocky moraine, streamlining such as that along the Kugaluk River (Fig. 3.8) would result. The expected evidence of meltwater release further up flow is, in fact, noted along the upper Kugaluk River (Fig. 2.4b) where s-forms, rock drumlins, and transverse troughs and rises occur (cf. Kor et al., 1991). The hummocky moraine probably acted as a marginal zone of cold ice which helped to trap and store subglacial meltwater prior to release. Thus, an erosional origin for the Penny Bay drumlins may explain both the discrepancy between landform and clast orientation and the fact that drumlins cut through stagnating, hummocky moraine. Irrespective of the agent of formation, the drumlins are clearly younger than the hummocky moraine and, thus, were formed late in the glacial episode.

3.3.5 Banning Lake

Banning Lake drumlins form a prominent field, set on low coastal plains and converging towards the head of Prince

Albert Sound (Fig. 3.2). The field displays a narrow central zone (2-3 km wide) of short, sinuous forms (10-20 m high and 200-300 m long) which cut through (at $\sim 310^\circ$) a broad zone (10-20 km) of long spindle forms (15-30 m high; 1-3 km long) oriented at 340° . Large crescentic depressions mark parts of the broader field (Fig. 3.9). Rock is exposed between drumlins but not as extensively as in other fields (e.g. eastern Read Island).

A series of washed and measured sections (cut in a long spindle form) and pits in long and short forms at the border of the two fields provide data on the Banning Lake field (Fig. 3.10). The drumlin slopes show signs of minor mass movement and marine washing has affected the surface.

Drumlin sediments, about 5 m thick, consist of compact, massive diamicton with subrounded and subangular carbonate clasts. Striated and faceted stones are common within the diamicton. Some striated and faceted stones are oriented parallel to the long axes of the drumlins (Fig. 3.11). Striae on oriented clasts also parallel the long axis of the drumlin. Fabric studies indicate strong preferred clast orientations parallel to drumlin long axes at several levels (Fig. 3.10a). Longitudinal exposures and surface pits show variable and dispersed orientation; some were parallel to large and small drumlin axes and some showed slope or surface effects (Fig. 3.10b,c).

Orientation data, massive diamicton, faceting and striae on axis-parallel clasts favour subglacial deposition by lodgement (but do not rule out deformation) for diamicton in both the spindle forms (broad field) and the sinuous forms (narrow field).

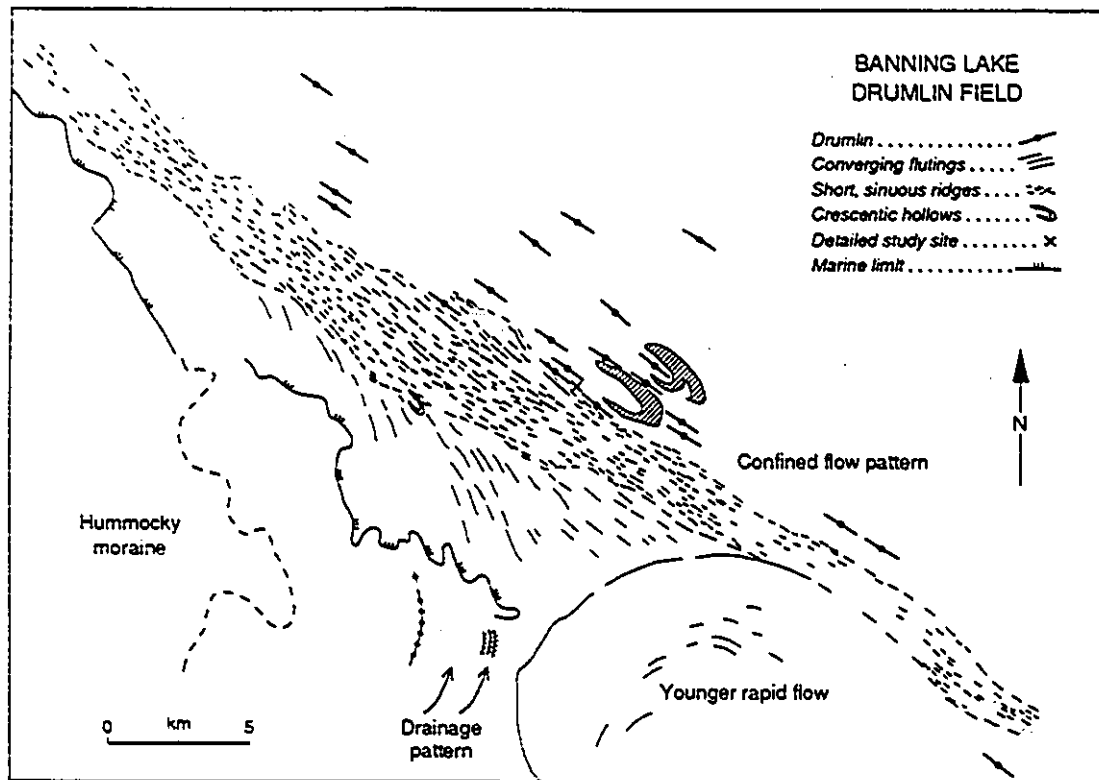


Figure 3.9. Banning Lake study site. Note long spindle drumlin and adjacent small sinuous drumlin ridges at the study site and the crescentic lake depression to the north.

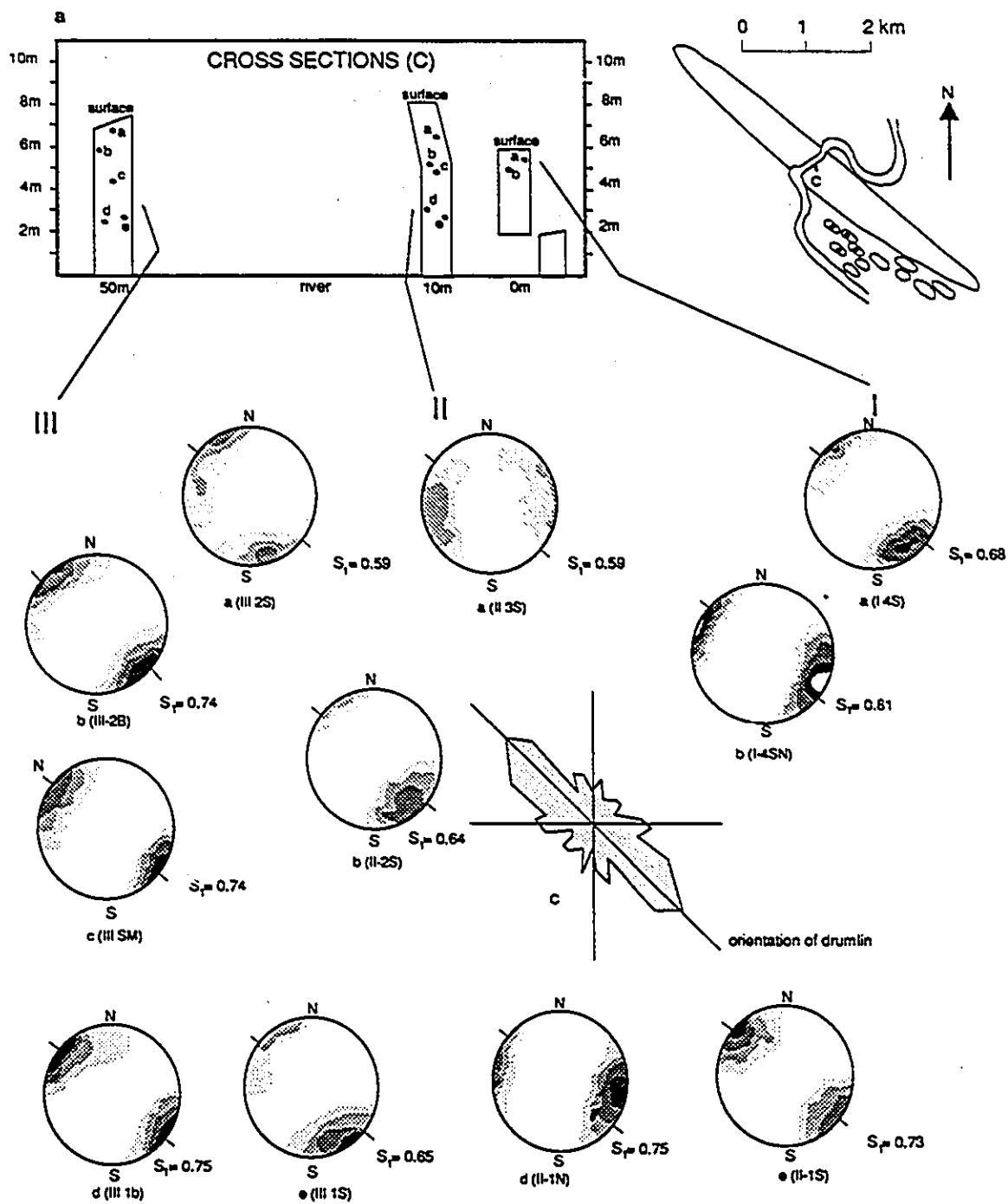


Figure 3.10 Fabric plots at Banning Lake.
a) trenches in cross-section.

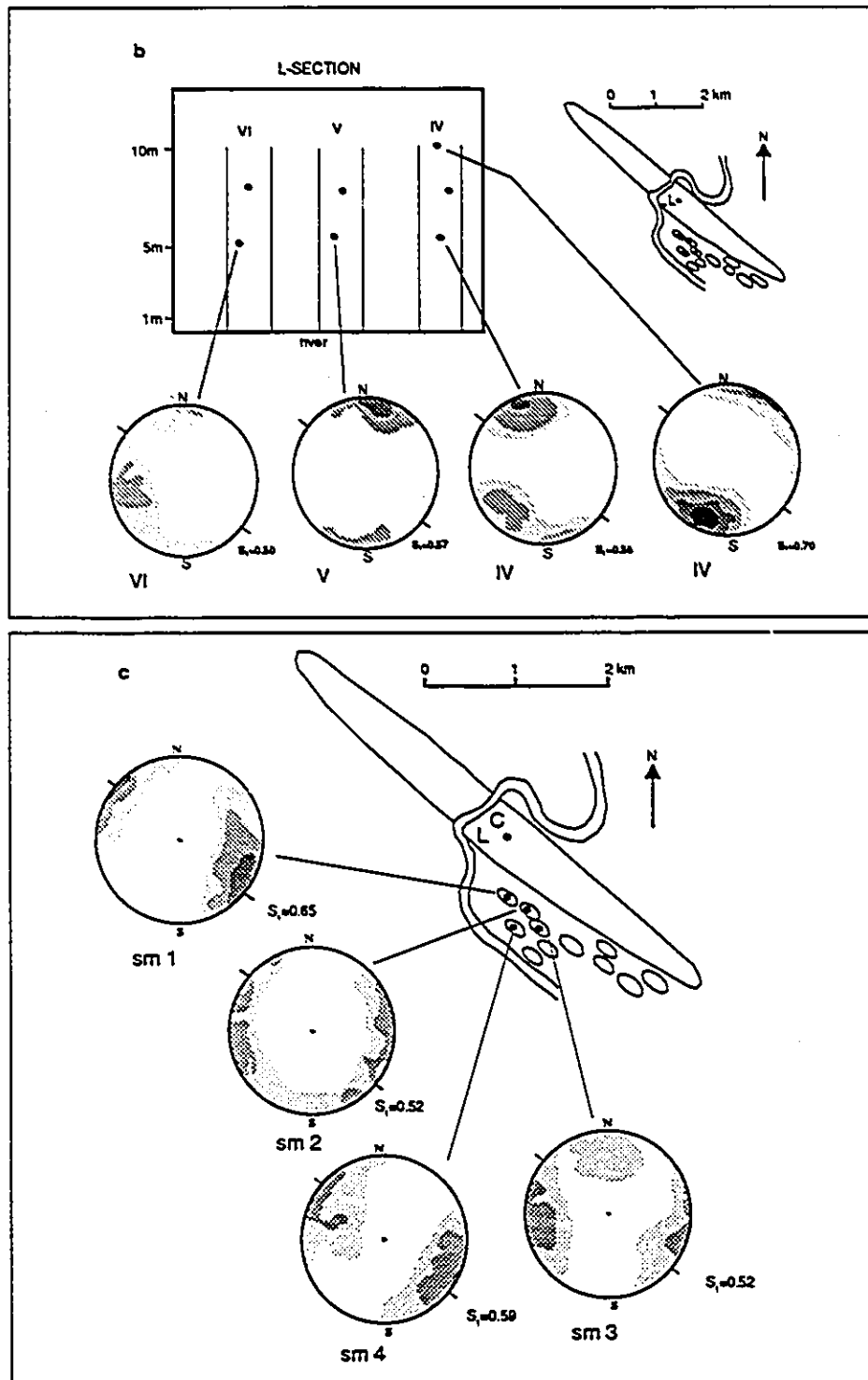


Figure 3.10 Cond't b) trenches in longitudinal section
c) surface pits on small drumlins. Drumlin axis marked with tick.



Figure 3.11. Elongate, faceted and striated stones which parallel the orientation of drumlin long axis at Banning Lake. GSC 1991-169A.

The distinct central drumlin zone with its sinuous flow pattern suggests secondary modification of the drumlin field with a variable but restricted flow. It appears to correspond to the thalweg of a broader flow which was apparently funnelled from east to west (Fig. 3.9). Confined subglacial fluvial erosion may apply to this relatively narrow zone because ice streams of this width are unknown (Bentley, 1987). There is evidence that the confined flows evolved from broader flows as flutings converge from the south (Fig. 3.9). This converging flow contains crescentic hollows indicative of subglacial meltwater flow (Shaw and Sharpe, 1987). Subglacial sheetflow is also suggested to explain the broader flow based on the large crescentic erosion patterns (Fig. 3.9) (Shaw and Sharpe, 1987; Kor et al., 1991). In summary, aspects of drumlin form and arrangement do not accord with observed properties of drumlin sediments and they may represent different processes; for

example, whereas the sediment may reflect deposition by active ice, the landforms reflect later subglaciofluvial erosion.

3.4 Erosional features in streamlined terrain

Erosional features within streamlined terrain appear significant to the genesis of drumlin fields. They include large sediment-free areas, tunnel valleys, transverse ridges, rocks drumlins, and large-scale and small-scale erosional forms on exposed rock surfaces. These features are best displayed at the upflow end of the Read Island field, south-central Victoria Island (Figs. 3.2 and 3.12).

3.4.1 Forms and features

Large areas of bedrock in streamlined terrain lie in linear zones 1-10 km wide and several km long (S, Fig. 3.13 a,b). Crescentic erosional zones occur directly upflow of large, remnant sediment ridges (T, Fig. 3.13b) and upflow-facing rock scarps (Figs. 3.12 and 3.14) defined by scoured depressions, many of which are lake-filled. Scour zones may surround broad (3 km), streamlined remnant ridges (T, Fig. 3.13a,b), smaller (up to 1 km) spindle-shaped ridges of drift (d, Fig. 3.14) and broad-nosed (5 km) rock ridges or rock drumlins (Fig. 3.14). Scour zones may also display small (200 m-300 m wide) rock drumlins with superimposed s-forms (Fig. 3.15). Scour forms occur at various scales (cm to km).

Broad sediment ridges (T) contain hummocky (ice-cored?) moraine, with large (100 m) polygons (Fig. 3.13 a,b), standing 10-20 m above the scoured terrain. These remnant ridges show a vague transverse pattern on their surfaces. Broad-nosed rock ridges carry smaller (0.5 km wide)

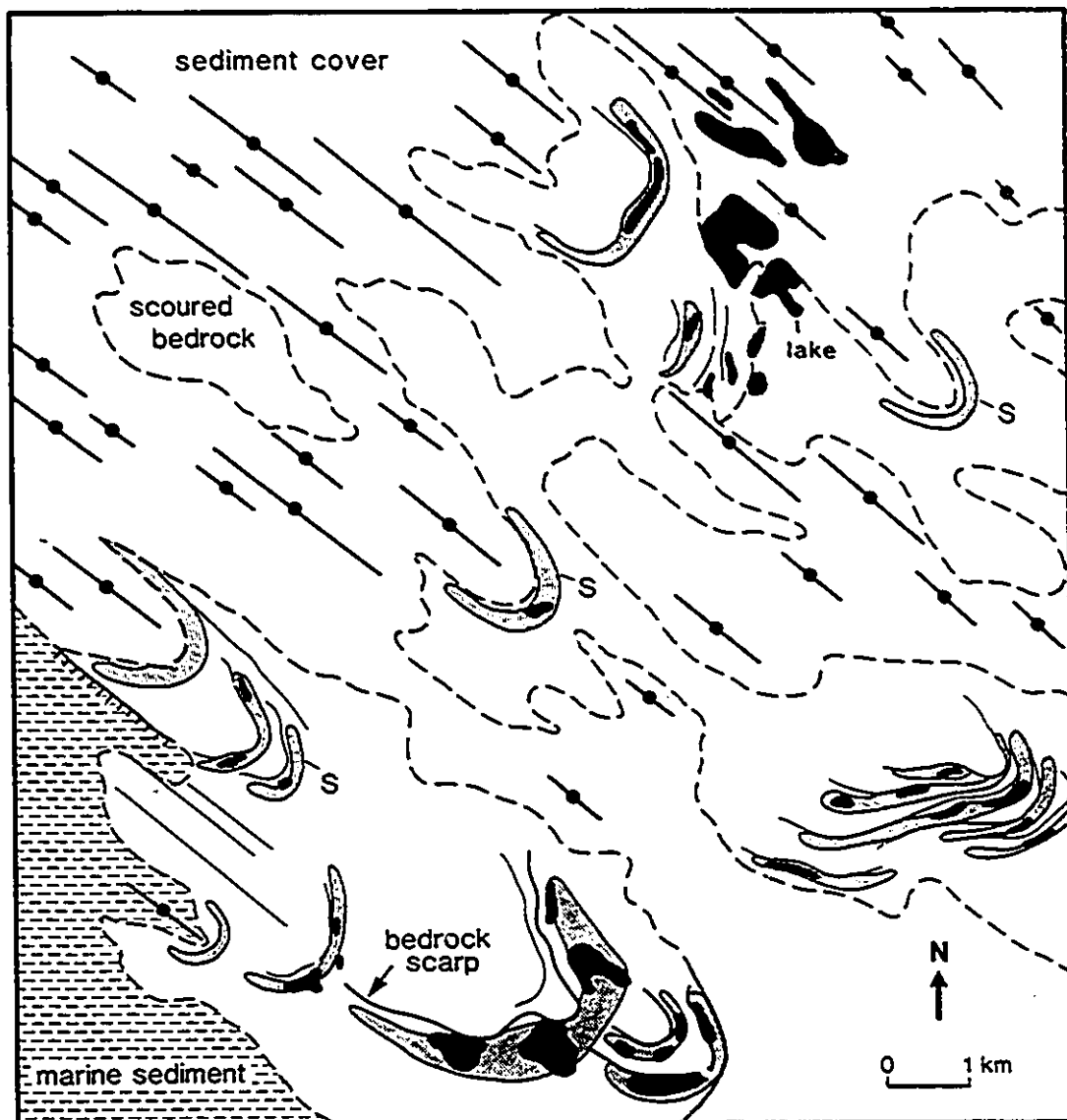


Figure 3.12. Drumlins and areas of erosion south of Banning Lake drawn from NAPL photograph A15704-22. Note crescentic scours (lakes) and bedrock escarpment noses, with sediment drumlins as residuals. Paleoflow was to northwest.



Figure 3.13. Aerial photographs of sediment-free bedrock areas.
a) Scoured bedrock areas (S) surround a streamlined zone of thick drift cover (T). Flow from bottom to top (1) is cut by a later flow (2) separated by a marginal ridge (M). NAPL A14900-41.



Figure 3.13 b) Scoured bedrock (S) reveals joint patterns (j), rock drumlins (rd) and eroded lakes (e) on the upflow position (arrows) of a remnant ridge (rr) of thick drift (T). This is a streamlined ridge of drift with hummocky forms (hm) and large polygons (p) indicating ice-cored terrain and a stagnant state of the resulting overlying ice. NAPL A14900-43. Flow bottom to top (arrow).

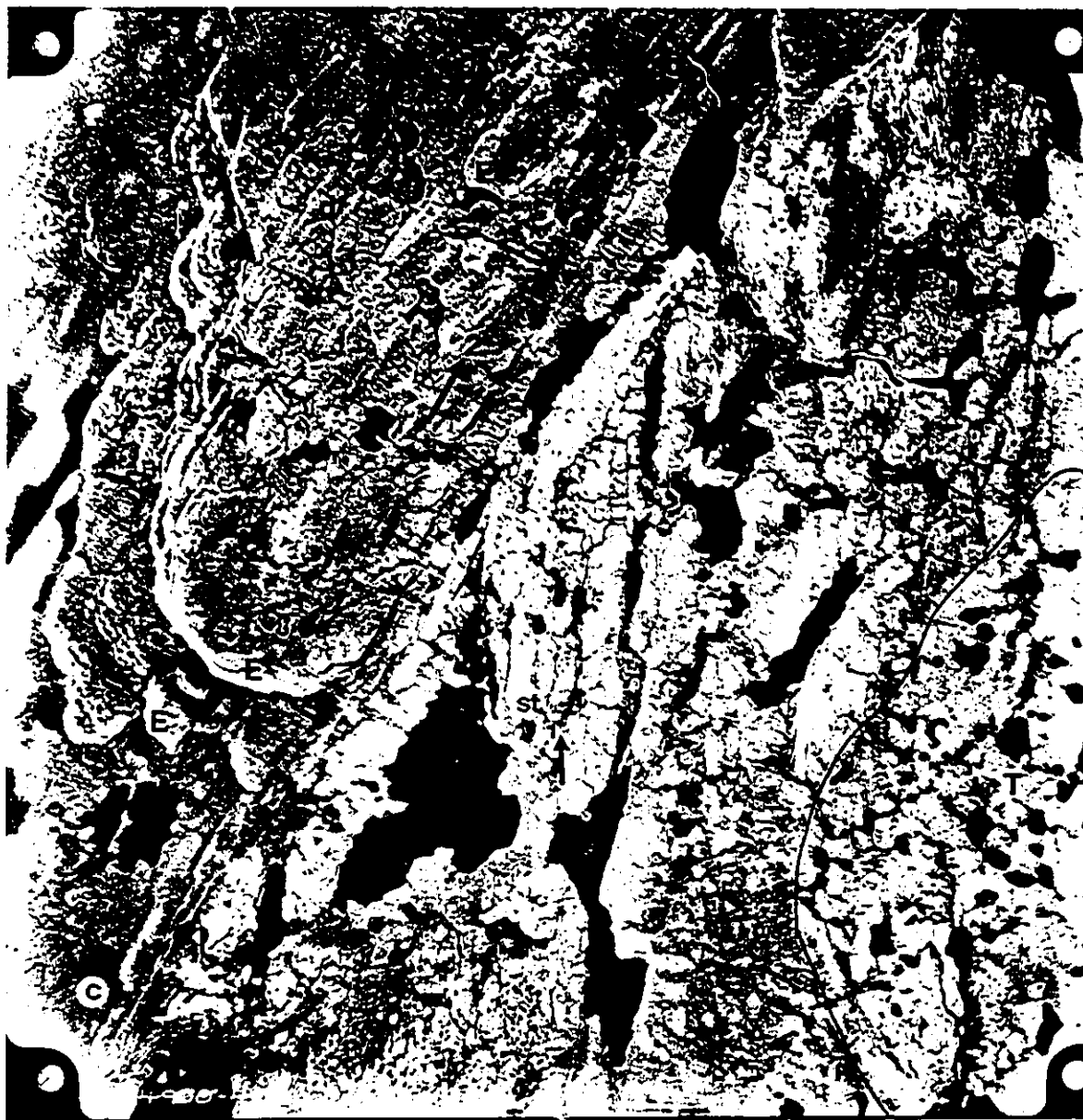


Figure 3.14. Escarpment nose (E), consisting of gently-dipping carbonate rocks, facing upflow. Note drumlins (d) on lee side of large escarpment forms. NAPL A14900-40. Note older flow scoured terrain (st), 1 cross-cut by flow 2. Crescentic scour(s) indicating paleoflow to the northwest. T is thick drift area.

streamlined forms on their surfaces (Fig. 3.15) and linear furrows (1-3 m wide and 100-200 m long; Fig. 3.15d).

These bedrock areas carry few striations, partly because rock weathering has destroyed the glaciated surface, though there are many erosional forms such as flutes, troughs, and crescentic scours (cf. Kor et al., 1991), collectively termed, s-forms (Fig. 3.15). These forms occur at scales ranging from centimetres to hundreds of metres; comma forms (eroded lake basins) are up to several kilometres long (Fig. 3.3, CF). Some s-forms are superimposed on rock drumlins. Bare rock areas indicate semi-continuous scour zones, possibly tunnel valleys, with small eskers (Fig. 3.2). The best example of such tunnel valleys is northeast of Cambridge Bay where a 2 km wide scour zone cuts across a drumlin field at a low angle. Rock drumlins occur within the tunnel valley scour zone (see Fig. 4.4), as they are in the Read Island field (Fig. 3.15e). Scour zones may contain lags of scattered boulders 15 cm to 1 m or more in diameter.

These scour zones and erosional forms are common in the upflow end of the Read Island drumlin field where a topographic rise met confined flow from Coronation Gulf and Dease Strait (Fig. 3.1). Erosion forms cover a 30 km width at Dease Strait, where gravelly ridges define the flow margin (Fig. 3.1).

3.42 Interpretation

Large scours

Large areas of exposed bedrock in areas with abundant drift represent zones of selective erosion rather than areas of non-deposition. Crescentic forms indicate enhanced erosion of the upflow side of topographic rises or obstacles to flow (Allen, 1982). This form, and the presence of a lag of coarse clasts on the eroded surface, indicate the action



Figure 3.15. Sculpted forms on carbonate bedrock, Wollaston Peninsula (arrow gives flow direction): a. small p-form (sichelwannen, highlighted); GSC 1991-167



Figure 3.15 b. small rat-tails (Prest, 1983). GSC 204535-L



Figure 3.15 c. Rock drumlin with superimposed sichelwannen. Flow left to right. GSC 204763-Y

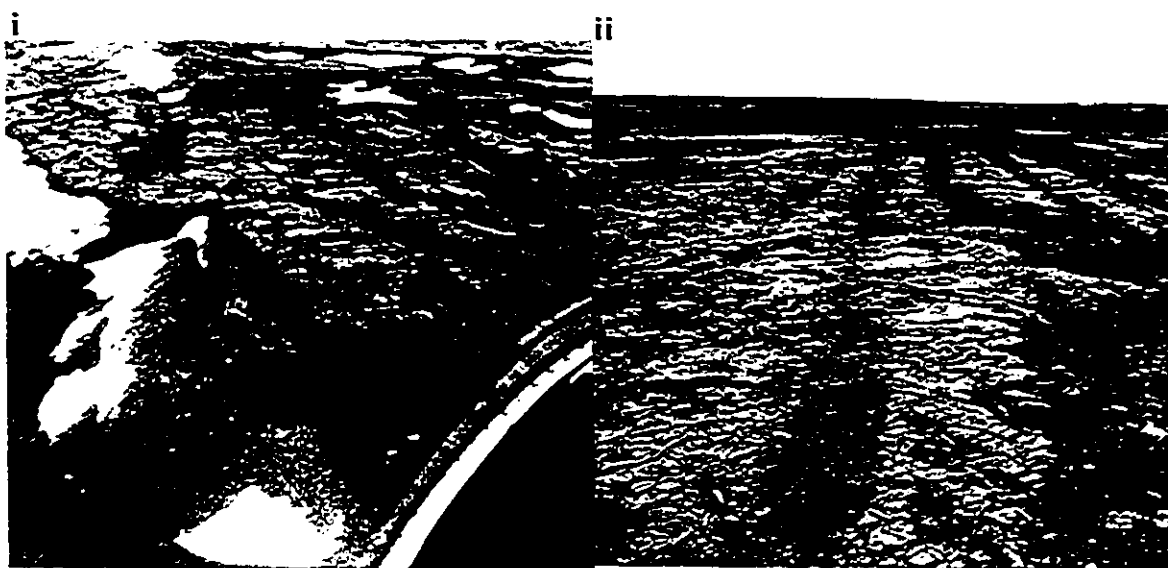


Figure 3.15 d. Linear troughs or grooves on escarpment nose. i) aerial view GSC 204763-R, ii) Ground view; note boulder lag on surface. Flow left to right (i) and away from viewer (ii). GSC 204764-M.



Figure 3.15 e. Rock drumlins at the two scales. i) outcrop scale, 5 m long (note s-form near notebook = 20 cm) GSC 204764-H



ii) large form ~250 m long (person for scale). GSC 204764

of powerful turbulent flows (Kor et al., 1991). Subglacial meltwater flow is indicated because no ice-marginal outwash deposits are associated with any of the scoured zones. Most eroded sediment was apparently removed from the field. Linear erosion zones form extensions of crescentic scours and, where elongate parallel to flow, may represent tunnel channels. These forms add to the evidence for subglacial meltwater flow.

Remnant ridges

Elongate ridges composed of sediment or rock and found in the lee of crescentic scour zones are best interpreted as scour remnant ridges (Allen, 1982) (Figs. 3.13b; 3.15c). The large (3-5 km), streamlined ridges and smaller spindle forms are interpreted as drumlins. These formed in response to the same erosive meltwater flows which scoured the rock surfaces. The ridges are remnant ridges (Allen, 1982; Shaw and Sharpe, 1987; Sharpe and Shaw, 1989) and correspond to the erosional drumlins of Shaw and Sharpe (1987).

Large up-flow-facing scarps also occur downflow of crescentic scours and in many cases are streamlined (Fig. 3.14). These features are also interpreted as erosional, in part from their relationship to the crescentic scours. Similar features, termed escarpment noses or flow-noses in Ontario (Shaw and Gilbert, 1991), are also interpreted as the products of large subglacial meltwater flows.

Many of these up-flow-facing scarps have sculpted erosion forms (s-forms), particularly flutings, on their upper surface as well as a lag of large clasts (Fig. 3.15). Small-scale erosion forms, s-forms, are the product of subglacial meltwater flow (Ljugner, 1930; Dahl, 1965) and are compatible with the evidence for broad regional flow (Shaw, 1988a; Kor et al., 1991). Flutings on the upper surface of

escarpment noses are linear due to the higher velocities in confined subglacial flow over the escarpment compared to flows sweeping around the escarpments (Kor et al., 1991). In addition, more erosive flow and large escarpment noses (deeper flows) are prominent at the east end of the Read Island field, possibly due to the fact that this area created a large topographic rise and confined meltwater flows which swept across the Dease Strait depression from the mainland (Fig. 3.1).

In summary, there is clear evidence for meltwater flow which produced complex erosional patterns (e.g. east end of the Read Island field). These patterns are attributed to turbulent flow (Allen, 1982). Because these features cover a 30 km wide zone and escarpment noses are 20-30 m high, the flows were about 30 km wide and overtopped high escarpments. Thus the scour zones and associated erosion features make a strong case for the erosion of associated drumlins by broad, deep, and powerful meltwater floods.

3.5 Discussion

Landform association is a powerful aid to landform interpretation. A possible genetic link between, hummocky moraine and drumlins is suggested by the arrangement of landforms on Wollaston Peninsula. The Penny Bay drumlin field dissects the western arc of hummocky moraine (Figs. 3.2; 3.8). This indicates powerful local flow late in deglaciation. Hummocky moraine generally indicates ice stagnation during final formation (e.g. Lawson, 1979), so it is unlikely that a dome of active ice in central Wollaston Peninsula formed the drumlins there; especially since a central dome would have been completely surrounded by wasting ice and cut off from the main ice masses occupying the

adjacent marine troughs. These drumlins more likely resulted from the erosion of existing drift and icy sediments by the release of stored meltwater from a wasting ice mass that covered Wollaston Peninsula. The fact that two streamlined fields at Penny Bay converge from level uplands in the centre of the Peninsula supports this argument, as does the fact that the drumlin characteristics favour an erosional origin (e.g. streamlined forms and scouring, Fig. 3.8; sediment-stripped bedrock, and rock and sediment drumlins immediately adjacent to hummocky moraine). Similar dissection of hummocky moraine is seen at a local scale (Fig. 3.13) where flooding formed streamlined remnants.

Truncation of hummocky (ice-cored) terrain by drumlin fields may relate to rapid ice flow if there was sufficient volume of ice in local caps to feed ice streams or surging glaciers. Recent geophysical data from Antarctica (Alley et al., 1986) indicate that ice streaming may occur by deformation of thick (~6 m) of subglacial sediment. Theoretical considerations suggest that such deformation could occur in soft sediment about 10 m thick, possibly eroding their substrates (Alley, 1991). Whereas sub-ice deformation may be offered as an alternative means of eroding the above drumlin fields, it is unlikely that it explains erosion of ice-cored terrain tens of metres thick. Deforming flows have insufficient power to melt thick ice cores.

A further link between drumlins, hummocky moraine and episodic meltwater drainage is the dissection of hummocky moraine on southern Wollaston Peninsula by a number of channels. These channels cut through a broad marginal moraine belt bordering the large Read Island drumlin field. This landform setting which is very similar to tunnel valleys and moraines in Wisconsin described by Attig et al. (1989).

In the Wisconsin example, the authors called on marginal permafrost and frozen bed conditions to trap subglacial meltwater prior to flooding. On Wollaston Peninsula, end moraine and hummocky moraine may have provided a frozen toe or a topographic barrier to meltwater drainage. The fact that the moraines are presently largely ice-cored favours freezing near the ice margin during deglaciation. This view is supported by the presence of coarse sediment in some sections and the predominance of sorted sediments in most sections in hummocky moraine. That drumlins and tunnel valleys may be glaciofluvial landforms representing a sheet flow to channel flow transition (Shaw and Kvill, 1984) is a clear possibility along southern Victoria Island. In addition, the final element in this glaciofluvial transition, eskers, forms an overlapping, divergent array with the Read Island drumlins (Fig. 1.14).

The time for formation and age of streamlined landscapes is poorly constrained. Hodgson (1987) suggested several late Wisconsinan cross-flows took place within ~500-1,000 years on northeast Victoria Island. Boulton and Clark (1990) suggested a record of rapid cross-cutting (induced by climate) to explain cross-flow patterns produced by the Wisconsinan Laurentide Ice Sheet. Dyke and Morris (1988) suggested preservation of streamlined terrain from earliest Wisconsinan time on nearby Prince of Wales Island (Fig. 3.1). Strong evidence shows preservation of Mid-Wisconsinan eskers and hummocky moraine that survived the last major ice advance (Lagerbäck, 1988). This makes it difficult to unravel complex landform patterns. Nevertheless, if streamlining occurred late in a glacial episode (Dardis and McCabe, 1989), then cross-cut forms on southern Victoria Island indicate rapid switching of flow direction (3-6 per 1,000 years).

Release of stored subglacial meltwater from changing storage basins appears plausible as an explanation of the frequency of discharges and the variability of flow directions. Rapid ice flows, surges and ice streams are also possible explanations, but they do not so easily explain radical changes in flow direction, scoured terrains, and transitions to glaciofluvial landforms (tunnel channels and eskers), and the strong evidence for formative flows with separation structures.

Recent developments on modern glaciers may help constrain our models of drumlin formation. Surging glaciers in Alaska showed a clear relationship with high subglacial water pressures and post-surge flooding (Kamb et al., 1985). No drumlins were produced, but the storage and sudden release of large subglacial reservoirs may apply to former conditions on southern Victoria Island (Sharpe, 1988a). Rapid flow in Antarctic ice streams appears to relate to deforming sub-ice slurries (e.g. Alley et al., 1986) which some argue produces drumlins (Boulton, 1986; Menzies, 1987; Boyce and Eyles, 1991). Sediment fills in many drumlins (e.g. Dardis, 1983; Shaw and Kvill, 1984; McCabe and Dardis, 1989a,b) and the Read Island examples can not be accounted for by this depositional mechanism. Similarly, for rock drumlins with little sediment (e.g. Kor et al., 1991; and, Read Island examples), sediment deformation explains neither the sculpted forms nor the absence of sediment.

3.6 Summary

In this section I identify several characteristics of drumlin form and composition and make inferences concerning streamlined landforms on southern Victoria Island, in an attempt to improve understanding of drumlin formation.

First, the spatial arrangement of drumlins is important. The landforms occur as sets of forms ranging from drumlins to flutes. The set of transitional landforms represents spatially varying flow strengths in flows the scale of drumlin fields. The forms are analogous to bedforms produced by separated, turbulent flow (Allen, 1982; Shaw and Kvill, 1984; Shaw and Sharpe, 1987). Some forms show a composite or en echelon arrangement in parts of the landform set (Fig 3.3c). This organization is again characteristic of complex, turbulent flows (Allen, 1982). Streamlined fields cross-cut by inset narrow fields (with sinuous forms) reflect a transition from broad flow to more confined flow (sheetflow to tunnel flow; e.g. Banning Lake, Fig. 3.9).

The composition of the drumlins provides information on their formation if the drumlins are depositional in origin. In this regard the stratified cores of western Read Island drumlins favour a glaciofluvial cavity fill origin, although a conformable fill is not clearly demonstrated. The occurrence of meltwater erosion forms (s-forms) and sediment-free zones (tunnel valleys) in interdrumlin areas, support the erosive action of subglacial meltwater on the eastern zone of Read Island drumlin field. Rock drumlins require erosion for their formation. Rock drumlins with superimposed s-forms suggest that subglacial meltwater flows were powerful agents of erosion (e.g. Shaw and Sharpe, 1987; Sharpe and Shaw, 1989; Kor et al., 1991).

The presence of till in the exposed (top) sections of Penny Bay and Banning Lake drumlins favours ice processes for the deposition of drumlin sediment, probably lodgement and less likely deformation till. The combination of cross-fabrics, flow scouring (Penny Bay), and large crescentic

erosion forms parallel to drumlin axis (Banning Lake) indicates that erosional processes may also provide a viable explanation for drumlin formation at Penny Bay and Banning Lake. In particular, subglacial meltwater erosion may have sculpted drumlins in and adjacent to the up-field scour zones which display crescentic depressions and remnant streamlined ridges, and in places, dissected hummocky ice-cored terrain.

Finally, it may be suggested that different processes operated in different parts of the same drumlin field. The large scour zones and associated small erosion features in the eastern end of the Read Island field contrast with the western end where smaller divergent forms, en echelon forms, and more limited sediment stripping are found. Whereas the eastern end of the field appears to be dominated by confined erosional flows which rose from Dease Strait, the western end may reflect complex lateral flow expansion beyond the Richardson Islands and possible deposition in cavities downflow (Fig. 3.1).

4.0 ESKERS

Eskers are important landforms in south central Victoria Island that record the late glacial plumbing system of the wasting ice sheets. Eskers record former meltwater discharge that responded to pressure gradients defined by the profile of the ice sheet (Shreve, 1972). This relationship allows reconstruction of the ice sheet and its drainage. The arrangement of eskers defines the distribution of late ice masses while their form identifies the nature of the conduits of the final drainage system. Esker sediments record the depositional environment of these conduits.

This chapter attempts to relate esker form and sediments to the conditions of formation. This is carried out by mapping landform elements of the esker system (scours, tunnel valleys, and ridges), by determining their spatial arrangement (Fig. 4.1), and by examining available sections. The hydrology and sedimentology of eskers is briefly reviewed to aid interpretation of the esker sediments and to aid in making inferences about late-glacial environments of deposition and the state of the melting ice masses.

4.1 Esker pattern

Two patterns of esker systems are prominent on southern Victoria Island. In southwestern Victoria Island, eskers form long continuous or semi-continuous networks that branch or are arborescent, suggesting a former connected drainage system. In southeastern Victoria Island, esker patterns are fragmented or disconnected, suggesting a former less connected or disrupted drainage system. Eskers are virtually absent from Wollaston Peninsula plateau (Fig. 4.1).

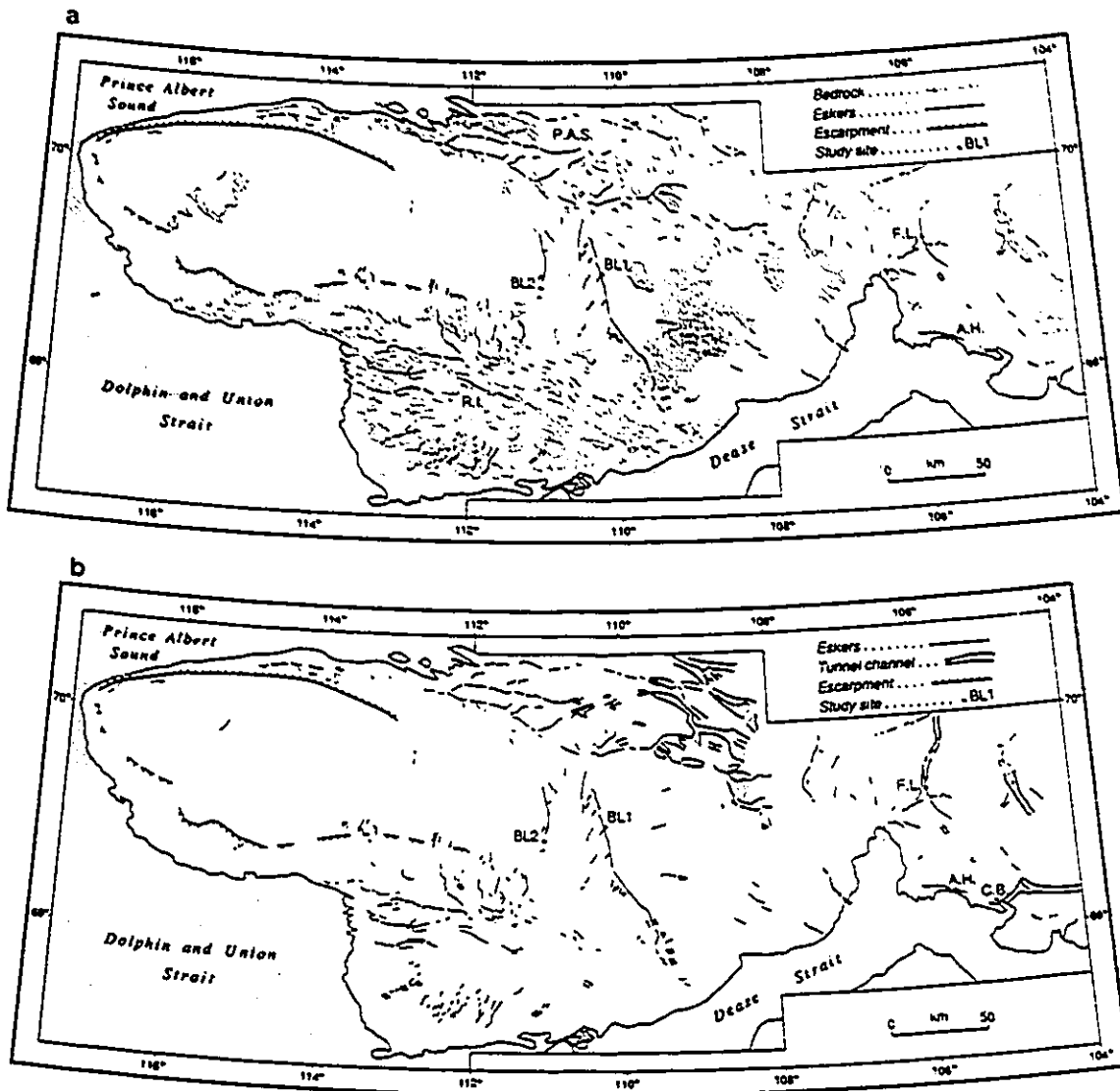


Figure 4.1 a. Distribution of eskers and exposed bedrock, southern Victoria Island showing topographic position of major esker networks; R.I. = Read Island, P.A.S. = Prince Albert Sound; B.L.=Banning Lake, A.H.=Augustus hills, and F.L.= Ferguson Lake; b. eskers and tunnel valleys.

Continuous esker systems

Read Island esker system, 100 km long, is a continuous network forming a divergent pattern, particularly on its northern flank (RI Fig. 4.1). Eskers are 5-10 m high, sharp-crested ridges flanked by sediment-scoured zones. The esker pattern follows that of the large diverging Read Island drumlin field (Fig. 3.2). The Prince Albert Sound esker system (P.A.S., Fig. 4.1) is about 125 km long, continuous and displays a converging pattern north of Wollaston Peninsula. Eskers are large and broad in main segments, and smaller and sharp-crested in joining segments. A series of tunnel channels is concentrated at the east (source area) end of this system (Fig. 4.1). This esker system also mirrors the pattern of converging drumlins in the same area (Fig. 4.1b). A third esker system, Banning Lake, comprises two long (up to 125 km) segments with few joining ridges (west flank of BL 1, Fig. 4.1). Esker trunk segments are mainly long, broad ridges but joining segments are short and sharp-crested. A large sediment-free (scoured) area occurs at the southeast end of this system (Figs. 1.14; 4.1b). This esker system lies at the margins of several streamlined fields.

Fragmented esker systems

Eskers in southeastern Victoria Island are fragmented and some appear to cross-cut one another (F.L. Fig. 4.1). Eskers commonly are broad, up to 100 m high (A.H.) and are associated with tunnel channels (Fig. 4.1). Some eskers show no association with streamlined forms (A.H.), while in other cases landforms are gradational (F.L., Fig. 4.1; Sharpe, 1985).

4.2 Scours and tunnel valleys

Eskers on southern Victoria Island are often flanked by

bare or thinly covered bedrock areas (Fig. 4.1a) that are readily mapped on aerial photographs (Fig. 3.13). These exposed bedrock areas are 1-2 km wide and occur as short, scoured, lengths 0.5-1 km (Fig. 4.2). There are also scoured channels, 5-25 km long, termed tunnel valleys (Wright, 1973). In places the scour zones and channels are cut as rock trenches, but more commonly they are eroded into sediment several metres to ten or more metres thick (Fig. 4.3). Scours and channels are common in streamlined terrain of southern Victoria Island (Fig. 4.1b). Similar sediment-poor zones have been noted in arctic streamlined terrain (St-Onge, 1984a,b; Shilts, 1984; Aylsworth and Shilts, 1989b).

Scoured surfaces carry a lag of scattered boulders, 0.5 m to 1.0 m diameter (Fig. 4.2b). In places, channel scours contain transverse, wave-like ridges several metres high and hundreds of metres across. Trains of these transverse ridges occur adjacent to the well-defined scoured zones (e.g. Fig. 2.5). The ridges generally have stony surfaces but they are sandy in some areas.

Broad scoured zones, tunnel channels (Figs. 4.1b, 4.3) and linear rock ridges (rock drumlins) up to 20 m high (Fig. 4.4) occur near Cambridge Bay. These rock ridges and adjacent sediment ridges (drumlins) have crescentic depressions containing lakes around their upflow ends. This assemblage of scour features links the streamlined terrain (crescentic scours, sediment residuals, or drumlins) and tunnel channels (crescentic scours, rock drumlins, and transverse forms) (Fig. 4.5). Several other extensive scours contain fields of rock drumlins (Fig. 4.4).

Scoured zones in areas of thick sediment (e.g. Ferguson Lake, Fig. 4.1) occur as a series of arcuate lakes, some of which are modified by secondary thermokarst. In most

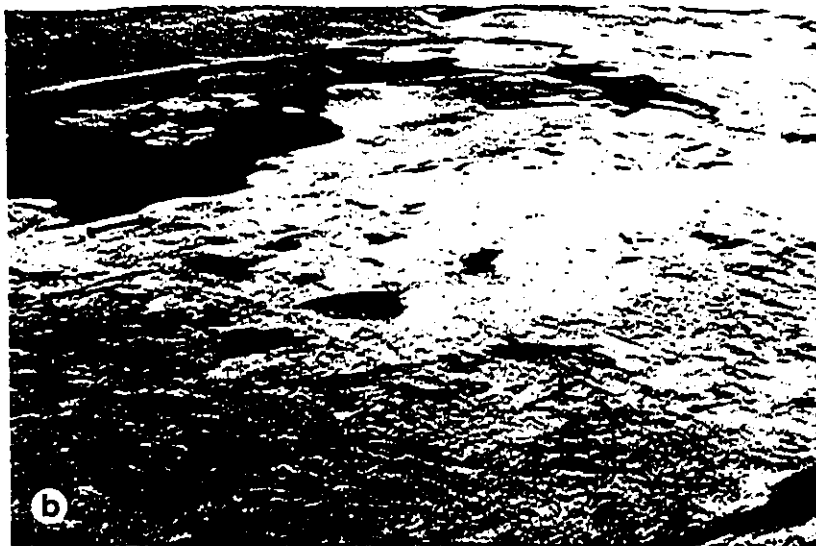


Figure 4.2. Oblique aerial photographs show scour zones:
a) general view; scale bar=1 km

b) closer view of scour zone with lag of large boulders in
view; scale bar=100 m.

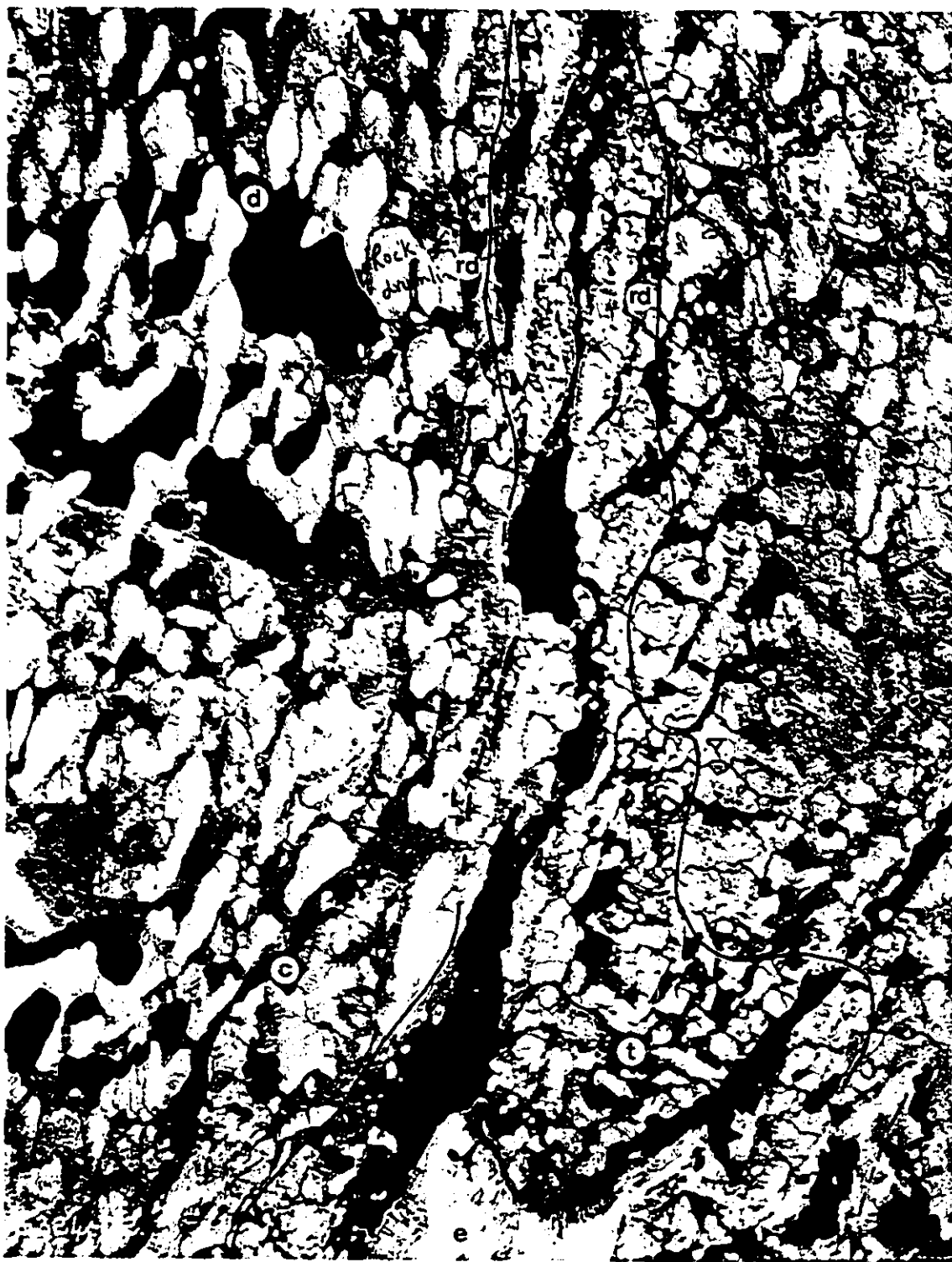


Figure 4.3. Photograph of tunnel valley scour zones (outline) on carbonate bedrock, northwest of Cambridge Bay. NAPL Photograph A16324-72. Note en echelon sediment drumlins (d), rock drumlins (rd), esker (e) and transverse ridges (t).

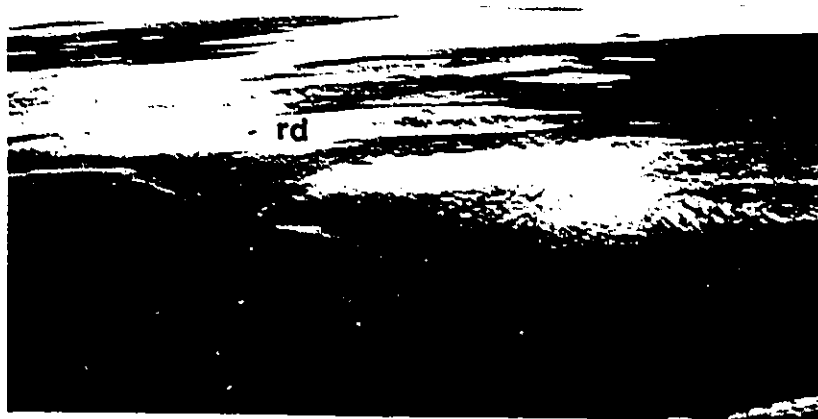
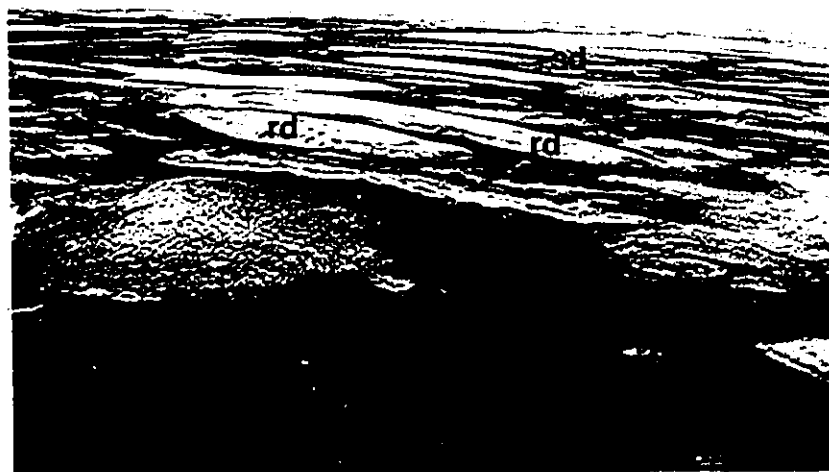


Figure 4.4. Erosional remnants, rock drumlins in tunnel valley northeast of Cambridge Bay, a) crescentic depression (c) upflow of rock drumlin (rd); GSC # 204764-C.



b) rock drumlin (rd) with sediment drumlin (sd) in background GSC # 204764-G.

ESKER AND TUNNEL VALLEY LANDFORMS

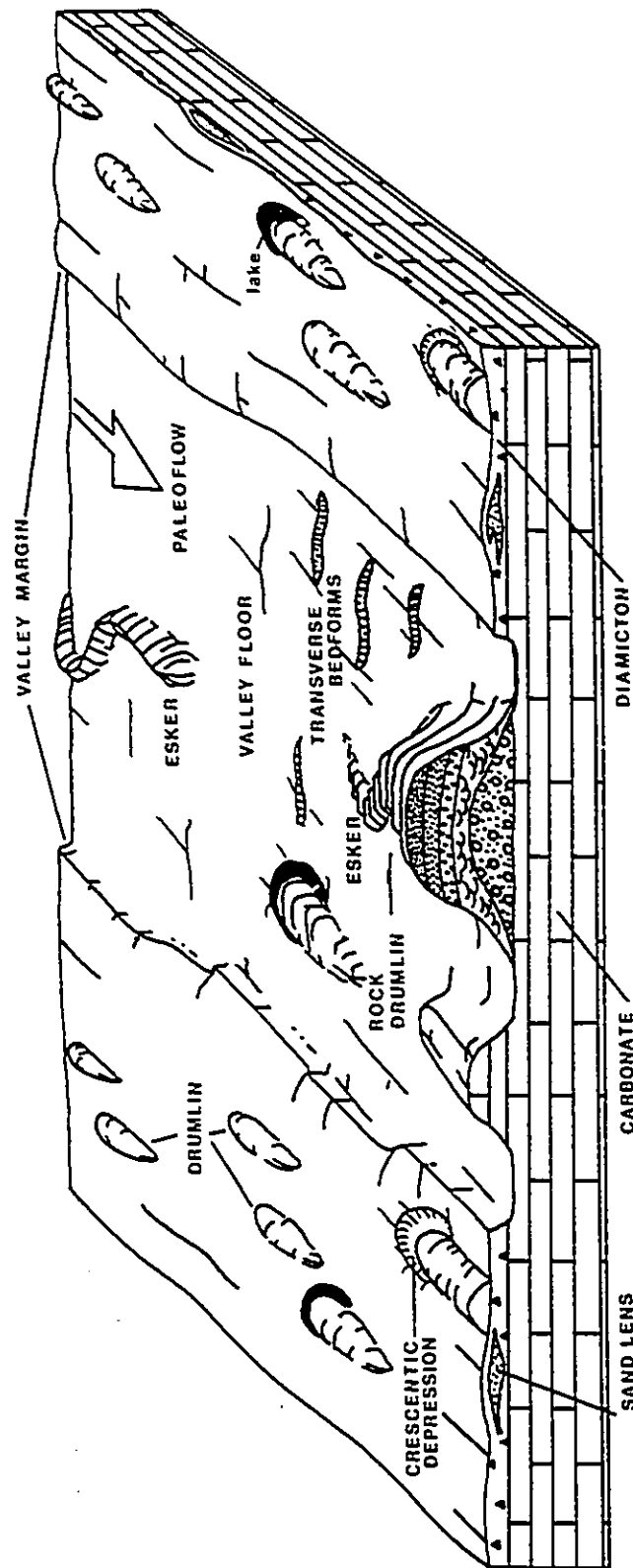


Figure 4.5. Schematic model of a tunnel valley eroded into a streamlined (drumlin) field.

settings, eskers and scoured landforms are sub-parallel to the trend of, and cut through fields of drumlins or drumlinoids (Fig. 1.14). The fact that eskers are found within scours and the scours have fluted margins indicates complex sequences of formation, discussed later for the Ferguson Lake esker complex.

4.3 Ridge form and sediments

Two basic esker forms are identified: broad ridges, and, narrow ridges. Broad-ridged eskers are subdivided into flat-topped and hummocky forms. These distinctions emphasize the lobes, terrace features and multiple ridges of broad eskers. Broad, hummocky eskers (0.5 km-1.5 km wide) are most common while broad, flat-topped (0.2 km-1.0 km wide) eskers are also present. Broad eskers (termed extended deposits by Hebrand and Åmark, 1989) extend for tens of kilometres in an evenly-spaced dendritic pattern (Banning Lake and Prince Albert Sound, Fig. 4.1a) covering up to 50% of the total esker length. Flat-topped esker segments are 0.5 km-5 km long and occur within broader eskers regions.

Broad eskers are 10-30 m high and gullies expose sandy sediments with few stones. In places, for example at Banning Lake, a braided, channelled surface marks flat lobate ridges. Broad eskers trend sub-parallel or oblique to flow trends in streamlined terrain.

Esker networks also contain sharp-crested, narrow segments. Narrow ridges are commonly less than 100 m wide and they may have lateral scours. These segments are hundreds of metres to several kilometres long and they comprise 30% of the total esker network. Sharp-crested, narrow eskers are 5-25 m high and are commonly covered with

coarse, rounded, cobbly to bouldery gravel. Narrow eskers tend to run parallel to streamlined forms often reproducing their diverging patterns (e.g. Read Island Fig. 4.1a).

Eskers in the region have very low gradients and only their general slopes can be defined. Some Banning Lake eskers have an elevation range of about 10 m over a 30 km length of esker. Some esker systems slope to a common connecting point or drainage node; for example, north of Banning Lake area. The general location of eskers on southern Victoria Island favours topographic lows set around Wollaston Peninsula plateau (Fig. 4.1), a point that may support a subglacial interpretation.

4.4 Measured sections

Several sections in broad, flat-topped eskers were measured, mainly in gullies. Smaller, sharp-crested eskers have coarse, stony surfaces and fewer sections were measured. This may bias the evaluation of esker sediments but various types of eskers are represented. Three areas are recorded; Banning Lake, Augustus Hills, and Ferguson Lake (Fig. 4.1a).

4.4.1 Banning Lake

Introduction

The Banning Lake esker complex comprises a prominent north-south esker system (BL 1) which joins a major east-west esker system to the north (Fig. 4.1a). A second north-south ridge (BL 2), to the west of the main Banning Lake ridge, is of enigmatic origin but consists of an esker ridge for at least part of its length. The Banning Lake ridge is part of a network of broad and narrow esker ridges and scours adjacent to study site 124 (Figs. 4.6a). The ridge is more

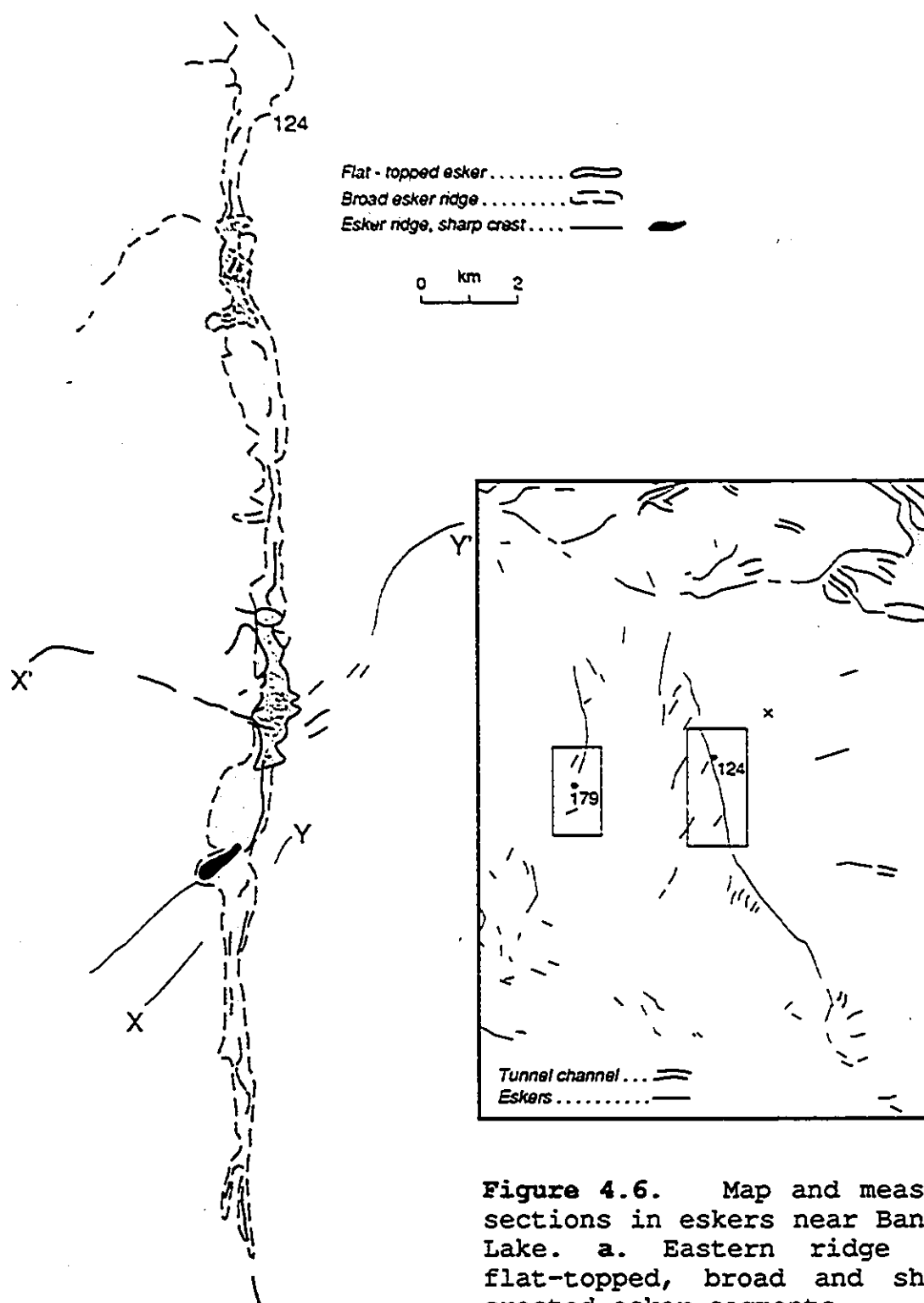


Figure 4.6. Map and measured sections in eskers near Banning Lake. a. Eastern ridge with flat-topped, broad and sharp-crested esker segments.

than 100 km long and has a series of feeder ridges, particularly on its western side (Figs. 4.6a; 4.1). Flat-topped fan or delta features formed where feeders connect to the main ridges. These feeders record sediment dispersal from west to east. The flat-topped, broad features are better defined and younger than the hummocky, broad ridges, and in places, their connecting ridges appear to cross-cut (they may be tributaries) the major north-south ridge (Fig. 4.6a, X-Y; X'-Y').

A large portion (~75%) of the Banning Lake east ridge (Fig. 4.6a) is broad and marked in places by braided channel patterns, "deltaic lobes" and fans, or it is interrupted by short segments of sharp-crested, narrow ridges, gaps and/or tunnel valleys. Broad ridges show large polygons (coarse sediment) on flat-topped segments whereas sandy sediments occur in gullied, dissected hummocky stretches. Small ice-block lakes are noticeable within broad segments of the features.

A separate western ridge has only a few connecting eskers (Fig. 4.6b). The main ridge has an adjacent ice-cored terrain and a short, broad segment with exposed sections (site 179). Connecting eskers join from the east. Fyles (1963) considered a series of arcuate ridges and the ice-cored terrain on the east flank (Fig. 4.6b) to be a moraine ridge. Several discrete, broad lobate segments indicate an esker ridge at section 179 and their sedimentary characteristics support this identification. A short, parallel esker ridge, which occurs to the northwest of section 179 (E on Fig. 4.6b), ends in a subaqueous fan, indicating paleoflow towards the southwest.

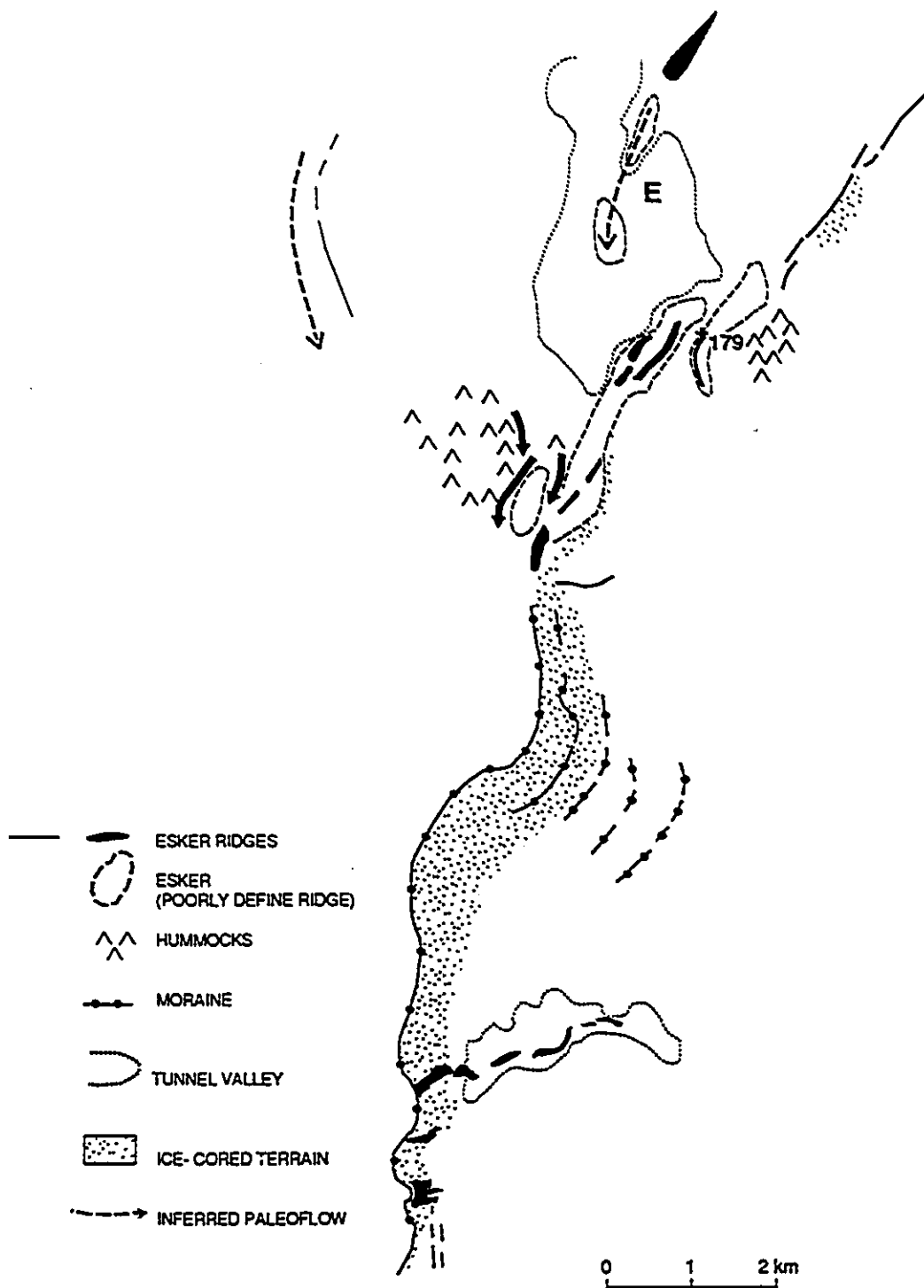


Figure 4.6b. Esker moraine complex on west side of Banning Lake (B.L.2, Figure 5.1). Based on NAPL A15787-109.

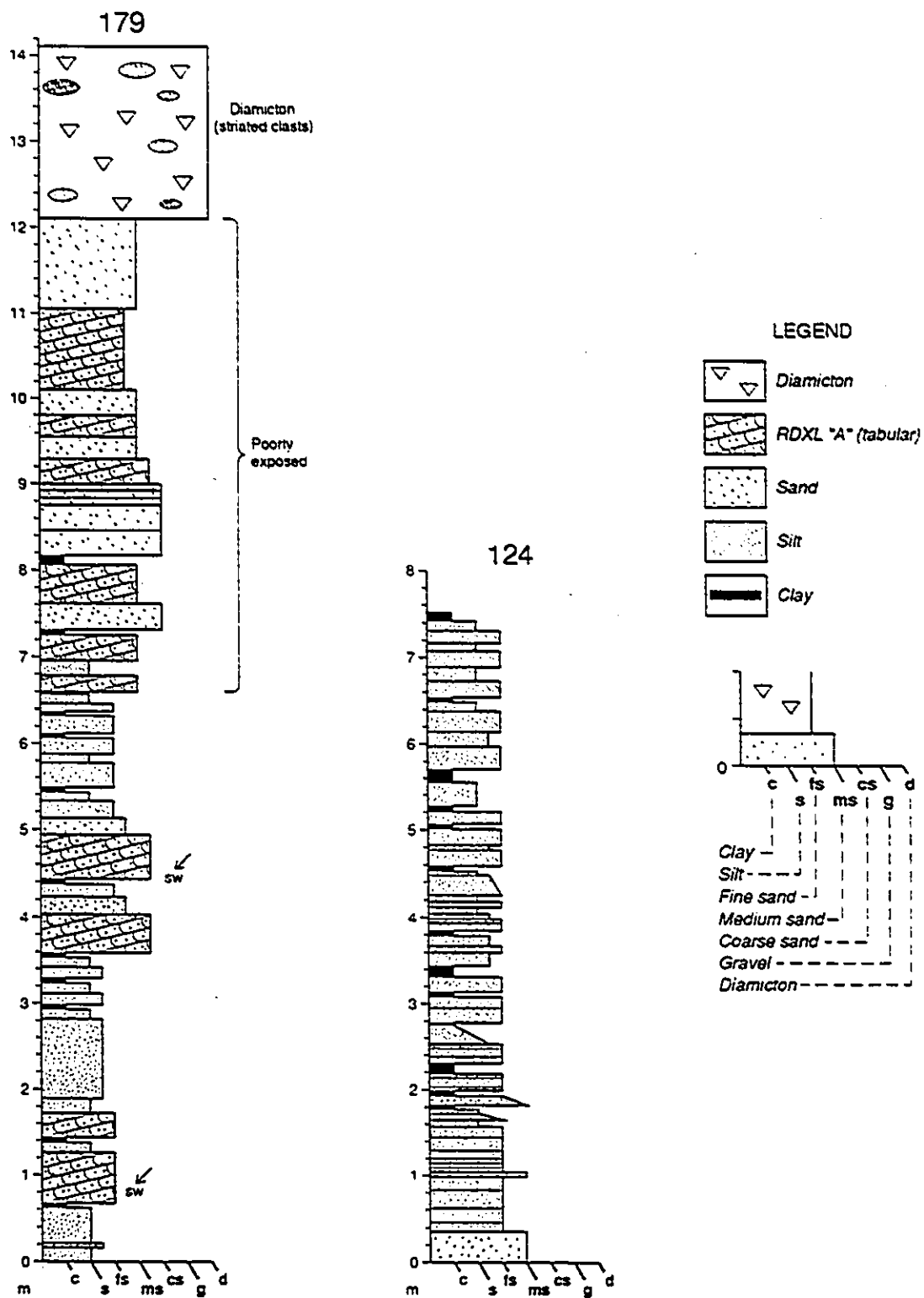


Figure 4.6 c. Measured section at 87-124; at 84-179

Both Banning Lake esker systems occur as north-south ridges defining the divergent pattern of an oriented landscape west of Banning Lake. The divergent pattern of landforms includes: streamlined forms, oriented ridges, minor moraines, and transverse ridges (Fig. 4.1; Sharpe, in preparation). A third north-south esker ridge cross-cuts the divergent landform pattern (Fig. 1.14; 4.6 a).

The measured sections provide only a cursory, but important view of the ridge sediments. Whereas the Banning Lake ridges were considered end moraines (Fyles, 1963), their internal structure and paleoflow directions allow one to distinguish moraines and eskers.

Sections

The two measured sections, 84-179 and 87-124 (Fig. 4.6c), reveal only a portion (7 m and 15 m) of the sediment in the 25-30 m high landforms. The sites are both above the limit of marine washing but lacustrine washing may have slightly affected the top surfaces. Braided channel forms are evident on the ridge surface 10 km to the south of site 124. Both measured intervals reveal thick sequences with horizontal, rhythmically bedded, fine sand, silt, and clay (Fig. 4.6 c). Coarser beds, diamictos, occasionally cap sequences. Cross-laminated fine to medium sand predominates (site 179) in fining-upwards beds, 10-50 cm thick. Silty clay beds, 2-10 cm thick, cap these graded units.

Massive sand and plane-bedded fine to medium sand are rare and trough and tabular cross-bedded beds were not observed; although they may occur deeper in the landform. Minor loading structures, soft-sediment deformation, and small faulting structures are present; no major faulting or deformation was recognized. Paleoflow in the western ridge is southwesterly, parallel to flow in an adjacent esker (E,

Fig. 4.6b). Paleoflow in the eastern ridge is northward based on the orientation of small lobes and the braided channel patterns (Fig. 4.6a).

Interpretation

The arrangement and form of these ridges need explanation. The Banning Lake network was apparently controlled by topography which directed the divergent, late-glacial flow around Wollaston plateau (Fig. 1.4). Small, sharp ridges connect to larger broad ridges: the former defined lobate drainage networks; the latter defined enlarged creases between ice lobes (Fig. 4.1).

Broad flat-topped ridges are considered to represent open-channel eskers, possibly formed under subaerial conditions (Hebrand and Åmark, 1989): the braided pattern south of site 124 provides evidence of subaerial channels and northward paleoflow. Level topography and similar elevations along the feature is in accord with flows under atmospheric rather than hydrostatic pressure. Fans to the east suggests standing water there, but ponding was only local because few fans are found along the sides of the north-south ridge (site 124). The fine-textured rhythmites with no gravel (site 124) are interpreted as annual (varve) beds in glaciolacustrine sediments (Fig. 4.6c) (Saunderson, 1975).

Narrow, sharp-crested ridges, within larger broad esker complexes, represent deposits younger than the broad ridges, which appear to be degraded flat-topped forms or forms affected by erosion, possibly thermokarst. If the broad ridges represent degraded forms and the narrow ridges represent subglacial conditions (Shreve, 1972), then we have evidence of periodic return to subglacial sedimentation following a period of open-channel flow of meltwater which, on the evidence of varve sedimentation, was primarily

supraglacial in origin.

Small eskers which cross the main north-south ridge are interpreted to have formed before the main ridge (Fig. 4.6a, X-Y; X'-Y'). This implies a westward shift in the drainage network or a later development of the north-south drainage, followed by confined drainage to and along the main ridge axis or lobe junction. More simply, the cross ridges may have been tributaries of the main esker.

4.4.2 Augustus Hills

Introduction

Several sections were measured in a broad, flat-topped feature about 25 km west of Cambridge Bay (AH, Figs. 4.1; 4.7). This ridge is one of several large, very prominent eskerine ridges in the area. It is less than 30 km long, 2-3 km wide, and 50-100 m high (Fig. 4.7). It is flat-topped in the central 1-3 km, steep on the north slope, and gentler sloping and broader on the south (Fig. 4.7 c). A longitudinal profile comprises a series of swales and rises which decrease in amplitude to the west (Fig. 4.7 c). In places, deep gullies dissect the sandy sediment comprising the ridge. The surface of the ridge indicates washing by marine waters; raised strandlines are common, particularly on its south side (marine cover reached 170 m a.s.l. about 9,000 years B.P). Rare surface stones, including some ventifacts, form a coarse lag.

The Augustus Hills ridge is broad in comparison to its length, and thus it is similar to the Mount Pelly and Lady Pelly esker ridges to the east (Fig. 4.1; 4.7 b). Although modified by littoral and eolian processes, the 100 m height of the ridge makes it an order of magnitude larger than any littoral feature found elsewhere on southern Victoria Island.

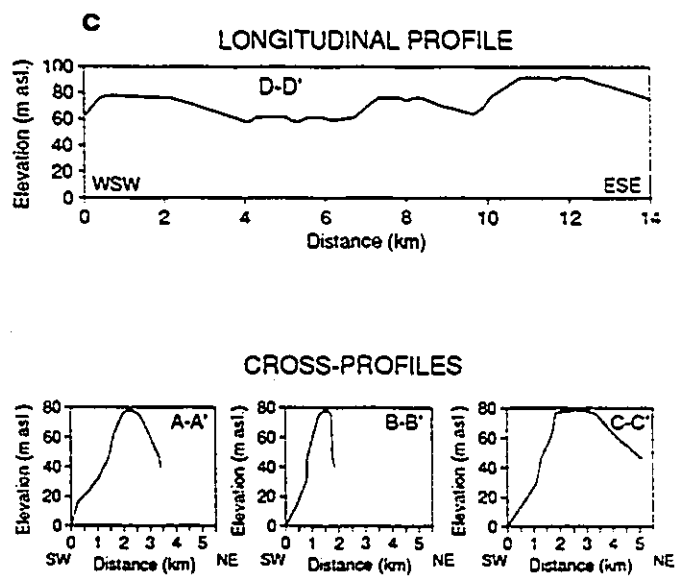
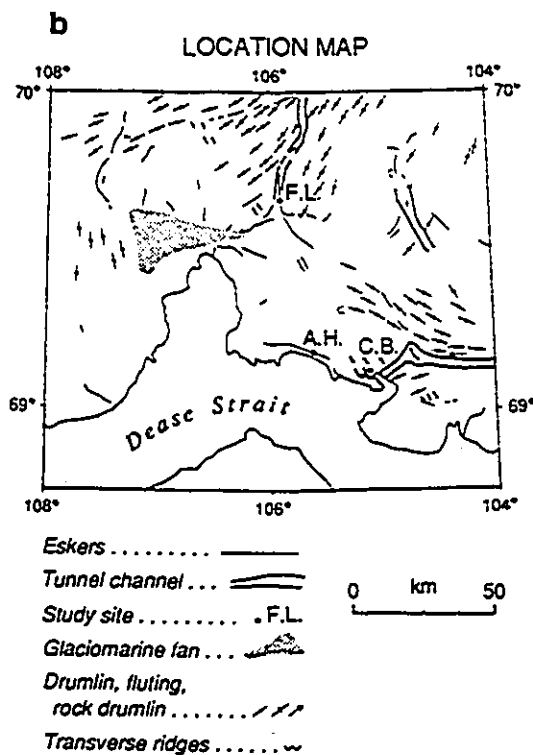
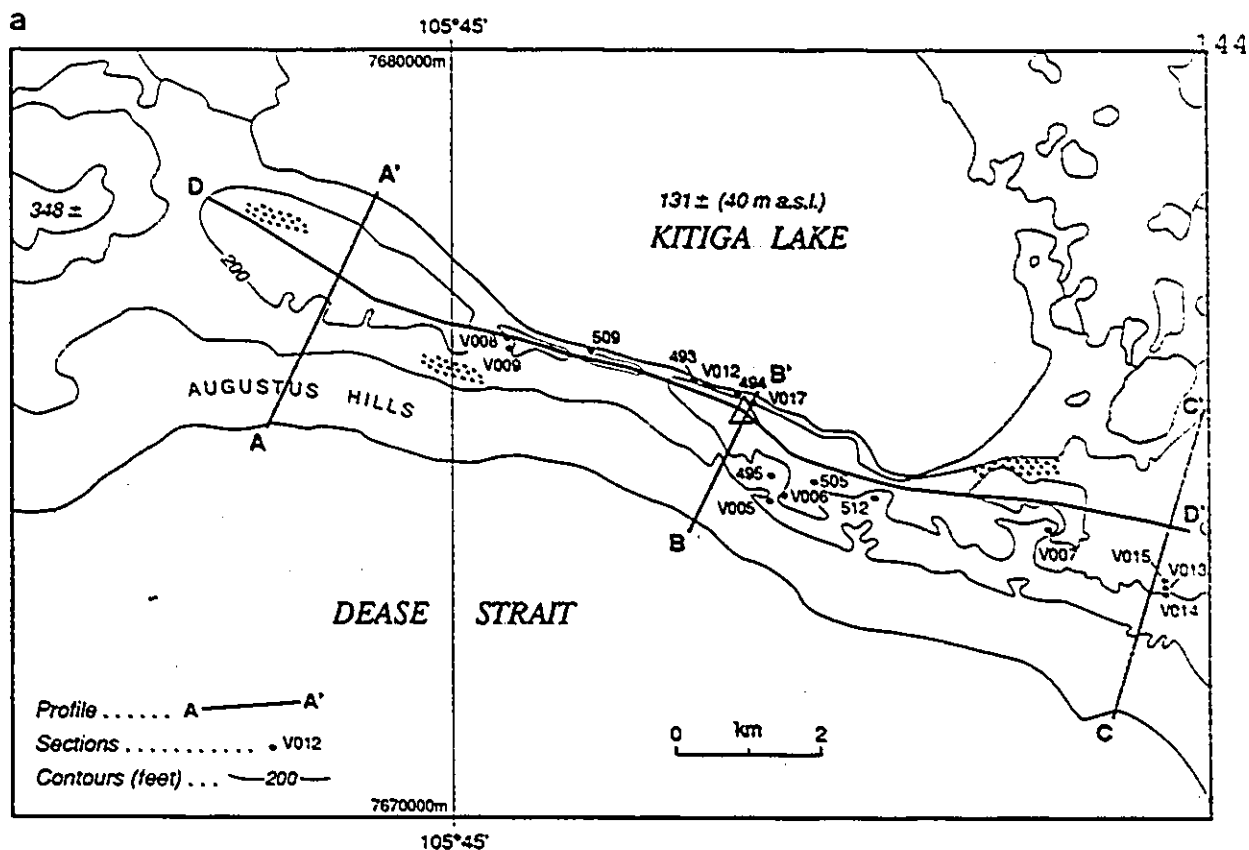


Figure 4.7. a. Map of Augustus Hills ridge and study sites.
b. location map; see Fig. 5.1. c. profiles.

Setting

The Augustus Hills ridge is oriented roughly east-west on the northern margin of Dease Strait (Figs. 4.1; 4.7). This near-shore setting raises the possibility of littoral origins which are discussed below. The ridge is one of several large prominent eskerine ridges. It extends from the east through Cambridge Bay as part of a tunnel valley (Fig. 4.5; with small esker) trending west and southwest for 40-50 km (Fig. 4.7b). A ridge on the west side of Wellington Bay may be the westward extent of Augustus Hills or it may have curved northwest to join a group of eskers connected to a large glaciomarine fan deposit (Fig. 4.7 b). Augustus Hills are sub-parallel to part of an esker complex north of Ferguson Lake (Fig. 4.1) which also converges on the glaciomarine fan north of Wellington Bay (Fig. 4.1).

No nearby streamlined landforms, except those adjacent to the Cambridge Bay tunnel channel, parallel Augustus Hills (Figs. 4.1; 4.5). The mainly rock-formed tunnel channel, contains transverse bedforms, and rock drumlins with up-flow crescentic scours (Fig. 4.2). Rock drumlins within the tunnel channel parallel sediment drumlins on flanking terrain which also show up-flow crescentic scours (Fig. 4.5).

Sections

Many sections were measured over 10 km of the Augustus Hills ridge, although most were found in a 5 km stretch adjacent to Kitiga Lake (Fig. 4.7). Measured sections are 10-35 m high and individual sections reveal about half of the ridge sediments (Fig. 4.7c); in total the sections cover the top 50 m of a 100 m thick package (Fig. 4.8). Bed geometry appears to be planar along Kitiga Lake bluffs (Fig. 4.9).

The sections comprises rhythmic sequences: ripple cross-laminated fine to medium sand beds, grading to silty

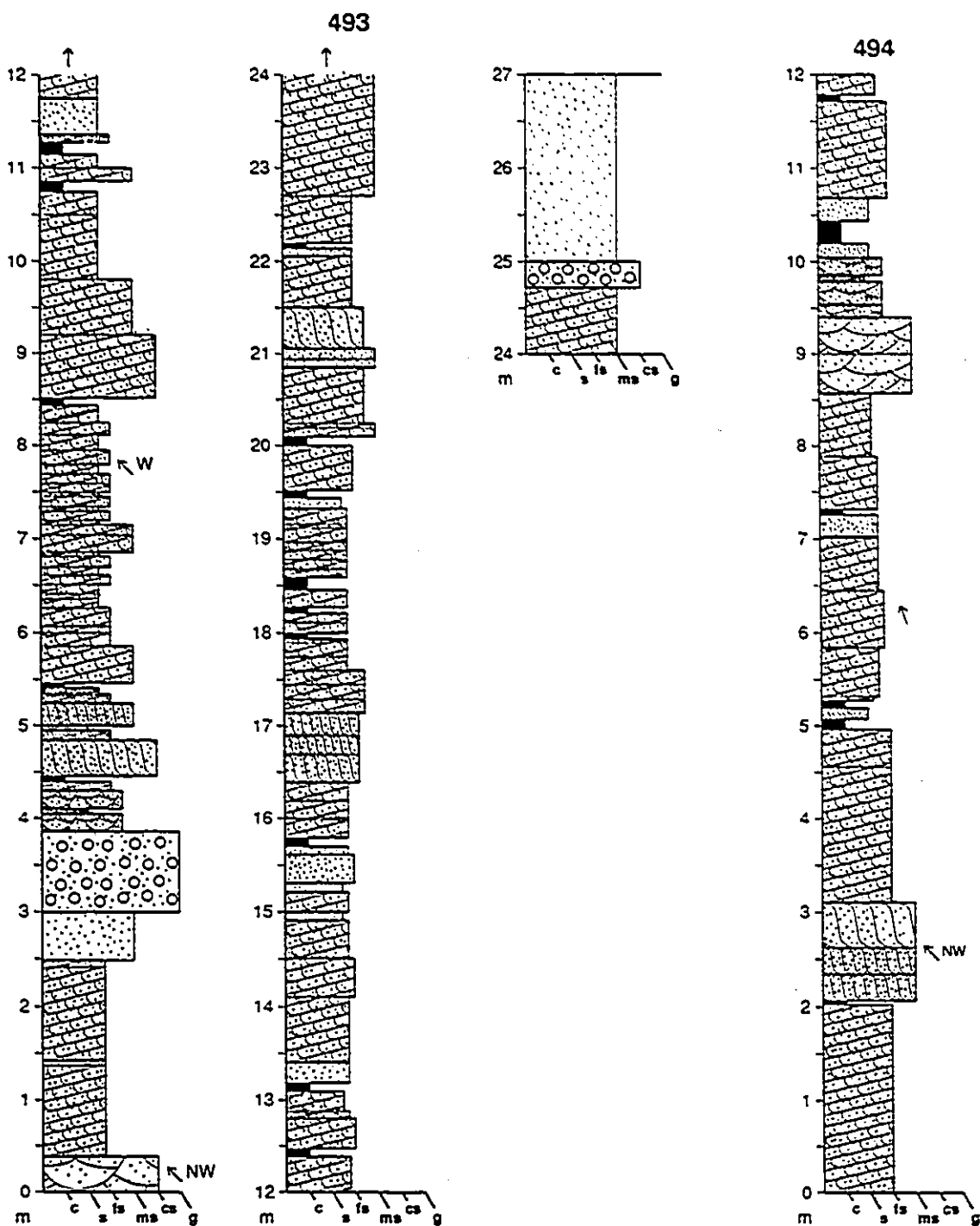


Figure 4.8. Measured section in esker at Augustus Hills.
 a) 493 and 494 measured by D. Sharpe and R. Kelly

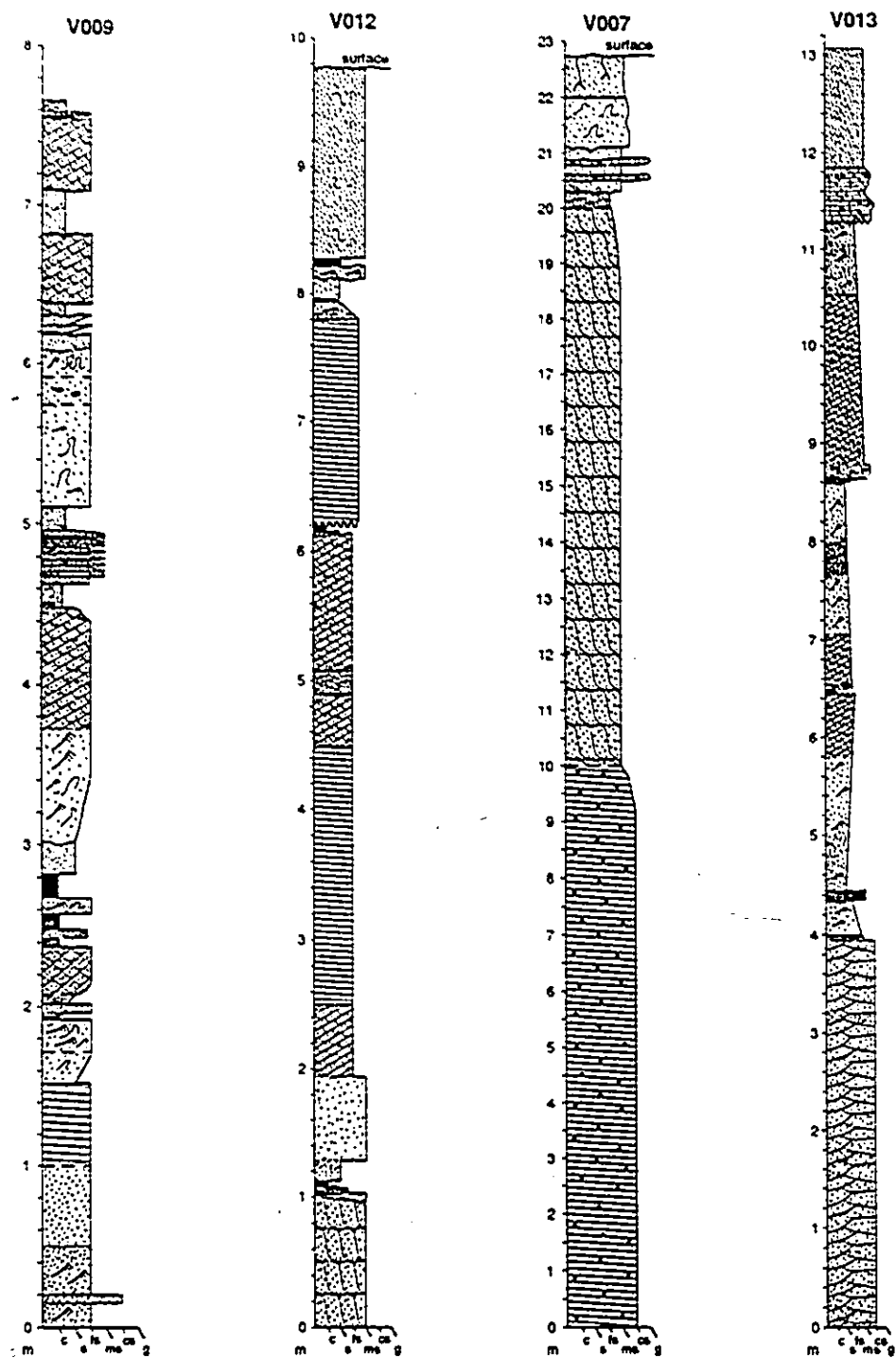


Figure 4.8 b. V007, 9, 12, -V013 measured by T. Brennand and M. Byerley. See next page for legend. Note scale change.

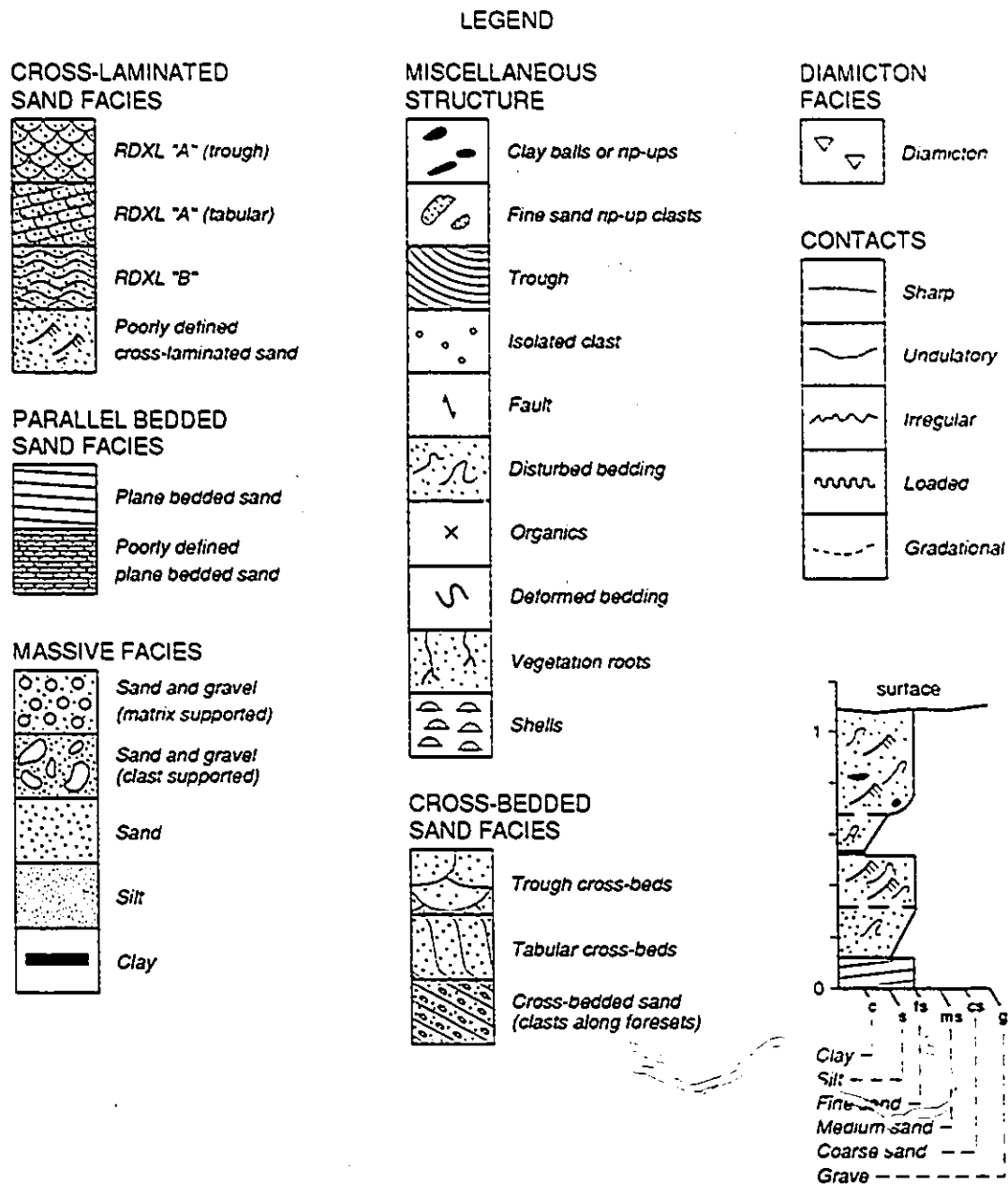


Figure 4.8 c. Legend for figure 4.8b

fine sand, capped by clay or silty clay (Fig. 4.8c). The rhythmic sequences are 0.2 m to several metres thick. Cross-laminated sets show low angles of climb (5° - 10°). Lower parts of the section include trough and tabular cross-bedded sets of medium sand and minor amounts of massive coarse to pebbly sand.

The most remarkable aspect of these sediments is the absence of gravel. Pebbly sand occurs at the top of several sections and rare cobbles or boulders are found at the surface. Paleoflow was predominantly towards the west to northwest, yet there is no noticeable change in texture east to west, except that clay or silty clay beds are rare in eastern sections, while clayey beds are common further west (Fig. 4.8).

Bedding in eastern sections consists of large-scale planar sets and smaller scale, tabular and trough cross-beds, with minor amounts of ripple drift. Thick trough cross-bedded cosets occur in south central areas, while climbing ripple and plane bedded sets are more common in westward sections (Fig. 4.8). Whereas beds are typically one to several metres thick in the east, thinner (5-25 cm) beds are common in the west. This decrease in flow regime for bedform stratification and set thickness represents a transition to a lower energy environment from east to west.

Fining-upwards sets are common, while coarsening upwards sets are less common. This graded character is found in eastern sections while a more rhythmic pattern is found in western sections.

Interpretation

The geomorphology of the long, straight to curvilinear ridge at Augustus Hills identifies it as an esker or, alternatively, a raised littoral feature, beach ridge;



Figure 4.9. Planar tabular sediments, Kitiga Lake. GSC 1991-601. Section is ~30 m high. White line is ~50 m long.

baymouth bar or a spit. The coastal location, beach terraces and the presence of marine fauna in upper sediments confirms the action of marine processes. The consistent west to northwest paleoflow direction and parallel to ridge orientation argues for an esker rather than a spit or bar. Bedforms within the ridge, mainly ripples and some dunes, are a product of unidirectional flow; no bidirectional forms, which may result from waves or tidal currents, were noticed. In addition, marine fauna (bivalves), common in littoral sediments on Victoria Island, were restricted to the upper few metres of reworked sediment in the ridge (V008, Fig. 4.8). Aeolian reworking probably explains the massive sand with roots found at the top of most sections. The ventifacts (e.g. Plate II) on the ridge surface also indicate aeolian reworking.

The most significant sedimentary attribute is the rhythmic character of most if not all of the ridge sediments. Each fining-upward sequence indicates waning flow, the succession of rhythmic beds indicates pulsating flow. Many fining-upward sequences are capped with clay indicating that

sediment transport came to a halt as clays fell from suspension. Lack of transport points to deposition in still water at the end of each rhythm. Still water may occur at the ice-margin, subglacially, or even supraglacially over short periods. The presence of thick clayey caps to rhythmic sets suggests the possibility of annual controls on sedimentation, and such annual cycles are common in esker sediments (Saunderson and Jopling, 1980; Hebrand and Amark, 1989).

The change in dominant sedimentary structure from cross-stratification to cross-lamination and the decrease in set thickness from east to west represents a bedform change from dunes to ripples in unidirectional flow. This trend is accompanied by a decrease in texture from east to west as shown by an increase in clay. The decrease in bed thickness and grain size from east to west indicates dissipation of energy along the flowline as a consequence of flow expansion westward.

Minimal disturbance of the ridge sediments rules out supraglacial and englacial deposition, and the absence of coarse sediment and sharp ridges makes a subglacial environment of deposition less likely. The solitary nature of the Augustus Hills ridge raises questions about the source of meltwater and sediment: the connection to the Cambridge Bay tunnel channel likely provided discharge and sediment and deposition probably relates to local flow expansion. The landform associations of the tunnel channel indicates subglacial, sheet flooding prior to channel and esker formation (Fig. 4.5).

The undulating nature of the ridge top and the coincident presence of a wider ridge (Fig. 4.7c) may mean that three main pulses of sedimentation (ice-margin

positions) produced the ridge; however, it more likely suggests deposition in a subglacial setting.

The lack of coarse sediment in extensive esker exposure, may be explained by burial in covered intervals. It is also possible that coarse sediment was not available for transport or was transported through the system at an earlier and higher energy stage of flow.

4.43 Ferguson Lake

Introduction

The Ferguson Lake esker complex consists of an orthogonal set of ridges and channels within streamlined terrain north of Ferguson Lake (Fig. 4.1). An east-west ridge, which leads into a glaciomarine fan to the west (Sharpe, 1988b), is flanked by thin marine sediments (Fig. 4.10). The main north-south esker complex displays a sinuous course and in part occupies a major channel system. Regional flow was to the southwest during landscape formation as indicated by striae, s-forms and crescentic scours (comma shapes) around elongate ridges (Shaw and Sharpe, 1987). Marine limit crosses the area from the northwest at about 150 m a.s.l., so marine processes modified surface form and obscured streamlined forms and parts of the esker complex in the south (Fig 4.10).

Landform associations

Associated landforms are important to understanding this esker complex; they consist of fluted and streamlined areas, channels and thermokarst lakes (Fig. 4.11a). A small group of transverse ridges is found down-flow of the streamlined forms (Fig. 4.10).

Channels underlie and are aligned with main esker ridges; thus, the tunnel channels are of subglacial origin.

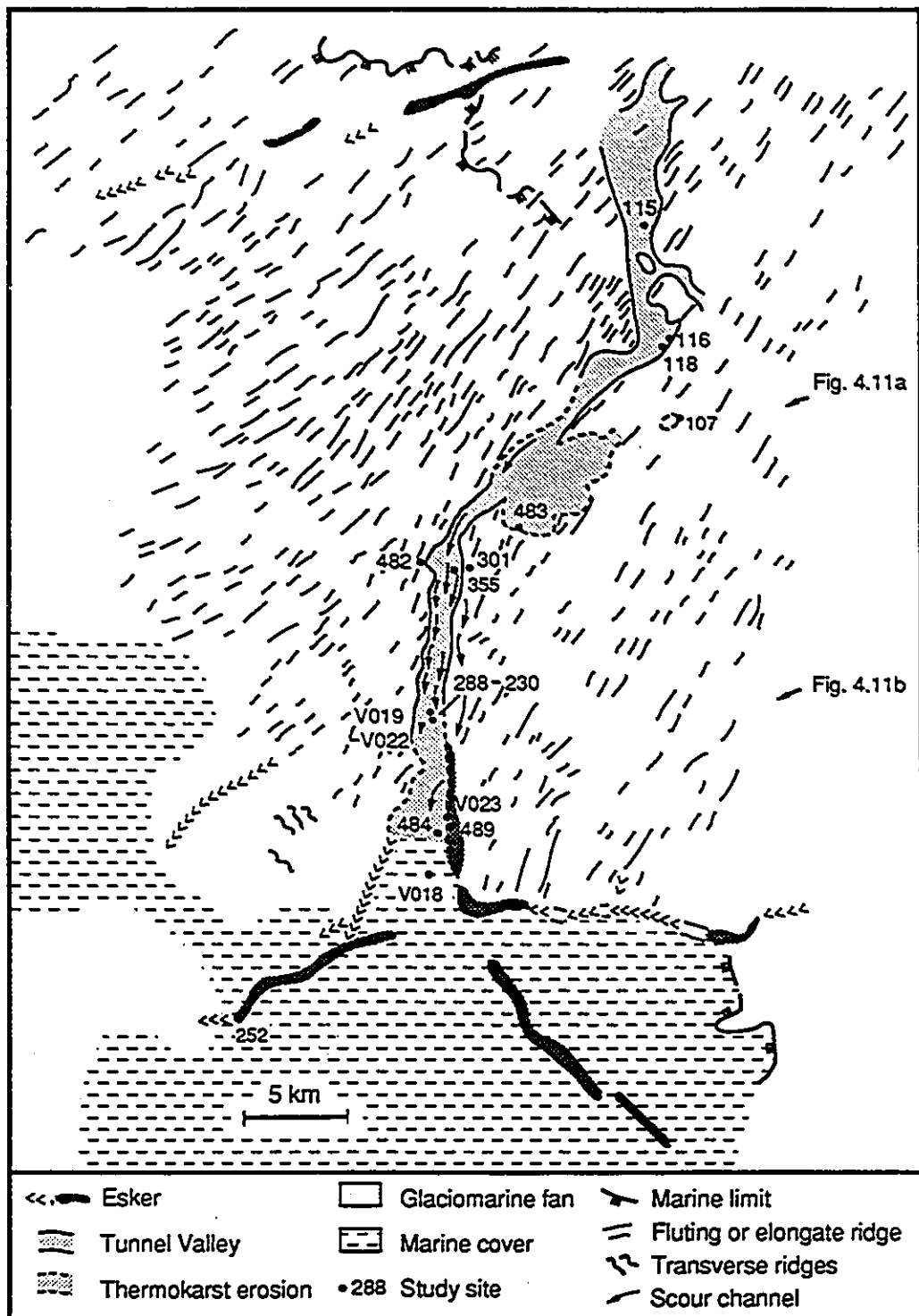


Figure 4.10. Map of North Ferguson Lake esker complex and related landforms.

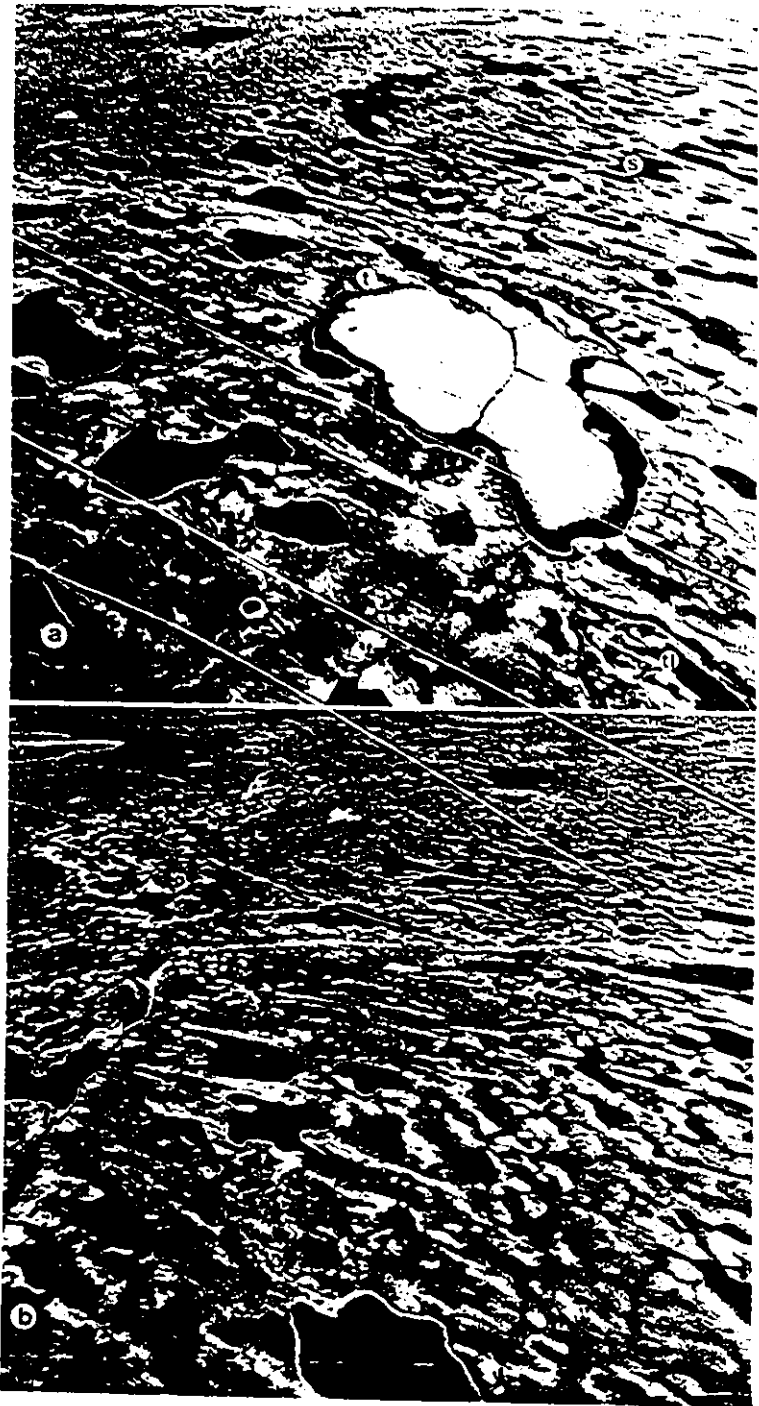


Figure 4.11. Aerial view of Ferguson Lake esker complex: a) North, note streamlined landforms (s) in background channel scours (arrows) thermokarst (-) eroded channel walls, (t) and thaw lake (tl); NAPL T466L-125, b) south, note glaciomarine fan main esker ridge (e) with no tunnel valley and (gf) complex. NAPL T466L-121.

The main channel is shallow, broad and sinuous to the north, but becomes narrower and straighter (with intricate sub-channels) to the south (Fig. 4.11a,b). Scalloped and enlarged margins (especially the west scarp) and large polygonal forms mark northern channels, whereas, esker ridges are common in the southern channel. The main channel appears to predate fluting (Fig. 4.11c) but later channel cutting also clearly post-dates fluting (Fig. 4.10, scour channels).

The esker complex is parallel or sub-parallel to the streamlined landforms, but overall, it cross-cuts the field at a low angle (Fig. 4.10). However, the esker system was apparently overridden by glacial ice, which also formed the flutings, according to Fyles (1963).

The relative age of crossing esker systems north of Ferguson Lake is complex but it is doubtful that the orthogonal systems operated simultaneously as there is little if any sediment at the juncture of crossing ridges or channels. The north-south esker complex appears to link with Lady Pelly and Mount Pelly ridges to the southeast (Fig. 4.1); the latter ridge was cross-cut by a tunnel channel (Fig. 4.7b). Flutings post-date formation of the north-south channel because they have eroded the south or southwest margin of tunnel channels (Fig. 5.11c). These flutings (1 km-2.5 km long; 100 m wide; 5 m-8 m high) start at the level of the tunnel channel and rise southwards to meet the level of adjacent fluted terrain. On the northeast side, the channel margin forms a sharp edge or step (Fig. 5.11c). These characteristics, added to crescentic scours, suggest an assemblage of landforms (flutings, tunnel channel and eskers) related to subglacial fluvial processes of landformation (Shaw, 1983; Shaw and Sharpe, 1987; Shaw, et al., 1989).

Landform associations show cross-cuts, but a transition

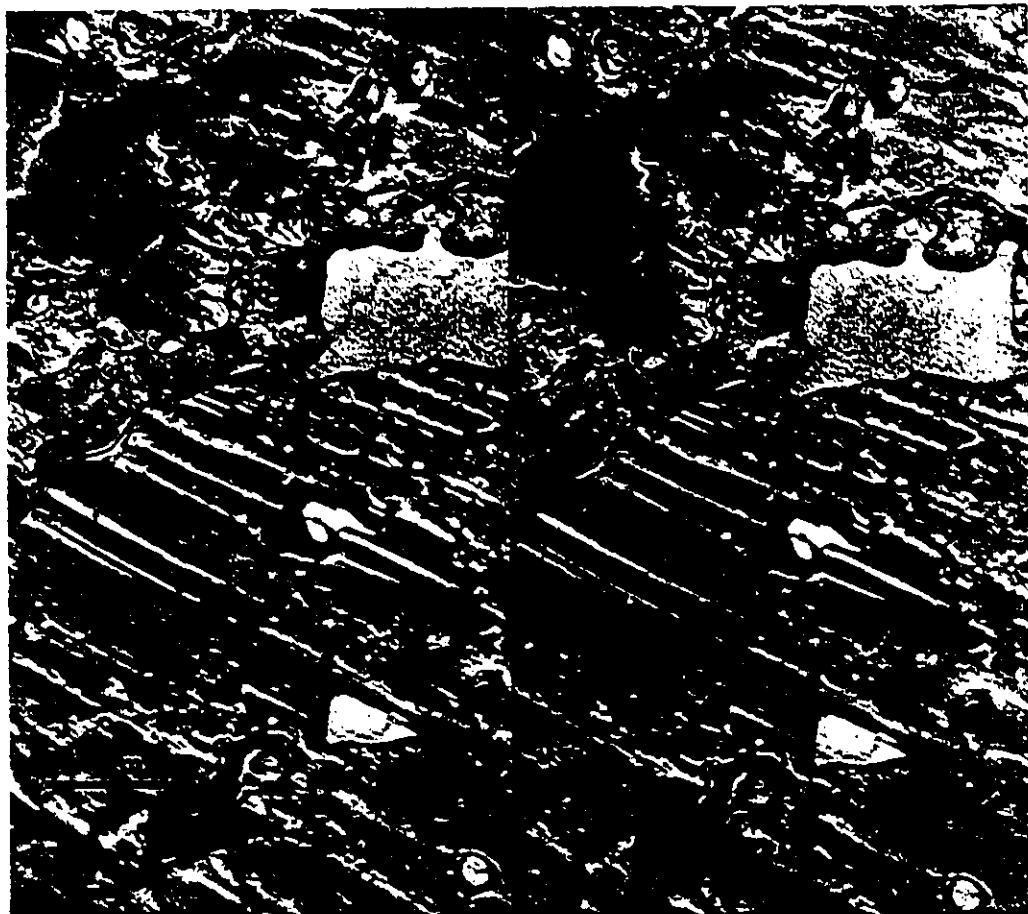


Figure 4.11 c) Fluted forms (f) on southwest margin of tunnel channel (tc). NAPL A16170-155.

in forms is also apparent as fluting adjacent to the esker complexes gives way to sinuous elongate ridges farther away (Figs. 4.10, 4.11a). Also, the main esker merges eastward with subdued streamlined forms. A sinuous streamlined form is contiguous with a small esker east of the main complex (SE, Fig. 4.11b) and sorted sandy sediments occur in these streamlined ridges (Sharpe, 1985).

Esker ridges form only a small portion (~ 20%) of the whole feature and channels are more common. A broad (1 km-2 km) gently sloping asymmetrical ridge occupies a 5 km portion

of the complex. The 25 m high ridge has gentle slopes to the east, and steeper western slopes with secondary cusate erosional scars. There are also small lateral ridges within the tunnel channel (site V018 and 484; Fig. 4.10).

Sections

More than fifteen measured sections (Fig 4.12) describe sediments of the main esker and small tributary ridges along the 40 km long esker complex (Fig. 4.1). It is unclear if all the sediment relates to ridge-building events; some measured sections apparently record sedimentation during the formation of adjacent fluted terrain (Fig 4.12c; sections 107, 301, 355, 482, 483 and sections 115 and 118, Sharpe, 1985). Sections clearly related to esker formation (Fig. 4.12a) are discussed first.

1. Main ridge

The longest section exposes about 50% of the exposed height of the 500 m wide main ridge. Most noticeable is the lack of sediment coarser than sand: there is virtually no sediment in the gravel range except for a surface lag (Fig. 4.12a; section 489). Rare limestones occur in sandy sequences. Most sediment is medium to fine sand which commonly grades up to mud (silty clay and clay). These fining-up sequences impart a rhythmic pattern to the sediments. Most contacts are gradational or conformable.

Coarse sand only occurs in occasional thick sets (20-30 cm; may be up to 1 m) of tabular cross-beds, found in the lowest measured intervals (Fig. 4.12a, sections 228, V019). Less common trough cross-bedded medium sand occurs in 25-100 cm thick cosets. Most common are thick cosets (1-2 m) of climbing ripple cross-lamination in fine sand, found stacked one on another. Angles of climb range from 5-20°; higher angles indicate increasing suspension sedimentation. Thick,

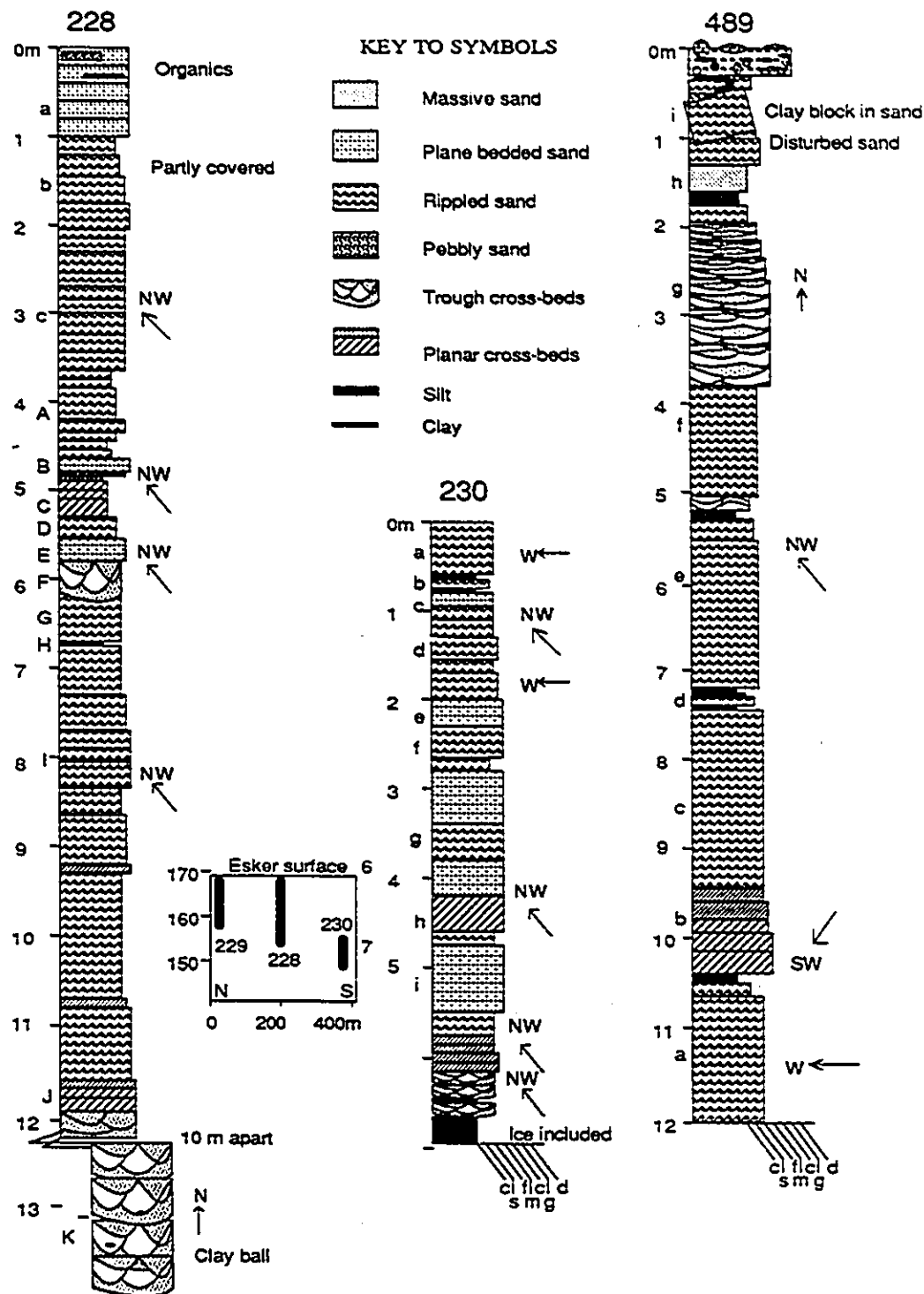


Figure 4.12. Measured sections along the North Ferguson Lake esker complex. a) within esker ridges, (228, 230 489). Diamicton at top of section 489; see legend figure 4.12 b.

1 m to 5 m packages of fine sediment mainly represent suspended load deposition (Fig. 4.12a, Section 489). In places, plane-bedded sand is found.

Little deformation of the sediment sequences was noted, except minor soft sediment deformation (e.g. ball and pillow structures, and convolute bedding in thin units e.g. 489; Fig. 4.12a) between undeformed beds. Soft, red and green shale fragments occur at the surface of the main esker ridge (section 489).

2. Lateral ridges

Lateral ridges are small single, double-crested or anastomosing esker ridges or remnants, 0.1 km to 2 km long, found west of the main esker and within wide tunnel channels (Fig. 4.10: 4.12b).

Thin diamicton beds cap sedimentary sequences of some small lateral ridges (484 and V018; Fig. 4.12b). Two distinct sequences occur: (i) Section 484 consists of undeformed, rhythmically bedded, fine to medium sand. Graded and massive beds are more common than rippled beds towards the top of the section. Minor soft sediment deformation is common in the forms of flame and ball and pillow structures. The rhythmic bedding and fine grain sizes of this sequence and the main esker are similar. (ii) Section V018 contains, coarser sediments, mainly pebbly sand. Reverse faults and other deformation structures indicate displacement towards a central depression in the ridge.

3. Channel margins and associated landforms

Sections on the margin of the main esker expose gravel, gravelly sand, and diamicton (sites 355, 483 and 482, Fig. 4.12c). Site 355 lies within a scour channel which may contain older esker or flute sediments. Tabular cross-beds are common and show paleoflow towards east to northeast.

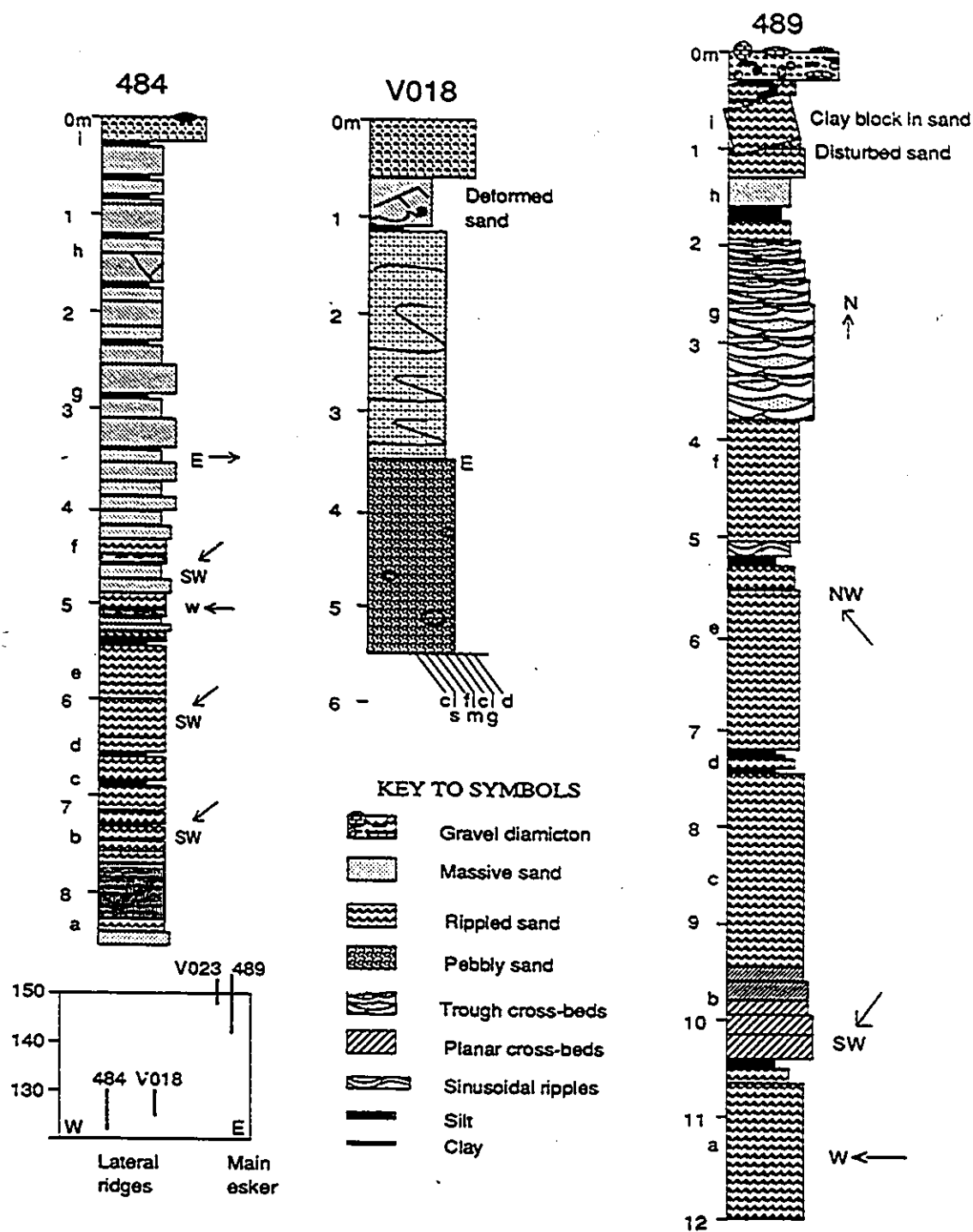


Figure 4.12 b. lateral ridges (484, V018), 489 for comparison. Elevation in landforms shown in m a.s.l.

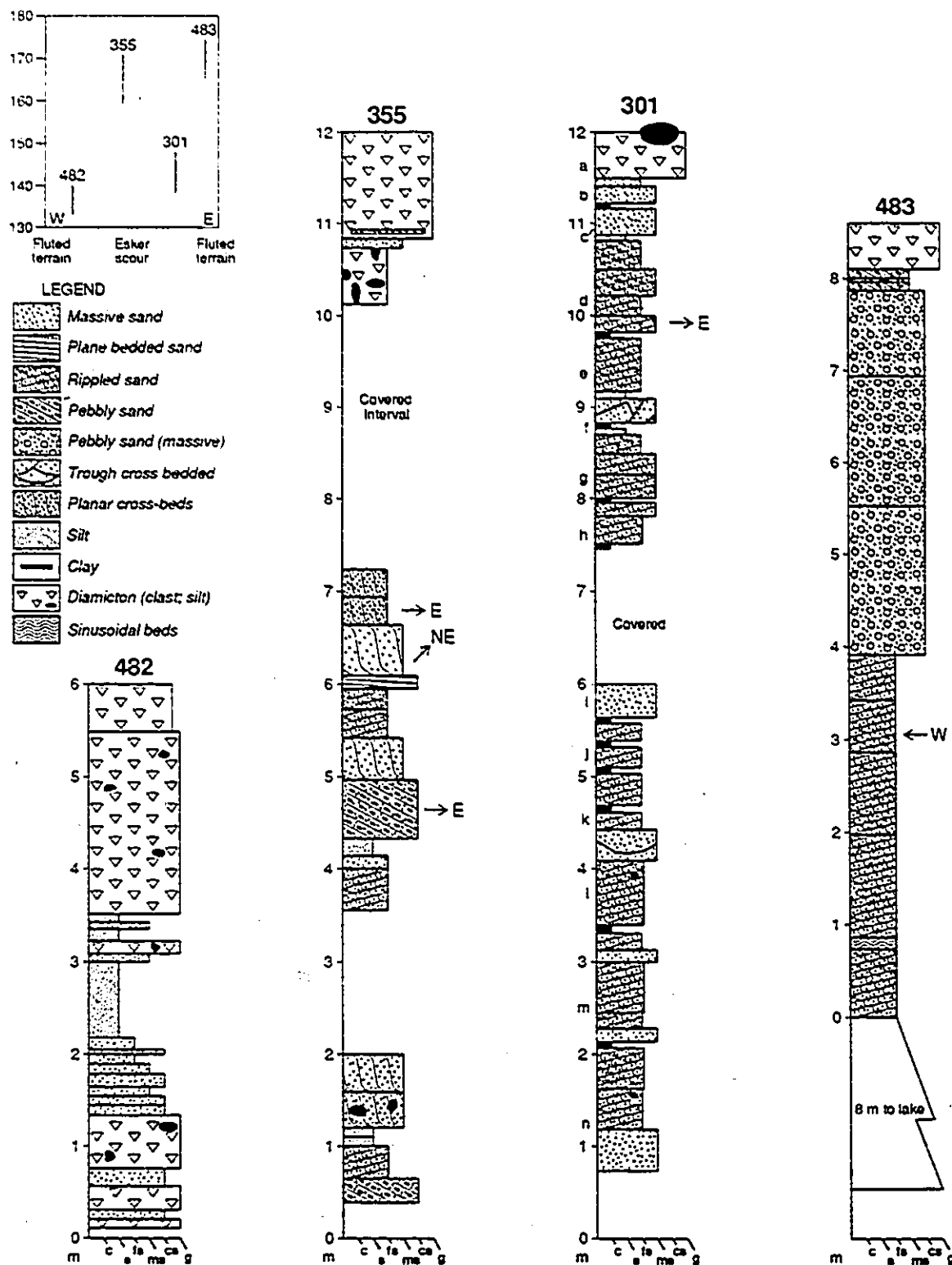


Figure 4.12 c. Other sections in esker complex (482, 355, 301, 483). Elevation in landforms shown in m a.s.l.

Sites 482 and 483 are clearly in streamlined terrain adjacent to the esker system. Diamicton and sorted beds occur as interbeds in the lower west ridge (482); cobbly gravel, cross-cutting underlying rippled sand, and diamicton occur within the higher fluted ridge (483). Ripples show westward paleoflow direction. Section 301 comprises 10 rhythmic sequences (0.2 m-1.5 m thick) of cross-laminated fine to medium sand, similar to those within esker ridges. Paleoflow is east towards a lake (scour?) depression. The landform and sediments at site 301 appear to be lobe or fan extensions to the esker. Paleoflow direction suggests lateral channel filling by an off-shoot from the main esker.

One section (107) within fluted terrain shows coarse sediment, with scour and fill structures, and considerable fault displacement (Fig. 4.13). Cobble and pebble gravel and pebbly sand at this site are the coarsest sediments in the Ferguson Lake area. Bed contacts are mainly erosional. Channel scours (with gravel lags) and cross-sets show paleoflow direction parallel to fluting.

The pattern of thermokarst lakes and channels within and marginal to the esker system indicates extensive ground ice beneath this fluted landscape (Fig. 4.11c). Ground ice may also be inferred beneath the fluted landscape from the faulted and displaced sediments on the margin of a drained thermokarst lake (Fig. 4.13). The drained lake displays lateral gullying above a terrace of eroded sediments. The sediments show vertical displacement of >6 m across a normal fault (Fig. 4.13b). Displacement occurred prior to deposition of the top, unfaulted 2 m of coarse sand, which shows weak bedding and dips to the southwest. Subsidence of more than 6 m of sediment is attributed to melting of ground ice from beneath more than 9 m of disturbed sand and gravel.

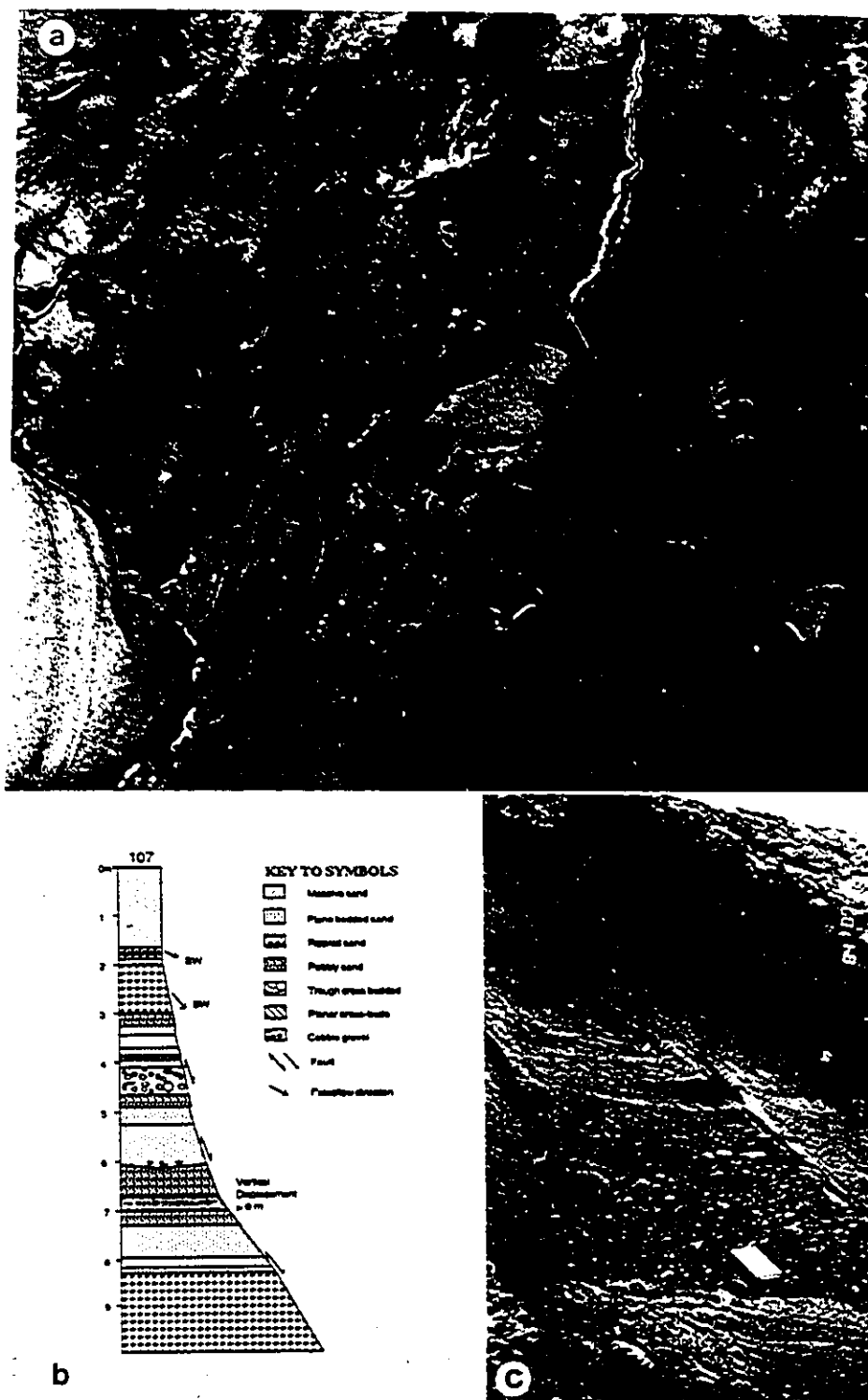


Figure 4.13 a. Thermokarst Lake (T.L) showing strandlines (s) formed during subsidence. Section 107 locates displaced sediments, in fluted terrain (f). b.) section measured on fluted terrain at thermokarst lake. c. photograph of faulted sediments. Notebook is 20 cm. GSC 1991-601A.

The drained thaw lake was about 15 m deep when completely full and >10 m deep when at the level of the highest lake terrace (s, Fig. 4.13a); these figures indicate that surface sediment loss of >10 m is required to induce thaw subsidence. Other sections in the area reveal thick (>10 m) of undisturbed sediment (115, 118, Sharpe, 1985); this indicates an erratic pattern of ground ice or a sufficient sediment cover to protect underlying ground ice from thermokarst erosion.

Paleoflow directions

Paleoflow directions in sediments marginal to the esker complex indicate powerful flow to the southwest, parallel to fluting trends (site 107, Fig. 4.13b). Other paleoflows however, are variable in marginal sites (Fig. 4.12c). Paleoflow data for all esker sections are variable: flow directions of southwest, west, northwest, and north are found at one site (section 489, Fig. 4.12a). These paleoflows are progressive up section from west, southwest to northerly. Section 228 shows a consistent north to northwestward paleoflow. Adjacent sections have west to northwest paleoflows. These paleoflow directions are contrary to those inferred from regional geomorphological data (streamlined terrain, striae, s-forms) which indicate southwest paleoflow. Sediments in small lateral or remnant ridges show west to southwest paleoflow trends.

Interpretation

Esker sediments predominantly indicate low-energy environments typical of glaciolacustrine sedimentation. The maximum grain-size is coarse sand and most sections show rhythmic bedding. Each sequence fines upward from medium or fine sand to silt or clay. The structure (cross-beds to cross-lamination to drapes) within each rhythm also

represents a waning flow regime, except for minor variations in velocity or sediment supply. Climbing ripples and occasional sinusoidal ripples mark the increase and dominance of suspension load deposition over bedload deposition, which when all currents ceased, give way to clay deposition. Such sequences, while common in eskers (Ashley, 1985), are thought to imply low-energy.

The relatively low energy conditions inferred from these sections (Fig. 4.12a) may be explained either as lateral deposition within a subglacial conduit during waning or low discharge events (Hebrand and Åmark, 1989; Gorrell and Shaw, 1991), or as the product of subaqueous deposition within an ice-walled open channel (Saunderson and Jopling, 1980). Although no extensive faulting was observed within the exposed sediments, this does not preclude sediment disturbance within the main esker ridge, since exposures are scarce. Slow melting of ground ice and permafrost conditions, however, could explain minor sediment disturbance. The rhythmic bedding of the sedimentary packages suggests some seasonal control on sedimentation, and thus a supraglacial to subglacial connection in the meltwater system. Such a connection is expected where ice was thin (Clayton and Moran, 1974; Hooke, 1984; Gustavson and Boothroyd, 1987), and thin ice was suggested for the area by Fyles (1963) and Sharpe (1988a).

The small size of the lateral ridges and their position within wide tunnel channels suggests that they involved late-stage glaciofluvial deposition at this site. The coarse pebbly sand and thick beds at one site are of note because they indicate that higher energy flow in a confined conduit followed low-energy deposition inferred for the main ridge. Deposition apparently occurred on or against ice, which lead

to pervasive faulting as melt took place (McDonald and Shilts, 1975), although lateral support was removed from some ridges with only minor disturbance. Periglacial deformation from ground ice activity may also account for the faulted ridges (French and Dutkiewicz, 1976).

Alternatively, the coarse section (V018, Fig. 4.12b) may represent a remnant ridge, as it lies well below the main esker ridge, where coarse core sediments may be buried and the paleoflow trend parallels that of fluting. However, the section lies within scoured terrain which has no surface lag deposits and coarse sediment is rarely found anywhere in the esker complex (e.g. 352, east-west esker; Fig. 4.10).

Erosional features

The scour channels and other scour features (comma lakes, Fig. 4.11c) indicate that powerful erosive flows occurred prior to low-energy sedimentation in the main esker ridge. The channels and depressions adjacent to the esker ridges are thus tunnel channels (tunnel valleys of Wright, 1973). The landform association, streamlined terrain, tunnel channels, and eskers strongly suggests widespread meltwater flow (Shaw, 1983; Shaw and Kvill, 1984; Shaw and Sharpe, 1987; Shaw et al., 1989). Thus, a subglacial meltwater sheet may be invoked to explain the streamlined forms (Shaw and Sharpe, 1987; Shaw et al., 1989; Kor et al., 1991). The arcuate geometry of the channel margins may also relate to erosive action of meltwater bursts during early stages of esker activity. However, the enhanced scalloping of the west tunnel wall requires meltwater flow that underwent flow separation in a pre-existing channel (Brennand and Sharpe, in prep). The development of vortices within this channel may have sculpted the channel wall and left remnant ridges as flutings on the upper surface (Fig. 4.11c). With a decline

in discharge, the meltwater would have been confined in the tunnel channel.

The geometry of these channels indicates further erosion by thermokarst processes, but the manner in which thermokarst erosion took place is not clear because the sediment walls are not ice-rich. During the late stages of meltwater flow in the esker conduit, thin marginal ice was not able to protect sub-ice ground from freezing and these marginal, permafrost sediments were eroded by meltwater. The presence of extensive ground ice may be inferred from various landform associations (Figs. 2.10; 4.13) and this is of importance to the sequence of events at Ferguson Lake.

Variable paleocurrent directions are typical for fine sandy cross-laminated sequences ($\sim 270^\circ$) (Saunderson, 1975). Reverse flow in eddies or gyres, (some regressive ripples are noted), distributary flow over lobes, and changing flow with time may explain the variability of flow directions. It is also possible that north and northwest flow may relate to a former link with Lady Pelly and Mount Pelly esker segments and flow from the southeast. This creates complications as northward flow occurs at the top of the esker sequence (Fig. 4.11), and Lady and Mount Pelly ridges predate and were later eroded by east to west streamlining and tunnel channel flow (Fig. 4.7b). A simpler (?) explanation is to suggest that earlier flow in esker conduits was to the southwest (paleocurrent data and regional flow indicators), and that enlarged conduits in thin marginal ice were back-filled by low-energy sedimentation from the southeast, driven by a regional gradient of the Keewatin Ice (Fig. 1.1).

Discussion

Sections adjacent to the esker system (Fig. 4.12b, 284 and Fig. 4.12c, 301), but not clearly within the esker, also

show rhythmic ripple sequences with clay caps. If these are all early esker sediments, then low-energy sedimentation was the rule during deglaciation and esker formation, interrupted by high-energy scouring events. Rhythmic sedimentation indicates a seasonal control on meltwater input and a probable contribution of supraglacial meltwater during this stage of deglaciation.

It is also probable that meltwater cut upward into the ice at this time, forming a combined Nye-Röthlisberger channel (Hooke, 1988). As meltwater discharge declined further, and there was insufficient power to erode the basal substrate, meltwater would have occupied simple Röthlisberger-channels. It was into these that the esker sediments were deposited. Finally, ice stagnated and smaller glaciofluvial ridges were deposited in cracks in the ice.

4.5 Glacial hydrology

A brief review of glacier hydrology with respect to eskers is necessary to help place these observations on esker form and sediments in a regional perspective.

Glacial meltwater has three source areas: supraglacial (~90%), englacial (low) and subglacial, all of which contribute to the hydrologic regime of the glacier (Paterson, 1983). The ways in which this water are known to move through the glacier contribute to our ability to interpret eskers and other glaciofluvial landforms.

Supraglacial water may flow in sheets, but these are unstable and rapidly develop into a network of rivulets and channels ending abruptly in crevasses or moulins (Shreve, 1972). Much like a karst system these may transmit water rapidly to the bed, particularly in crevassed glaciers. Other englacial water moves in a network of veins at ice-

grain boundaries linked to passages (Nye and Frank, 1973). Passages may also intercept surface water, but either way, they tend to grow in a tree-like system of conduits connected to the basal drainage. Drainage is normal to equipotential lines which dip up-glacier relative to the ice surface at $1/11$ th the slope of the ice surface (Shreve, 1972). Although subglacial drainage is generally more complex due to the effects of sediment load and bed topography, it facilitates research by preserving a permanent record in the forms of tunnel channels and eskers. Flow follows the slope of the potential surface (slope of the ice) but bed topography becomes important when ice slopes are low.

Steady-state subglacial discharge moves in four ways: R-channels (Röthlisberger, 1972) cut up into the ice; N-channels cut into the sub-ice bed (Nye, 1973); thin films form under great pressure (Weertman, 1973); and, linked cavities develop from moderate pressure (Kamb, 1987). R-channels form as a result of the balance between the melting of passageways by flowing water and closure of those passages by plastic ice flow. Pressure in the channels is inversely related to discharge so that the larger channels have the lowest pressure. This means that flow in smaller channels under higher pressure is captured by larger channels under lower pressure, creating an integrated, arborescent network. Eskers are deposited in R-channels. N-channels are formed when melting (at least rapid melt) does not occur and water pressure is greater than ice pressure, and, when discharge varies directly with pressure. N-channels form the common landform known as tunnel channels or valleys. Films form when passageways are not connected and when subglacial flow is low. They are unstable when they are less than 4 mm in thickness. Linked cavities form tortuous routes when ice

profiles are low and slope towards the margin. They are under high pressure that varies directly with discharge and with increased pressure the linked cavities are replaced by straight channels. Extreme, non steady-state flows, which cannot be contained in these channels, produce hydraulic-jacking in the ice (Bindshalder, 1983) so that broad flows, possibly sheet flows, flood the glacier bed (Nye, 1976). Such episodic flows may have been important on Victoria Island and observations presented in this thesis may help guide the evolving theory (e.g. Shoemaker, 1991, 1992a,b).

4.6 Discussion of southern Victoria Island eskers

4.6.1 Setting, regional features and controls

Esker ridges currently situated within scoured terrain or channels. Such channels relate to earlier, stronger flows that were primarily erosional. The associated features (stripped bedrock, scour channels, rock drumlins and transverse bedforms) form wide landscapes that suggest broad powerful erosional flows (Fig. 4.1). To remove sediment of slightly finer size than the coarse channel lags (~1 m) requires flow velocities in the order of 10 m/s (Elfström, 1987). Therefore, violent meltwater discharges must have modified the landscape prior to the meltwater flows recorded by esker sediments.

Thus, scour zones, channels and tunnel valleys record early and powerful drainage events. These erosional features mark the release of stored meltwater which could not have been carried by smaller channels now filled with sediment. These floods probably occurred as jökulhlaups (Nye, 1976; Shaw et al., 1989), or as outbursts following the disruption of subglacial drainage by surging (Kamb et al., 1985).

Eskers on southern Victoria Island are arranged in large arborescent networks set in major topographic lows (particularly Read Island and Banning Lake; Fig. 4.1). They represent a fully integrated glacier drainage system. It may be argued that, if ice was thick, such integration developed supraglacially from a surface drainage network (Alysworth and Shilts, 1989b). But the eskers formed at a late stage when ablation produced thin, possibly stagnating ice, and in the absence of restrictive tunnel closure, lengthy subglacial esker systems developed. In addition, esker follow the subglacial erosional (channel) network that were already in place. Consequently, although they depended on supraglacial drainage for their discharge, the esker tunnel pattern was controlled by preexisting subglacial channels that may well have influenced supraglacial drainage through its control on moulins. Crevasses and moulins are best developed during and after surges which cause glacier thinning and longitudinal extension (Clarke, 1987). Surging, in turn, has been linked to subglacial flooding (Kamb et al., 1985) and sheet floods (Shaw and Sharpe, 1987; Shaw et al., 1989; Sharpe and Shaw, 1989).

Evidence that the Banning Lake eskers developed subglacially is given by the eastward paleoflows for the sharp-crested eskers, towards the main north-south trunk esker (Fig. 4.6a). The lack of deformation in observed sediments (Fig. 4.12) also rules against supraglacial formation of eskers. The network of eskers at Banning Lake also represents a connected drainage system with a low pressure gradient, probably related to slow-moving or stagnant ice. This ice may have wasted in place following fast-glacier flow in a surge or an ice stream which produced a thinner and flatter ice sheet.

4.6.2 Sediments

The rhythmic sequences in many of the esker sedimentary successions indicate strongly rhythmic meltwater discharge. Annual cycles of glacial meltwater result mainly from seasonal variation in supraglacial meltwater. Consequently, the dominant rhythmicity in esker sediments is probably seasonal (Banerjee and MacDonald, 1975; Hebrand and Åmark, 1989). Esker spacing in low terrain is too wide to drain the subglacial water of the catchment beneath thick active ice, so that supraglacial sources were more important. In addition basal melt was probably small compared to surface melt. Meltwater however, may be released from basal storage as the seasonal melt progresses (Paterson, 1983); this adds to the flow variation caused by surface meltwater supply. The broad pattern of scour zones and exposed bedrock may attest to the sudden release of such stored subglacial meltwater prior to the formation of the eskers.

The predominant sediment within most esker sequences described in this thesis is fine sand with caps of silty sand and clay. These sediments are found in esker ridges elsewhere (Saunderson, 1975), lacustrine deltas (Ashley et., 1985), and subaqueous fan ridges (Rust and Romanelli, 1975; Rust, 1977, 1988; Cheel, 1982). They clearly have more affinities with standing bodies of water than subaerial streams because graded units, massive sand, ripple-drift sequences (climbing and sinusoidal ripples), and, silt and clay are mainly deposited from suspension. Such fine sandy sediments are considered to be deposited in ice-marginal, deltas or subaqueous fans, but not in closed-tunnels (Aario, 1972; Saunderson, 1975; Hebrand and Åmark, 1989). Deposition may have been in partially-filled, subglacial conduits (Saunderson, 1975) but presumably this requires a stagnant

ice mass to prevent tunnel closure. The cutting of tunnel channels may have produced wide tunnels in which there was deposition of broad, fine-textured eskers and which weakened the ice causing subsequent roof collapse and open-channel sedimentation.

4.6.3 Esker-glacier reconstructions

The cross-sectional geometry of eskers may be used to reconstruct ice surface profiles (Shreve, 1985a.b), assuming a steady state, full pipe flow. A limited heat supply for melting the steep sidewalls of ice tunnels is thought to produce broad-crested ridges (Shreve, 1985a). The level terrain and low esker gradients are evidence against esker formation beneath thick, high-profile ice sheets on southern Victoria Island. Broad, low-gradient eskers relate to low ice profiles, particularly if they were deposited subaerially or supported in open ice-walled passageways. Meltwater moved at low velocities, producing low-flow-regime bedforms and fine sediment.

Thin ice and advancing permafrost conditions may have caused interruption or damming of flow to modify the esker system in thick drift areas such as Ferguson Lake. This may account for multiple flow systems in such areas.

From the lack of gravel, it appears that esker sediment transport and deposition were restricted to grades finer than coarse sand. This raises the question of what happened to the coarse component of the till which served as a source for the glaciofluvial sediment.

4.7 Summary

Scoured bedrock zones that flank eskers on southern Victoria Island (Fig. 4.1) are the equivalent of tunnel

valleys cut into thick sediment areas such as Ferguson Lake (Fig. 4.10). Scour zones contain s-forms, transverse forms, lags and rock drumlins (Fig. 4.5) which allow links to be made to the streamlined terrain commonly associated with esker networks. Thus, esker systems develop from early erosional phases having common origin with drumlins which may also have resulted from glaciofluvial activity.

Esker ridges may be broad or narrow. Narrow ridges parallel streamlined forms and are sharp-crested and smaller than broad ridges. Broad ridges are composite forms located at the margins of lobate landform sets and former ice-lobe boundaries (Fig. 4.1).

Broad ridges contain planar sets of rhythmic fine-grained sediments with little gravel. They represent deposition under atmospheric conditions primarily from supraglacial meltwater. Poorly exposed narrow ridges, which have coarse gravel to bouldery surfaces and interiors, are interpreted to have been deposited subglacially.

Broad ridges are contiguous with narrow ridges (e.g. Augustus Hills/ Cambridge Bay esker, Fig. 4.7b) and indicate the discontinuous or restricted nature of open channel conditions during regional melt. Broad esker segments may represent broad conduits in the ice, inherited from earlier, more vigorous (erosional) flows; the broad eskers represent a waning flow deposit. The record of earlier powerful flows is either buried or these flows did not deposit sediment in the broad ridges.

Some eskers near Ferguson Lake show links with permafrost where thermokarst enlargement of tunnel valley margins has occurred. Thinning ice and marginal cold zones interrupted esker drainage during final decay by regional stagnation.

5.0 QUATERNARY HISTORY

5.1 Events prior to last glaciation

Little is known of events predating the last glaciation of Wollaston Peninsula. Ice apparently covered all but northwest Victoria Island in Early Wisconsinan time (Vincent, 1989). Vincent (1982, 1984) correlated till from Banks Island and argued that the last ice did not cover all of Wollaston Peninsula. However, it appears more likely that Late Wisconsinan ice covered Wollaston Peninsula, but did not cover all of Diamond Jenness Peninsula (Fig. 1.1b) to the north (Vincent, 1989) where middle Wisconsinan or older deposits lie at the surface.

Fyles (1963) reported organic-bearing sediments underlying probable Late Wisconsinan till on adjacent Diamond Jenness Peninsula north of Wollaston Peninsula. Matted leaves and herbaceous plants were dated >32 000 400 BP (GSC-388, Blake, 1974), >37 000 BP and >38 000 BP (GSC-3613 and GSC-3592; Vincent, 1989). The organics were considered to be interstadial, or of last interglacial or older age based on their apparent similarities to present-day vegetation (Fyles, 1963). Vincent (1989) suggested that the organics may be Middle Wisconsinan based on an Early Wisconsinan interpretation for the age of the underlying red till.

On Wollaston Peninsula (Kugaluk River), disseminated organics (small fragments of wood), within sand at 95 m a.s.l. underlying a surface till, have been dated at 45 100 years BP (TO-95; accelerator, mass spectrometer, Appendix VII). Although this may indicate that nonglacial conditions were present on Wollaston Peninsula about 45 000 years ago,

the sandy sediments themselves may not have been deposited at that time. It is unlikely that shallow lacustrine or fluvial sediments would have been deposited in thick sequences in this setting at 95 m a.s.l. during a nonglacial interval. It seems more likely that the organic material was reworked by glaciofluvial erosion and redeposited prior to the late Wisconsinan glacial advance which deposited the overlying till. Accelerator dates of this antiquity are suspect and may give only minimum ages.

The dated wood fragments are spruce, a very unlikely species to be growing on Victoria Island during Pleistocene times (J. Matthews, pers. comm., 1986). Tree fragments have not been found in the organic beds from Diamond Jenness Peninsula or in the interglacial beds on Banks Island (Matthews et al., 1986). Among several possible explanations for this anomaly, three seem particularly plausible: 1) Pleistocene organic beds on Wollaston Peninsula or up ice off the area may have been glacially eroded and transported; 2) Spruce wood may have floated across Dolphin and Union Strait during a high sea stand; or 3) aeolian transport by winter storms may have carried organic fragments into the area, during periods of little or no snow cover, prior to Late Wisconsinan ice occupation. In any case, the underlying glacial diamicton, resting on bedrock at the Kugaluk River (Fig. 3.2) section, probably represents glacial conditions predating a non-glacial interval. Glacial deposits in a similar stratigraphic position are found in a section near Cape Baring. These two sections in western Wollaston Peninsula (ice flowed from the southeast) give evidence for complete ice cover of Wollaston Peninsula at least once, and possibly twice. Evidence for three full glaciations is preserved (down ice-flow) on Banks Island (Vincent, 1989).

The glacial deposits above the dated organics have been traced by surface mapping and are clearly late Wisconsinan in age (Sharpe, 1984 and following section). This mapping supports the concept that Late Wisconsinan ice covered all of Wollaston Peninsula. The implication is that Late Wisconsinan ice limits lay 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). Lower glacial deposits, depending on the actual age of the recovered wood, may indicate Early Wisconsinan ice cover too.

5.2 Late Wisconsinan glacial activity

Discussion on ice sheet growth and the limits of the last glacial maximum is most reliable when based on accurate ice limits, ice flow patterns related to ice build-up, and an understanding of deglaciation. The extent of the last glaciers on Victoria Island and the northwestern margin of the Laurentide ice sheet have been in debate. Briefly, whereas Vincent (1982, 1984) and Dyke et al. (1982) argued for restricted Late Wisconsinan ice on western Victoria Island, Sharpe (1984) and later Dyke (1987) and Vincent (1989) argued for full ice cover of the island, extending to eastern Banks Island.

The spatial relationship of landforms outlined earlier helps establish the limits of glacial activity during late Wisconsinan time (Fig. 5.1). The landform pattern is continuous, gradational and seems to have formed in association with a single, large ice-stream which moved from east to west along Coronation Gulf and Dolphin and Union Strait (Sharpe, 1984 and 1988a). Ice also occupied Prince Albert Sound but apparently did not produce a similar

landform complex there because of strong topographic control by the bedrock scarps on the north side of Wollaston Peninsula. Alternatively, portions of a landform set may be submerged in the sound. The strongly-shaped topography of the Kagloryuak River valley suggests ice flow similar to that to the south. Therefore, an ice stream may have split and flowed as two streams around the central plateau. Thus, the regional ice flow from the southeast was redirected westwards along the major marine troughs (Coronation Gulf, Dolpin and Union Strait, and Prince Albert Sound). No evidence exists, however, for major east to west flow across Victoria Island itself as proposed by Dyke (1984).

The glacial limit during Late Wisconsinan time is based on the concept of full ice cover on Wollaston Peninsula (Sharpe, 1984). At 12 000 year BP ice extended well into Amundsen Gulf (Fig. 5.1), but its exact margin is difficult to define from information on Wollaston Peninsula alone. Instead, landform patterns south of Dolphin and Union Strait were used to determine this limit (Sharpe, 1984; St-Onge, 1987). The marine limit at Cape Baring (115 m) implies an approximate ice thickness of 350 m (based on rock/ice density ratios of 3:1 and no eustatic effect (Andrews, 1970). If the ice profile was parabolic (mean slope, 0.3 m/km) its position would fall close to the minimum limit of Vincent (1989). A lower ice profile (mean slope, 0.1 m/km), possibly occurring as an ice shelf or a surged ice margin zone, would have extended somewhat beyond that shown (Figs. 2.7 and 5.1). Hodgson and Vincent (1984) suggest there was a late glacial ice-shelf in Viscount Melville Sound, adjacent to northern Victoria Island. Ice shelves generally form at the margins of polar or cold-based glaciers (Meier and Post, 1987) rather than warm-based glaciers. Streamlined landforms and warm

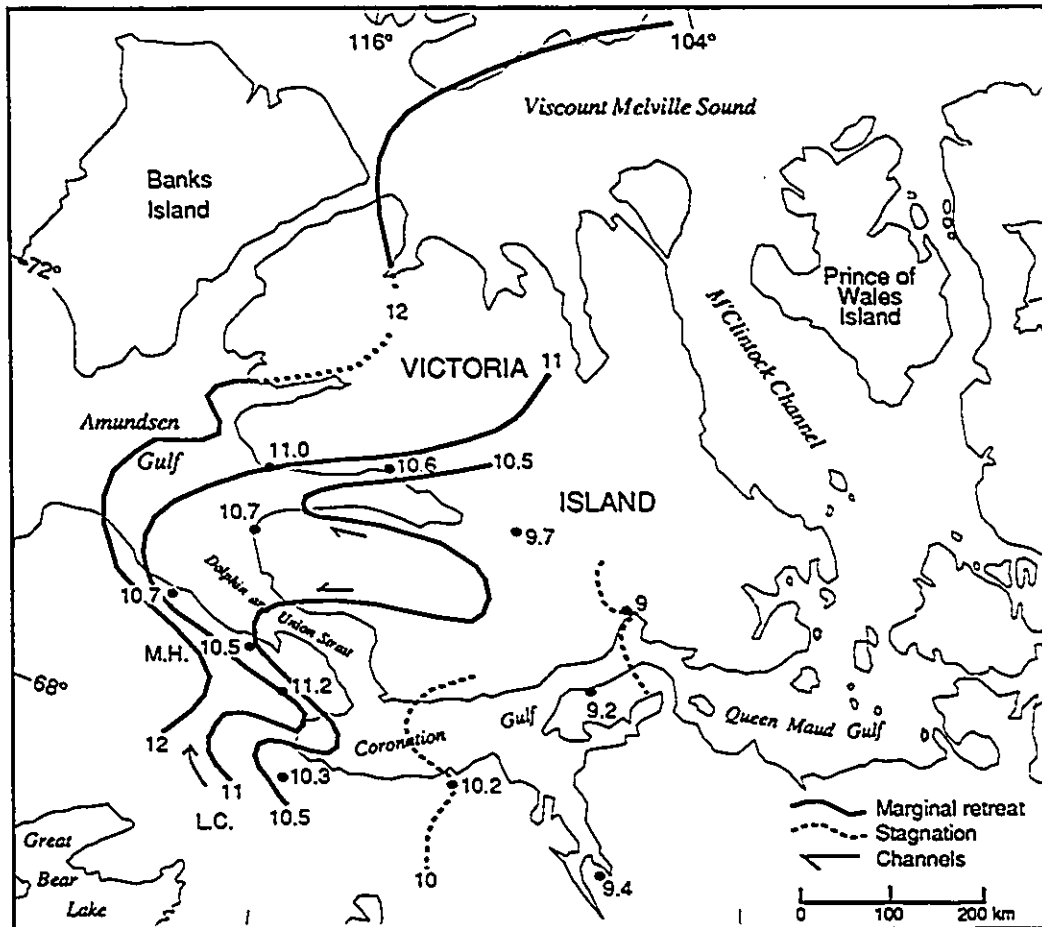


Figure 5.1. Approximate ice-marginal position on southwestern Victoria Island during marginal retreat (solid lines). Dashed lines are vague ice margins during regional stagnation. Time lines approximate in 1000 years B.P.

basal conditions were likely dominant on Wollaston Peninsula except during the latest glacial time when marginal zones were likely frozen. For this reason extensive ice-shelves are not inferred off Wollaston Peninsula.

An ice limit of >500 m a.s.l. was suggested for the Melville Hills region on the mainland southwest of the study area (Sharpe, 1984). It was based on relating the distribution of streamlined landforms, drift distribution and ice-marginal drainage features to the similar landform set on Victoria Island (Sharpe, 1984; 1988). This limit has been confirmed (St-Onge and McMartin, 1987; St-Onge, 1988;), but dating control is still required to establish the ice limits as Late Wisconsinan.

Ice-flow directional indicators, drift dispersal, oriented landforms, and striations allow one to reconstruct flow patterns. Landform patterns and, to some extent, striations may be demonstrated as products of ice flow during the last deglaciation, although their patterns may be better models for initial ice flow conditions. However, without dating control on landforms, only relative ages, determined from cross-cutting relationships, can be discussed with assurance (cf. Lagerbäck, 1988). Solution weathering of carbonate bedrock limits collection of striation data. Streamlined landforms preserved on Victoria Island are considered to be late events formed during the waning stages of glaciation rather than formed as initial last glaciation flow patterns (Fyles, 1963; Sharpe, 1985, 1988a; Hodgson, 1987) on Victoria Island.

Drift dispersal patterns based on the distribution of Precambrian erratics show ice flow from the southeast. Local dispersal patterns (cf. carbonate dilution or zone of low values; Fig. 2.2), in the lee of Precambrian inliers, also

show a southeast to northwest trend. This presumably reflects an ice-dispersal center on the Keewatin Ice Divide (Lee, 1959; Prest et al., 1969; Prest, 1984). As discussed above, streamlined patterns run east-west in the region of northwest drift dispersal, which may support the view that the drumlins represent later erosional events following deposition of basal diamicton.

Fyles (1963) also suggested a dominant southeast to northwest last glacial ice movement across Victoria Island based on regional ice-flow indicators. He supported this ice-flow history by reference to the general northwest to southeast pattern of deglaciation and increasing elevation of raised marine features (to the southeast) related to postglacial emergence. Deleveling trends on the mainland to the south also indicate a northwest flow pattern (Andrews, 1987).

A predominant east to west ice-flow pattern was later portrayed by Dyke (1984) as part of his hypothesis for an ice centre or dome in M'Clintock Channel. The reasoning for the M'Clintock dome depended on an anomalous pattern of isobases and a west to east ice-flow pattern across Prince of Wales Island to Boothia Peninsula. Dyke (1984) used streamlined landforms on Victoria Island to propose ice flow (of inferred long duration) trending westward from his proposed M'Clintock Dome and crossing Victoria Island. He considered these features indicating westward flow to be older (based on apparent cross-cutting relationships) than other indicators, for example the northward to northwestward flow patterns on Prince of Wales Island (Fig. 3.1). Dyke assigned westward flow to the last Laurentide advance stage, but northwest ice flow to the deglaciation phase.

If deglaciation effects are slight, features relating

to glacial advance may be preserved in the landscape. But it is also possible that the majority of landforms and deposits are formed during deglaciation. Fyles (1963) considered that the complex pattern of streamlined landforms on Victoria Island related to changes of ice-flow direction during thinning (or retreat) of the ice-sheet. This view was supported by studies of cross-cutting landform relationships in northwestern Victoria Island (Hodgson, 1987) and in eastern and southwestern Victoria Island (Sharpe, 1985; Sharpe, 1988a). Dyke and Morris (1988) also proposed multiple ages for streamlined flow patterns on Prince of Wales Island to the east (Fig. 3.1).

The concept of long-term, last glacial ice flow towards the west from M'Clintock Channel as proposed by Dyke (1984) and Dyke and Prest (1987) does not fit geological evidence from Victoria Island for the following reasons:

1. Sustained flow from M'Clintock Channel does not explain the widespread and abundant distribution of Precambrian erratics in the youngest drift of Victoria Island, unless they were recycled from earlier advances.
2. East to west flow patterns in Prince Albert Sound area occurred because ice, derived from flow across Kent Peninsula (Keewatin Ice Divide), streamed around Wollaston Peninsula.
3. East to west flow patterns on eastern Victoria Island resulted from late flow of thick ice that resided in the marine troughs. Such late flow patterns derived from similar thick ice masses situated in Coronation Gulf and Dolphin and Union Strait (Fig. 3.1a).
4. Almost all surface landforms (streamlined forms) appear to be related to late deglacial flows, caused by ice streaming and/or late-glacial floods (Sharpe, 1988a).

5. High elevations of postglacial emergence in the Bathurst Inlet-Kent Peninsula area are related to thick ice (that in part straddled Coronation Gulf, see 3 above) rather than early encroachment of the sea as proposed by Dyke and Dredge (1989).
6. Postglacial tectonics (Dyke et al., 1990) adequately explain the anomalous uplift pattern in M'Clintock Channel, nullifying the need for M'Clintock Dome (Dyke and Prest, 1987).

5.3 Deglaciation

Landforms on southern Victoria Island, important for determining ice limits, also allow mapping of Late Wisconsinan deglaciation. The pattern of deglaciation for Wollaston Peninsula, which can be clearly traced (Figs. 5.2a-d, 5.3), is summarized (Fig. 5.1) so as to illustrate its relationship with adjacent areas. This deglaciation history is supported by McMartin and St-Onge (1989) and by Vincent (1989), who earlier had suggested incomplete ice cover of Wollaston Peninsula (Vincent, 1984).

The pattern and style of deglaciation on Wollaston Peninsula are marked by initial ice-marginal retreat, followed by marginal stagnation, surging and regional stagnation (Sharpe, 1988a). In addition to the arrangement of the landforms several lines of evidence are available to suggest this pattern. The age relationships of the landforms on Wollaston Peninsula indicate formation within Late Wisconsinan time as defined by the date of marine submergence (about 11,000 years, e.g. 10 700, BP; GSC-3566) on northwestern Wollaston (Cape Baring) and a minimum date (~9 360 BP; GSC-4622) on the Colville moraine (Appendix VI).

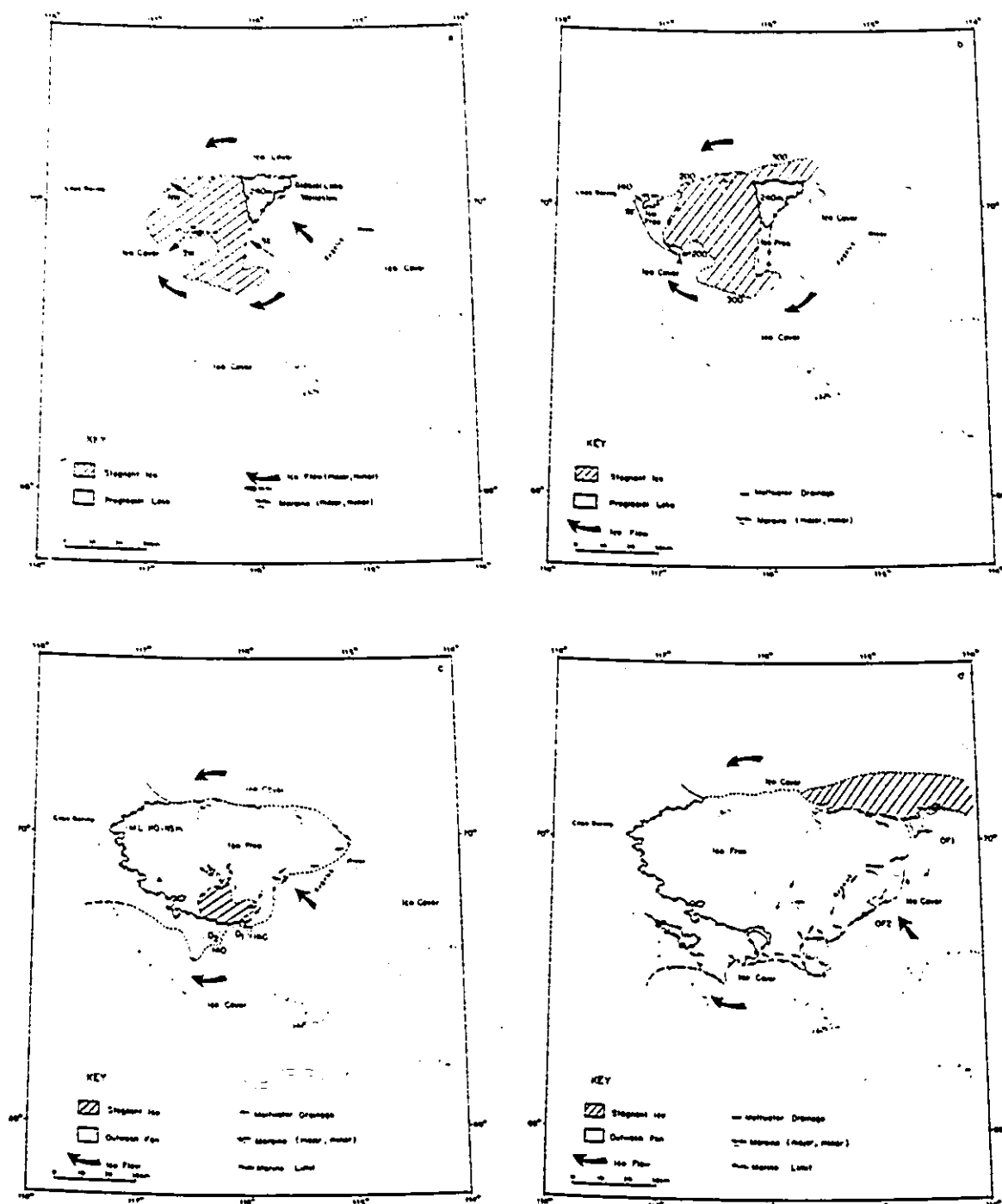


Figure 5.2. Ice-marginal positions at Wollaston Peninsula: a. Phase 1 (ponding) - ice margins and conditions during formation of glacial Lake Wollaston. b. Phase 2 (ponds to streams) - ice conditions during formation of southeast-sloping glaciofluvial terrace sequence at Cape Baring. c. Phase 3 (ice-marginal streams) - development of large meltwater drainage systems with ice margin positioned east of central Kugaluk River. d. Phase 4 (outwash fans) - development of large alluvial fans and extensive meltwater drainage system.

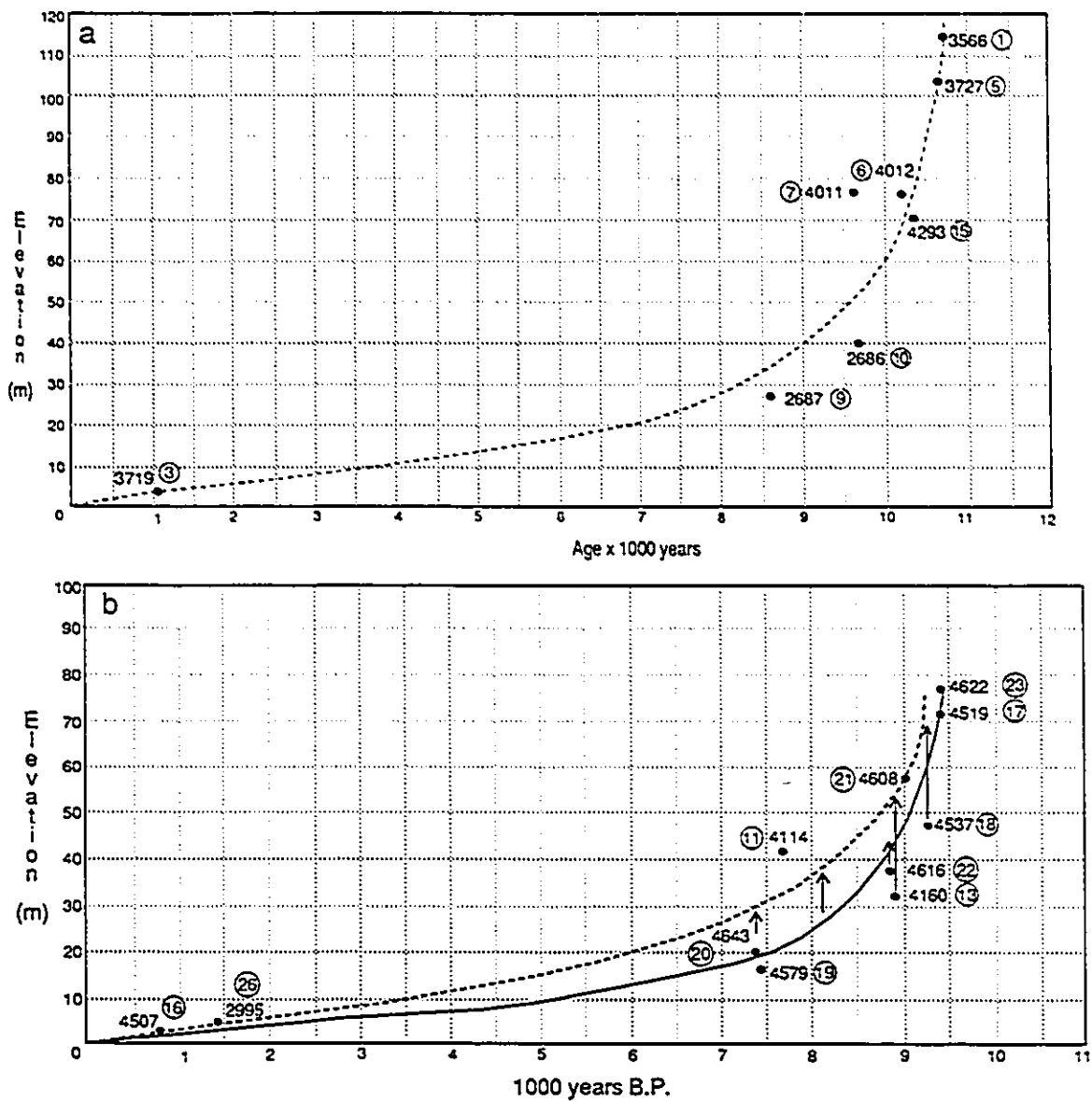


Figure 5.3 Preliminary emergence curves from Wollaston Peninsula: a. Cape Back-Cape Baring; b. Point Caen. Locations (circled numbers) shown on Figure 6.4; other numbers (eg. 4622) are G.S.C. laboratory numbers (see Appendix VII).

The major reasons for suggesting full ice cover during the last glacial event are summarized from Sharpe (1984) as follows:

1. Projected glacial profiles from the highest fresh (late Wisconsinan) landforms (elevations close to 400 m a.s.l. in eastern Wollaston) indicate full ice cover of Wollaston Peninsula where the main peninsula is about 200 m a.s.l. (Fig. 1.4).
2. There are no significant ice-marginal accumulations of glacial drift or lacustrine deltaic sediments marking a prolonged ice margin corresponding to the minimum ice extent on Wollaston plateau.
3. Large outwash (alluvial) fans rather than deltas indicate free-drainage during and following ice retreat rather than large-scale ponding required by minimum ice cover (Fig. 5.1c and Sharpe, 1984).
4. The deglacial features on central Wollaston Peninsula, considered Early Wisconsinan by Vincent (1984) are interpreted as thin deposits and subtle landforms (ground moraine) on level terrain, rather than features of subdued, old glacial terrain (Fig. 1.4).
5. Hummocky moraine which encircles Wollaston Peninsula is cross-cut by drumlin fields.
6. Soil data show no significant difference in properties between alleged older and younger terrain.
7. The glaciological model proposed by Sharpe (1988a) provides an alternative age explanation (landform zonation) for surfaces that were previously considered to be older.

Deglaciation was underway prior to 11 000 years BP as indicated by the marine shell dates at Holman on adjacent

Diamond Jenness Peninsula (Vincent, 1989). Wollaston Peninsula was deglaciated by at least 11 000 BP; marine shells from near Cape Baring (found close to marine limit) were dated at 10 700 BP (Sharpe, 1984). Prior to this and for 1000 years or more following 11 000 years BP, ice marginal ponds were common on Wollaston Peninsula (Fig. 5.1a). At the same time (11 000 ka to 10 500 ka), Glacial Lake Coppermine and a series of smaller lakes in the Richardson River valley on the mainland south of Coronation Gulf were dammed by ice occupying Dolphin and Union Strait and Coronation Gulf (St-Onge and Bruneau, 1982; Mercier, 1984; McMartin and St-Onge, 1989).

Shortly after 11 000 years B.P., ice-marginal retreat across Wollaston Peninsula was underway. Extensive meltwater channels suggest that abundant discharge of meltwater along the ice margin accompanied retreat (Fig. 5.2 c, d). At about the same time meltwater was overflowing to the sea from glacial Lake Coppermine to the south (St-Onge, 1987; Fig. 5.1). The ice margin was in contact with the sea and meltwater was also apparently accumulating subglacially in Dolphin and Union Strait. This stored water caused rapid movement in the ice stream in Dolphin and Union Strait. This ice configuration and development apparently led to compressive flow, and hummocky and lateral moraine formation (Sharpe, 1988a). The deformed lateral moraines are complex and their development may imply several closely-spaced surge events. These rapid flow events may also have produced cross-cutting relationships of the adjacent streamlined forms, but *en echelon* drumlin patterns at Read Island are best explained as complex forms resulting from turbulent meltwater flow. The rapid flow events are attributed to the former action of ice streams that were apparently enhanced by

subglacial water and drawdown (Hughes et al., 1985; Sharpe, submitted).

It is further suggested that large floods may have followed surging flow and produced the array of streamlined landforms (Sharpe, 1988a). This apparently radical interpretation is based on the internal structure of some drumlins (Read Island), meltwater erosion forms on bedrock (including rock drumlins), evidence of complex formative flow (Read Island; Banning Lake), and tunnel valleys cross-cutting the streamlined forms.

One of the most significant deglacial features is the Colville moraine system that encircle Wollaston Peninsula. Features of this system have limited direct dating control (marine limit sediments), but their formation can be attributed to glaciologic, dynamic and structural control (Sharpe, 1988a) in addition to solely climatic control (as proposed for other large arctic moraines, e.g. Dyke, 1984 and Dyke and Dredge, 1989).

After glacial surges and possible large post-surge flood events (Kamb et al., 1985), the ice was so extended and thin that the role of active ice on Wollaston Peninsula effectively ended. The lack of ice-marginal landforms, deposits and drainage systems suggests that the remaining ice stagnated and melted in place. The hypothetical ice-marginal position shown at 10 000 years BP (Fig. 6.2) refers only to the minimum time of marine incursion and does not imply an active ice front. The structure of large open-channel eskers in the area shows a predominance of low-energy sediments commonly including stacked, probably annual rhythmic sequences (varves). The input of supraglacial water seems to have been important at this stage of ice sheet wasting on southern Victoria Island.

In eastern Victoria Island complex landform patterns relate to former complex flow patterns that probably resulted from local ice-surface or topographic gradients. These flows are thought to represent late-glacial surges or ice-streams that redistributed ice into flatter profiles (Sharpe and Shaw, 1989). Flatter profiles are partly inferred in the discussion of regional ice stagnation that could have resulted from thinner, inactive ice. If the landform patterns are related to surges, they were possibly propagated by large-scale subglacial water storage and discharges. This may help to account for the short-lived nature of the flows (Hodgson, 1987) and the propensity for cross-cutting within the streamlined landforms of eastern Victoria Island (Sharpe, 1985; Fig. 3.1). These complexities need further study and testing; their resolution is beyond the scope of this report.

The pattern of deglaciation outlined above is based on the assumption that streamlined landforms represent glaciofluvial responses to the rate of melting (probably warm-based ice) and subglacial water storage within the Laurentide Ice Sheet, in addition to mass balance responses to climatic amelioration. This interpretation places important constraints on paleoclimatic reconstruction. Mass disintegration of large portions of the ice sheet is a predicted result of the preceding hypothesis. Regional disintegration is thought to have occurred late in the deglaciation sequence for Wollaston Peninsula, and to have resulted from glacio-dynamic events (large storage and discharge events) as a positive feedback response to direct climatic warming. Alternatively, the pattern of deglaciation has been argued to represent deglaciation by marginal retreat, in places enhanced by calving margins controlled by sea level control. This may explain some flow events, but

others are not clearly tied to sea level control.

5.4 Marine limit

Marine limit on Wollaston Peninsula, as in other areas is variable and time transgressive because the ice sheet applied a variable load to the crust and it melted in an irregular fashion during regional stagnation. Marine limit as traced on map 1650A rises to the southeast, reflecting the dominant ice movement from the Bathurst Inlet area to the southeast. The Bathurst Inlet area is a region of high marine limit (Dyke, 1987); so it seems reasonable that it represents an area of thick ice accumulation and possibly formed an extension of the Keewatin Ice Divide recognized to the south (Prest et al., 1968). Alternatively, the Bathurst Inlet area may have been an area of rapid initial marine incursion rather than an area of ice loading (Dyke and Dredge, 1989). Neither the history of ice retreat for southern Victoria Island presented here nor that for south of Coronation Gulf (St-Onge, 1987) appear to support this latter interpretation, because ice occupied the Coronation Gulf and Dolphin Union area in late glacial time as indicated by late landward flow and the wide distribution of ice-supported lakes (Map 1650A). The sea appears to have entered the area erratically, accompanied by restrained rebound in areas where melting ice masses prevented transgression in a regular west to east succession.

5.5 Postglacial emergence

The pattern of postglacial emergence is defined along southern Victoria Island, particularly Wollaston Peninsula, by dating marine deposits at various altitudes (Table 6, Fig. 5.4, 5.5). This dating is incomplete, and only preliminary

emergence curves are presented for Cape Back on northern Wollaston Peninsula, and Point Caen on southwestern Wollaston Peninsula (Fig. 5.4a and 5.4b). The more tightly-controlled curve is the one presented for a small valley in southwestern Wollaston Peninsula at Point Caen (Fig. 5.4b). This emergence curve has a form similar to the preliminary emergence curve for Holman on Diamond Jenness Peninsula to the north. These curves compare to curves compiled for the south-central Arctic (Dyke and Dredge, 1989).

The curves appear to represent normal glacioisostatic emergence areas depicting relaxation of the Earth's crust related to the size, thickness and duration of the applied glacial load. For example, there was very rapid rise in sea level in the early part the sea level history related to ice removal. The curves flatten rapidly below 30 m or for the last 8,000 to 9,000 years. The curves are nested and indicate the earlier deglaciation of Cape Back relative to Point Caen. The curve for the Bathurst Inlet area to the south is nested inside the Wollaston curves and suggests continued normal emergence in that area.

The data for the lower portions of the Wollaston curves are insufficient to test for possible ice forebulge effects. Where these are significant, land may have been undergoing submergence since middle Holocene time. A wood date on low beach terraces near Cape Back may relate to the emergence of a zone formerly submerged during Holocene times (Appendix VI). The curve is drawn to show continuous uplift at a negative exponential rate, because there are no observed submerged geomorphic features that might support this interpretation. The dated wood certainly drifted across the sea from the mainland and, hence, it may only provide a maximum age for the sea level feature itself.

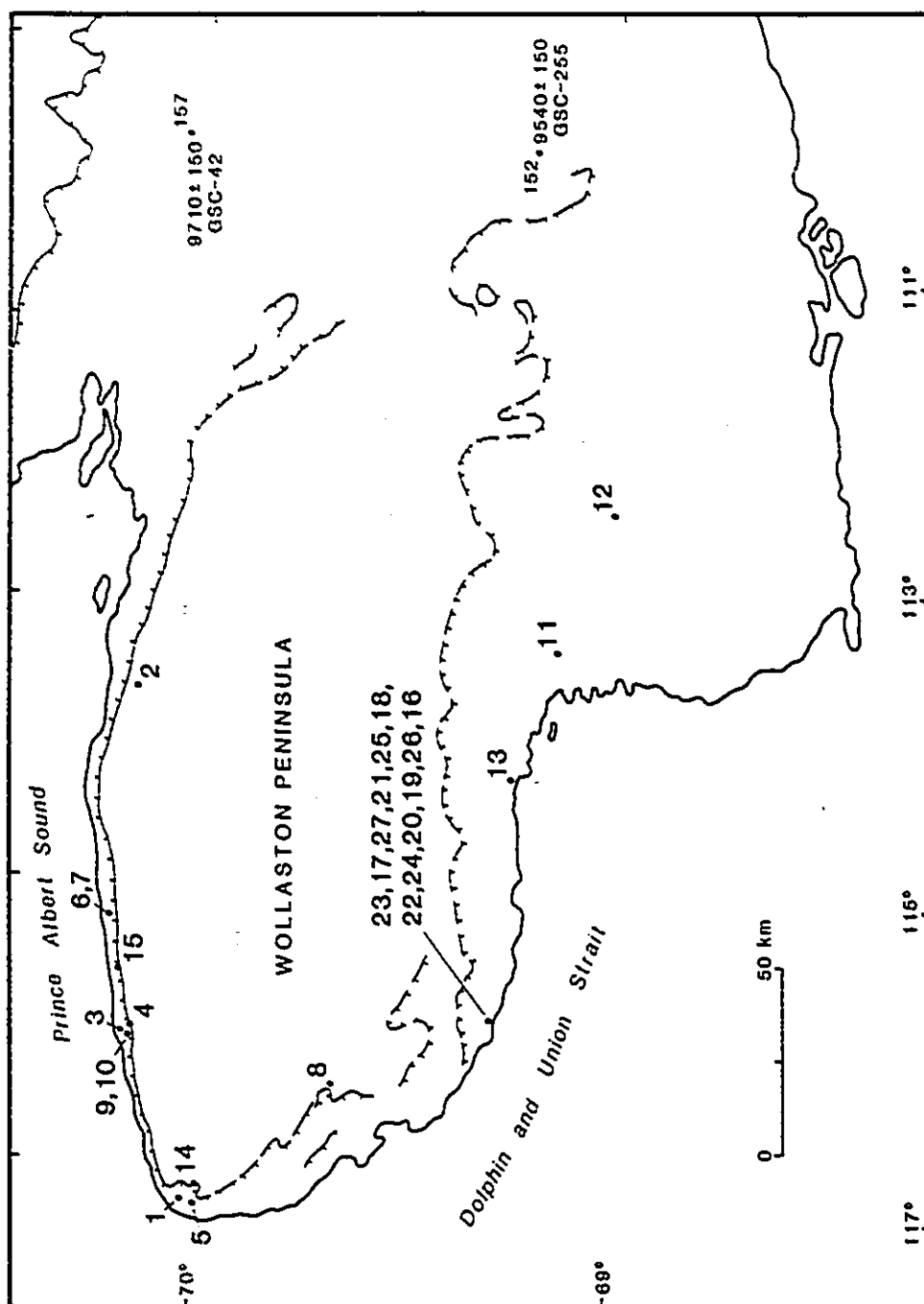


Figure 5.4. Location of radiocarbon date sites on Wollaston Peninsula. Numbers correspond to Appendix VII. Hatched line is marine limit.

Extrapolation of the Point Caen curve to a marine limit about 100 m a.s.l. gives estimates of deglaciation at about 9,700 years B.P. This age provides an interesting comparison with organic remains found in morainal sediments near Point Caen. Wood found in this location has been dated at $9,360 \pm 110$ years BP (GSC-4622), which points to early colonization by plants in areas adjacent to the recently departed ice sheet. Emergence curves in the adjacent Bathurst Inlet area give estimates of deglaciation in southeastern Victoria Island prior to 8000 BP. This accords with preliminary data from north of Cambridge Bay. More detailed surveys may further resolve the ice load and emergence history in the area.

5.6 Summary

Transitional, contiguous landform sets and dating allow Late Wisconsinan glacial limits to be established on southern Victoria Island and on the adjacent mainland. These ice limits support the inference of extensive ice-cored terrain resulting from relict glaciers. Glaciological constraints argue against these limits being associated with extensive ice shelves; warm-based and temperate ice breaks up readily.

Marine limit, with a steady northwest to southeast rise, is obscured by high lacustrine limits. Late-lying ice in the large marine straits trapped lakes against the island uplands, prior to marine incursion. This incursion was erratic because ice wasting was by regional stagnation rather than marginal retreat. Deglaciation between ~11,000-10,000 B.P. was by ice margin retreat as shown by outwash and ice-marginal landforms; between ~10,000-9,000 B.P. ice melted in place following widespread glacier thinning caused by surging. Extensive eskers developed in the stagnant ice.

6.0 CONCLUSIONS

There is variable drift thickness, and even local absence of drift, across southern Victoria Island, where underlying bedrock forms a uniform carbonate platform. Topography appears to have exerted an important control on sediment thickness, and subglacial erosion of sediment is thought to have played a much more significant role than non-deposition or bedrock erodibility in causing variations in drift cover.

The idea that lithology exerts a major control on glacial processes and their resultant landforms is a major theme in glacial geology (e.g. Shilts et al., 1987) and glaciology (e.g. Boulton and Jones, 1979). The relative amounts of debris in glaciers is confidently correlated with the geology of the glacier bed and, according to some workers (Aylsworth and Shilts, 1989a), has little to do with regionally changing dynamics of the ice-water system. This view is challenged in this thesis because the erodibility of carbonate sediment across southern Victoria Island is uniform and is known to be high (e.g. Vagners and Dreimanis, 1971). Yet the amounts of glacial sediment vary spatially. In addition, major erosional events clearly influence the sediment thickness and landforms of southern Victoria Island.

6.1 Hummocky moraine

Areas of apparently thick sediment, with large polygonal patterns, and thermokarst slumps in hummocky and lateral moraine are probably underlain by extensive and massive ground ice. These hummocky moraine deposits lie close to a carbonate escarpment where flow resistance was greatest; these are surrounded by areas of thin, streamlined

drift and bedrock scouring along the adjacent lowlands.

The hummocky moraine on Victoria Island may be compared with similar topography and deposits in the prairies, Keewatin, and elsewhere in the western Canadian arctic. Prairie forms are mainly subglacial in origin produced by ice press and marginal thrust (Stalker, 1973; Moran et al., 1980). Some supraglacial deposited debris occurs, but thermokarst is of minor importance. Keewatin forms are low-relief till hummocks likely formed as subglacial bedforms related to adjacent streamlined forms and Rogen moraine (Aylsworth and Shilts, 1989a,b). In the western arctic, however, there is both direct and indirect evidence that the terrain is largely ice-cored. Some hummocky terrain is simply thermokarst formed as extensive segregated ice bodies melted (Mackay, 1971; Rampton, 1988). Thick morainal areas, particularly within Late Wisconsinan limits (e.g. southern Victoria Island; southern Banks Island), remain ice-cored because sediment protects underlying glacier ice (e.g. Sharpe, 1988a; French and Harry, 1988; 1990).

6.2 Transitional landforms

A suite of landforms can be related to a single flow event (Fig. 2.1). It is also the basis for the analysis of landform associations and genesis. Landform transitions include eskers gradational to sinuous streamlined forms, and ground moraine gradational to hummocky moraine. Hummocky moraine and streamlined forms are also end members of a continuum (Fig. 1.14). Ground moraine includes diamictons and interbeds of sorted sediment in contrast to the classical models in which it is composed of lodgement till. Hummocky moraine contains complex deposits including ice and may be related to a variety of factors such as topography, marginal

thinning and thrusting, permafrost, and meltwater migration or flooding. The escarpment topography of Wollaston Peninsula was of major significance to hummocky moraine formation, but because of potential buried ice it may simply be hummocky terrain.

Although landform models are not very dependent to lithology on southern Victoria Island they are influenced nevertheless by topography, and especially permafrost. The abundance of streamlined landforms may favour the relative importance of warm-based glacial regimes compared to cold-based conditions. However, the formation of aggrading permafrost around the retreating ice margin, and the presence of stagnant dead-ice bodies controlled the development of hummocky terrain and ice-cored lateral moraines.

6.3 Drumlins

Several inferences are made concerning streamlined landforms on southern Victoria Island, particularly concerning drumlin form and composition.

First, the continuum of streamlined forms includes forms ranging from drumlins to flutes. This landform continuum represents varying flow strengths, best developed in the main flow of what were likely single flow events. A complex or *en echelon* arrangement occurs in parts of the field (Fig 3.3c). The forms and form organization are analogous to bedforms produced by complex, separated, turbulent flows (Allen, 1982; Shaw and Kvill, 1984). Streamlined fields cross-cut by inset narrow fields (with sinuous forms) reflect a transition from broad flow to confined flow (Fig. 4.9) suggestive of sheetflood to channel flow transitions.

Second, the composition of drumlins provides

information as to whether the drumlins are depositional or erosional in origin. In this regard, the stratified sediments within western Read Island drumlins favour a glaciofluvial cavity fill origin, although conformable fill was not clearly demonstrated. The occurrence of meltwater erosion forms (s-forms) and of sediment-free zones (tunnel valleys) in interdrumlin areas supports the concept of subglacial meltwater erosion on the eastern side of Read Island drumlin field. Rock drumlins definitely require erosion for their formation and, with superimposed s-forms, suggest that subglacial meltwater flows were powerful erosion agents on southern Victoria Island (e.g. Shaw and Sharpe, 1987; Sharpe and Shaw, 1989; Kor et al., 1991).

Third, the presence of till in the exposed (top) sections of drumlins favours ice-depositional processes for the drumlin sediment, probably lodgement and less likely deformation till. The combination of inset cross-cut forms, cross-fabrics, flow scouring (e.g. Penny Bay), and large crescentic erosion forms parallel to drumlin axis (e.g. Banning Lake) indicate that subglacial meltwater erosion may also have sculpted drumlins composed of diamicton. The upflow scour zones display crescentic depressions, streamlined remnant ridges, and in places, dissected hummocky ice-cored terrain.

Finally, it is suggested that different processes operated in different parts of the same drumlin field. The large scour zones and associated small erosion features in the eastern end of the Read Island field contrast with the western end where smaller divergent forms, *en echelon* forms, and less sediment stripping are found. Whereas the eastern end of the field appears to have been dominated by confined erosional flow, the western end reflects complex lateral flow

expansion beyond the Richardson Islands and possible deposition in cavities downflow (Fig. 3.1).

6.4 Eskers

Esker systems on southern Victoria Island are associated with scoured bedrock or sediment zones that flank small, sharp-crested narrow ridges and/or broad flat-topped ridges. Scour zones contain s-forms, transverse forms, lags and rock drumlins. The narrow, sharp-crested ridges parallel streamlined forms; broad ridges are composite in form, run oblique to streamlined trends and are found at the margins of lobate landform sets and former ice lobe boundaries. The narrow ridges have coarse gravel to bouldery surfaces and interiors, and are interpreted as subglacial deposits. The broad ridges contain rhythmic sequences of fine-grained sediments with little gravel. They represent deposition under subaqueous conditions, primarily from annual supraglacial meltwater discharges.

Some broad ridges are contiguous with narrow ridges, indicating discontinuous open channels of restricted length during regional deglaciation. Broad esker segments appear to represent major channelways in the ice, inherited from earlier, more vigorous erosional flows; the broad eskers are ice-contact glaciolacustrine landforms. Records of earlier powerful flows are either buried or represented only as erosional surfaces.

Some eskers show links with permafrost conditions where thermokarst enlargement of tunnel valley margins has occurred. Thinning ice profiles and marginal cold zones interrupted esker drainage during final decay which occurred primarily during regional stagnation rather than marginal retreat.

In summary, esker systems in the area possess scour features (tunnel channels and inset s-forms), transitions with flutings and association with drumlins fields which suggest that eskers are the later, variable but waning-stage glaciofluvial features following probable widespread flooding.

6.5 General

The arctic setting, within the zone of continuous permafrost, has an impact on landform development. The widespread distribution of large-scale polygonal patterned ground on Victoria Island suggests continuous permafrost and frost-induced processes. Massive buried ground ice apparently induces widespread thermal contraction cracking in the overlying sediments under the present climatic conditions. The exact distribution of ground ice, although charted with respect to large polygons, thaw lakes and thermokarst slumps, is related to the processes of moraine formation which must take into account the origin of the ice itself. The buried ice is likely of Pleistocene age and glacial origin rather than of Holocene age and segregation origin. The lack of Holocene water source in addition to other factors means that the extensive massive ice is likely buried glacier ice or refrozen meltwater from steady state or episodic meltwater release.

The distinction between periglacial and glacial landforms may be difficult to make using only surface morphologies. The melting of ground ice produces secondary landforms on Victoria Island that may be confused with primary glacial landforms. Thaw of massive ground ice produces ridged terrain which gives the false appearance of nested minor moraines; elsewhere, it gives the appearance of

kames or small eskers. The Pleistocene age and probable glaciogenic origin of the ice rules out a periglacial origin for landforms with extensive buried ice. In essence, large areas of hummocky moraine on southern Victoria Island should more properly be termed hummocky terrain; because the landforms are characterized by thermokarst and not necessarily by drift accumulation, which is the usual case for moraines.

Glaciolacustrine deltaic deposits are coarse, more prominent and more extensive than corresponding glaciomarine proximal sediments. Presumably this reflects the proximity of glaciers during glaciolacustrine deposition and enhanced rates of sedimentation.

End and lateral moraines comprise subaqueous (interbedded sediment flows) or subaerial (massive diamictos or thrust structures) ridges. Lateral shear, marginal thrust, and a permafrost fringe were important to formation of deformed moraine ridges.

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Appendix I

Terrain characteristics identified from LANDSAT images, Victoria Island

<u>Colours</u>	<u>Feature</u>	<u>Geomorphological unit</u>	<u>Comment</u>
Dark green, black	Low relief or level	Featureless till plain or distal outwash sediments (fine sands); coastal plain	fine-grained soil; poor drainage
Dark green/yellow	a) Upland morainic with many small lakes (transitional)	Hummocky moraine	Relief provides drainage; potential buried ice
	b) Lowland	Drumlin field (individual/multiple)	Bedrock or silt between drumlin
Predominantly white	Upland crests and ridges with a few lakes, kettle lakes	End moraine; discrete moraines visible; hummocky moraine in smaller areas; felsenmeer	Joint pattern in carbonate bedrock
Bright white	a) Uniform silt, fine sand, or gravel; marine, alluvial, or outwash b) Thin drift and bedrock areas	Predominantly marine silt (distal deltaic) with less alluvial silt and marine flats; some uniform outwash gravel	Eolian influence Joint patterns
Red	Lake ice, ocean ice saturated (isothermal)	Warm, isothermal ice	Late July image
Pink (white mottles)	River icings, ocean ice (cold)	Cold, coherent ice	Seasonal groundwater

Note: marine limit defined by boundary between dark green and predominantly white. See Figures 1.6 a and b.

Appendix II

Ground temperature profile, Wollaston Peninsula

<u>Depth (m)</u>	<u>Temperature (C)</u>
0.8	-5.3
1.5	-1.2
2.3	-11.2
3.05	-10.1
3.8	-8.0
4.6	-11.9
6.1	-12.0
7.6	-11.9
9.1	-10.6
12.2	-16.2

Site: Kagloryuak River (I 16-6)

Date: 10.8.1980

Air T (C): 5.0

Source: E.B.A. Engineering Consultants Ltd., 1981,
personal communication.

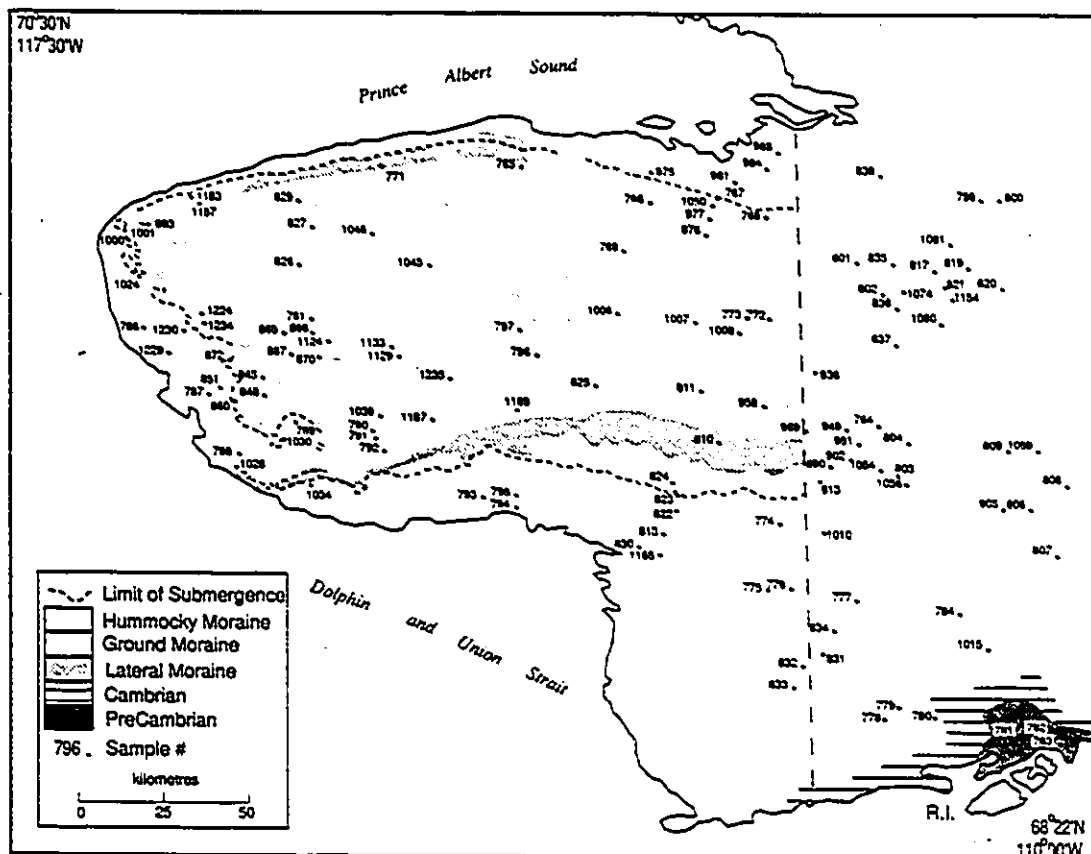
Appendix III
Summary of geological units

<u>Geological unit</u>	<u>Landform</u>	<u>Sediments</u>
<i>Terrestrial deposits</i>		
Acolian deposits silt	Dunes, loess cover	Fine and medium sand; minor
Organic deposits	Wetlands, swamp, fen	Peat, muck, minor mineral soil
Colluvial deposits	Solifluction lobes, aprons, debris flow lobes	Sandy silt diamicton with organic debris locally
Fluvial deposits sand	Floodplain, terraces	Gravel, gravel sand, and silty
Felsenmeer	Rock domes, blisters, block fields	Angular cobbles and boulders
<i>Marine deposits</i>		
Littoral deposits	Raised beaches	Sand and cobbly gravel
Tidal deposits	Tidal flats	Silt, sand, very minor gravel
Sublittoral deposits	Marine shelf	Sandy silt, silt, clay, minor gravel
Deltaic deposits	Marine delta	Gravel, sand
<i>Glacigenic deposits</i>		
Glaciomarine deposits	Fans and deltas	Gravel, sand, silt, minor clay
Glaciolacustrine deposits	Fans and deltas	Gravelly sand, silt, clay
Glaciofluvial deposits	Sandurs, terraces	Gravel, gravelly sand
Hummocky moraine deposits	Hummocky moraine ¹	Diamicton, gravel, sand, silt
Subglacial deposits	Ground moraine (stream- lined forms, drumlins, drumlinoid, and flutings)	Diamicton, sand, silt, and gravel
Ice marginal deposits	End moraine (lateral moraine), shear moraine	Diamicton, sand, gravel, boulders
Bedrock	Carbonate plateau, escarp- ments; Precambrian hills	Felsenmeer (carbonate)

¹Highlighted landforms are discussed in detail.

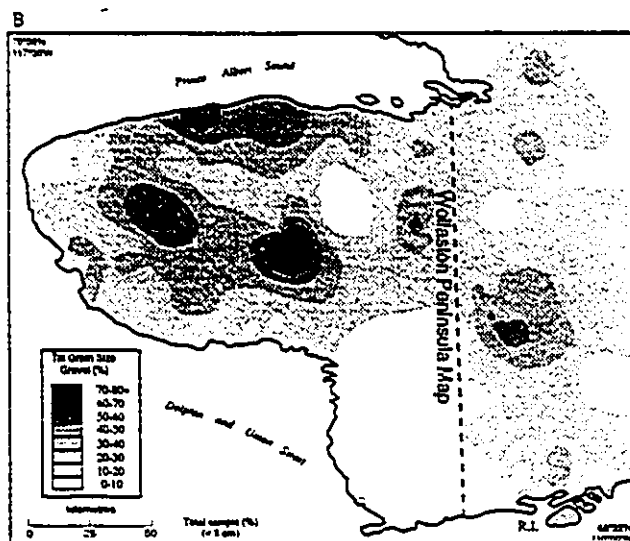
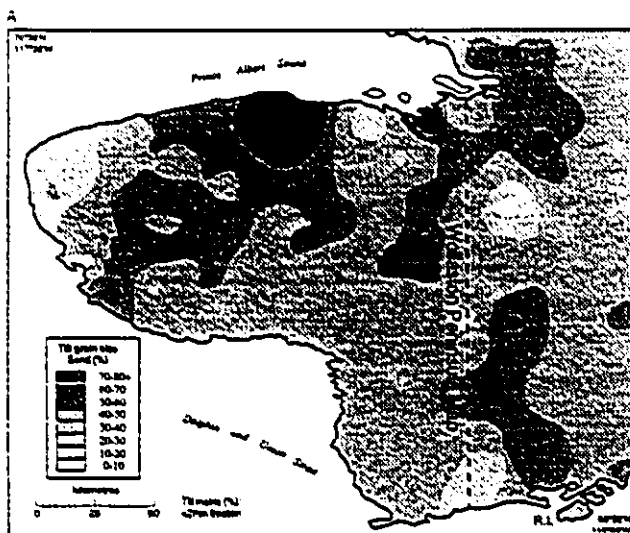
Appendix IV

Map of till sample locations

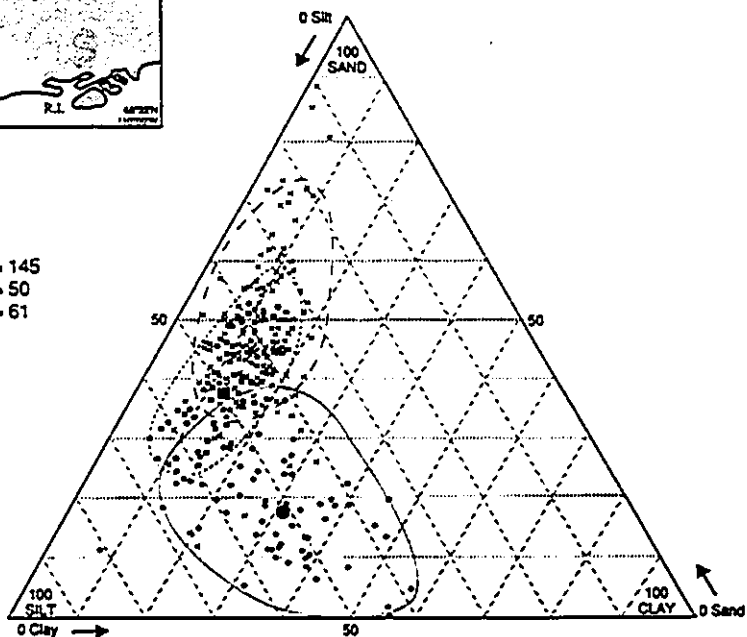


Appendix V

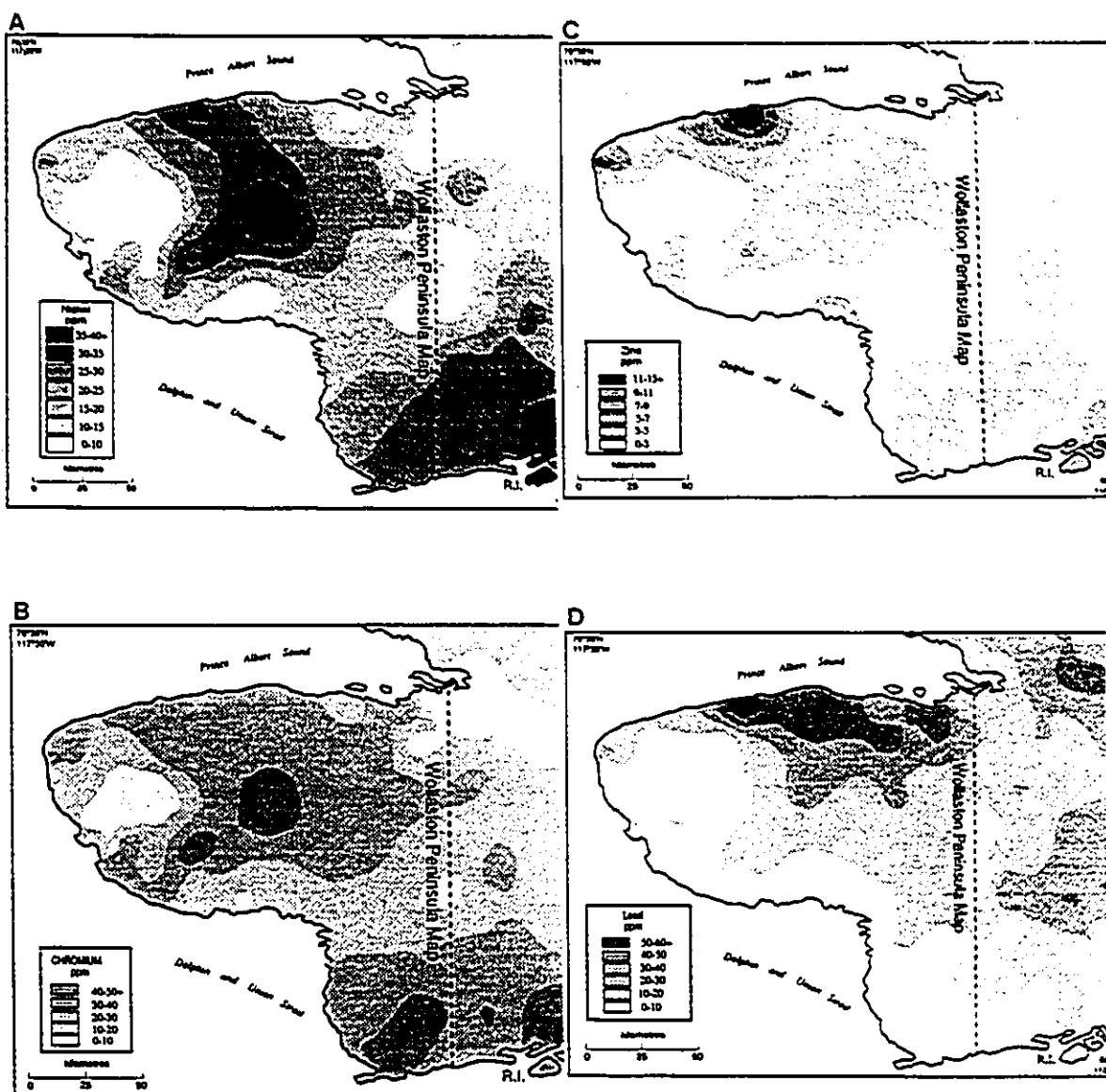
Till grain size plots; A. per cent sand, B. per cent gravel, C. till grain size Victoria Island (Wollaston) and southern Ontario (Elma Till).



- C
- WOLLASTON (carbonate substrata) N= 145
 - ELMA TILL (carbonate substrata) N= 50
 - ST. JOSEPH TILL (lacustrine substrata) N= 61
- (• ■ × mean value)



Appendix VI
Trace element plots



Appendix VII
Radiocarbon age determinations, Wollaston Peninsula and adjacent areas

Location Map	Sample (Map #)	GSC #	Date (BP)	Location	Elevation (m)	Material	Comments
1	82-SBB-SH34	3566	10 700 100	Cape Baring	91	shells	May date marine limit (115-120 m) and deglaciation of northwest Wollaston Peninsula
2	82-SBB-SH 11A	3580	9150 120	Lindaluk Island	84	shells	Dates marine limit and deglaciation.
3	83-NJ-0040	3719	1150 60	Cape Back	4	wood	at east extent of Wollaston Peninsula
4	83-NJ-0035	3725	8980 80	N. Wollaston	47	shells	Dates sea level at 4 m above present sea level; wood buried in beach gravel
							Dates related to shells in littoral sediment and therefore date on sea level at or about 47 m a.s.l.
5	82-SBB-SH56 (694)	3727	10 600 100	Cape Baring	105	shells	Dates marine limit and deglaciation as sample found 5-10 m below marine limit; supports GSC-3566
6	82-SBB-SH37 (1005)	4011	9240 150	E of Cape Back	77	shells	Appears to be too young; may be slightly contaminated by encrustation
7	82-SBB-SH37 (1005)	4012	10 200 280	E of Cape Back	77	shells	Dates appear to be good estimate of age of marine limit (115-120 m a.s.l.)
8	82-SBB-27 (677)	10-95	45 100	Kugalluk River	100	detrital	Dates from ice free period on Wollaston ¹
9	83-NJ-0039	S-2687	8565 220	Cape Back	28	whalebone	Date appears to date sea level at 28 m a.s.l.
10	83-NJ-0039	S-2686	9780 220	E Cape Back	40	whalebone	Date appears to be too old
11	83-NJ-0002	4114	7660 70	Forsyth Bay	40	shells	Dates formation of beach ridge at 40 m a.s.l.
12	82-SBB-SH-43 (774)	4134	9290 170	S of Mount Burzup	93	shells	Date provides minimum estimate of marine limit and deglaciation
13	82-SBB-SH-75 (794)	4160	8800 130	S coast	33	shells	Date is too old for this surface and probably relates to closer to marine limit 130 m a.s.l.
14	82-SBB-SH-50	4203	10 600 90	S Cape Baring	80	shells	Date corroborates GSC-3566 and GSC-3727
15	82-NJ-0038	4293	10 400	E of Cape Back	71	shells	Date relates to marine limit of about 120 m a.s.l.
16	87-NJ-15f	4504	740 50	Point Caen	3	wood	Date provides age of low level sea
17	87-NJ-2f	4519	9380 120	Point Caen	69	shells	Dates marine limit at >100 m a.s.l.
18	87-NJ-2f	4537	9280 80	Point Caen	46	shells	Dates sea level closer to marine limit at >100 m a.s.l.
19	87-NJ-26f	4579	744 90	Point Caen	15	shells	Dates a higher sea level than 15 m a.s.l.
20	87-NJ-22f	4643	7350 90	Point Caen	19	shells	Dates a higher sea level than 19 m a.s.l.
21	87-NJ-10f	4608	9020 80	Point Caen	56	shells	Dates sea level close to get above 56 m and possibly close to marine limit
22	87-NJ-12f	4616	8840 100	Point Caen	36	shells	Dates surface higher than 36 m a.s.l.
23	87-NJ-18f	4622	9360 110	Point Caen	100	wood	Dates slump event in huzrocky moraine above marine limit at time following deglaciation i.e., minimum date on deglaciation and plant colonization
24	87-NJ-7f	4648	460 50	Point Caen	27	wood	Date on wood that has been moved (compare GSC-4504)
25	87-NJ-8f	S-2993	9190 135	Point Caen	51	whalebone	Date relates to higher sea level than 51 m a.s.l.
26	87-NJ-14f	S-2995	1390 70	Point Caen	4	whalebone	Date provides estimate of low level sea; possibly moved
27	87-NJ-9f	S-2994	9600 146	Point Caen	57	whalebone	Date marine limit and deglaciation

¹Dated at Isotracer, University of Toronto by mass accelerator.



Plate I. a. Aeolian deposits, E, with colluvial stones, over diamicton, D. GSC 204441A.

Plate I. b. Ventifact showing characteristic sharp flute edges relating to flow separation in turbulent flow; flow (arrow) was from top right to bottom left. GSC 204441-A.

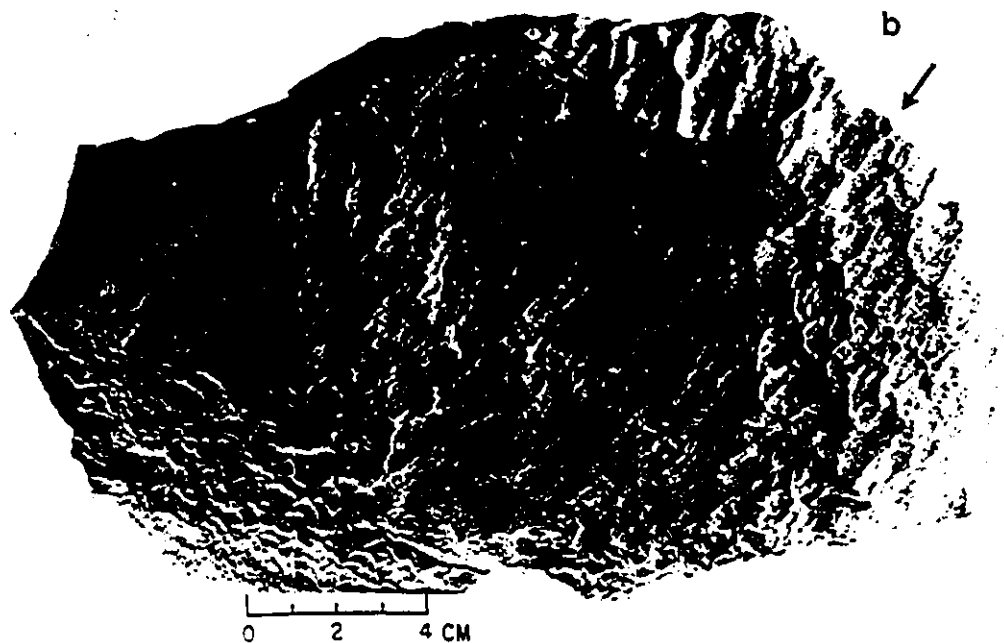
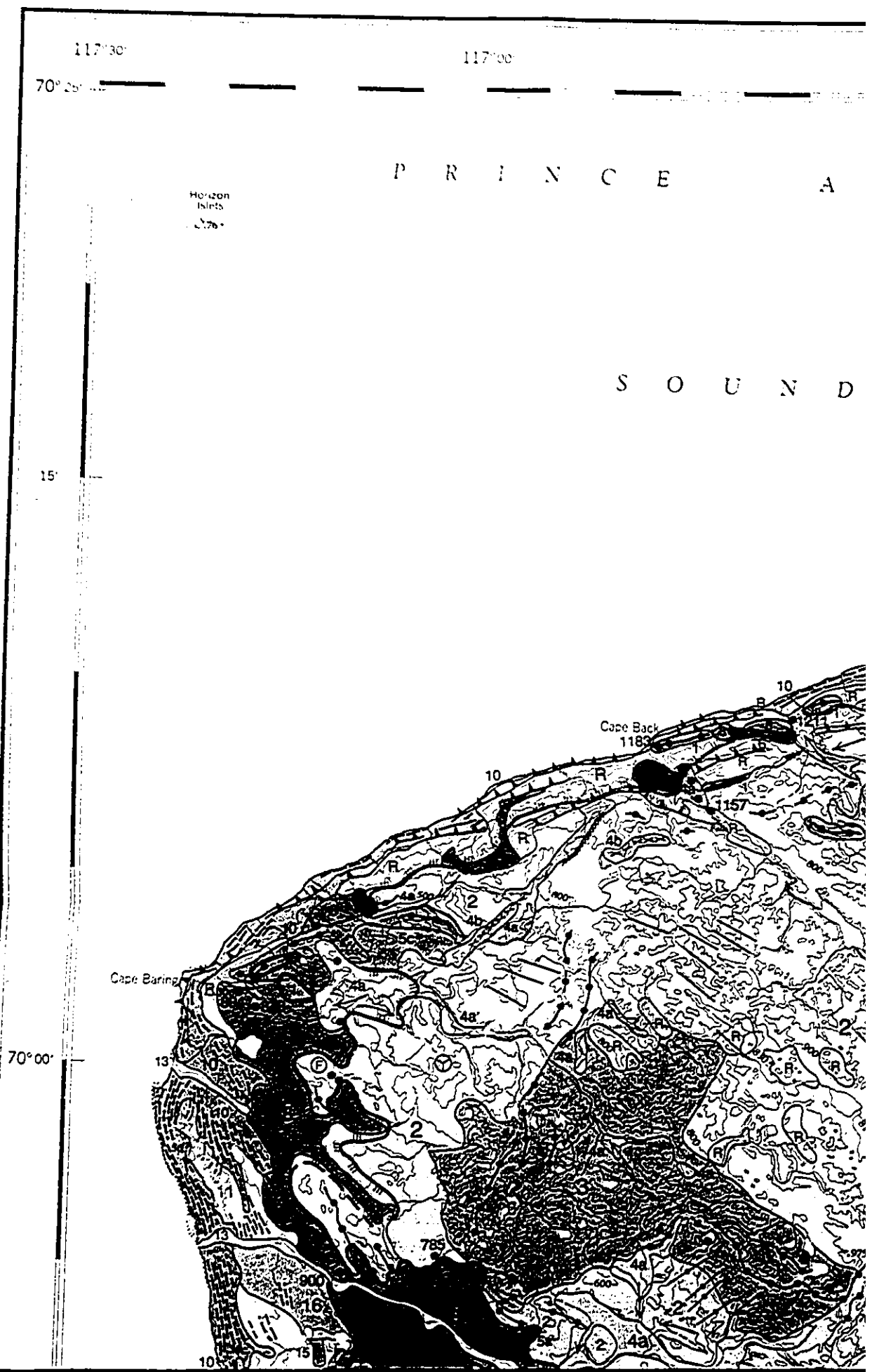


Plate II. Photograph of drained lake showing ring of lacustrine deposits (arrow) around lake margin within hummocky moraine. GSC 205193.



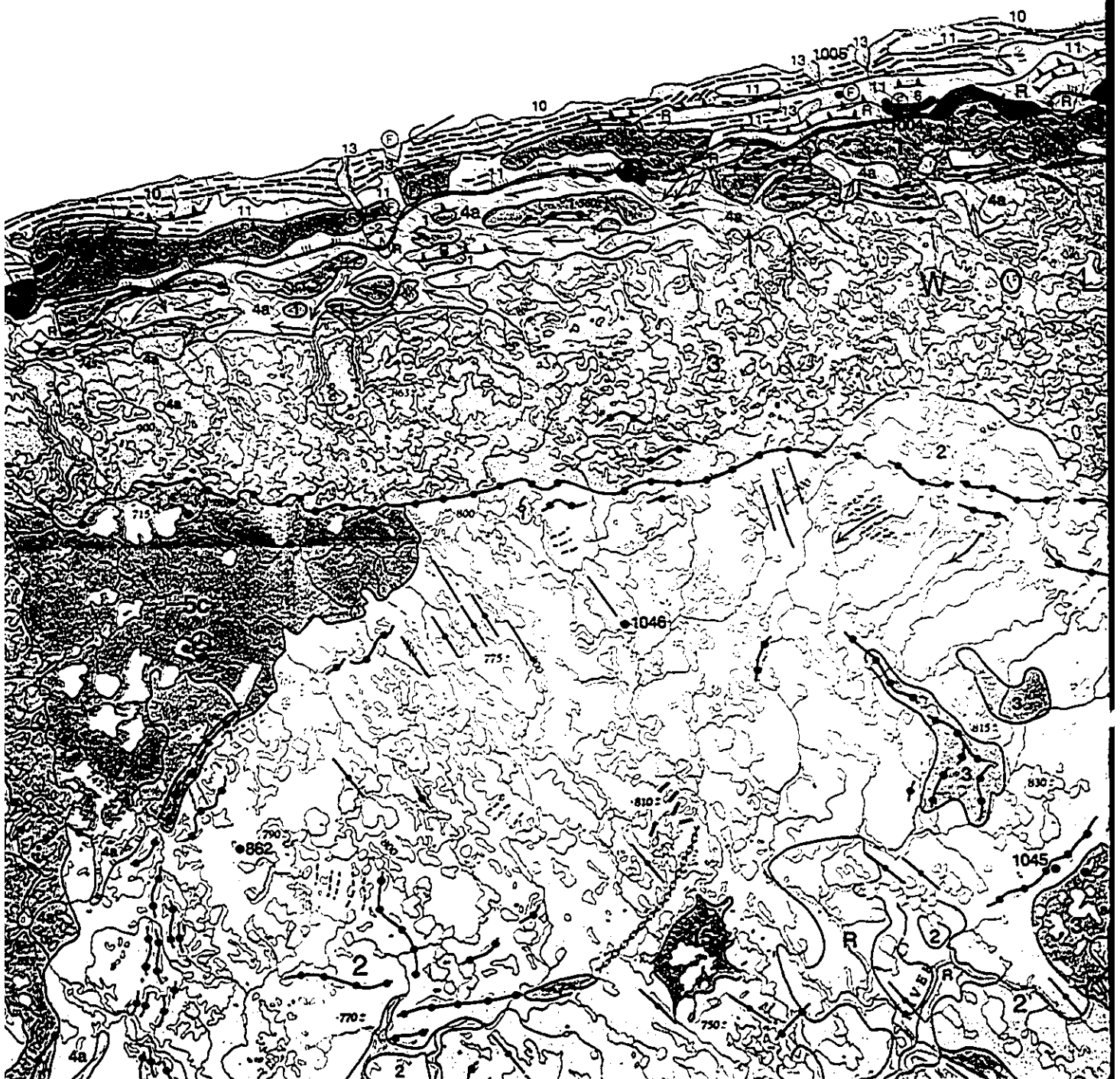


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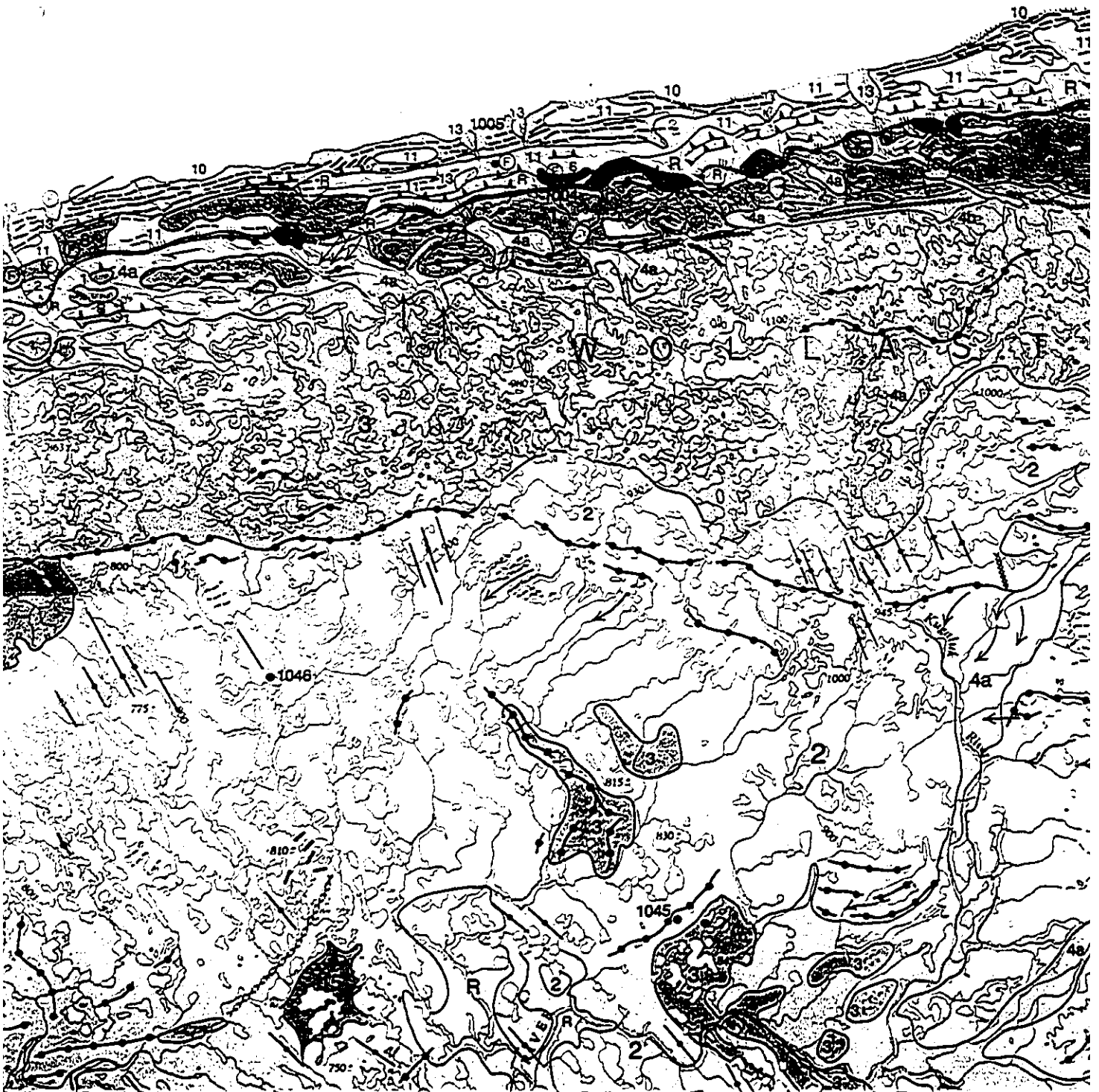
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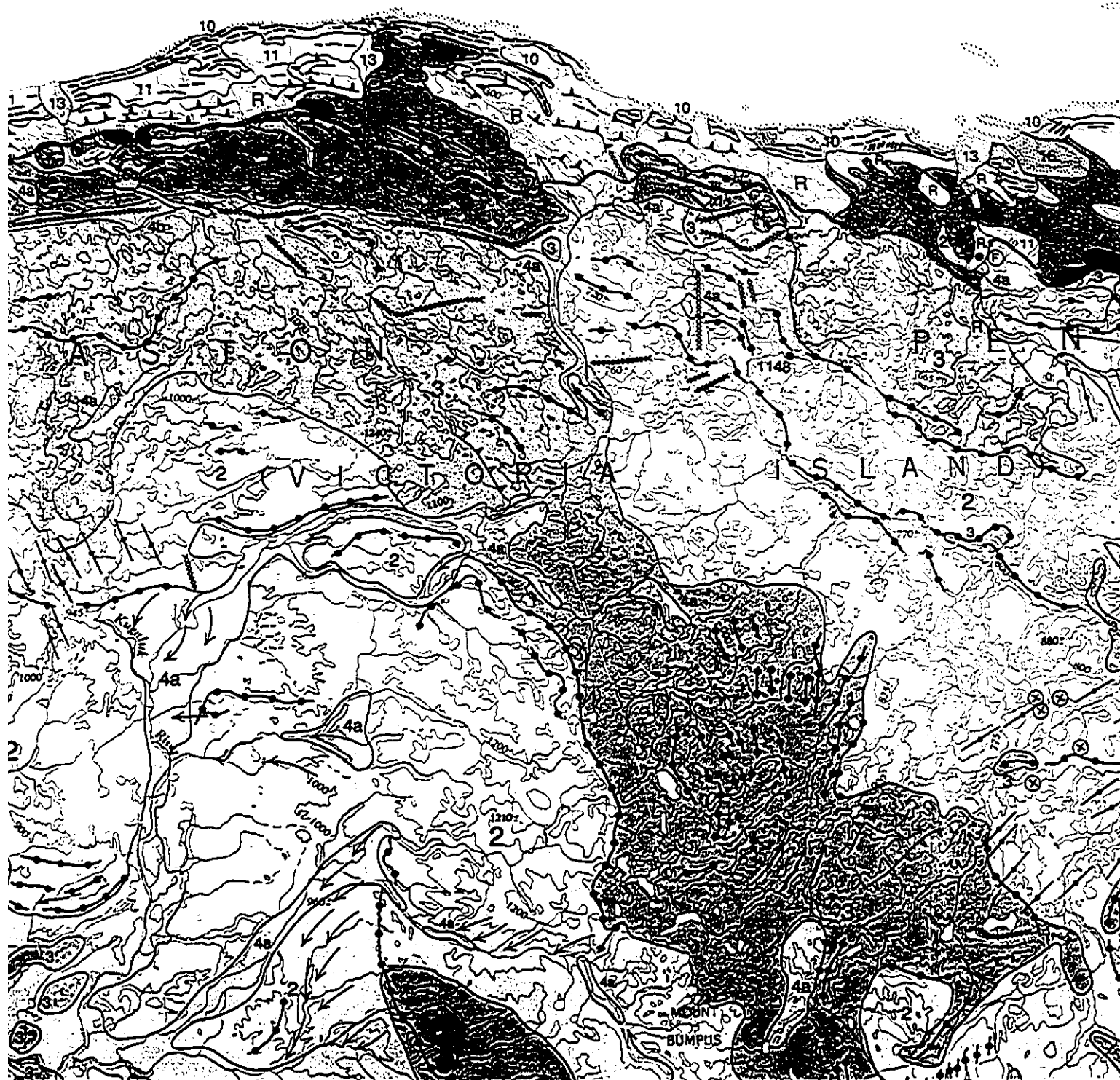
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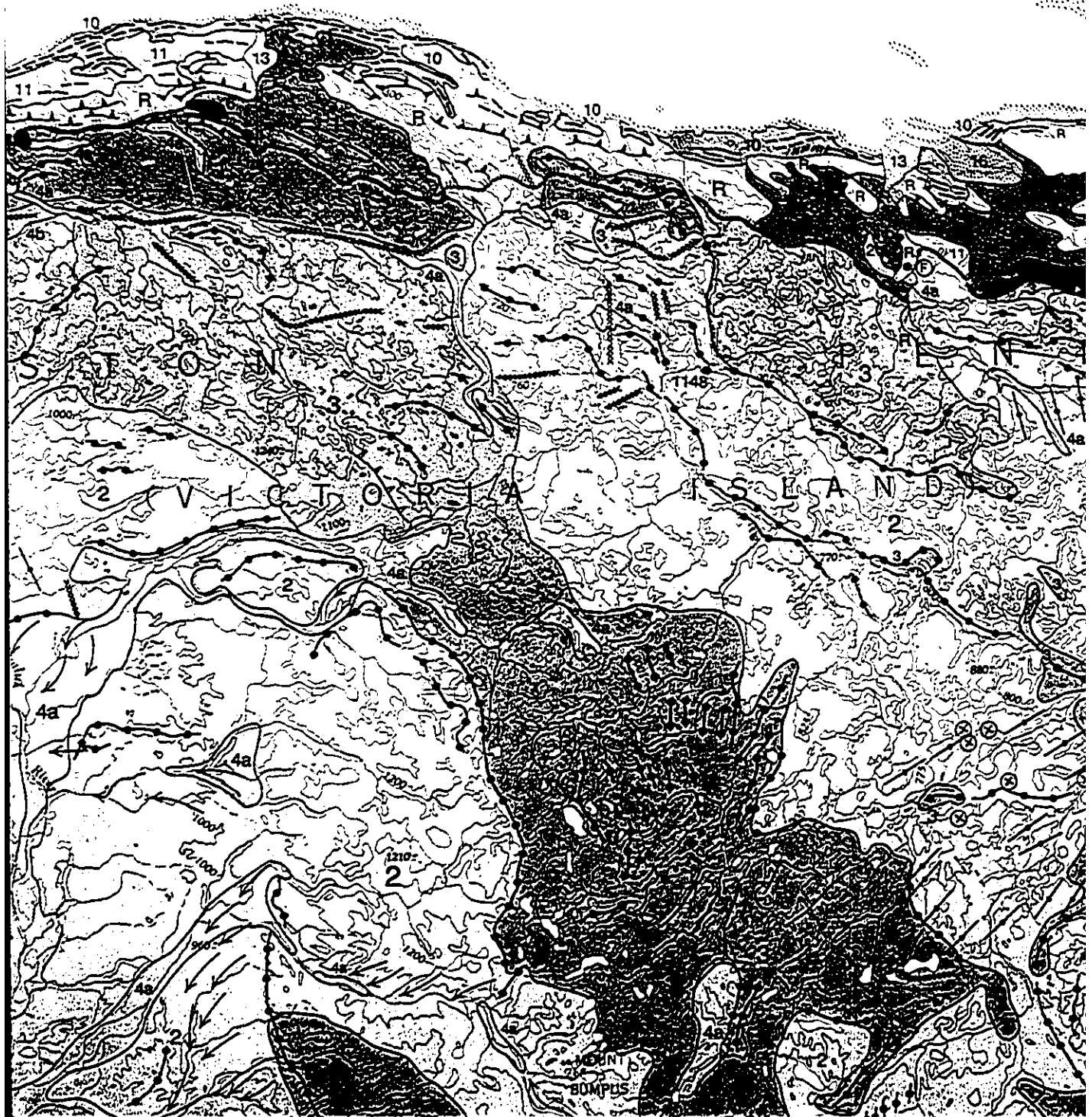
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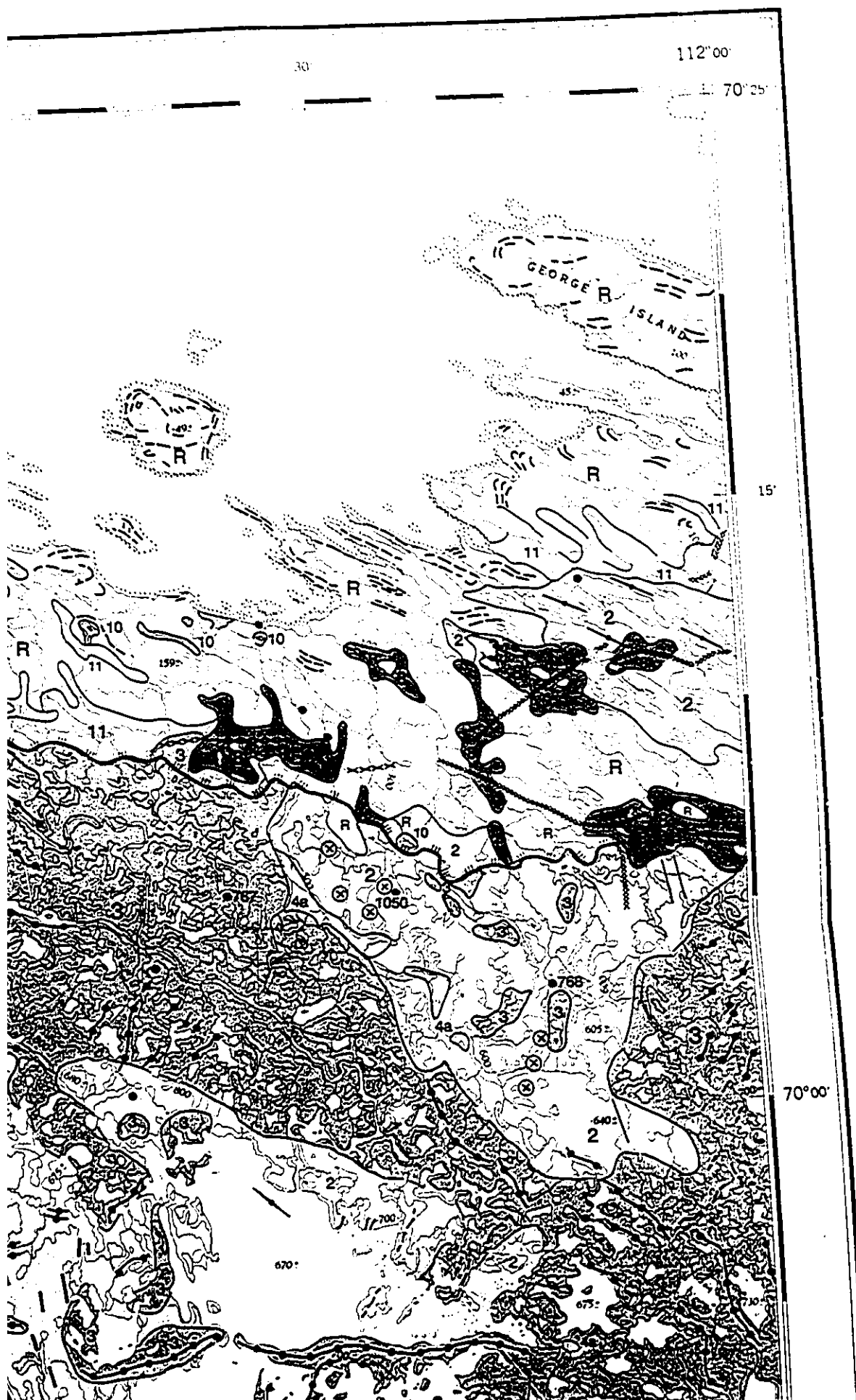


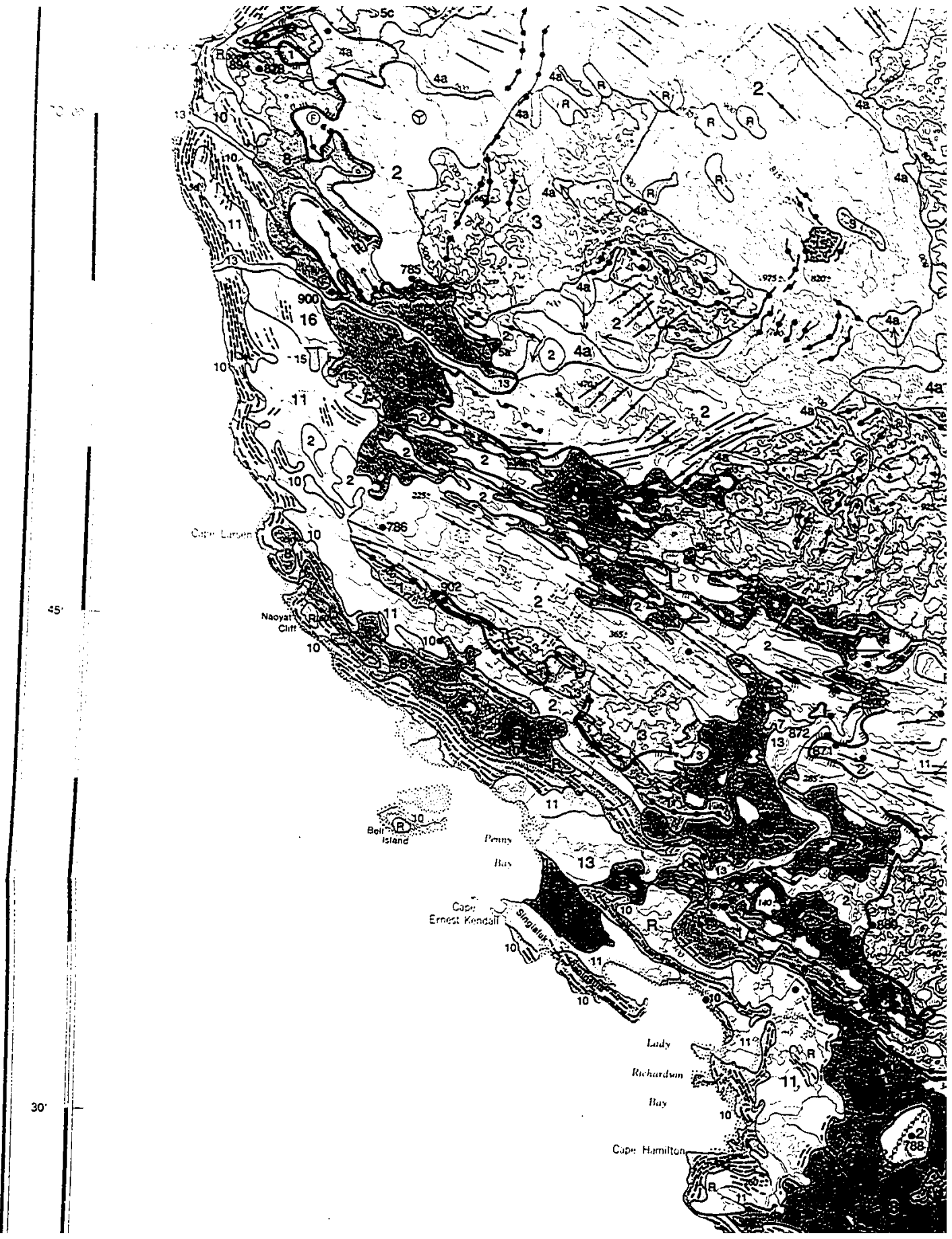
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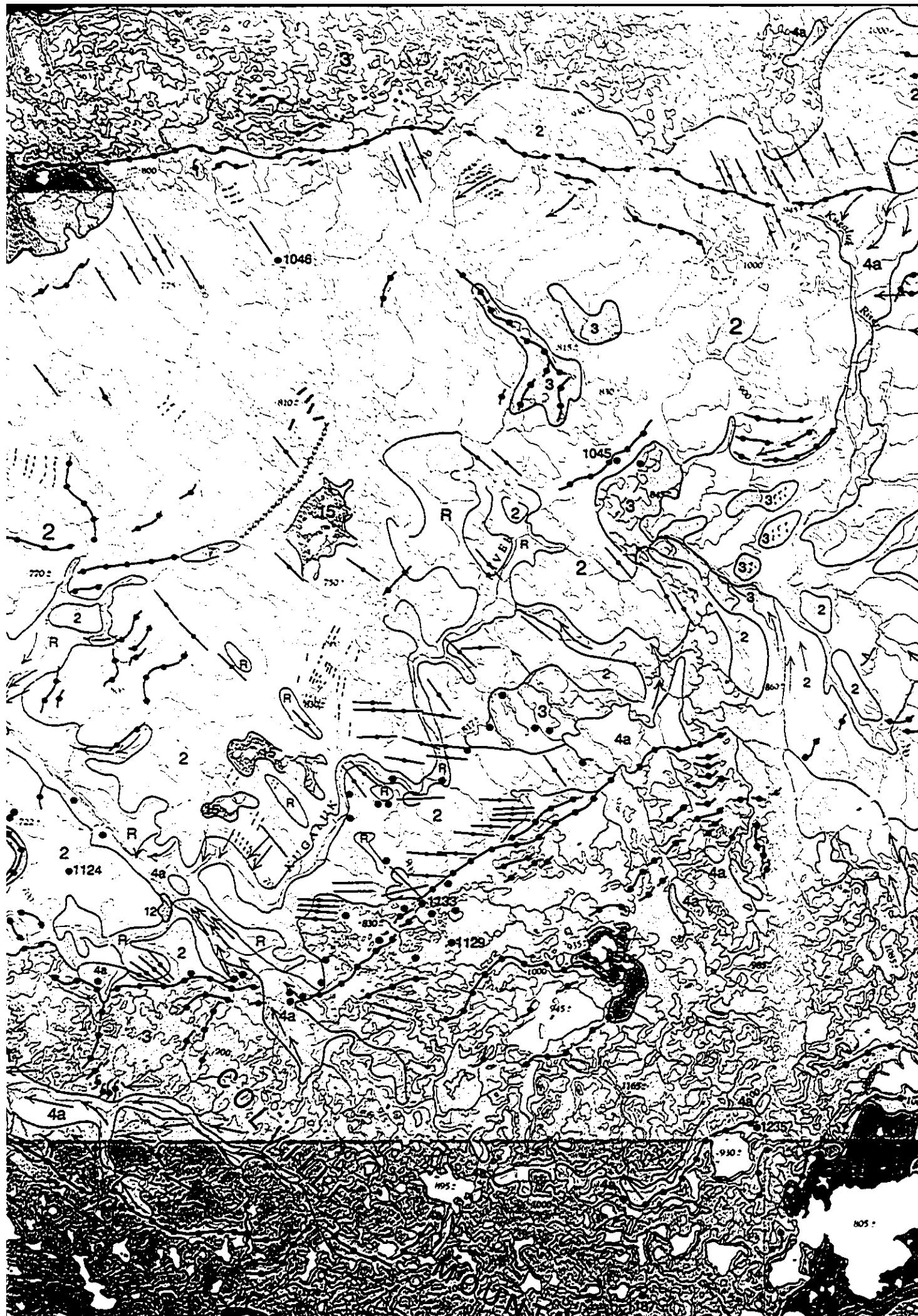






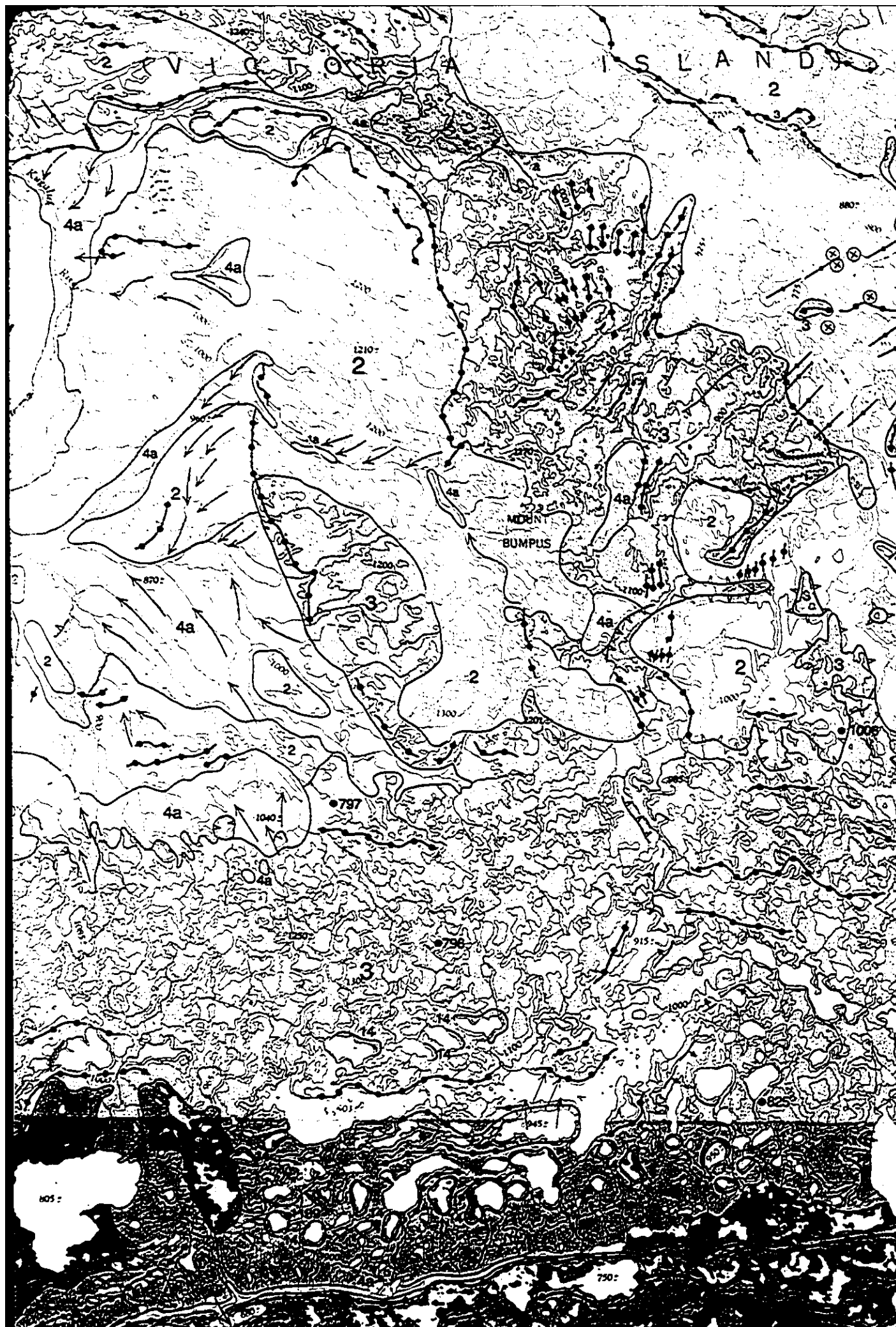






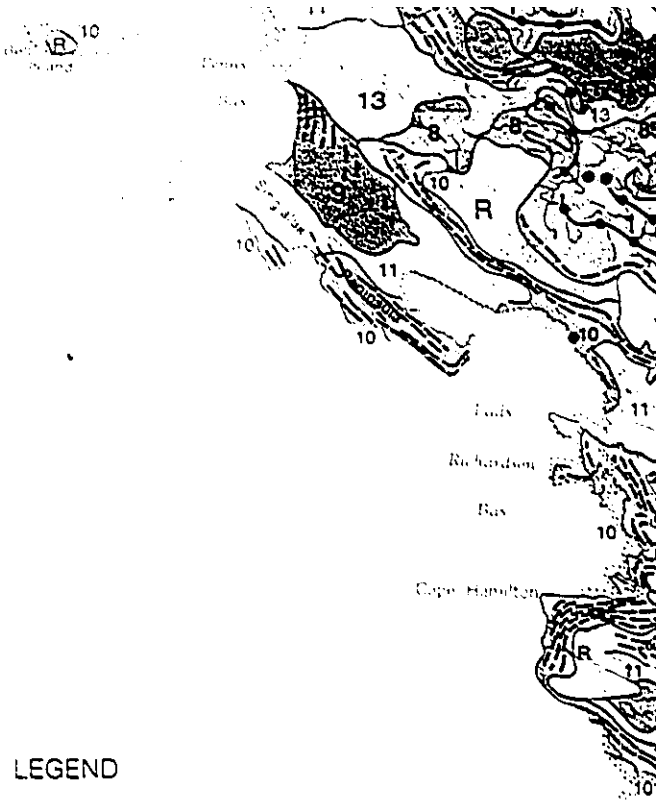












LEGEND

QUATERNARY HOLOCENE

TERRESTRIAL DEPOSITS: deposited by wind, gravity (colluvial), running water (fluvial), and in standing water (organic); these processes are active today and their deposits usually include organic debris

16

Eolian Deposits: thin veneer (1m) or dunes (2-3m) of fine sand covering other deposits; wind erosion of marine deposits limits plant stabilization

15

Organic Deposits: silty to sandy organic sediments; 1-3m thick; resting on a variety of poorly drained substrates; peat is present locally; mudboils, hummocks, and willow thickets are common

14

Colluvial Deposits: massive diamicton consisting of compact, stony, sandy silt, calcareous debris and rubble; 1-2m thick; consists of reworked primary sediment; occurs as solifluction lobes, terraces, or slump bowls, particularly in thermokarst terrain; only large colluvial occurrences that can clearly be distinguished from till are shown

13

Fluvial Deposits: gravel to gravelly sand near channels and silty sand and minor silt or clay, together with eolian and colluvial deposits, on floodplains; 10-20m thick on alluvial terraces, and thin veneer (1-2m) on strath terraces; frost wedges (1-2m deep) are common in gravelly sediments

12

Felsenmeer: platy gravel size fragments, occurring as a blanket less than 1m thick on Paleozoic rock; found predominantly on upland plateaus and along strath terraces; rock heave is common where deposit is thin

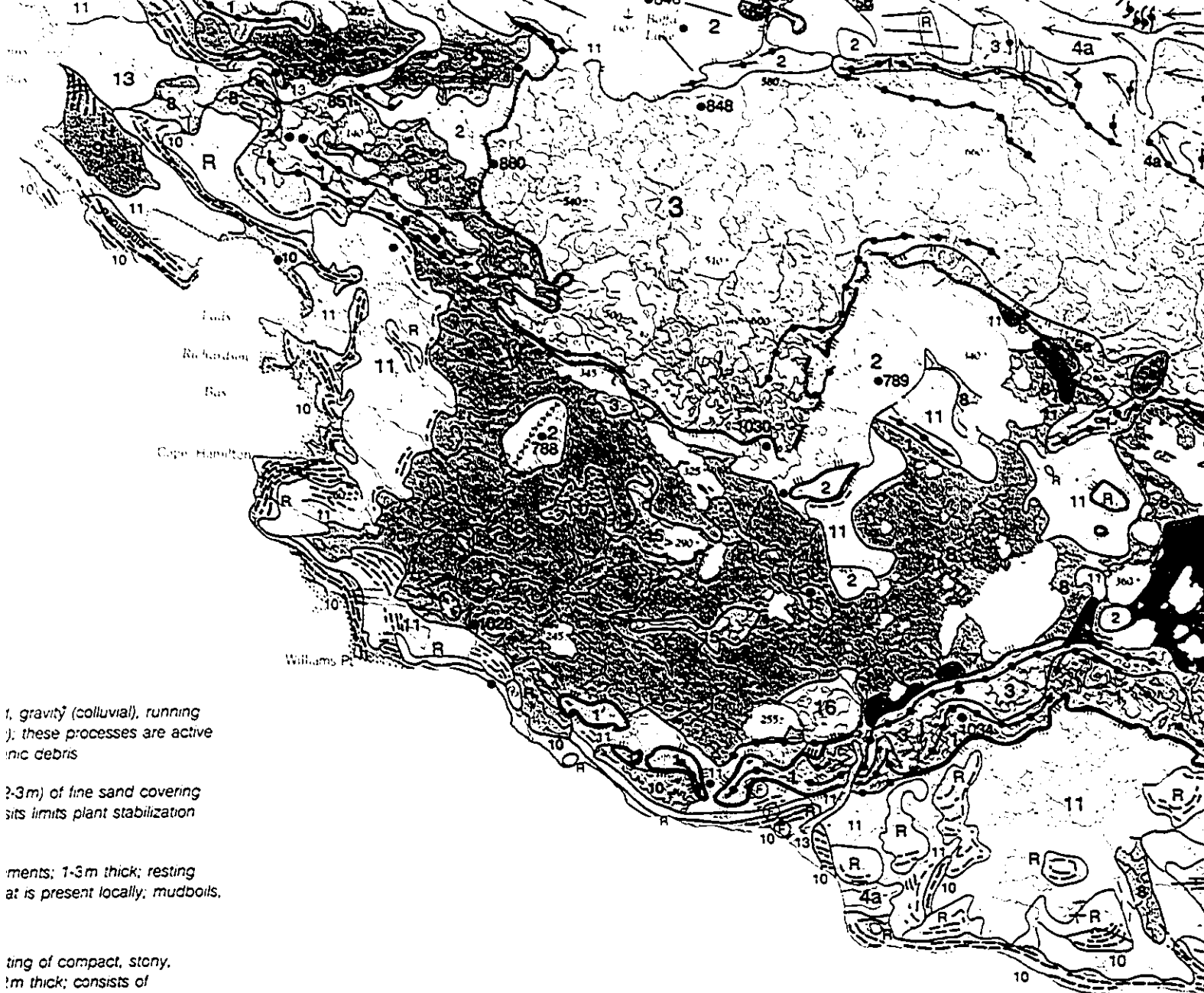
MARINE DEPOSITS: well sorted and stratified to massive gravel, sand, silt, and clay deposited during regression of the postglacial sea; occurs as ridges and blankets on large coastal platforms and terraces (south coast) marking regressive sea level events, and as narrow and terraced forms grading from glaciomarine outwash terraces (north coast). Sediments may have been disturbed by pack ice and iceberg scour

11

Undifferentiated: complex of silt and sandy silt on bedrock, diamicton, or locally gravel; commonly a veneer 1-3m thick

10

Littoral Deposits: gravel and gravelly sand; 1-4m thick; occurs as flights of raised strandlines, beaches (shingle) and spits



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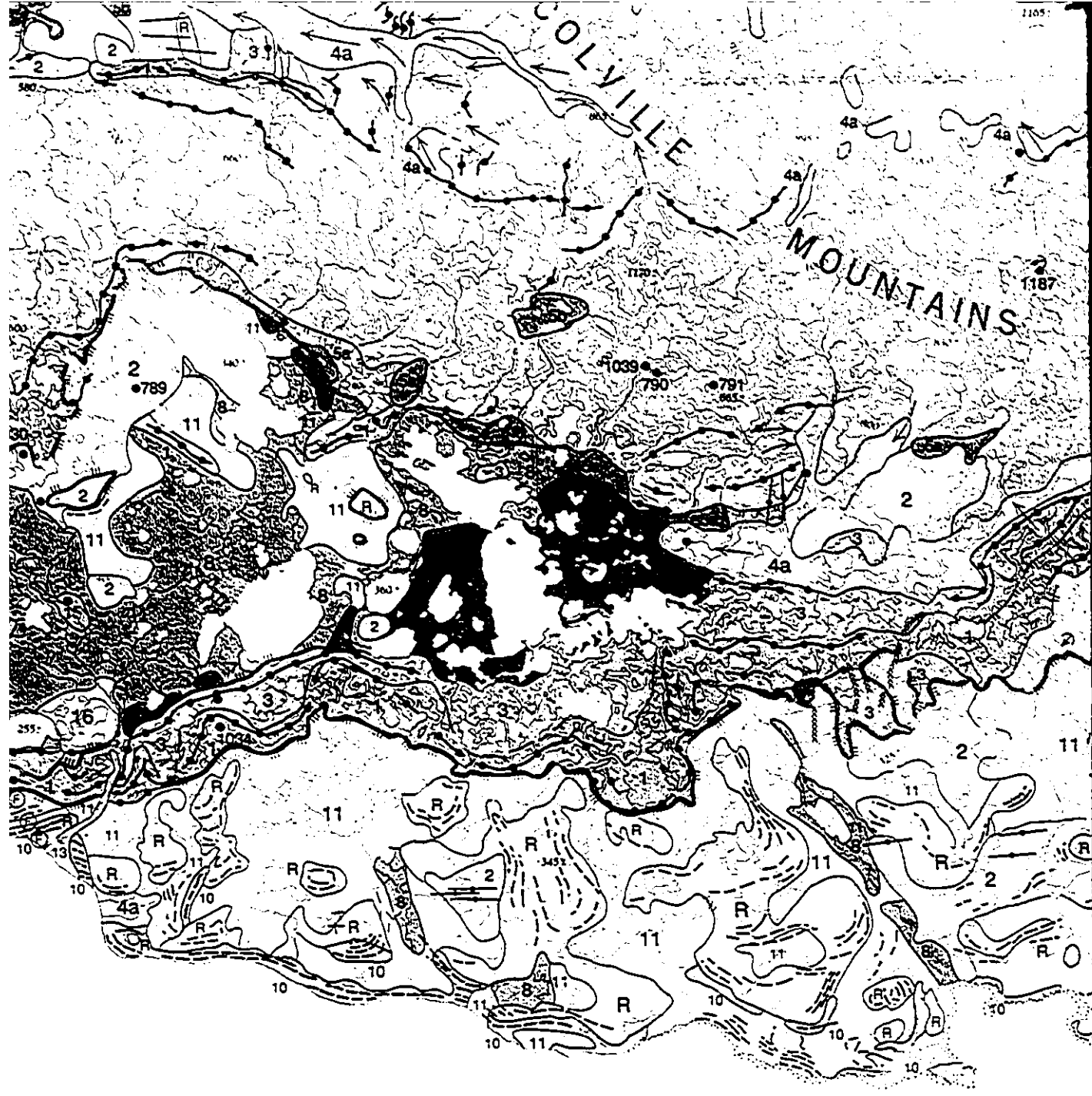
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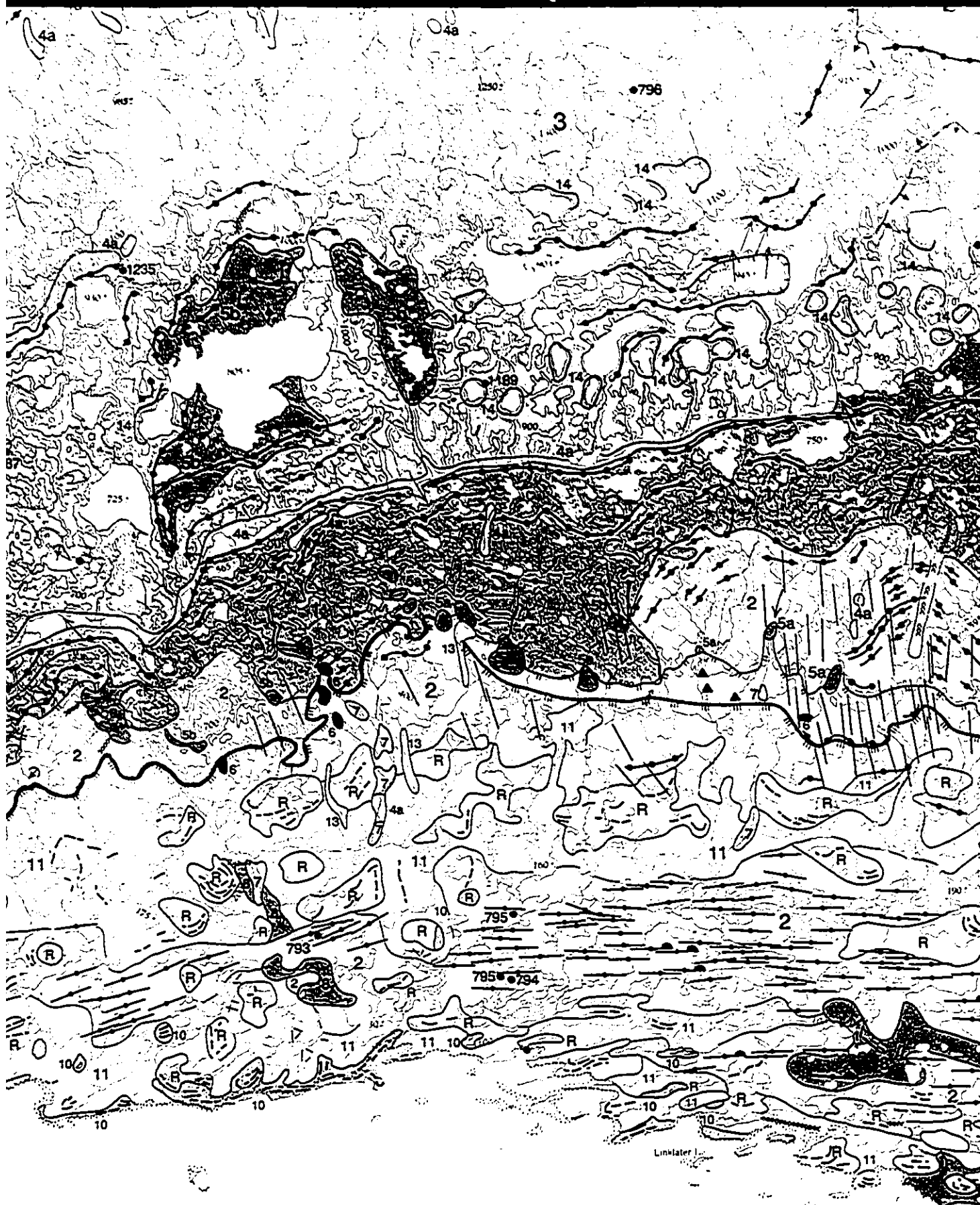
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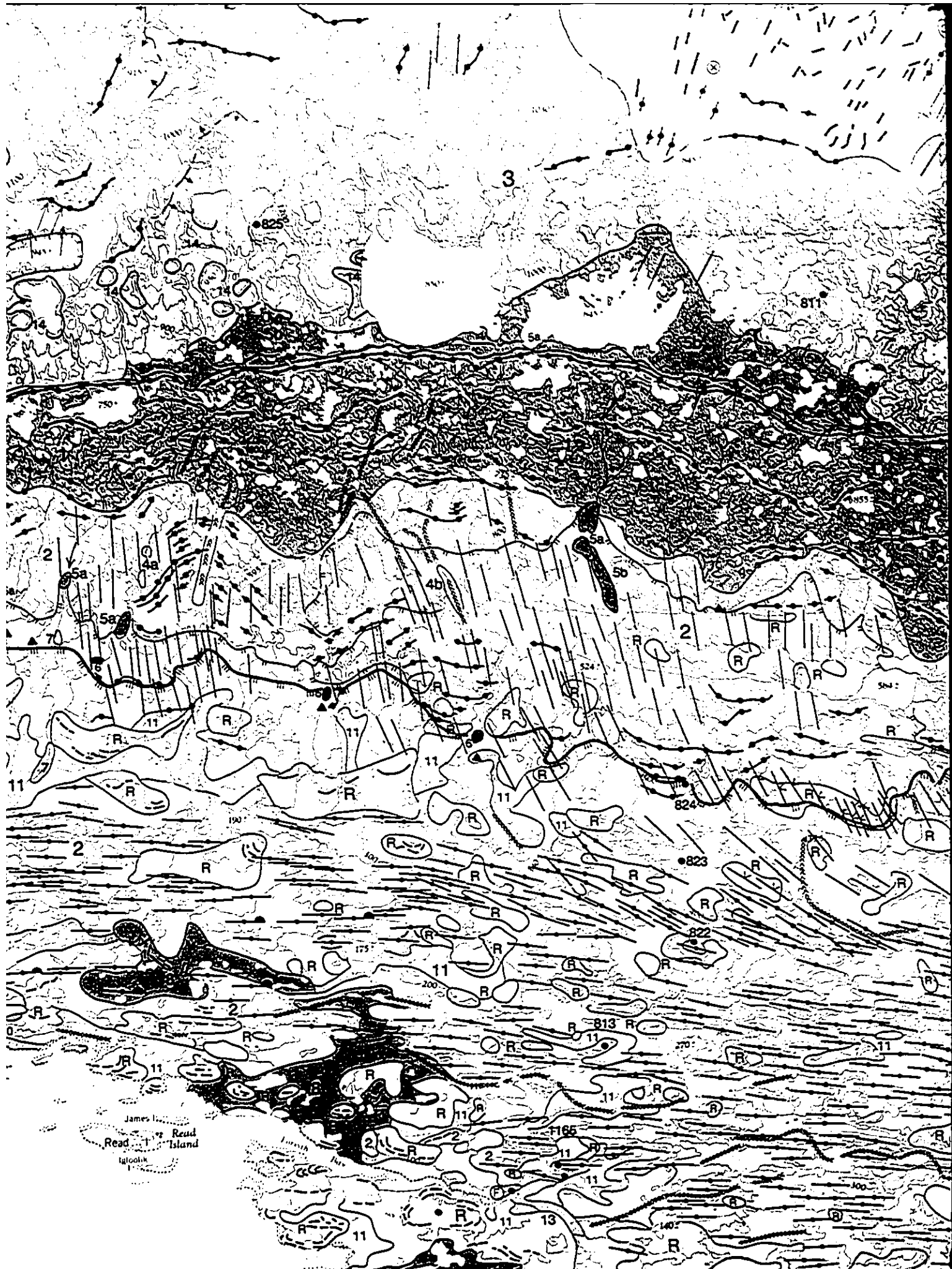
1-4m thick; occurs as flights
t spits

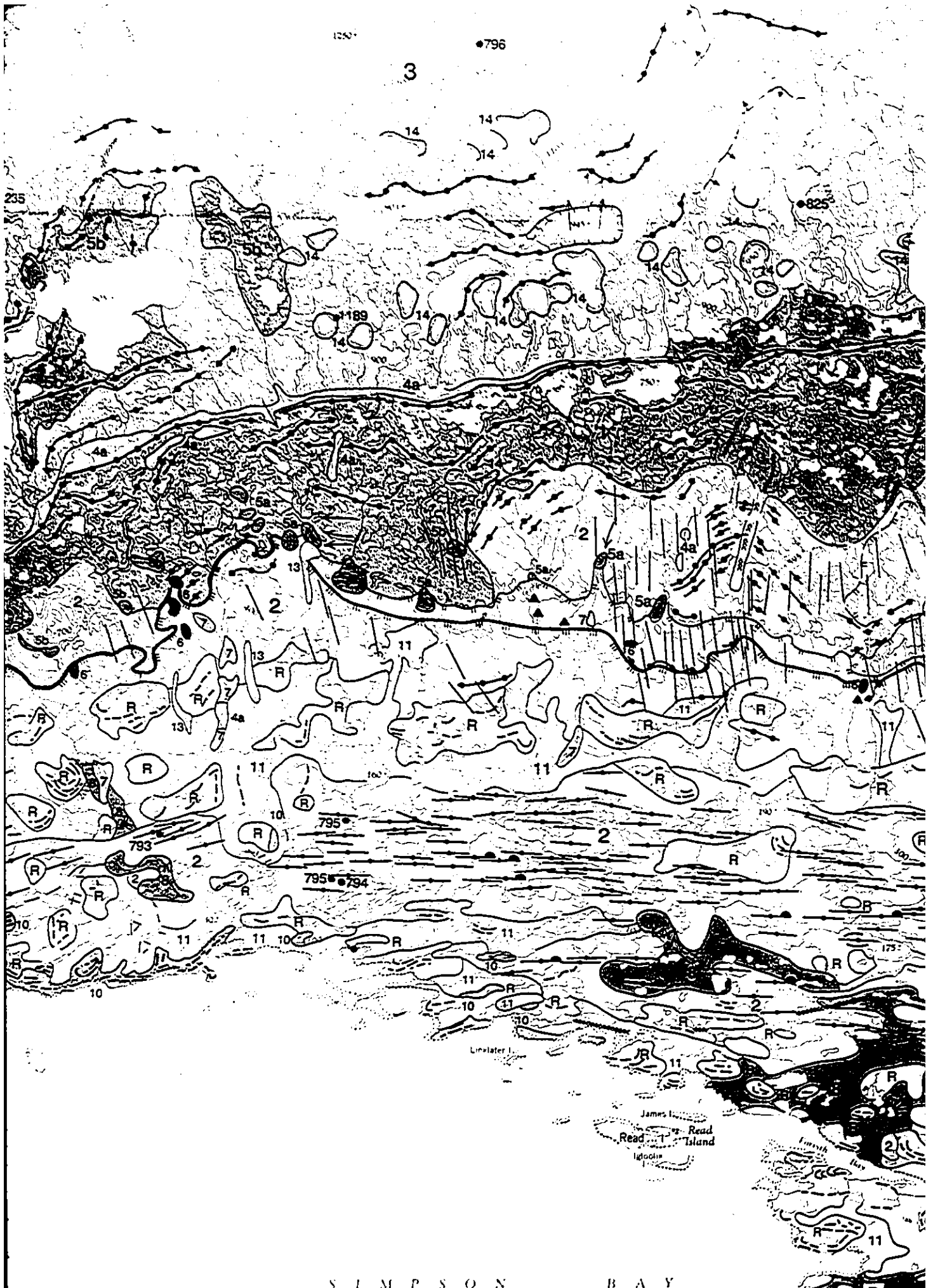






James I.
Read
Read
Island
Igloodik









30'

15'

69°00'

45'

13

Fluvial Deposits: gravel to gravelly sand and clay, and silt, sand and minor silt or clay, together with eolian and colluvial deposits on floodplains, 10-20m thick on alluvial terraces, and thin veneer, 1-2m on strath terraces; most wedges (1-2m deep) are composed of gravelly sediments

12

Felsenmeer: platy gravel size fragments, occurring as a blanket less than 1m thick on Paleozoic rock, found predominantly on upland plateaus and along strath terraces; rock heave is common where deposit is thin

MARINE DEPOSITS: well sorted and stratified to massive gravel, sand, silt, and clay deposited during regression of the postglacial sea; occurs as ridges and blankets on large coastal platforms and terraces (south coast) marking regressive sea level events, and as narrow and terraced forms grading from glaciomarine outwash terraces (north coast). Sediments may have been disturbed by pack ice and iceberg scour

11

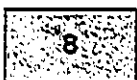
Undifferentiated: complex of silt and sandy silt on bedrock, diamicton, or locally gravel; commonly a veneer 1-3m thick

10

Littoral Deposits: gravel and gravelly sand; 1-4m thick; occurs as flights of raised strandlines, beaches (shingle) and spits



Tidal Deposits: silt, sand, with minor gravel and clay, about 1m thick; occurs locally at modern coast



Sublittoral Deposits: silty to sandy silt and clay, locally stony; 1-5m thick; occurs as offshore facies downslope from beach terraces; forms a blanket scoured by drifting ice

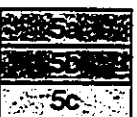
7

Deltaic Deposits: gravelly sand and sand; 1-5m thick; occurs as raised forms below marine limit; occurs as terraces along modern rivers; these deposits are areally much less significant than their glaciomarine counterpart

LATE WISCONSINAN



GLACIOMARINE DELTA/FAN DEPOSITS: coarse (10-30m thick) bouldery gravel to gravelly sand, and sand compose deltas; massive (10-20m thick) silt and clay in prodelta and fan settings; deltas occur as large raised features that grade from braided outwash systems and end moraines to steep-sloped delta/fan toe areas; glaciomarine deltas define marine limit whereas fans occur below marine limit



GLACIOLACUSTRINE DEPOSITS: 5a, gravelly sand; 1-5m thick; occurs as fans or deltas in shallow water; found in flat, dissected areas. 5b, bedded sand, silt, clay, and redistributed moraine (diamicton); 1-2m thick; deposited in short-lived lakes on hummocky moraine. 5c, interbedded sand and silt; 10-20m thick; present in plains and exposed in large pingos

4a,4b

GLACIOFLUVIAL DEPOSITS: 4a, gravel to sandy gravel and sand; 2-20m thick; occurs in terrace remnants and deltaic surfaces. 4b, gravel, gravelly sand, minor silt and clay; 10-20m thick; occurs as sharp-crested and flat-topped eskers



HUMMOCKY MORaine DEPOSITS: interbedded diamicton (till flows), sand, gravel, and silt; 10-50+(?)m thick; horizontal bedding common; diamicton is usual surface sediment displaying large polygons; large areas of hummocks and depressions are common; less common are linear (rim) ridges and moraine plateau which contain sand, gravel, and coarse bouldery gravel; originates both as subglacial and supraglacial sediment gravity flows or meltout; this redistribution of sediment by thaw slumping over buried (glacial?) ice continues today

2

SUBGLACIAL DEPOSITS: massive diamicton (till); 1-5m thick; commonly interbedded with (or underlain by) sand and gravel; fluting present

Geologic
Cliff in b
Flutings
Drumlins
Washove
Striae ice
Linear gl
Moraine
Ice-conta
De Geer
Moraine
Esker (di
Isolated
Meltwater
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Limit of
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Delta (sn
Pingo
Thermoka
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Fossil loc
Ice scour
Ground
Sample l
Ice press

fine sand and silt, sand
and colluvial deposits on
bed and the veneer, 1-2 m on
are common in gravel.

occurring as a blanket less
redominantly on upland
leave is common where

bed to massive gravel, sand,
of the postglacial sea; occurs
platforms and terraces (south
and as narrow and terraced
terraces (north coast).
back ice and iceberg scour

y silt on bedrock, diamicton,
1 m thick

; 1-4 m thick; occurs as flights
d spits

il and clay, about 1 m thick;

clay, locally stony; 1-5 m
from beach terraces; forms a

1-5 m thick; occurs as raised
as along modern rivers; these
than their glaciomarine

coarse (10-30 m thick) bouldery
e deltas; massive (10-20 m thick) silt
as occur as large raised
systems and end moraines to
nine deltas define marine limit

elly sand; 1-5 m thick; occurs
flat, dissected areas. 5b,
moraine (diamicton); 1-2 m
nmocky moraine. 5c,
resent in plains and exposed

sandy gravel and sand;
rd deltaic surfaces. 4b,
0-20 m thick; occurs as

added diamicton (till flows),
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ich contain sand, gravel, and
subglacial and supraglacial
distribution of sediment by
continues today

on (till); 1-5 m thick; commonly

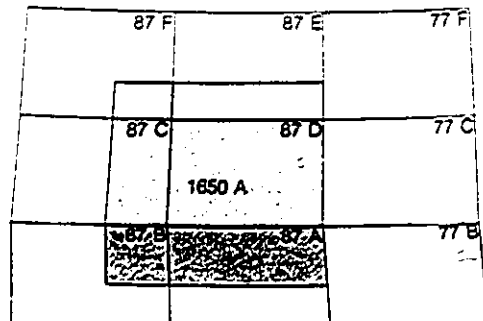
Geological boundary	
Cliff in bedrock	
Flutings	
Drumlins	
Washover bar (on drumlin) - current direction towards semi-circle	
Striae ice flow (direction known, unknown)	
Linear glacial mark on bedrock	
Moraine ridge	
Ice-contact face (teeth on ice side)	
De Geer moraine	
Moraine plateau	
Esker (direction of flow inferred; unknown)	
Isolated Kame	
Meltwater channel (ice controlled, ice not controlled)	
Limit of submergence (marine)	
Limit of submergence (lake)	
Raised strandline feature	
Delta (small)	
Pingo	
Thermokarst depression (stable)	
Active thaw slump	
Fossil locality (marine)	
Ice scour	
Ground observation	
Sample location	
Ice pressed ridges	

Geology by D.R. Sharpe and F.M. Nixon, 1982, 1983

Geology compiled by D.R. Sharpe 1984, 1985

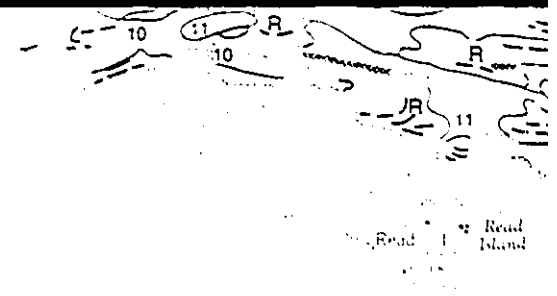
Geological cartography by Paul St-Amour Geological Survey of Canada

Any revisions or additional geological information known to the user
would be welcomed by the Geological Survey of Canada



NATIONAL TOPOGRAPHIC SYSTEM REFERENCE

10 10



N I M P S O N B A Y

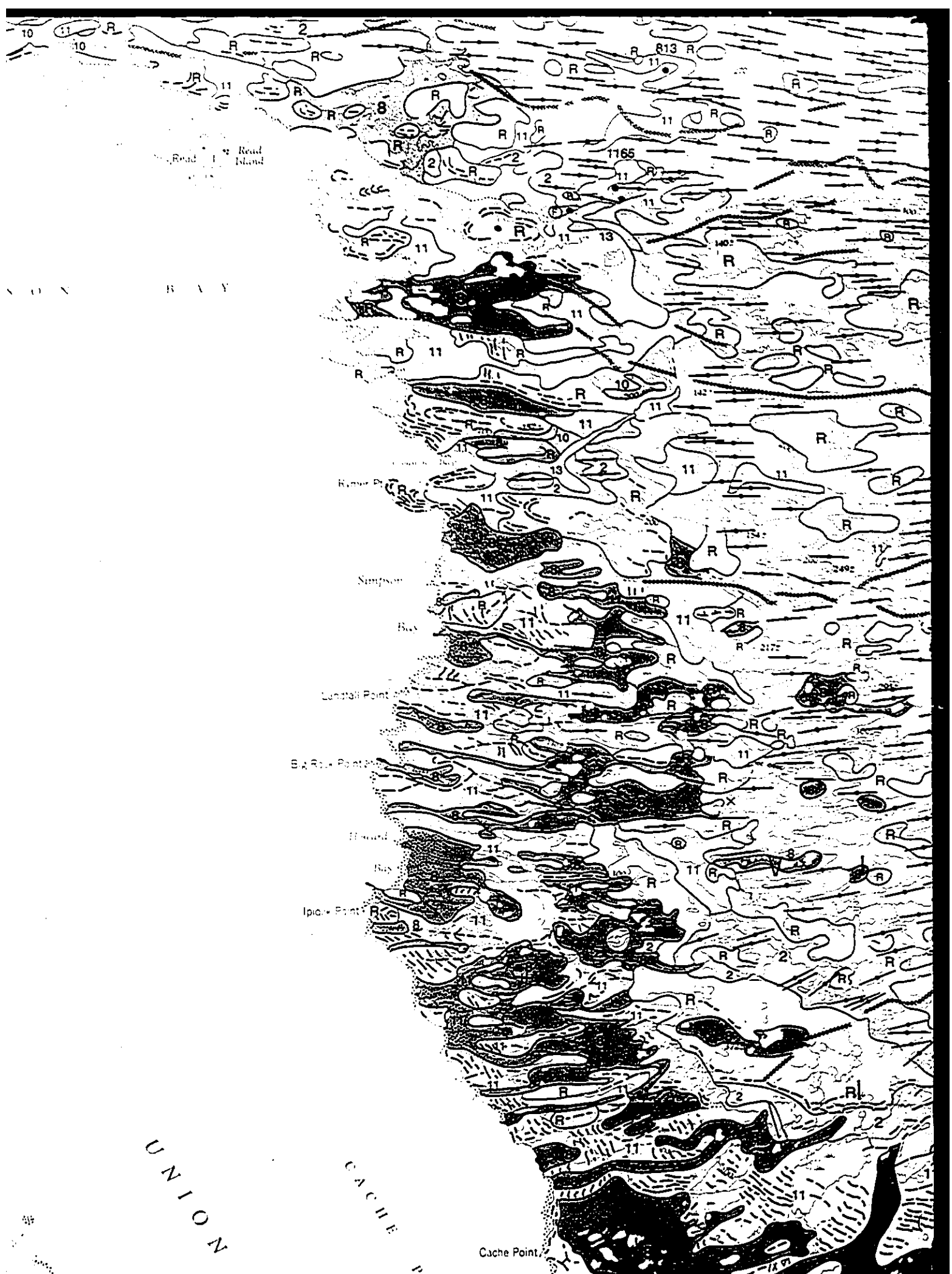
D O L P H I N

Banxland Inlet
Harems
Island
Sutton
Island
Liston

I N D



U N I O N



SIMPSON BAY

PHILIP

Bahia del Inlet

Waters Island

Sutton Island

Liston

AND

Ryder Pt

Simpson

Bay

Landfall Point

Big Rock Point

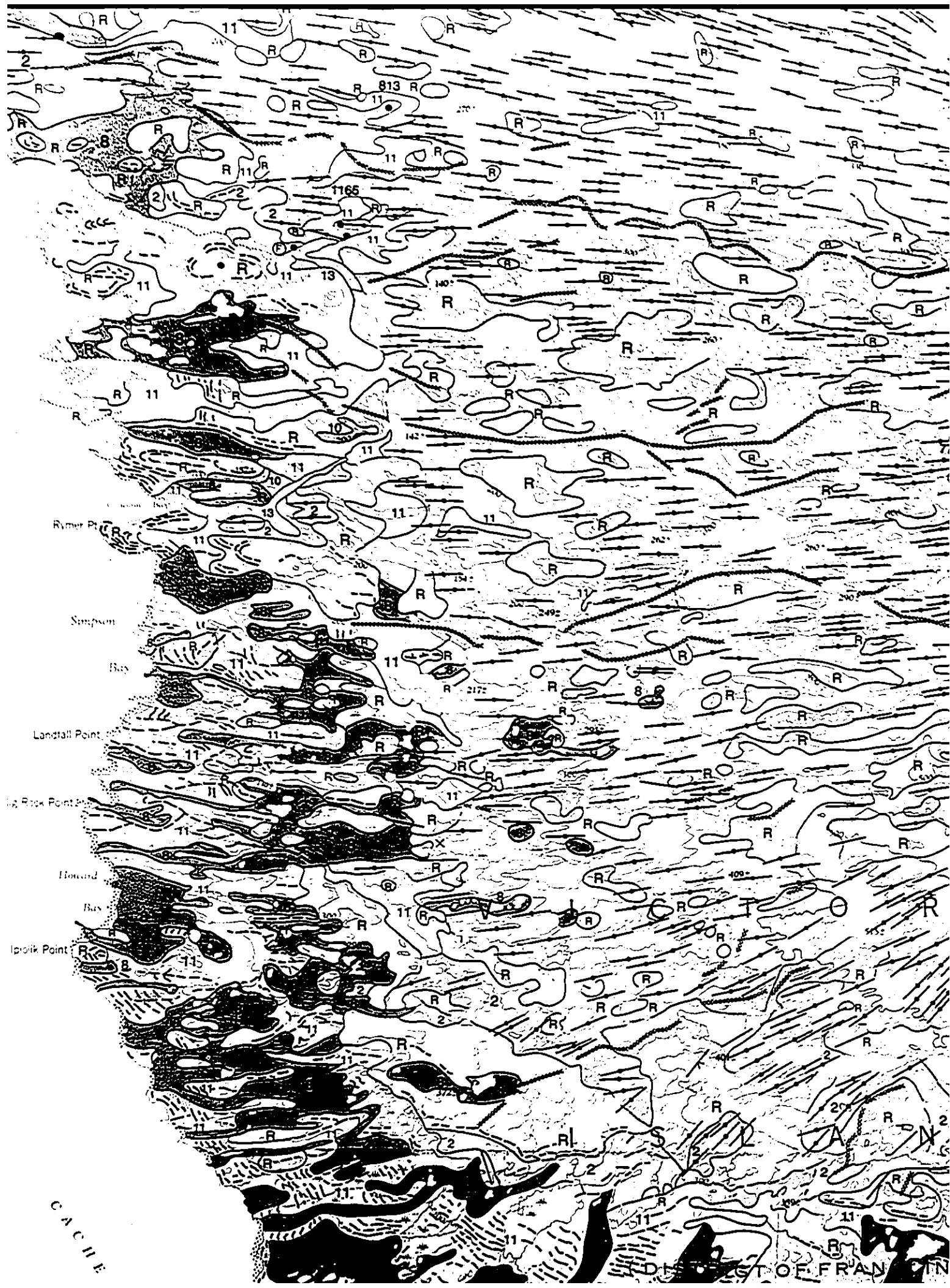
Holland

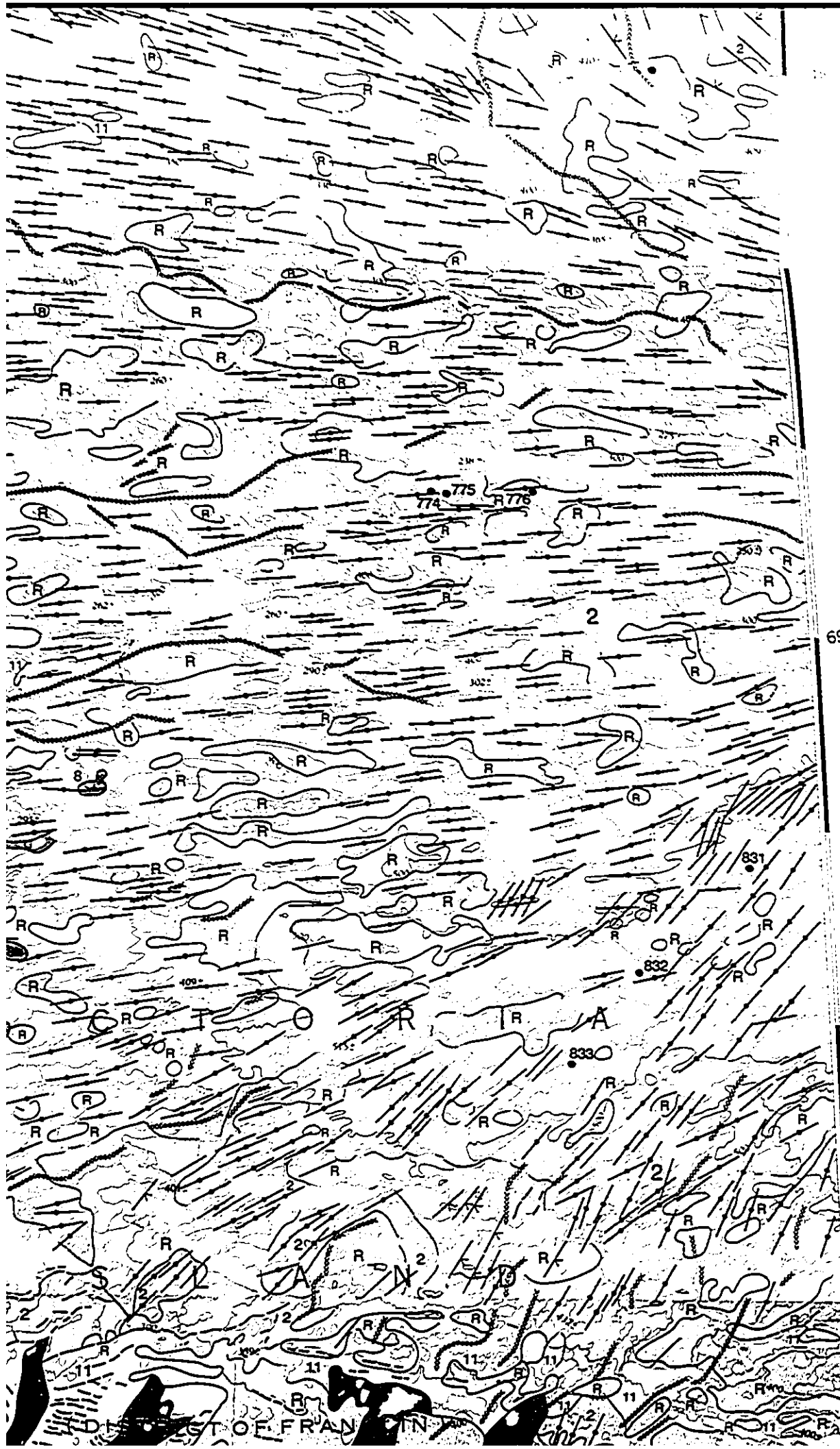
Bay

Ipswich Point

UNIO

CACHU



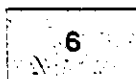


69°00'

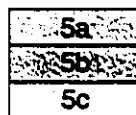
45'

DISTRICT OF FRANCONIA

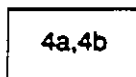
FIG. A. BOONSEN ET AL.



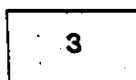
GLACIOFLUVIAL DEPOSITS: 6a, gravel to sandy gravel and sand; 2-20m thick; occurs in terrace remnants and deltaic surfaces. 6b, gravel, gravelly sand, minor silt and clay; 10-20m thick; occurs as sharp-crested and flat-topped eskers. 6c, gravel, gravelly sand, minor silt and clay; 10-20m thick; occurs as sharp-crested and flat-topped eskers. 6d, gravel, gravelly sand, minor silt and clay; 10-20m thick; occurs as sharp-crested and flat-topped eskers. 6e, gravel, gravelly sand, minor silt and clay; 10-20m thick; occurs as sharp-crested and flat-topped eskers.



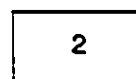
GLACIOFLUVIAL DEPOSITS: 5a, gravelly sand, 1-5m thick; occurs as fans or deltas in shallow water, found in flat, dissected areas. 5b, bedded sand, silt, clay, and redistributed moraine (diamictic), 1-2m thick; deposited in short-lived lakes on hummocky moraine. 5c, interbedded sand and silt; 10-20m thick; present in plains and exposed in large pingos.



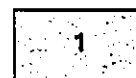
GLACIOFLUVIAL DEPOSITS: 4a, gravel to sandy gravel and sand; 2-20m thick; occurs in terrace remnants and deltaic surfaces. 4b, gravel, gravelly sand, minor silt and clay; 10-20m thick; occurs as sharp-crested and flat-topped eskers.



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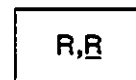


SUBGLACIAL DEPOSITS: massive diamictic (till); 1-5m thick; commonly interbedded with (or underlain by) sand and gravel; fluting present where drift is thin (1-2m), and drumlins occur where drift is thick (10-15m); large areas of flat terrain, commonly bedrock controlled; deposited subglacially by meltout, sediment gravity flow, meltwater flow, and minor lodgment.

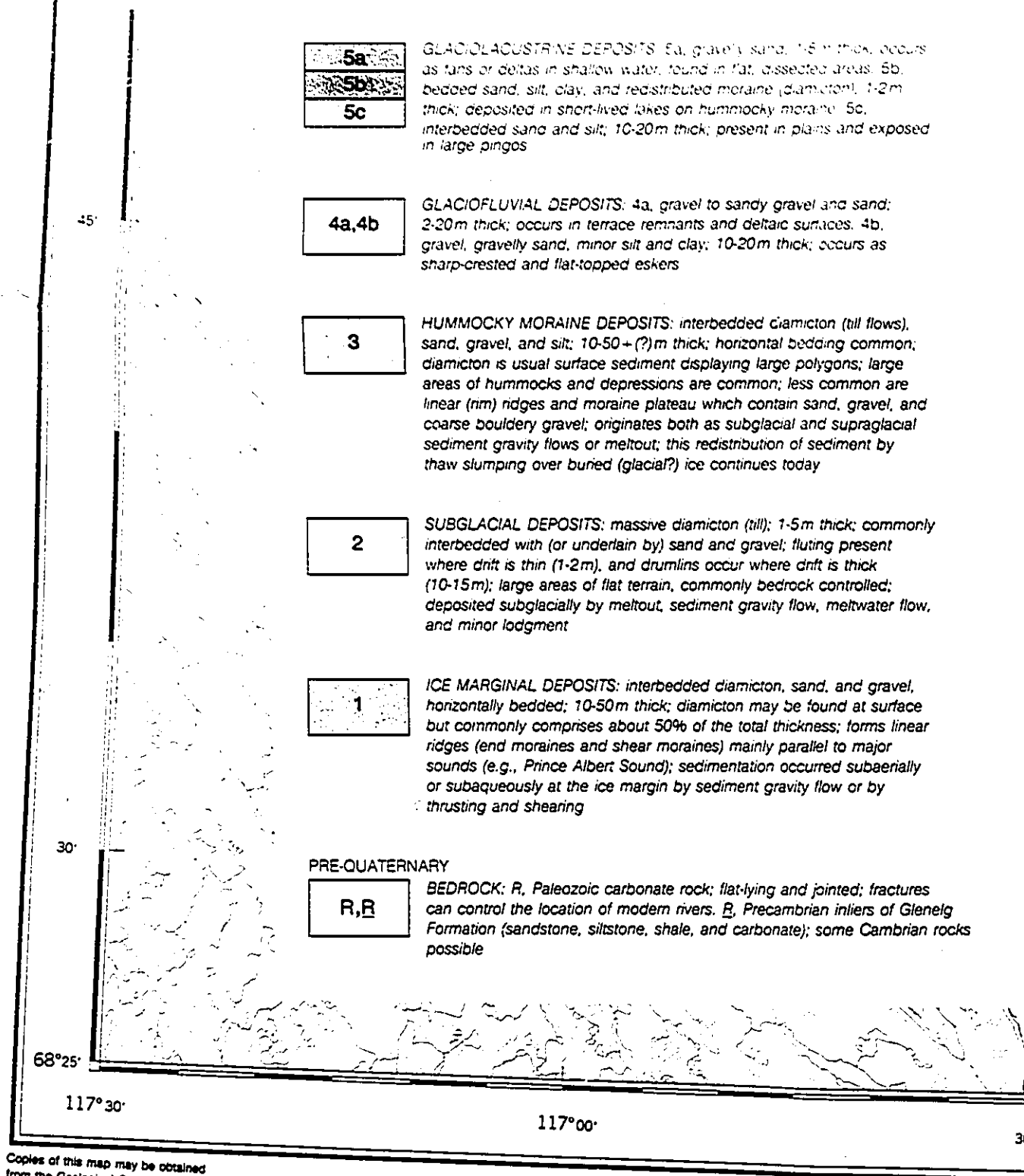


ICE MARGINAL DEPOSITS: interbedded diamictic, sand, and gravel, horizontally bedded; 10-50m thick; diamictic may be found at surface but commonly comprises about 50% of the total thickness; forms linear ridges (end moraines and shear moraines) mainly parallel to major sounds (e.g., Prince Albert Sound); sedimentation occurred subaerially or subaqueously at the ice margin by sediment gravity flow or by thrusting and shearing.

PRE-QUATERNARY



BEDROCK: R, Paleozoic carbonate rock; flat-lying and jointed; fractures can control the location of modern rivers. R, Precambrian inliers of Glenelg Formation (sandstone, siltstone, shale, and carbonate); some Cambrian rocks possible.



Copies of this map may be obtained from the Geological Survey of Canada: 601 Booth Street, Ottawa, Ontario K1A 0E8 3303-33rd Street, N.W., Calgary, Alberta T2L 2A7

Canada

to sandy gravel and sand;
and deltaic surfaces. 4b.
; 10-20m thick; occurs as

tion, sand, and gravel, may be found at surface the total thickness; forms linear es) mainly parallel to major imentation occurred subaerially ediment gravity flow or by

s. B, Precambrian inliers of Glenelg and carbonate); some Cambrian rocks

Method used _____
 Date of sampling _____
 Name of collector(s) _____
 Host(s) (plant, animal, etc.) _____
 Distribution _____
 Range _____
 Thermohaline depression station _____
 Active thaw slump _____
 Fossil locality (manner) _____
 Ice scour _____
 Ground observation _____
 Sample location _____
 Ice pressed ridges _____

Geology by D.R. Sharpe and F.M. Nixon, 1982, 1983

Geology compiled by D.R. Sharpe 1984, 1985

Geological cartography by Paul St-Amour Geological Survey of Canada

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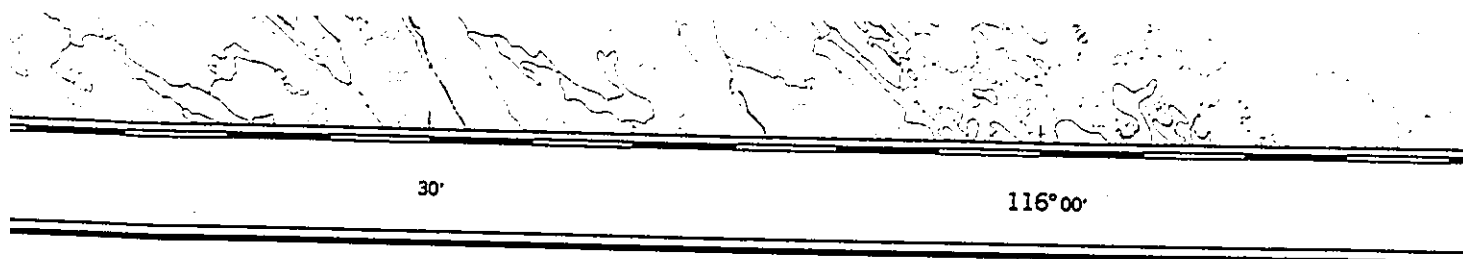
Base map from parts of maps published at the same scale
by the Army Survey Establishment,
R.C.E. in 1957-1962, 1966, 1982

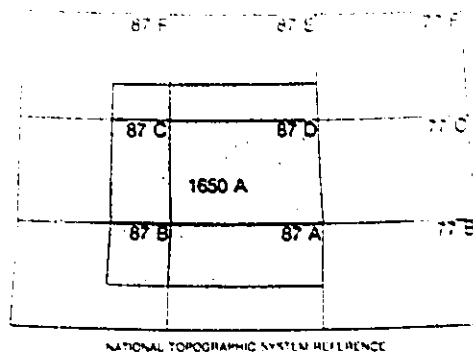
Copies of the topographical editions covering this map area may be obtained from the Canada Map Office, Department of Energy, Mines and Resources, Ottawa, Ontario, K1A 0E9

Mean magnetic declination 1988, 38°42' East, decreasing 27.3' annually. Readings vary from 38°50' in the SW corner to 40°34' in the NE corner of the map area.

The proximity of the North Magnetic Pole causes the magnetic compass to be erratic in this area

Elevations in feet above mean sea level





NATIONAL TOPOGRAPHIC SYSTEM REFERENCE

Wixon, 1982, 1983

1964, 1985

Geological Survey of Canada

Information known to the user
Geological Survey of Canada

Map at the same scale
Sheetment,
1982

Map of this map area may be
obtained from the Department of
Geology, Ottawa, K1A 0E9

Map 42' East, decreasing
10' in the SW corner to
the map area.

Map of Pole causes the
map in this area

Map in sea level

MAP 1650A

SURFICIAL GEOLOGY

WOLLASTON PENINSULA

(Victoria Island)

DISTRICT OF FRANKLIN

NORTHWEST TERRITORIES

Scale 1:250 000 - Échelle 1/250 000

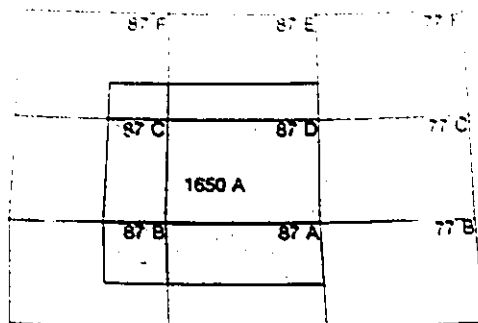
Kilometres 5 0 5 10 15 20 Kilom

Transverse Mercator Projection
CM 117°. Scale Factor 0.9996
© Crown copyrights reserved

116° 00'

30'

115° 00'



NATIONAL TOPOGRAPHIC SYSTEM REFERENCE

MAP 1650A

SURFICIAL GEOLOGY

WOLLASTON PENINSULA

(Victoria Island)

DISTRICT OF FRANKLIN

NORTHWEST TERRITORIES

Scale 1:250 000 - Échelle 1/250 000

Kilometres 5 0 5 10 15 20 Kilomètres

Transverse Mercator Projection

CM 117°, Scale Factor 0.9996

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30°

115° 00'

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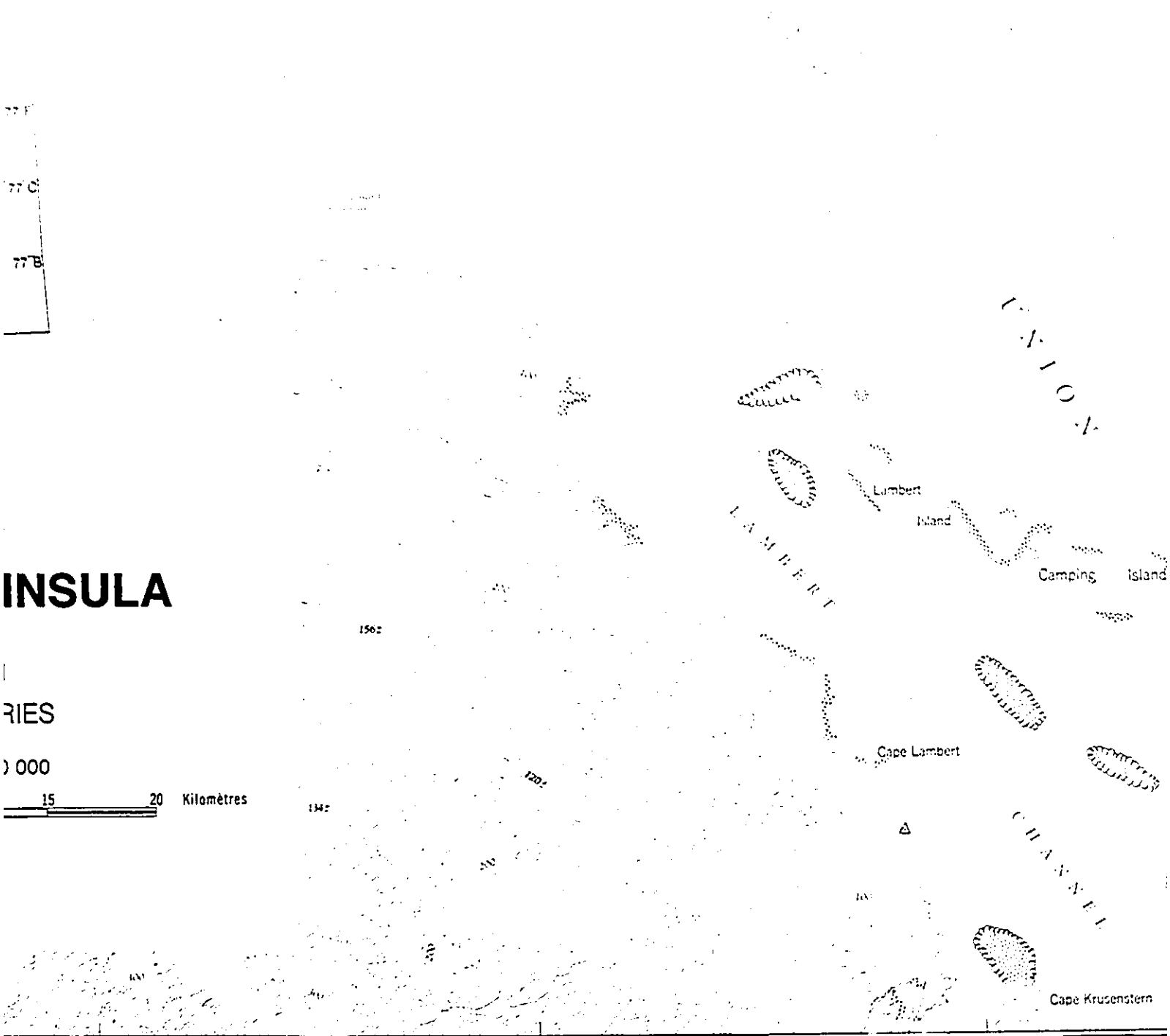
77° E
77° C
77° B

INSULA

RIES

0 000

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115° 00'

30'

114° 00'



NOTE: Islands off the shore of the District of Mackenzie are with the District of Franklin.

114°00'

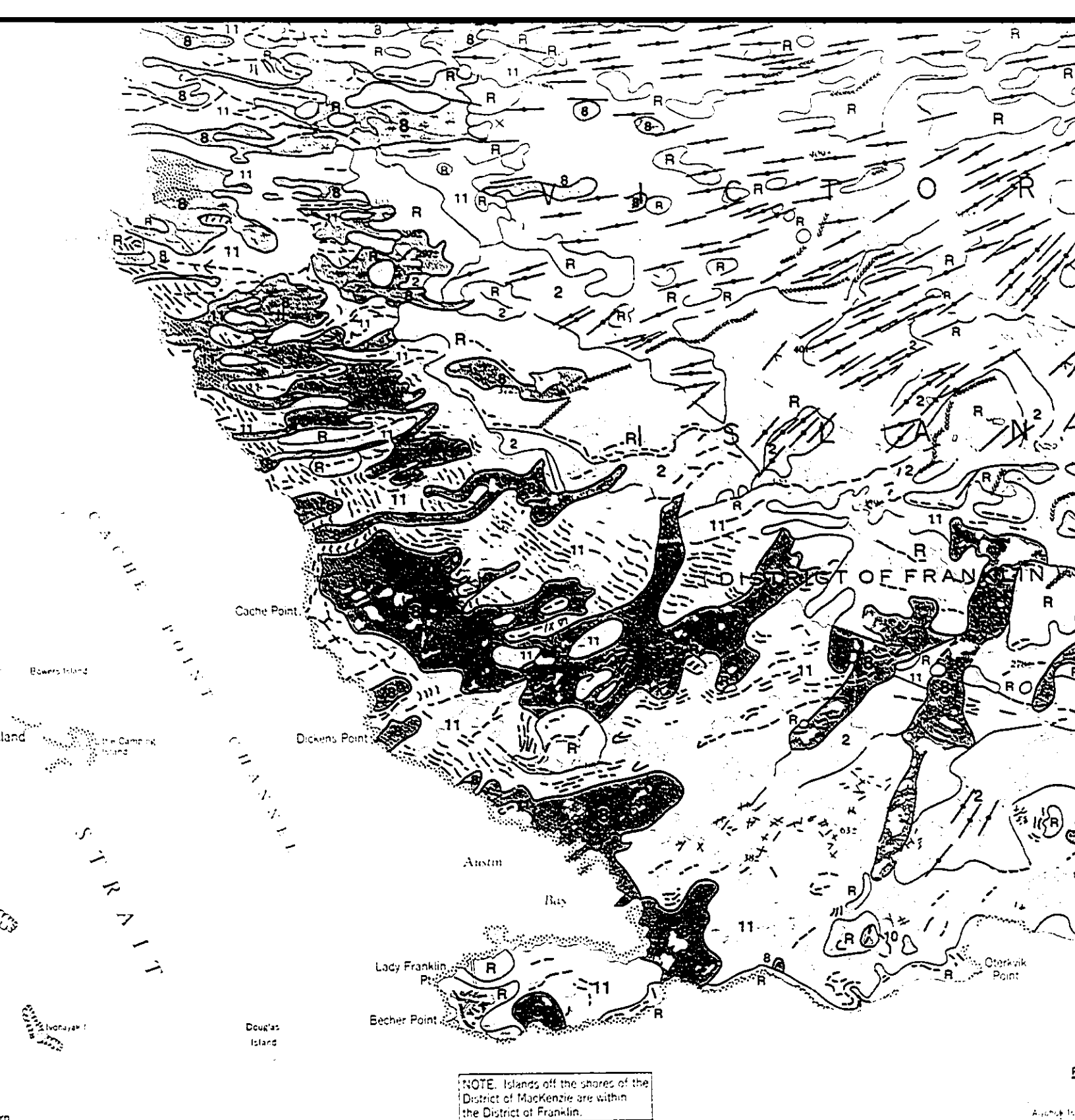
30'



30'

114°00'

30'



NOTE. Islands off the shores of the District of MacKenzie are within the District of Franklin.

30°

113° 00'

30°



Recommended citation:

Sharpe, D.R. and Nixon, F.M.

1989: Surficial geology.

Wollaston Peninsula, (Victoria Island),
District of Franklin, Northwest Territories,
Geological Survey of Canada,
Map 1650A, scale 1:250 000