QUATERNARY GEOGRAPHY AND SEDIMENTOLOGY OF THE HUMBER RIVER BASIN AND ADJACENT AREAS

M. J. Batterson

Report 03-02

St. John's, Newfoundland
2003

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GOVERNMENT OF
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Cover

Aerial view looking east along the Humber River valley below Deer Lake. The valley captures the major elements of landscape modification during the Quaternary. The valley was a conduit for westward-flowing ice during the late Wisconsinan. During deglaciation, ice receded up the valley by calving into a raised postglacial sea that inundated the Deer Lake basin. The ice retreat halted long enough at the mouth of modern Deer Lake to form the ice-contact delta seen in the background. Falling relative sea levels eventually isolated Deer Lake from the sea and the lower Humber River was established.

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MAP

Map 2001-42. Surficial Geology of the Humber River Basin and Surrounding areas (parts of NTS 12H and 12A), Newfoundland. . . . . back pocket

ABSTRACT

During the Quaternary, the Humber River basin in western Newfoundland was completely glaciated. Glacial erosional features suggest an early southward flow from a source north of the basin, which covered the coastal margins in the western part of the basin, including the Harrys River valley. Subsequent regional ice flow was southwestward to northwestward from a dispersal centre on The Topsails. South- to southwestward-flowing ice from the Long Range Mountains occupied the Upper Humber River valley; this flow was confluent with ice from The Topsails flowing northwestward toward Bonne Bay.

Ice retreated from the inner coast about 13 000 BP. During retreat, the ice occupying the Deer Lake valley dammed a proglacial lake in the adjacent Grand Lake basin, up to 85 m above the present lake level, as indicated by strandlines found on the west side, and deltas on the east. This lake, named glacial Lake Howley, drained through its western end into the Harrys River valley and subsequently into northern St. George's Bay. The lake was lowered by exposure of the South Brook valley outlet, and finally drained, catastrophically, through a spillway at Junction Brook.

Marine incursion accompanied glacial retreat in the Deer Lake valley. Marine limit at the coast was 60 m asl, based on the elevation of a delta in the Hughes Brook valley. Inland deltas found at the head of Deer Lake and fine-grained sediment exposed within the Deer Lake valley show inundation below 45 m modern elevation. Dated marine macrofossils in Humber Arm and the Lower Humber River valley suggest the deltas at the head of Deer Lake formed at about 12 500 BP.

INTRODUCTION

OVERVIEW OF THE HUMBER RIVER BASIN

This report represents the culmination of several years of study into the Quaternary geology of the Humber River basin and is part of an ongoing systematic investigation of the Quaternary geology of the Island of Newfoundland. Initial research began in 1989 as a reconnaissance project to resolve the reported opposing interpretations of ice-flow history in the Upper Humber River basin. It quickly became apparent that the basin had a very complex glacial history, potentially being a conduit for ice from dispersal centres on the Long Range Mountains to the north, The Topsails to the east and remnant ice centres on Birchy Ridge. Impetus for the research resulted from, a) the recognition that a large inland proglacial lake had existed during deglaciation, and b) the discovery of thick sequences of waterlain sediments up to 50 km inland of the present coast. The Quaternary stratigraphy, and resolution of the ice-flow history of the Humber River basin has allowed significant reinterpretation of the deglacial and early Holocene history of this part of the west Newfoundland coast. It also has broader implications for the Quaternary chronostratigraphic framework of Atlantic Canada.

LOCATION

The study area extends between 48°30' and 49°47'N, and from 56°30' to 57°54'W, and includes all, or parts, of 17, 1:50 000-scale NTS map areas (12A/11, 12, 13, 14, 15; 12H/2, 3, 4, 5, 6, 7, 8, 10, 11, 12, 13, and 12B/9, Figure 1). It includes the area defined by the Humber River basin and smaller catchments along the western margin of the basin, including Corner Brook, Hughes Brook, Old Mans Brook, and Goose Arm Brook (Figure 1). The Humber River basin covers an area of about 4400 km^2, has a maximum length of 145 km, from Little Grand Lake in the south to within 10 km of the coast near St. Paul's Inlet in the north and is 105 km wide. The Humber River valley occupies the largest part of the basin and is divided into the Upper Humber River occupying that area north of Deer Lake between Birchy Ridge and the Long Range Mountains, and the Lower Humber River occupying that area between Deer Lake and the mouth at Corner Brook. Other components of the basin are the Deer Lake valley, Grand Lake valley, and Birchy Lake valley, all occupied by their respective lakes (Figure 1). Much of the basin lies below 100 m elevation, and includes a series of interconnected subbasins that impound several of Newfoundland's major lakes, including Birchy Lake, Deer Lake, Grand Lake, Hinds Lake, Sandy Lake and Sheffield Lake.

Much of the area is easily accessible by a network of paved and gravel roads. The Trans-Canada Highway (TCH) traverses the central part, and other roads extend north and south from it. Well-maintained gravel roads, constructed to support logging operations, cover many of the intervening areas. The shores of Grand Lake are accessible by boat from Howley and Northern Harbour. Part of The Topsails may be reached by vehicle along the bed of the now-defunct Newfoundland railway. Some areas, such as the summits of the Long Range Mountains, are accessible only by foot or helicopter.

An earlier map of Newfoundland by Cram (1900; Figure 2) shows the communities of Deer Lake, Cormack, Harvey (now Pasadena), and Bay of Islands (now Corner Brook). Birchy Pond (now Birchy Lake) at an elevation of 38 m was joined by a narrow channel to Sandy Pond (now Sandy Lake). Sandy Lake was considerably smaller than its present equivalent, and was joined to Grand Pond (now Grand Lake) at 35 m by The Main Brook, a 13-km long stream entering Grand Lake near Howley. Howley lay approximately 5 km north of the 1900 shores of Grand Lake. Raising lake levels resulting from construction of the Junction Brook dam enlarged Sandy Lake to include the Main Brook valley and the lowlands northeast of Howley. The dam also produced a slight increase in the surface area of Grand Lake, by inundating the lowlands that lay west of Howley. Although some elevations on the Cram map are accurate, such as Notched Mountain at about 490 m, many are inaccurate (e.g., Mt. Sykes was mapped about 122 m lower than its actual elevation), including the surface elevations of Grand, Birchy and Sandy lakes. Nevertheless, the distribution of land-surface features is generally consistent with recently mapped configurations.

REGIONAL GEOLOGY

A knowledge of the distributions of the bedrock (Figure 3) provides the data necessary to relate erratic clasts to their sources. Rock types that are visually distinctive and confined to a discrete source area are particularly useful in aiding determination of distances and directions of glacial transport.

The centre of the study area is an interior basin underlain by Carboniferous flat-lying terrestrial, fluvial and lacustrine rocks of the Deer Lake Group (Hyde, 1979, 1984). These consist of red conglomerate, sandstone, and siltstone (North Brook Formation), grading upward to well-bedded to laminated grey to red siltstone and mudstone containing thin limestone beds (Rocky Brook Formation). Coarse arkosic sandstone, pebble conglomerate and sandstone (Humber Falls Formation) compose Birchy Ridge, on the eastern margin of the Humber River basin.

The northwestern part of the Deer Lake basin is flanked by the Long Range Mountains and are mostly underlain by

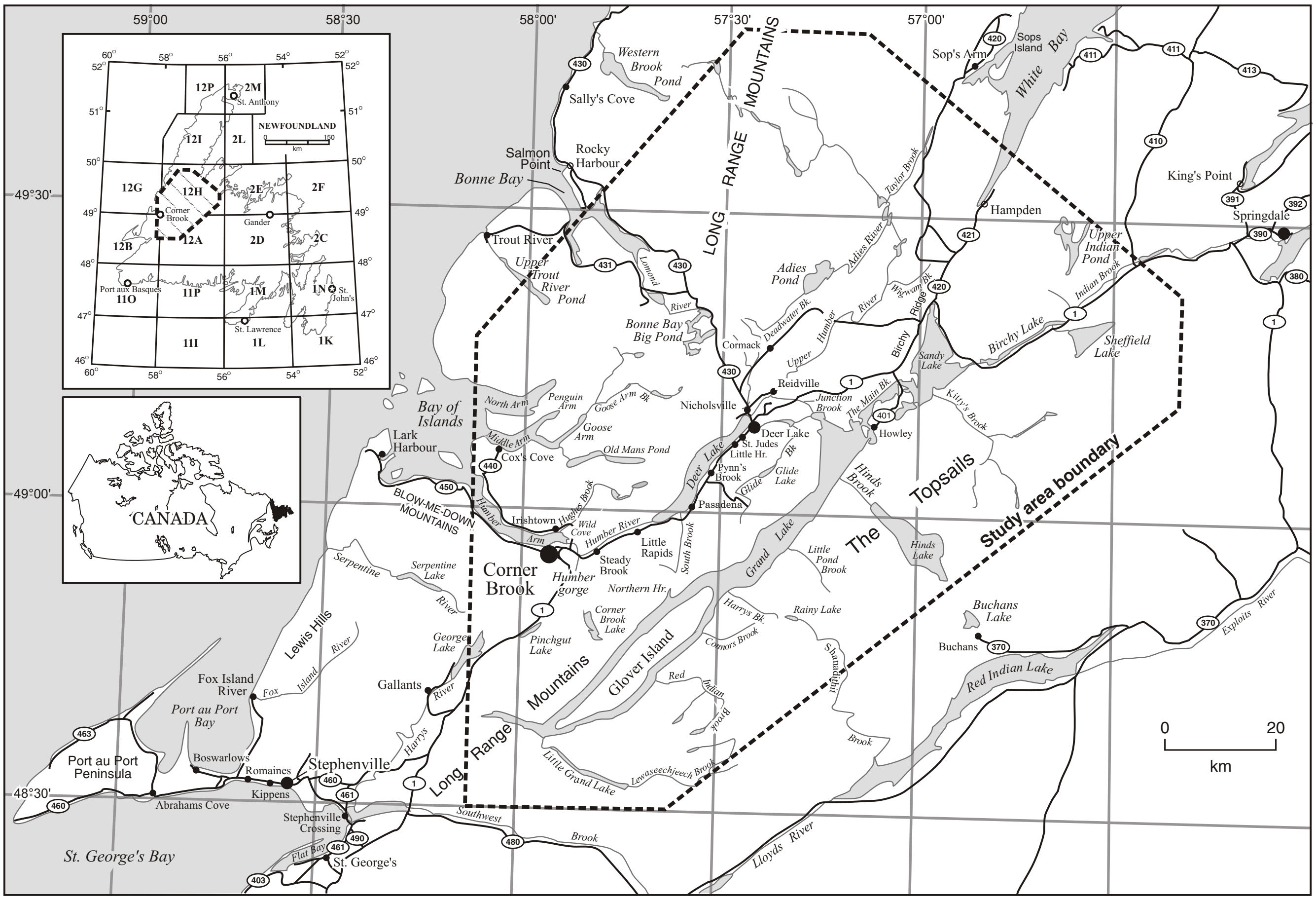


Figure 1. *Location of study area and map of places mentioned in text.*

Proterozoic gneiss and granitic gneiss (Erdmer, 1986; Owen, 1986). Proterozoic medium- to coarse-grained metagabbro outcrops northwest of Taylor Brook (Owen, *op. cit.*), and a small area of Paleoproterozoic to Middle Ordovician quartzite outcrops north of Adies Pond (Hyde, 1979). Proterozoic gneiss and granitic gneiss are also found at the western end of Grand Lake.

West of Deer Lake are Cambrian to Ordovician platformal rocks (Williams and Cawood, 1989) that include limestone and dolostone (St. George Group, Port au Port Group, Reluctant Head Formation), and shale, greywacke and conglomerates (Curling Group, Old Man's Pond group). Ophiolitic rocks of the Humber Arm Allochthon that formed during the Taconic Orogeny, compose the uplands along the outer coast that form the Lewis Hills, Blow-Me-Down Hills, North Arm Mountain, Table Mountain, and Lookout Hills massifs. Harzburgite, dunite, gabbro, and pillow basalt are the common rock types.

A large area of psammitic and pelitic schist (Mount Musgrave group) is exposed along the southwestern margin of the Deer Lake basin, and along the north shore of Grand Lake, east of Northern Harbour. Glover Island and the adjacent south side of Grand Lake are mostly basalt and tuff (Glover group). Silurian volcanic and intrusive rocks (Whalen and Currie, 1988; Saunders and Smyth, 1990) dominate the area to the east and north, including the east side of Grand Lake, Sandy Lake and White Bay (Figure 4). Units of particular interest include one- and two-feldspar granites (i.e., potassium-feldspar (e.g., orthoclase) and/or sodium-feldspar (e.g., albite)), some of which are peralkaline (Topsails intrusive suite); flow-banded rhyolite, tuff and basalt (Springdale Group); and gabbro, granodiorite and diorite (Hungry Mountain Complex).

Glacial movement from any of the centres surrounding the Deer Lake basin would be indicated by the dispersal pattern of clasts in sediments. Material transported by southward-flowing ice from the Long Range Mountains should contain gneiss and granitic gneiss clasts. Depending on the flow path, the sediment may also contain clasts derived from the gabbro near Taylor Brook, the quartzite near Adies Pond, or Devonian intrusive rocks that outcrop outside the Humber River basin, west of Sops Arm, such as the Gull Lake intrusive suite or the Devils Room granite.

Dispersal into the Deer Lake basin from The Topsails would, depending on the location of ice-dispersal centres, potentially include clasts from a variety of granites, includ-

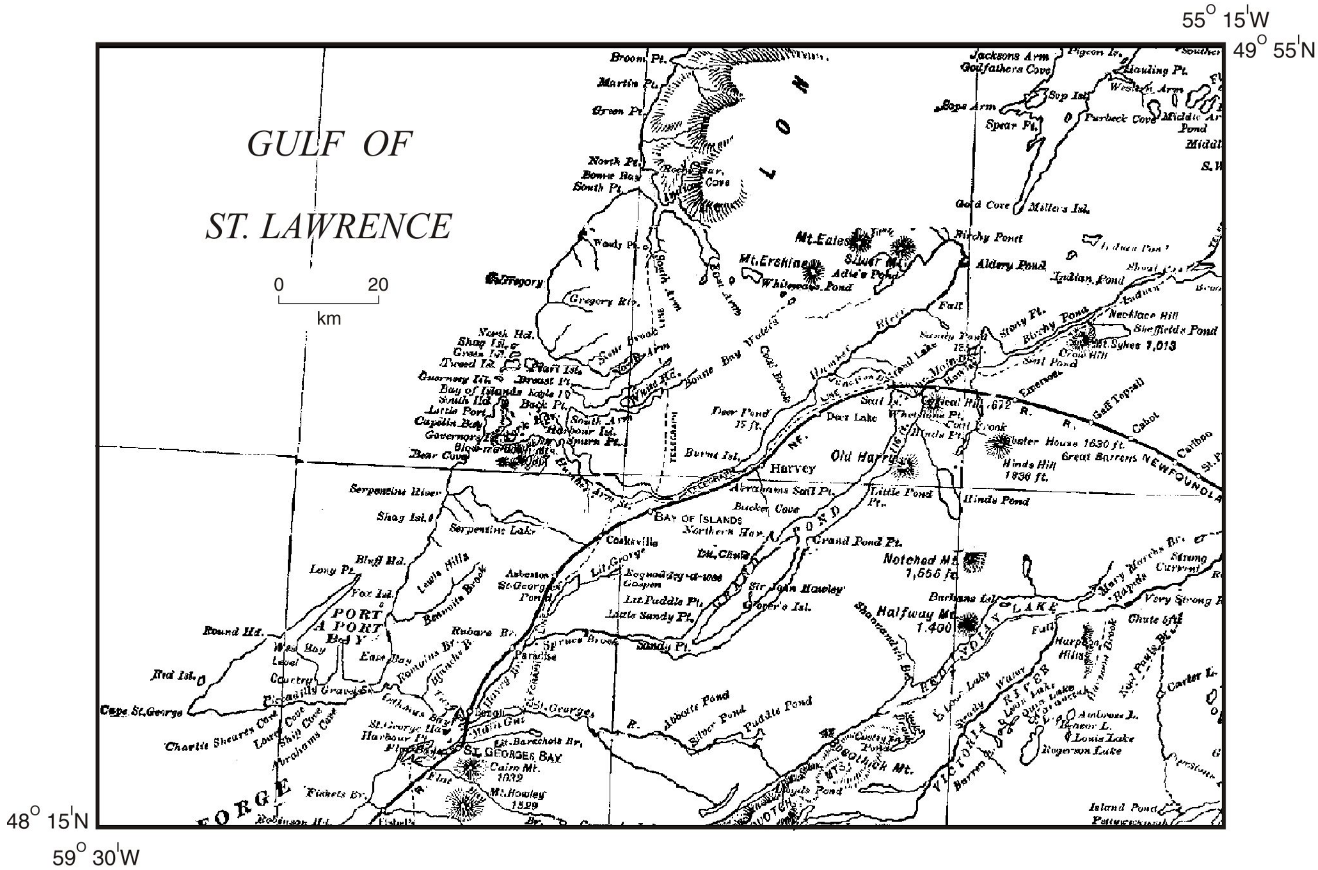


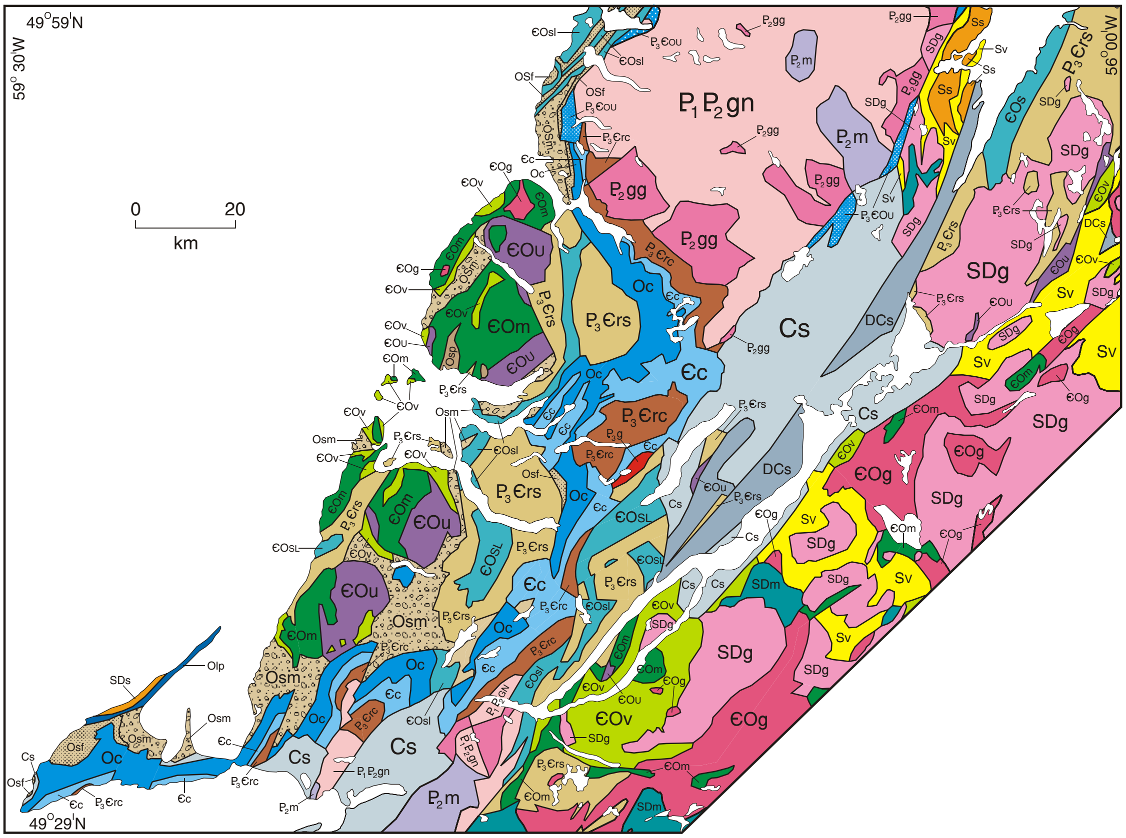
Figure 2. *Excerpt from Cram (1900) showing geography of western Newfoundland ca. 1900.*

ing peralkaline units. In addition, clasts from gabbro, rhyolite, basalt, gneiss, and other intrusive and volcanic rocks could be present.

There is a potential overlap in clast assemblages derived from these two source areas. Clasts derived from units that are, superficially and petrographically similar, are differentiated in hand specimen based on their mineralogical composition (e.g., gabbro exposed near Taylor Brook is dominantly a medium- to coarse-grained, mesocratic, pyroxene ± olivine ± amphibole metagabbro (Owen, 1986); in contrast, the gabbro of the Hungry Mountain Complex is a fine- to coarse-grained hornblende gabbro.) Gabbro of the Topsail intrusive suite has elongate clinopyroxene crystals having amphibole rims, euhedral plagioclase, and interstitial K-feldspar (Whalen and Currie, 1988).

Granites found in the igneous terrane west of White Bay have distinctly different physical properties from those granites found on The Topsails. In the White Bay area, the Moose Lake pluton is a pink to red, coarse-grained to megacrystic biotite granite (Saunders and Smyth, 1990). The Devils Room granite has pink to white K-feldspar megacrysts up to 25 cm long containing minor quartz (Saunders and Smyth, 1990). In contrast, granites exposed on The Topsails are extremely variable. The Hinds Brook granite, cropping out between Hinds Lake and Sandy Lake, is a white to pink, medium- to coarse-grained biotite–amphibole, K-feldspar, porphyritic two-feldspar granite. Granite rocks of the Topsail intrusive suite are white to red, fine-to medium-grained, equigranular, biotite ± amphibole, one- or two-feldspar granites, in large part peralkaline. In particular, the granite (Sp) that crops out over a large part of The Topsails from Lewaseechjeech Brook, south of Glover Island, to the vicinity of Sheffield Lake (*see* Figure 6) is a peralkaline, coarse-grained, amphibole granite, containing prominent quartz grains and a distinctive interstitial habit to the mafic minerals (Whalen and Currie, 1988).

Similarly, distinctive rhyolite clasts can be identified. Rhyolites found in the Natlin's Cove Formation (Smyth and Schillereff, 1982), south of Sops Arm, are pink to dark grey and contain flow-banding defined by opaque minerals, and quartz and feldspar phenocrysts. Rhyolites on The Topsails are assigned to the Springdale Group (*see* Figure 6) that covers a large area extending northeast from Shanuainrit Brook toward White Bay. The Springdale Group rhyolites include red, almost structureless rocks having little layering or flow-banding; orange to red units of sparse, small phenocrysts; rhyolites mixed with globules of basalt in a lava flow; brec-



LEGEND

Carboniferous (360 to 286 Ma)

Cs — Fluviatile and lacustrine sandstone, siltstone, conglomerate and minor carbonate rocks (Deer Lake Group, Howley Formation)

Devonian to Carboniferous (408 to 286 Ma)

DCs — Fluviatile and lacustrine sandstone, shale, conglomerate and minor carbonate rocks (Anguille Group)

Silurian to Devonian (438 to 360 Ma)

SDs — Shallow marine sandstone, conglomerate, limey shale and thin-bedded limestone.

SDg — Gabbro-syenite-granite-peralkaline granite intrusive suites and unseparated volcanic rocks

SDm — Gabbro and diorite intrusions, including minor ultramafic phases

Silurian (438 to 408 Ma)

Sv — Felsic volcanic rocks (includes unseparated sedimentary rocks)

Ss — Shallow marine and non-marine siliciclastic sedimentary rocks, inclucing sandstone, shale and conglomerate

Middle Ordovician (480 to 460 Ma)

Olp — Shallow marine limestone, shale and sandstone

Osp — Breccia, shale and sandstone

Early to Middle Ordovician (505 to 460 Ma)

Osf — Sandstone and shale turbidites, and carbonate breccia. Some calcareous black shale

Osm — Shale-matrix mélange, containing blocks of ophiolite rocks

Oc — Limestone and dolostone (St. George Group)

Cambrian to Middle Ordovician (570 to 460 Ma)

ЄOsl — Continental slope facies sandstone, shale, carbonates, chert, mafic volcanic rocks (includes part of Curling and Mount Musgrave groups)

ЄOv — Submarine volcanic rocks (includes Glover group)

Cambrian to Ordovician (570 to 438 Ma)

ЄOg — Granitoid intrusions

ЄOm — Mafic intrusions, including granitoid rocks, gabbro and diabase

ЄOu — Ultramafic rocks of ophiolite complexes

Cambrian (570 to 505 Ma)

Єc — Marine shelf limestone, dolostone and shale

Neoproterozoic to Cambrian (900 to 505 Ma)

P_3Єrc — Rift and transitional shelf facies: non-marine conglomerate, sandstone and mafic volcanic rocks; marine shale, sandstone limestone and dolostone (includes part of Reluctant Head Formation, Old Man's Pond group)

P_3Єrs — Rift and transitional continental slope facies: sandstone, shale, conglomerate and mafic volcanic rocks (includes part of Curling Group)

Neoproterozoic to Middle Ordovician (900 to 460 Ma)

P_3ЄOu — Undivided siliciclastic and carbonate rocks (includes parts of Port au Port and St. George groups)

Neoproterozoic (900 to 570 Ma)

P_3g — Granitoid intrusions

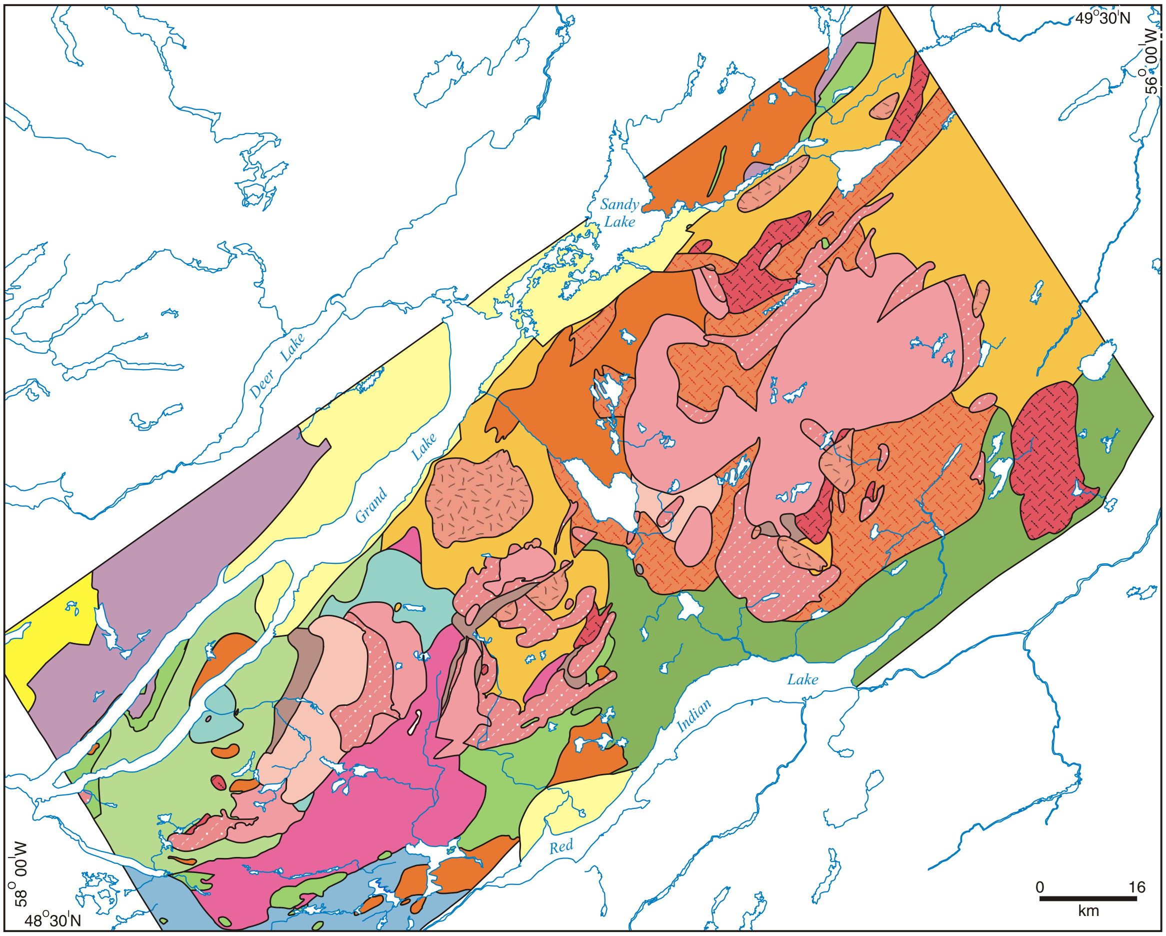
Mesoproterozoic (>1600 to 900 Ma)

P_2gg — Granitoid intrusions

P_2m — Mafic and anorthositic intrusions

$\mathrm{P}_1\mathrm{P}_2$gn — Granitic gneiss, mafic gneiss and paragneiss

Figure 3. *Bedrock geology of the Humber River basin and surrounding areas (*after *Colman-Sadd* et al., *1990).*



LEGEND

CARBONIFEROUS

Clastic sedimentary rocks, including sandstone, siltstone, mudstone, conglomerate and arkose. Forms parts of the Anguille Group, Deer Lake Group, and Howley Formation

SILURIAN

TOPSAILS INTRUSIVE SUITE

Quartz-feldspar porphyry (mainly peralkaline) rhyolite, porphyritic rhyolite, and minor basalt

White to red, fine- to medium-grained equigranular, one- and two-feldspar granite; minor quartz-feldspar porphyritic granite and aplite

White to red, medium- to coarse-grained equigranular amphibole sodic pyroxene one-feldspar granite

White to pink, medium- to coarse-grained biotite amphibole two feldspar granite

Red, medium-grained porphyritic two feldspar, quartz syenite to granite; grey to orange, medium to coarse grained gabbro to quartz syenite

Agmatite of mafic to ultramafic blocks in granitic to granodioritic matrix

SPRINGDALE GROUP

Red, flow banded rhyolite, rhyolite breccia; amygdaloidal and massive subaerial basalt

SILURIAN OR ORDOVICIAN

RAINY LAKE COMPLEX

Fine- to coarse-grained amphibole-clinopyroxene gabbro, diorite and quartz diorite and amphibole biotite granodiorite

ORDOVICIAN

White to red, massive to slightly foliated granite to granodiorite

BUCHANS GROUP

Basalt, rhyolite, tuff, diabase and gabbro

GLOVER GROUP

Pillow basalt, massive basalt, agglomerate, diabase, tuff and conglomerate

Platformal sedimentary rocks; includes Table Head Formation and St. George Group

Massive to moderately foliated granodiorite, biotite muscovite and biotite amphibole granite, and minor tonalite, containing many small mafic to ultramafic fragments

White to grey, medium- to coarse-grained hornblende biotite tonalite to diorite, moderately to strongly foliate

HUNGRY MOUNTAIN COMPLEX

Moderately to strongly foliated hornblende gabbro to granite containing many small to large mafic to ultramafic inclusions

Basalt, gabbro, hornblende, pyroxenite, ophiolitic rocks

NEOPROTEROZOIC TO EARLY PALEOZOIC

FLEUR DE LYS SUBGROUP

Semipelitic schist and psammitic gneiss

Figure 4. *Bedrock geology of The Topsails* (after *Whalen and Currie, 1988).*

Table 1. Rock types used in clast dispersal studies and their physical characteristics. Bold nomenclature in brackets corresponds to that used on Figure 4

Rock Unit	Location	Texture	Colour	Rock Type	Main Minerals	Distinctive Features
Hungry Mountain Complex (**Ohm**)	Around Hinds Lake, extending northeast to Sheffield Lake	Fine to coarse	–	Gabbro	Hornblende	Moderately to strongly foliated. One of few gabbro outcrops in area
Springdale Group (**Ssf**)	Northward on The Topsails, from Lewaseechjeech Brook.	Fine	Red	Rhyolite	–	Flow banded. One of few rhyolite areas
Rainy Lake Complex (**SOrl**)	Around Rainy Lake	Fine to coarse	–	Gabbro	Amphibole, clinopyroxene	Mildly saussuritized
Topsails Intrusive Suite (**Sqa**)	West of Hinds Lake, west of Sheffield Lake	–	Orange to green	Porphyry	Quartz, feldspar, peralkaline minerals (e.g. aegerine, arfvedsonite, aenigmatite)	Colour, porphyritic, peralkaline
Hinds Brook granite (**Oib**)	North of Hinds Lake toward Sandy Lake	Medium to coarse	White to pink	Granite	Biotite, amphibole, K-feldspar	K-feldspar porphyritic, contains 2 feldspars
Mount Musgrave Group (**Hf**)	West side of Humber basin	Fine to medium	Dark grey	Schist / psammite	Mica, quartz	Only micaceous schist in area
Deer Lake Group	Deer Lake basin	Fine to coarse	Red	Sandstone	–	Only source of red sandstone in area
Deer Lake Group	Deer Lake basin	Fine	Red	Siltstone	–	Only source of red siltstone in area
Devils Room granite	West of Sops Arm	Coarse	Pink	Granite	Quartz, plagioclase, biotite	Euhedral K-feldspar megacrysts
Gull Lake Intrusive suite	West of White Bay	Fine	–	Gabbro	Amphibole, plagioclase	–
Gull Lake intrusive suite (Moose Lake granite)	West of White Bay	Coarse	Pink to red	Granite	K-feldspar, plagioclase, quartz	Massive, contains microcline megacrysts
Grenville basement	Long Range Mountains	Fine to medium	Pink to grey	Gneiss, granite gneiss	Quartz, feldspar	Foliated rock, generally confined to north and south margins of area
Carbonate	West part of area	Fine	Various	Limestone, dolomite	Carbonates	Confined to west part of area

cias; and flow-banded rhyolite dykes (Whalen and Currie, 1988). Table 1 provides a summary of distinctive rock types and their characteristics.

In addition to the physical differences between individual clasts of gabbro, granite or rhyolite, the character of dispersal trains derived by a single glacial movement from sources in either the Long Range Mountains or The Topsails would be characterized by distinctive assemblages of clasts. Dispersal from The Topsails should be characterized by a wide range of clast types, reflective of the area's diverse bedrock geology. The Long Range Mountains assemblages should be more homogenous, unless ice flow was southward across the volcanic and igneous rocks west of White Bay. Clasts of Carboniferous sediments found outside the Deer Lake basin would reflect dispersal from that source area.

PHYSIOGRAPHY

Newfoundland has endured repeated cycles of denudation and subsequent uplift (for the last 245 Ma), most recently in association with the Quaternary glaciations. Thus, the physiography has erosional surfaces modified by more recent glacial and postglacial processes.

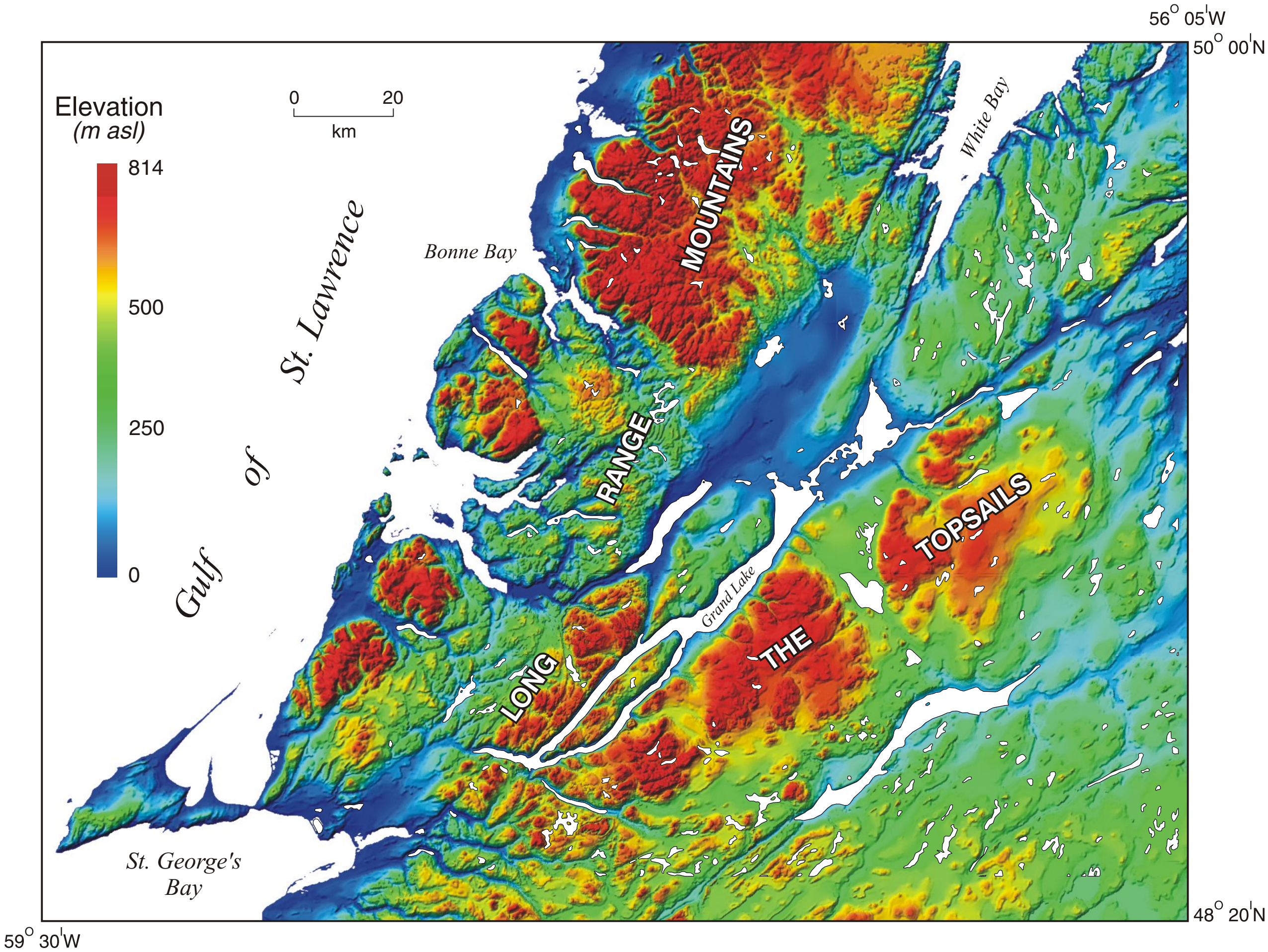


Figure 5. *Shaded relief map of the Humber River basin and surrounding areas.*

The western part of the study area is dominated by the Long Range Mountains (Figure 5), part of the Atlantic Uplands (Goldthwait, 1924 and Bostock 1970) that extends from the southwest tip of Newfoundland along the Great Northern Peninsula. The mountains are characterized by flat-topped peaks, having a maximum elevation of 814 m asl in the Lewis Hills; elevations decrease to the north and south. The upland plateau surfaces are commonly flat, featureless, deeply weathered surfaces having wide valleys and gentle slopes. These upper surfaces have been interpreted to represent uplifted erosional surfaces or peneplains (Twenhofel and MacClintock, 1940; Brookes, 1974; Rogerson, 1981; Grant, 1987). Erosional surfaces were identified at three elevations on the west coast, all of which tilt up toward the northwest (Twenhofel and MacClintock, 1940).

The upper surface is the Long Range peneplain that encompasses surfaces at between 600 and 650 m asl in the Long Range Mountains, dipping eastward to The Topsails, east of Grand Lake, where it is represented by four erosional remnants. These are the Gaff Topsail (573 m asl), Main Topsail (555 m asl), Mizzen Topsail (537 m asl) and Fore Topsail (491 m asl). Higher peaks in the Long Range Mountains, such as in the Lewis Hills (814 m asl), Gros Morne (806 m asl), and Round Hill in the Blow Me Down Mountains (762 m asl) may also be erosional remnants (Rogerson, 1981). The intermediate surface is the High Valley peneplain. It is represented by broad upland valleys at 520 m asl in the Long Range Mountains, dipping down to 400 m asl over much of the plateau surface of The Topsails. The lower surface, the Lawrence peneplain, is represented by flat-topped spurs in the Long Range Mountains at about 300 m asl, such as Table Mountain near Stephenville and the broad lowland on the southern part of The Topsails north of Red Indian Lake.

The age of these erosional surfaces is problematic. Brookes (1974) suggested they are pre-Quaternary, and probably formed during the Mesozoic or early Cenozoic. The surfaces truncate all rock types, and are overlain only by Quaternary sediment. Kerr (1994), in mapping the granites in The Topsails, noted that the commonly observed yellow and yellow-green granites are surface-alteration products of a primary green granite. The weathering is the result of alteration of potassium-feldspar crystals and extends to a depth of up to 10 m. Granite blocks quarried for construc-

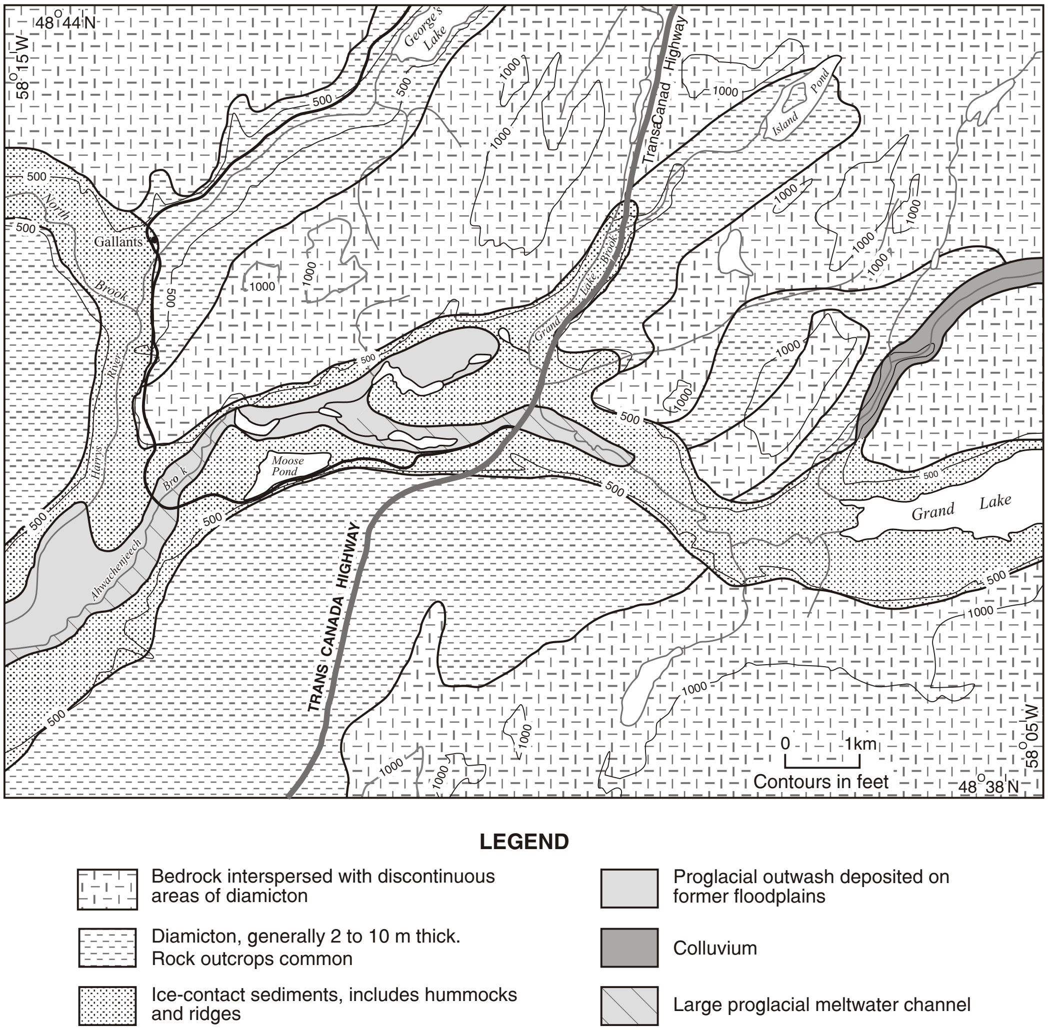


Figure 6. *Geography of the western end of Grand Lake.*

tion about 100 years ago have a less than 1-mm-thick weathering rind, and many granite surfaces are striated.

The Long Range Mountains are only breached in two places along their length. At the southwestern end of Grand Lake, a broad valley (up to 1500 m wide) extends to Harrys Brook (Figure 6). The valley is locally flat-bottomed and contains numerous exposures of sand and gravel. Modern drainage shows a poorly defined watershed at about 122 m asl in the vicinity of Gallants Junction on the TCH. To the east, Grand Lake Brook flows south through a narrow valley east of George's Lake, to the area of the watershed where it turns east and flows through a 5000-m-long channel into Grand Lake. The lower reach of this channel is incised through Quaternary sediment, and has a gradient of 1:95, compared to a 1:206 gradient upstream. To the west of the watershed, Ahwachenjeech Brook flows into Harrys River through a sinuous channel. Paleo-drainage was from Grand Lake southwestward into Harrys Brook, as indicated by the well-defined sinuous, flat-bottomed channel, 170 to 400 m wide, extending 11.5 km from the incised lower reaches of modern Grand Lake Brook to Harrys River valley (Figure 6). Drainage into Harrys River may also have been through Moose Pond, which has a paleo-channel to the southwest end.

The second breach of the Long Range Mountains drains most of the study area via the Humber River east of Corner Brook. The main channel is 125 km long, and drains through a broad, marshy lowland north of Deer Lake, from its headwaters in the southern part of the Long Range Mountains and flows into the Humber Arm at Corner Brook through the narrow Humber River gorge that is 4 km long. The gorge does not have the cubic parabolic shape typical of glaciated valleys (Sugden and John, 1976; Drewry, 1986); (Figure 7), suggesting it was probably produced by fluvial action. In contrast, the Wild Cove valley to the north has a cross-sectional profile closer to a parabola, and was likely glacially carved.

The modern drainage basin has an area of about 4400 km^2, and includes northeast–southwest-oriented lowlands

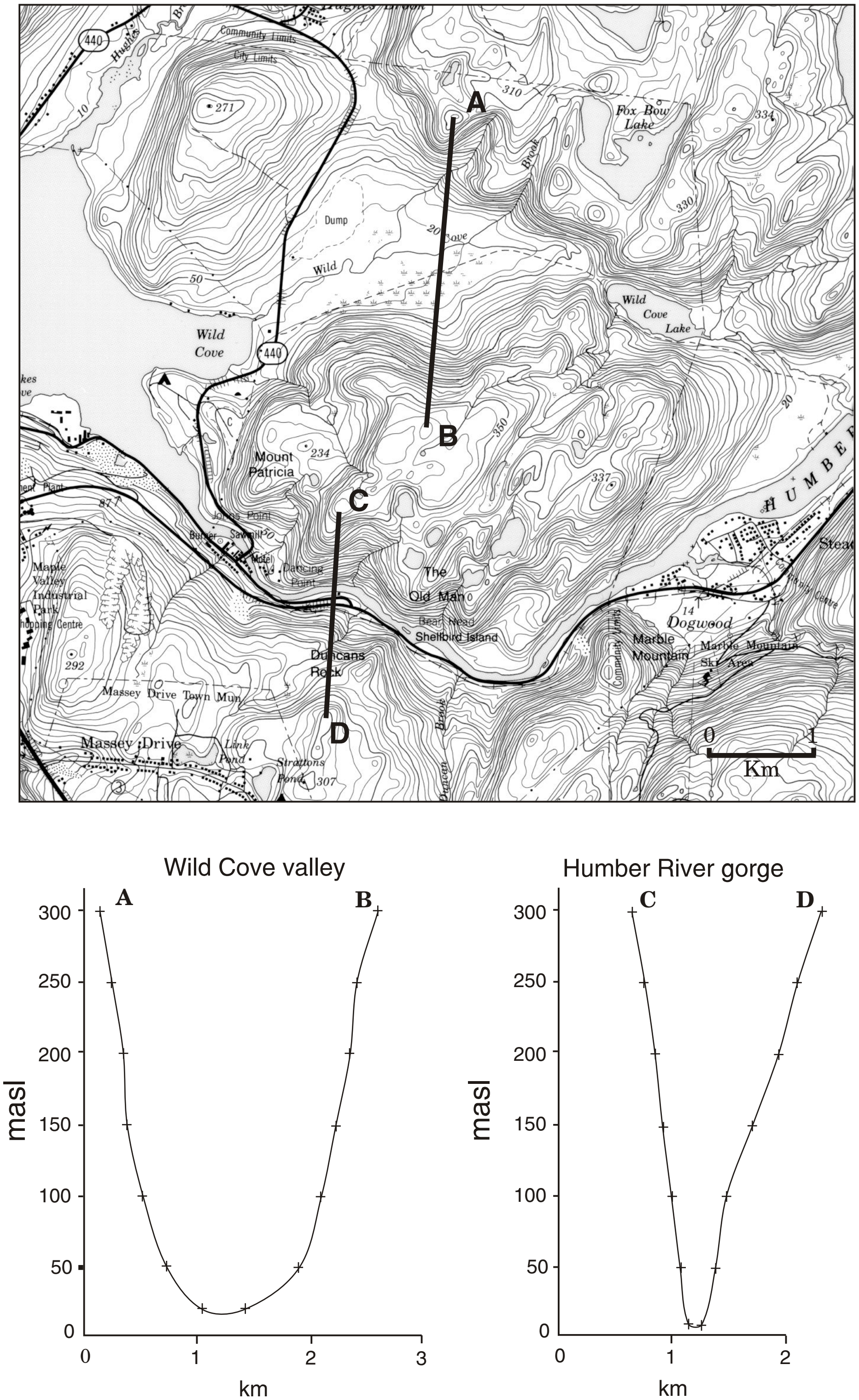


Figure 7. *Cross sectional profiles of the Wild Cove valley and Humber River showing differences in morphology.*

occupied by Deer Lake (elevation 5 m asl), and Grand Lake–Sandy Lake–Birchy Lake (elevation 82 m asl). Brookes (1970a) suggests that the Humber River gorge was cut since deglaciation and that the preglacial Humber River basin drained toward White Bay, citing differential isostatic rebound across the basin as evidence. It is unclear how the preglacial Humber River could have reached White Bay, but it must presumably have breached Birchy Ridge, however, there is no field evidence to support this suggestion.

The Humber River gorge appears to be structurally controlled. The lower part of Humber River valley is fault guided (Cawood and Van Gool, 1992), and consists of Paleoproterozoic to Early Cambrian psammite overlain by Early Cambrian carbonate rocks (Williams and Cawood, 1989). Subsurface water movement may initially have occurred along the contact, shown by the presence of cave structures and calcite veining indicative of open flow (I. Knight, Department of Mines and Energy, personal communication, 1996). Timing of the formation of the gorge is speculative, but there is no direct evidence that it was cut postglacially.

The Humber River basin contains several large lakes. Grand Lake and Deer Lake are two of the largest water bodies in Newfoundland having surface areas of 66 800 and 5698 ha, respectively; both trend northeast–southwest and are structurally controlled. Grand Lake contains a large island, occupying the southern half of the lake. Glover Island is 39 km long and 6.5 km wide, dominated by sheer cliffs up to 460 m high on all shorelines, except for the northeast shore. The origin of Glover Island is unclear, although Brookes (1970a) speculates that the island represents the interfluve between northeast-flowing stream channels that were overdeepened by glaciers.

The depth and bathymetry of both Grand Lake and Deer Lake are unknown. Jukes (1842) reported depths "greater than 3 fishing lines tied together" or greater than 100 fathoms (183 m) at the south end of Glover Island on Grand Lake. Murray (1882) reported two soundings on Grand Lake, one near Old Harry Mountain at 145 fathoms (265 m) and one south of Glover Island at greater than 184 fathoms (337 m). Soundings from Deer Lake indicate it has a maximum depth of 95 m (Seabrook, 1962). Other large lakes in the Humber River basin are Sheffield Lake, Adies Pond and Old Mans Pond (Figure 1). Each of these has a surface area of greater than 1000 ha.

PALEOBOTANY

Floral recolonization following deglaciation in southwest Newfoundland was described using the pollen record from Southwest Brook Lake (48°28'N, 57°59'W; 145 m asl) (Anderson and Lewis, 1992; Anderson and Macpherson, 1994) which records a basal radiocarbon date of 11 500 BP. Initial colonization was by a shrub-dominated tundra assemblage of willow (*Salix*), birch (*Betula*), juniper (*Juniperus*) and heather (Ericades), plus herbs (e.g., sage (*Artemesia*) and sedges (Cyperaceae)). Climatic amelioration continued until ~ 11 000 BP, at which time there was a return to cooler conditions that lasted until ~ 10 000 BP. At Southwest Brook Lake, this event is indicated by an increase in mineral lake sediment where the organic content drops as low as 1 percent (Anderson and Macpherson, 1994). The pollen record shows a rapid decline in shrub pollen during this cool phase in preference to a herb pollen assemblage, dominated by Cyperaceae, mountain sorrel (*Oxyria digyna*), and Artemesia. Anderson and Macpherson (1994) relate this period of climatic cooling to the Younger Dryas (Broecker *et al.*, 1988; Wright, 1989; Dansgaard *et al.*, 1993; Taylor *et al.*, 1993; Peteet, 1995).

Morphological and paleoecological evidence of Younger Dryas cooling is found in Newfoundland. Morphological evidence includes the moraines at Ten Mile Lake dated at 11 000 BP (Grant, 1969a), and fossil ice-wedge casts (Brookes, 1971; Eyles, 1977; Liverman *et al.*, 2000). Paleoecological evidence includes pollen data from numerous sites in northern and western Newfoundland (e.g., Macpherson and Anderson, 1985; Anderson and Lewis, 1992; Anderson and Macpherson, 1994), and diatom evidence from eastern Newfoundland (Wolfe and Butler, 1994).

Other periods of postglacial climatic cooling are recorded in the Maritime provinces, both pre- and post-Younger Dryas (Rawlence, 1988, 1992; Anderson and Lewis, 1992; Levesque *et al.*, 1993; Wilson *et al.*, 1993). In Newfoundland, evidence of the Killarney Oscillation between 11 200 and 10 900 BP (Levesque *et al.*, 1993) is suggested by a reduction in sediment organic content at Southwest Brook Lake (Anderson and Macpherson, 1994). A period of intense cooling at ~9600 BP is interpreted from an abrupt decline in the spruce population on the west coast (Anderson and Macpherson, 1994). This cooling phase may be associated with the release of meltwater into the Gulf of St. Lawrence from the drainage of glacial Lake Agassiz (Anderson and Lewis, 1992; Teller and Kehew, 1994; Teller, 1995). At Southwest Brook Lake, the end of the postglacial cooling episode at ~8500 BP is marked by an increase in spruce, tree birch and fir pollen, and the development of a boreal forest assemblage that includes the major components of the modern vegetation.

The time following 8000 BP is characterized by increased summer warmth and longer growing seasons. This is indicated by a gradual expansion of tree birch, and the arrival of black ash (*Fraxinus nigra*) recorded in cores from southwest and interior Newfoundland (Macpherson, 1995). Increased charcoal concentrations from forest fires indicate increased summer warmth. The temperature continued to warm until the Hypsithermal at about 6000 BP. This period is marked by expansion of red and white pine (*Pinus resinosa* and *Pinus strobus*) beyond modern limits, e.g., at Leading Tickles on the north coast of Newfoundland (Macpherson, 1995), and by a decrease in shrub birch, increased balsam fir, and increased sphagnum, indicating higher moisture levels. Mean summer temperatures were likely up to 1.5°C warmer than present during the Hypsithermal (Macpherson, 1995).

After about 4000 BP there was a slight decrease in mean annual maximum temperatures, indicated by the migration from the coast of red pine. Similarly, the length of the growing season was shortened, as shown by the expansion of spruce at the expense of birch, and relative moisture increased, as indicated by increased fir. Increasing moisture and decreasing temperatures also resulted in accelerated paludification, indicated by the cluster of basal bog dates following 4000 BP (Davis, 1984, 1993).

Most of the study area is presently within the Boreal Forest vegetation zone (Rowe, 1972), except for the summits of the Long Range Mountains where elevation and exposure produce a Tundra Forest vegetation zone.

Soil types are mostly podzols (humo-ferric and ferro-humic). Small areas of brunisols and gleysols are found throughout the area, as well as organic soils developed on wetlands. Button (1983), Kirby (1988) and Kirby *et al.* (1992) provide details on the distribution and characteristics of soils found within the area.

QUATERNARY REVIEW

There are several reviews of the Quaternary geology of Newfoundland, the most recent and thorough of which is by Grant (1989a). (*See* Prest (1970), Tucker (1976), Rogerson (1981), Dyke and Prest (1987) and Brookes (1982) for earlier reviews). The following discussion focuses on general issues concerning the areas surrounding the basin, including the west, north and southwest coasts, and the Red Indian Lake basin. A review of these areas is critical to discussions of Quaternary events within the Humber River basin through their effects on ice thickness and extent, ice-flow directions, and time of deglaciation. Finally, work specifically completed in the Humber River basin is considered.

Labrador Ice

The proximity of the Island of Newfoundland to Labrador has prompted considerable debate over the possible incursion of the Island by the Laurentide Ice Sheet during the Wisconsinan. Murray (1882) assumed that the Island was covered by ice moving down the Gulf of St. Lawrence, and crossing the Island in an east to northeast direction. He also recognized the development of local ice masses during deglaciation, as did Kerr (1870). Fairchild (1918), based on limited field observation including those recorded by Tyrell (*in* Fairchild, 1918, pp. 227-228), and relying heavily on data collected prior to, and subsequently published by Daly (1921), concluded that there was no physical evidence (striations) to suggest coverage by Labrador ice. Instead, crustal warping patterns on the west coast implied only the incursion of Labrador ice. Raised marine features increase in elevation northwestward along the Great Northern Peninsula, leading Daly (1902, 1921) and Flint (1940) to conclude that ice thickened toward Labrador, and that, at least, the west coast of the Island had been covered by ice from Labrador. MacClintock and Twenhofel (1940) and Tanner (1940) shared a similar view to Flint (*op. cit.*), although only southward-directed striations on the Port au Port Peninsula were cited as direct evidence (MacClintock and Twenhofel, 1940). These authors supposed that radial-flowing ice from a Newfoundland ice centre during deglaciation removed all evidence of the Labrador phase elsewhere. It was argued that Labrador ice covered the whole of the Island during the Wisconsinan maximum, possibly extending out as far as the Grand Banks (MacClintock and Twenhofel, 1940). During waning stages, radial flowing glaciers developed on highland centres, obliterating all evidence of earlier phases of flow. Betz (1939) observed striations oriented southeastward in the Canada Bay area, and Cooper (1937) recorded striations showing southward ice-flow at Bluff Head near Port au Port. Both these observations were used as evidence that Labrador ice covered at least part of the Island.

Direct evidence for the incursion onto the Island was found at the tip of the Great Northern Peninsula, where Cooper (1937) first identified erratics from Labrador and associated southwest striations. Grant (1969b, 1972, 1977a, 1987, 1992) provided detailed evidence that the Laurentide Ice Sheet covered the tip of the Great Northern Peninsula, north of a line between Margaret Bay and Canada Bay, up to an elevation of about 300 m asl, above which local island-based glaciation was dominant (Figure 8). There have been no erratics from Labrador found within the Humber River basin.

Nunataks

Following the general acceptance of the glacial theory to explain surface features on the Island, replacing the previously held diluvial view (e.g., Milne, 1874, 1876), the early workers on the west coast (e.g., Murray, 1882; Bell, 1884) were concerned with identifying evidence for glacier coverage. Bell (1884) considered that "...the glaciation appears to have been from the centre towards the sea on all sides" (p. 37). Murray (1882) found limited evidence for glacial coverage on summits of the Long Range Mountains and coastal highlands such as the Anguille Mountains, Blow Me Down Mountain and Lewis Hills. Twenhofel (1912) proposed that the Long Range Mountains were completely covered by glaciers, based on the distribution of striations on upland peaks. Coleman (1926) challenged this view, instead suggesting that many of the high plateaux were unglaciated during the late Wisconsinan, and possibly had never been glaciated. Evidence included the lack of glacial features such as erratics and striations, and weathered surfaces. Coleman (1926) argued that The Topsails, and the high peaks of the Long Range Mountains north from Port aux Basques, showed evidence of glaciations of pre-Wisconsinan (Kansan or Jerseyan) age. The tops of the southern Long Range Mountains and Blow Me Down Mountain in the Bay of Islands showed no evidence of glaciation, and Coleman (1926) concluded that these areas were never glaciated. Previously, Antevs (1922), in his compilation of the extent of Pleistocene glaciations, designated a large part of the Long Range Mountains on the Great Northern Peninsula as ice-free during Pleistocene glaciations, although suggesting that the southern part of the Long Range Mountains were ice-covered.

There are two contradictory views of glacial coverage for the west coast of Newfoundland. The ‘minimum' argu-

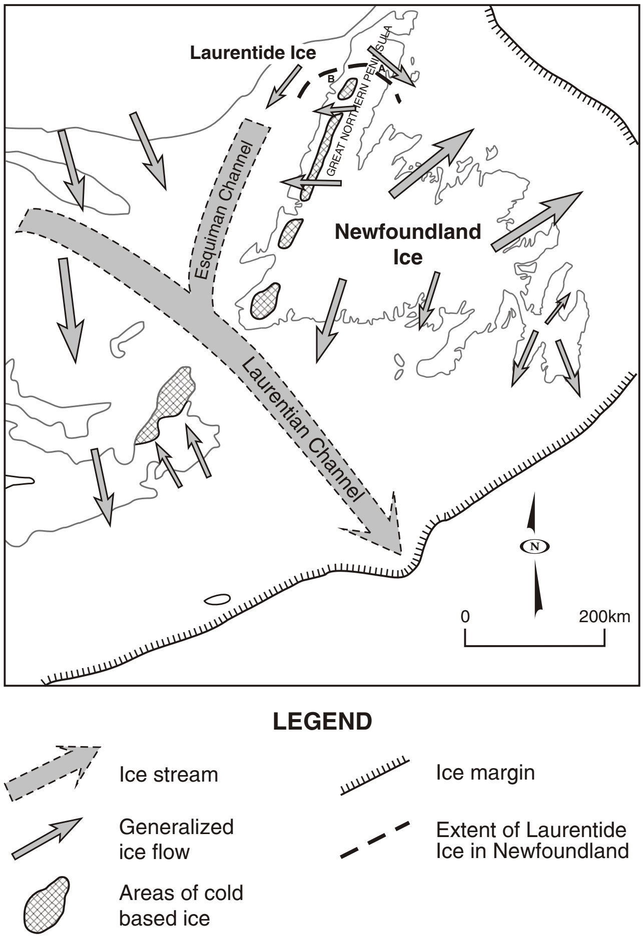


Figure 8. *Extent of Laurentide ice in western Newfoundland (*after *Grant, 1989b).*

ment contends that hilltops remained ice free during periods of glacial activity at lower levels, suggesting a restricted extent of ice. The ‘maximum' view argues that the hilltops were covered by glaciers, and that ice extent was well beyond the modern coast. Landforms such as tors and felsenmeer found on coastal highlands may have survived coverage by late Wisconsinan glaciers. Important contributions come from Baffin Island (Sugden and Watts, 1977) and the Torngat Mountains (Gangloff, 1983) where it was demonstrated that tors and felsenmeer survived the Laurentide Ice Sheet, and where incipient tafoni have developed during the Holocene. Ives (1978) provided a detailed history of the ‘minimum' versus ‘maximum' debate.

Apart from geological evidence, Coleman (1926) cited corroborative biological data from Fernald (1911, 1925) to support the minimum view. Based on the distribution of vascular plant species in upland areas, Fernald (1911, 1925, 1930) concluded that these areas must have remained ice free and acted as refugia, at least, during the Wisconsinan. Some of these plant species are now only found outside Newfoundland in the Rocky Mountains or in northeast Asia. The concept of refugia was supported by Lindroth (1963), who noted the preferred distribution of flightless beetles on the west coast (e.g., *Agonum bicolor, Carabus taedatus*, and *Bembidon morulum*). Other biologic evidence for refugia have been presented by Belland (distribution of mosses; 1981, 1987), Roberts (rare plants on serpentinized rocks; 1993), Liverman and Batterson (1985) and Bell *et al*. (distribution of *Cepaea* sp.; 1997).

The concept of west coast nunataks was disputed by Wynne-Edwards (1937, 1939) who argued the glacial climate was too harsh for plant survival, and rare plant distribution is better correlated with soil conditions related to underlying bedrock rather than Wisconsinan nunataks. Instead, Wynne-Edwards (1939) suggested that plants existed in areas marginal to the ice sheets, perhaps on the Labrador seaboard, or in areas to the south and west of Newfoundland.

Following 1960, the minimum viewpoint became favoured again, based initially on work in Baffin Island (e.g., Boyer and Pheasant, 1974; Miller and Dyke, 1974) and northern Labrador (e.g., Løken, 1962; Ives 1960, 1978). These arguments were extended to the Island by Grant (1969a, 1976, 1977a,b, 1989a) and Brookes (1970a, 1977a). This ‘minimum' viewpoint of late Wisconsinan ice is mainly dependent on the recognition of distinct weathering zones in the Long Range Mountains, Torngat Mountains and on Baffin Island, and equating them with periods on decreasing glacial extent.

Weathering zones in Newfoundland were first described by Coleman (1926). Grant (1977a) identified three weathering zones in the area of Gros Morne National Park. The highest zone (Weathering Zone 3) is characterized by an intensely altered surface, where weathering has removed most signs of glacial activity. Later, Grant (1989a) estimated this surface to represent oxygen isotope stage 12 (about 430 000 BP) on geomorphic grounds, and its potential correlation with the offshore record (cf. Alam and Piper, 1977). Below Zone 3 is an area of modified surfaces (Weathering Zone 2), where the morphology is clearly related to glacial activity. Grant (1989a) estimated the age of this surface to represent oxygen isotope stage 6 (about 140 000 BP), based on correlation with glacial deposition in the Gulf of St. Lawrence (Alam *et al.*, 1983). Weathering Zone 1 has the freshest evidence of glacial activity, and is correlated to the late Wisconsinan. Brookes (1977b) noted a similar arrangement of weathering zones in the Anguille and southern Long Range Mountains, and observed that several of the summits of the southern Long Range Mountains showed no evidence of glaciation, and hence were probably unglaciated. Coleman (1926) made similar observations in the Blow Me Down Mountains in the Bay of Islands.

Late Wisconsinan nunataks in Gros Morne National Park, such as Gros Morne Mountain and Big Level, were recently re-examined by cosmogenic radionuclide analysis of the ^{10}Be and ^{26}Al contents of quartz from veins and pegmatite dykes (Gosse and Grant, 1993; Gosse *et al.*, 1993).

Data indicate coverage by Wisconsinan ice, probably within the last 50 000 years. This work suggests a cold-based ice cap, with complete plateau ice cover, and suggests that the intensity of the features in the weathering zones does not necessarily correspond to the elapsed time since they were last glaciated. This hypothesis further suggests that cold-based ice overlies areas where these weathering features persisted – the supposition being that basal freeze-on would have protected (not destroyed) the surface-weathering features. Cold-based ice has been proposed as an explanation for the preservation of blockfields and patterned ground (e.g., Falconer, 1966; Gellatly *et al.*, 1988) and of pre-late Wisconsinan landscapes (e.g., England, 1987; Dyke, 1993; McCarroll and Nesje, 1993; Kleman, 1992, 1994) beneath the late Wisconsinan or recent ice masses.

The debate between complete plateau coverage and late Wisconsinan nunataks remains unresolved. Cosmogenic radionuclide analysis data shows that the sampled areas were ice covered during the late Wisconsinan. However, it does not necessarily imply that other coastal highlands, such as the southern Long Range Mountains and Anguille Mountains, were also ice covered during the last glacial. The elevation, and remoteness of these areas from the ice-dispersal centres affecting Gros Morne require a separate examination of each region.

North Coast and Exploits River Valley

A watershed at 104 m asl separates Birchy Lake from Indian River that flows into Notre Dame Bay (Figure 1), and a watershed at 145 m asl separates the Humber River basin from drainage into White Bay. The northeastern margin of the Humber River basin includes the area between White Bay and Notre Dame Bay, defined by the Baie Verte Peninsula. Grant (1977a, 1989a) suggested the northern parts of the peninsula were unglaciated, although Macpherson and Anderson (1985) noted that radiocarbon dates from the area suggest deglaciation at about 13 500 BP.

Detailed striation mapping by St. Croix and Taylor (1991) reconstructed the glacial-flow patterns, which show that at the late Wisconsinan maximum, the ice flowed northeast from an ice centre on The Topsails, and was deflected eastward by ice from the Long Range Mountains that occupied Notre Dame Bay. During deglaciation, the Baie Verte Peninsula hosted remnant ice centres from which ice flowed radially outward (Grant, 1974), including a probable southward flow toward the Indian Brook valley (Liverman, 1992).

Radiocarbon dating on marine molluscs from muds underlying ice-contact deltas along the coast provide minimum dates for deglaciation. Tucker (1974a) dated a 75 m delta at Springdale at 12 000 BP. Scott *et al.* (1991) demonstrated earlier deglaciation based on a radiocarbon date of 12 500 BP from marine shells located 10 km inland of the Springdale delta.

The late Wisconsinan maximum northeastward ice flow that affected the north coast originated from a source in The Topsails, identified by detailed striation and clast provenance studies (Vanderveer and Sparkes, 1982; Sparkes, 1985, 1987; Klassen, 1994; Klassen and Murton, 1996). This flow affected much of the Red Indian Lake–Exploits River valley area. The Topsails were also the source of southward-flowing ice that crossed Red Indian Lake during both the early and late Wisconsinan (Sparkes, 1985), separated by a glacial lake phase that occupied the Red Indian Lake basin up to 59 m above the present level (Vanderveer and Sparkes, 1982; Mihychuk, 1985). Individual ice-flow events were correlated with tills exposed in mine workings near Buchans (Vanderveer and Sparkes, 1982).

Southwest Coast

Southwest Newfoundland was deglaciated first. Radiocarbon dates from marine shells, above glacial diamictons, near St. George's Bay found at Robinsons (13 500 ± 210 BP, GSC-1200), Rope Cove (13 700 ± 340 BP, GSC-2942) and Abrahams Cove (13 700 ± 230 BP, GSC-1074) support early deglaciation (Brookes, 1974; Anderson and Macpherson, 1994).

The Humber River basin was connected to southwestern Newfoundland through paleo-drainage of Grand Lake into the Harrys River valley. Grant (1991) mapped a proglacial channel extending southwest from Grand Lake entering St. George's Bay near Stephenville Crossing. Brookes (1974) speculated that ice retreated up the Harrys River valley, and into the Grand Lake basin, presumably forming the channel through proglacial meltwater activity.

Southwestern Newfoundland is one of the few areas of extensive coastal exposures of Quaternary sediment that have a complex stratigraphy. MacClintock and Twenhofel (1940) described a sequence of sediments, interpreted as showing a lower regional till (St. George's River Drift), overlain by glacioisostatic marine onlap sediments (St. George's Bay Delta), and a local re-advance till (Robinsons Head Drift). Based on the distribution of striations, they suggested that ice flow was coastward from the interior during the St. George's River Drift stage, and found little evidence of Labrador ice, although they did not discount its influence.

Brookes (1969, 1970a,b, 1974, 1977a,b) accepted this tripartite stratigraphy and provided further details on the areal extent, character, and relation of the stratigraphy to postglacial relative sea levels. The St. George's River Drift overlies bedrock, and outcrops around St. George's Bay. It is a compact greyish pink to grey till, and is commonly overlain by delta bottomsets and foresets of the Bay St. George delta that was deposited up to about 43 m asl; the delta formed at about 13 500 BP (Brookes, 1987; Grant, 1989a). An exception occurs from Stephenville to west of Romaines Brook, where marine overlap was delayed until sea level regressed to 30 (±) m elevation. This delay was considered to be the result of a reactivation of the ice front, termed the Robinsons Head readvance. Brookes (*op. cit.*) mapped this re-advance as extending from near Romaines Brook in the north to near Highlands in the south, reaching the coast as several lobes. The Robinsons Head readvance was dated at about 12 600 BP, based on a single date from marine shells found within a sand bed interstratified in kame gravels at Kippens (Brookes, 1977a). Ice subsequently retreated up the Harrys River valley into the Grand Lake basin, its course marked by subaerial glaciofluvial sediments and eskers (Brookes, 1970a). Deglaciation of this part of Newfoundland may have occurred sometime after 12 600 BP.

The Humber River Valley

The first documented descriptions of the Humber River valley are those of Jukes (1842), reporting on expeditions he made through Newfoundland in 1839. Although he was influenced by the diluvial theory, Jukes (1842) observed ‘that fragments of rock, frequently of great size have been removed from their original position in all directions for a few miles' (p. 337). His descriptions of the Humber River valley include terrain as far north as the rapids upstream of the confluence with Junction Brook. Murray (1882) supposed that ice invaded the Island during the last glacial, and generally moved northeastward through the Humber Arm into Deer Lake, and from St. George's Bay through Grand Lake. Here, the ice masses merged to eventually flow out into Notre Dame Bay; the orientations of Deer Lake and Grand Lake were cited as evidence.

Coleman (1926) noted a ‘blue bowlder clay' in the Humbermouth to Curling area, close to modern sea level, deposited by ice moving seaward from the Humber Valley. A fossiliferous sediment at 43 m asl between two till units at Curling was inferred by Coleman (1926) to represent an interglacial deposit. Striations oriented east–west along the Humber Arm were consistent with westward flow, but northwest–southeast-directed striations at higher elevations were attributed to either a northwestward flow unconfined by the valley below, or southeastward-moving ice flow from a Labrador-centred ice mass.

MacClintock and Twenhofel (1940) noted the large accumulations of sediment at the head of Grand Lake (observed earlier by Murray, 1866), and suggested they were moraines formed during a stillstand or readvance, termed the Kittys Brook moraine stage. They also noted the large delta at the mouth of the Humber River, at about 46 to 48 m asl, and suggested that these and all other marine deposits in the area, including the ‘interglacial' deposit identified by Coleman (1926), were of late glacial age.

Lundqvist (1965) examined the eastern part of the study area and concluded that The Topsails were a dispersal centre. Subsequently, the area was overridden by ice flow from the main Wisconsinan centre in the Long Range Mountains. Rhythmically bedded sediments were observed in the eastern part of the canal between Birchy Lake and Indian Brook. These sediments were interpreted as ice-marginal glaciolacustrine deposits.

Three interpretations of ice-flow history for the Upper Humber River basin have been put forward by Rogerson (1979), Vanderveer (1981) and Batterson and Taylor (1990). Rogerson (1979) commented directly on ice-flow directions in the Upper Humber River basin (Figure 9a) and produced a, "speculative ice-flow map" for the Humber River basin based on till fabrics and regional observations. It shows a regional ice flow northward through the Upper Humber River valley out into White Bay, followed by topographically controlled flow off Birchy Ridge and the Long Range Mountains into the basin.

Vanderveer (1981) used striations, till fabrics, and topographic evidence to present a glacial chronology consisting of three separate ice movements (Figure 9b), each correlated with a distinctive till unit. The first event (pre-late Wisconsinan), indicated by a red, clay-rich till at the base of the Quaternary stratigraphy, originated from a centre to the northeast, overtopping parts of the Long Range Mountains and Birchy Ridge. The second event was an eastward to southeastward ice flow that covered the northeast part of the basin, and deposited a pinkish grey, sandy till over the red clay till. These two tills are, in places, separated by an interglacial or interstadial sand and gravel unit (Vanderveer and Sparkes, 1982). During the late Wisconsinan, the third ice advance was in a south to southwestward direction, down the Upper Humber River valley. Vanderveer and Sparkes (1982) suggested this flow was responsible for creating the major landforms in the area, including a series of drumlinoid ridges, recessional moraines and meltwater channels, as well as depositing a locally extensive immature and poorly comminuted till of local provenance.

Batterson and Taylor (1990) suggested that two separate ice-flow events affected the area (Figure 9c). The first was a regional, west to northwestward flow from a centre in The Topsails, interpreted from striations and clast provenance data. The second was a local, southwestward flow confined to the Upper Humber River valley, and recognised on the basis of striations and surface landforms. Further, they suggested that a large proglacial lake may have occupied the Grand Lake–Deer Lake–Sandy Lake–Birchy Lake basins during deglaciation.

Previously collected data on the Quaternary geology of the Humber River basin is fragmentary, and conclusions commonly contradictory. A clear ice-flow history and stratigraphic framework has not been established. In some areas, such as Birchy Ridge, researchers had recognized mutually contradictory ice-flow patterns. Although ice dispersal centres were identified in the Long Range Mountains and The Topsails, the number, timing, and extent of advances from these sources was poorly documented. Despite the volume of study, a detailed systematic regional analysis of the Quaternary history of the Humber River basin was lacking.

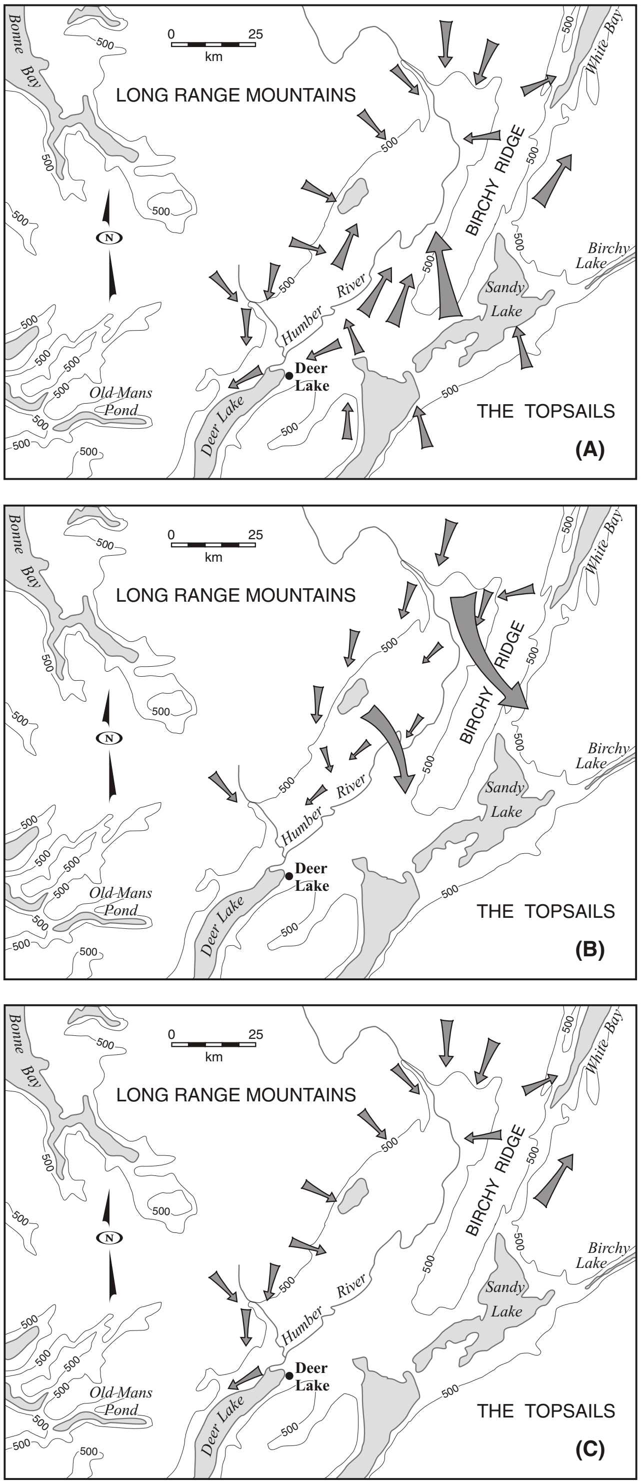


Figure 9. *Previously published maps of the ice-flow history of the Humber River basin.*

SEA-LEVEL HISTORY

The study of Quaternary sea-level history in Newfoundland is divided into three separate periods. The first was the identification and description of raised marine features prior to the 1900s; the observations were local, with little or no attempt to integrate data or define regional patterns. The second period included more systematic observations and integration of data; this prompted regional analysis and the production of isobase maps. The third, and current period from the 1970s onward developed, in conjunction with increasing knowledge of the rheology of the earth's crust, in response to loading by ice sheets. In particular, the concept that the relative sea-level fluctuations of areas marginal to waning ice sheets, are dominated by the effects of forebulge migration, produced an increasing discussion of local, as opposed to regional, relative sea-level history.

Raised sea levels were first noted by Jukes (1842) who, in describing clays found in the Exploits River valley, suggested that it was.... ‘highly probable that the country once stood at a lower level; that the arm of the sea formerly extended much farther up....' (p. 339). Jukes (*op. cit.*) also reported marine shell fragments about 10 m asl at Ship Cove, St. George's Bay, although he was unsure whether they were emplaced there by higher sea levels or by birds. Murray (1882) expanded the observations of raised marine features to the Baie Verte Peninsula, where he found shells near the Terra Nova mine site and at Tilt Cove, and in the Bonne Bay–Port au Port area. He went further and stated that Newfoundland was at one time depressed by 150 m, and that this had been caused by crustal deformation induced by mainland ice. This would have drowned the area between St. George's Bay and Hall's Bay through the Grand Lake–Indian Brook area, leaving the Great Northern Peninsula as an island.

DeGeer (1892) developed the first isobase map of Newfoundland (Figure 10a) which showed the zero isobase traversing the south and west coasts of the Island, with a distinct bulge over the island, suggesting an influence on crustal depression from local Newfoundland ice. Fairchild (1918) showed a much more prominent dome over Newfoundland, with the zero isobase farther south of the Island (Figure 10b), and maximum uplift of over 180 m. Based on observations from the coast, Daly (1921) concluded that the zero isobase crossed the west coast in the vicinity of Robinsons Head and extended out through the centre of Bonavista Bay (Figure 10c). Daly (1934) showed parallel isobases crossing the Island, suggesting that there were no local ice cap influence on isobase patterns (Figure 10d). Flint (1940) and Farrand and Gajda (1962) shows that there is no deformation in the shape of isobases crossing the Island, and used this as evidence to suggest that the Island was invaded by ice from Labrador, and that ice caps on Newfoundland had little impact on crustal deformation. However, Flint (*op. cit.*) had earlier identified a discontinuous wave-cut bench on the west coast, rising northward from sea level near Stephenville to over 75 m asl at St. Anthony; this was termed the Bay of Islands surface. He suggested that this surface resulted from a stillstand or re-advance of Labrador ice, although Grant (1994a) speculated that it may relate to the Ten-Mile Lake re-advance phase from piedmont glaciers

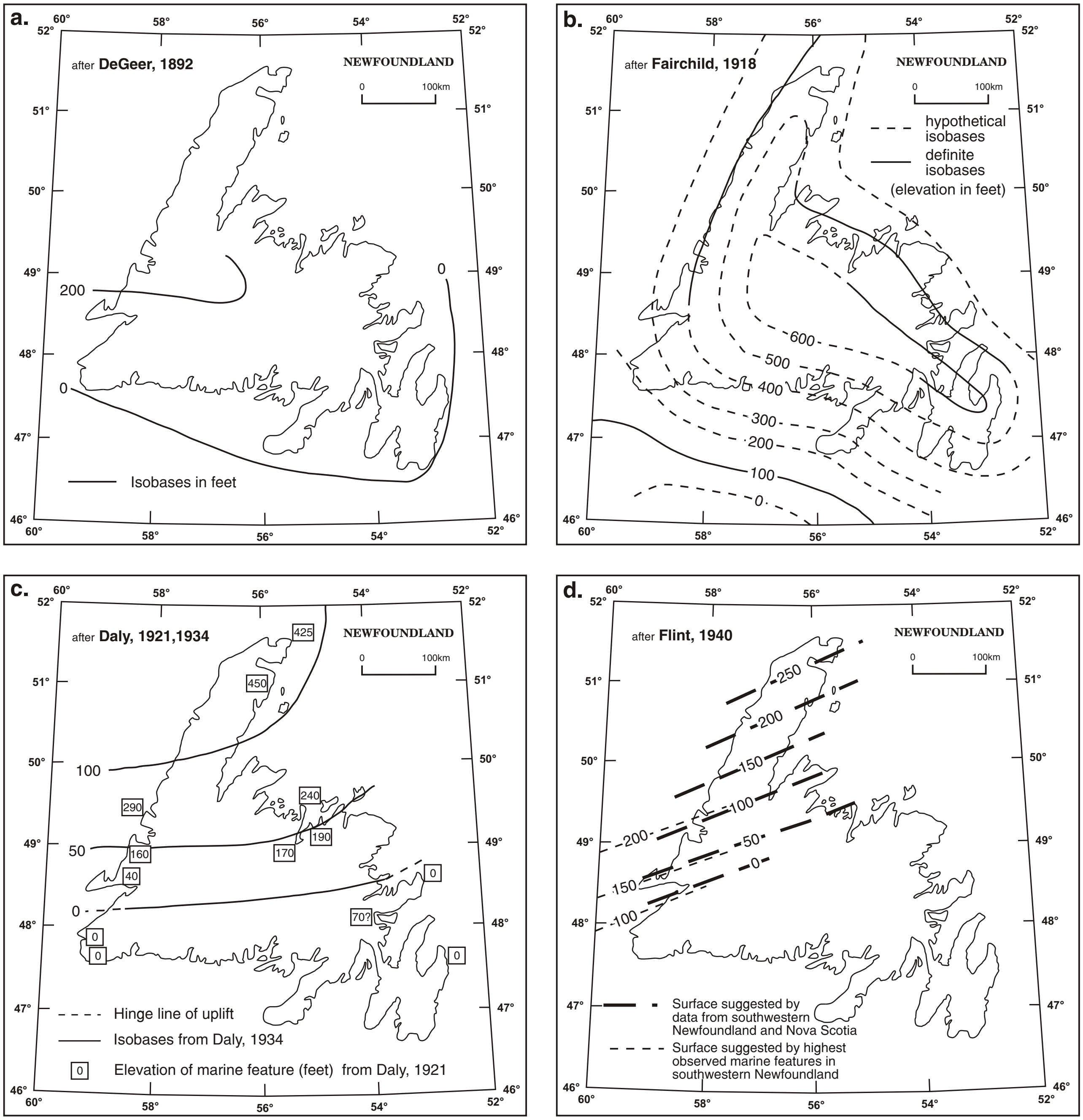


Figure 10. *Previously published isobase maps of Newfoundland.*

in the Long Range Mountains. Earlier, Grant (1980) showed isobases curving toward the interior of the Island in response to local ice centres, a pattern extended by Rogerson (1982) to show a central dome over the Red Indian Lake area. In scant recognition of this central dome, Grant (1987) shows interior curving isobases.

Apart from Flint (1940), there has been recognition of the affect of interior Newfoundland ice centres on isobase patterns, combined with the major influence of the Laurentide Ice Sheet on regional trends. The lack of raised marine features in the interior, and debate on late Wisconsinan ice extent, and by implication ice thickness, makes construction of isobases away from coastal areas speculative.

Although Daly (1921) speculated on the rheology of the earth's crust, and discussed the concept of a migrating forebulge, the lack of temporal control and data points meant that relative sea level (RSL) curves could not be determined. Walcott (1970, 1972) demonstrated that the effects of ice sheet loading on the earth's crust affected areas well removed from those areas underlain by ice.

The major affect on RSL, throughout the Holocene, is due to forebulge migration as a result of glacioisostasy, and relatively little influence from eustatic changes (Liverman, 1994). Sea level rose rapidly at the end of the last glacial, as water held in ice sheets was released to the oceans. It is uncertain whether this release was smooth and monotonic as suggested by Ruddiman (1987), or a step-wise release, punctuated by intervals of rapid rise as described by Fairbanks (1989, from Barbados coral reef records), or consisted of periods of rapid rises of about 0.2 m as a result of catastrophic outburst floods from the waning Laurentide Ice Sheet (Shaw, 1989; Blanchon and Shaw, 1993) separated by periods of little change. Regardless, sea level has continued to rise in some places and fall in others having a range of over 100 m (Lambeck, 1990). The larger changes are in areas underlain by, or proximal to, late Pleistocene ice sheets, and result from continued isostatic adjustment of the crust as a result of ice-sheet loading or non-uniform meltwater distribution. Large parts of the continental shelf off the west (Scotian Shelf), and south and east (Grand Banks) coasts of Newfoundland were exposed during the Pleistocene as a result of sea-level lowering of between 90 and 120 m (Piper *et al.*, 1990).

Newfoundland was peripheral to the Laurentide Ice Sheet and was influenced by the forebulge (Walcott, 1972; Peltier, 1974, 1976) produced by crustal loading to the north. Results of this loading (Figure 11) characterizes areas beneath the Laurentide Ice Sheet, and corresponds to Zone 1 (glaciated zone) of Clark *et al.* (1978) and Clark (1980), where isostatic adjustment has overwhelmed eustatic response. These curves are consistent in form with those reported from Labrador (e.g., Clark and Fitzhugh, 1992). At point B (Figure 11), the passage of the forebulge is marked by falling sea levels as the forebulge approaches, followed by rising sea levels as it passes. Point C shows an initial period of falling RSL followed by a rise to the present, and point D experiences a steady rise in RSL. Both points C and D should have no geomorphic features showing higher sea level above present. Within the areas defined by the Type-A curves, the record of falling RSL can be found both onshore (deltas, marine shells) or offshore (deep-water to shallow-water fossil assemblages from cores). Evidence of previous RSL changes in Types-B and -C areas can be found by freshwater to marine transitions in terrestrial settings (e.g., salt marshes) or in nearshore basins having a well-defined sill. No terrestrial evidence for Type-D areas should be found in the onshore or inter-tidal record.

Quinlan and Beaumont (1981) used these curves to develop models of RSL history having different ice-mass configurations, using an earth model similar to that of Peltier and Andrews (1976); two separate ice models were used. A ‘maximum' model has ice extending out to the modern coast around most of the Island, and is similar to that used by Peltier and Andrews (*op. cit.*). A ‘minimum' model, derived from Grant (1977a), has thin ice only reaching the inner coast. Both models showed increasing ice thickness toward the north-northeast reflecting the increasing affect of Laurentide ice. Each of these models allowed subdivision of the Island into distinct zones defined by the RSL curves. It was assumed that deglaciation was from an isostatically balanced earth at 18 000 BP, and ice and water loading was averaged over 1 by 1° grids. The models were then compared to known RSL curves, with the conclusion that the RSL history was best explained by a model falling between the minimum and maximum ice models. In a subsequent paper, Quinlan and Beaumont (1982) used the RSL record to reconstruct past ice configurations for Atlantic Canada. Their conclusions supported the presence of an ice dome on the north shore of the Gulf of St. Lawrence, suggested by a free-air gravity anomaly over the area, and limited ice in the Gulf of St. Lawrence. Despite some minor modifications (e.g., Scott *et al.*, 1987) the models proposed by Quinlan and Beaumont (1981, 1982) for Atlantic Canada have remained unchallenged.

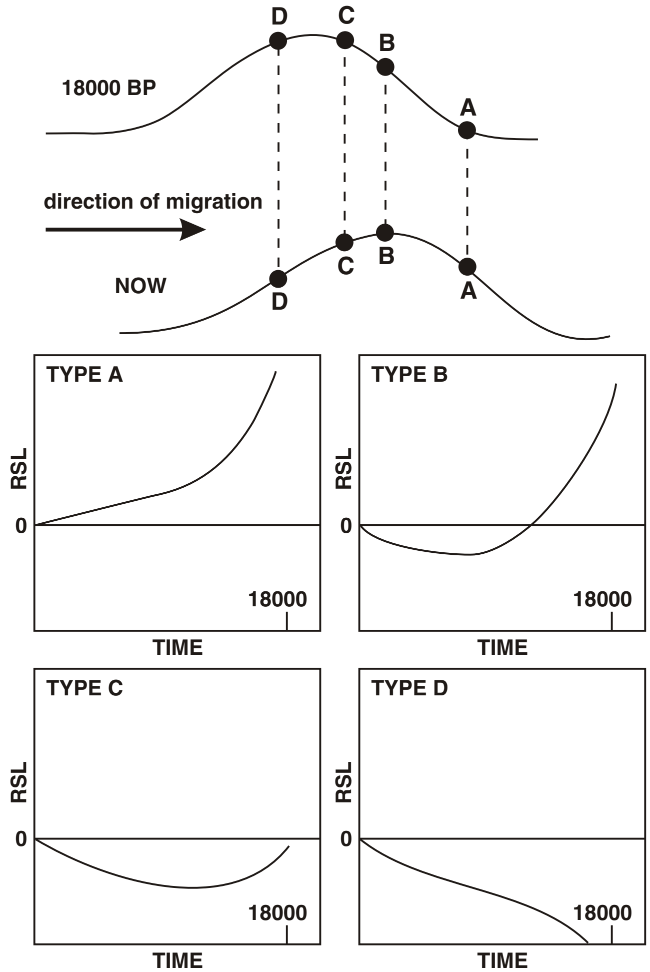


Figure 11. *Forebulge migration and resultant relative sea-level curves (*after *Quinlan and Beaumont, 1981).*

In western Newfoundland, RSL curves are only available from the Northern Peninsula (Grant, 1977a, 1992, 1994a, b); and St. George's Bay–Port au Port area (Brookes, 1977a; Brookes *et al.*, 1985; Forbes *et al.*, 1993). The Great

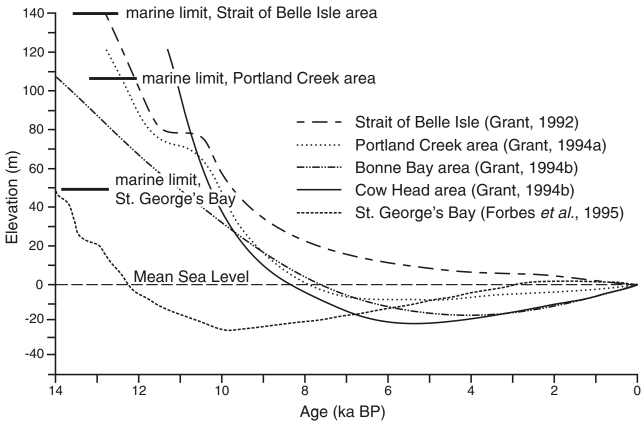


Figure 12. *Previously published relative sea-level curves for western Newfoundland.*

Northern Peninsula curves (Figure 12) from sites north of St. Barbe, such as L'anse aux Meadows are generally Type-A, and show continual emergence of the coast from marine limit. However, Grant (1992, 1994a) speculates that brief periods of submergence may have occurred in the last 3000 years, and Liverman (1994) suggests the tip of the Great Northern Peninsula may show a modified Type-B curve, marking the transition between coastlines showing Types-A and -B curves. Curves derived from south of St. Barbe, such as at Port au Choix, are Type-B, although the amount of emergence from sea level is minimal. Grant (1987, 1989a) showed northward-increasing marine limit from 75 m asl at Bonne Bay to 135 m at the tip of the Great Northern Peninsula. Marine limit in the Bay of Islands is shown at between 50 and 75 m asl, although Brookes (1974) had previously suggested a limit of less than 50 m asl.

The St. George's Bay (Type-B) curve shows emergence of the coast from a marine limit of about 45 m asl (Brookes, 1987; Grant, 1987), followed by a transition from emergence to submergence. The timing of this transition was initially placed at about 11 500 BP (Brookes, 1977a). Brookes *et al*. (1985) refined this curve based on radiocarbon dates (from marine shells, foraminifera and pollen), and showed the emergence–submergence transition occurred about 9500 ± 300 BP. A further iteration from Forbes *et al*. (1993), using additional radiocarbon dates and geomorphic data, demonstrated that the transition had occurred about 11 700 BP, with a 15 m lowstand at about 10 000 BP and RSL returning to near present at about 5500 BP (Figure 11). Shaw and Forbes (1995) subsequently redefined the postglacial sea-level lowstand at 25 to 30 m for St. George's Bay, decreasing to sea level just north of Bonne Bay. Humber Arm had a lowstand at about 6 m below present mean sea level. The timing of the lowstand was relatively late (~6500 BP) at the northern limit, and early (~9500 BP) in the southwest.

Liverman (1994) used all available radiocarbon dates on marine fossils to redefine the RSL history of the Island. The presence of *in-situ* marine shells shows that the RSL was higher than present when the shells were deposited. Areas characterized by Type-A curves should therefore have a range of radiocarbon dates from deglaciation to present (assuming no selective sampling), whereas Type-B curve areas should record only dates within a certain range, excluding the time when submergence occurred. The results showed that most of the Island is characterized by a Type-B curve. The exceptions are the tip of the Great Northern Peninsula, which show a Type-A curve involving either continual emergence or a modified Type-B curve with an emergence–submergence–emergence oscillation, as was speculated by Grant (1994a,b) and Liverman (1994); and the easternmost Avalon Peninsula, which may show a single Type-C curve. Similarly, the nearshore parts of the Scotian Shelf and Grand Banks record Type-C curves, and the outer margins record Type-D curves (Piper *et al.*, 1990). Liverman (1994) also showed that there was an earlier and more rapid sea-level transgression than that predicted by Quinlan and Beaumont (1982).

Shaw *et al.* (1995) conducted an offshore survey in the Humber Arm and adjacent fjords. The survey included a collection of a sediment core, taken from a depth of 92 m midway across the Humber Arm between Giles Point and Meadows Point (48°58.9'N, 58°03.1'W). The core shows at least 102 cm reddish brown 'buttery clay', overlain by 7 cm muddy gravelly sand, and 50 cm silty clay, the upper 25 cm of which is bioturbated. The muddy gravelly sand contained fragments of a marine bivalve, with a corrected age of 5360 ± 60 BP (Beta 81980), and was tentatively interpreted as being related to the postglacial lowstand (Shaw *et al.*, 1995). Table 2 shows grain size and foraminifera data from this core. The data shows the reddish brown buttery clay (mean 10.2 to 10.3ø) has a consistent grain size in, at least, the bottom 60 cm of the core. Foraminifera content is low below 80 cm, containing less than 5 species. One sample was barren (100 cm depth). Foraminifera species include *Cassidulina reniforme* and *Elphidium excavatum*. Both have a wide range of temperature and salinity tolerances, and are commonly the first to appear following deglaciation (Vilks *et al.*, 1989; MacLean *et al.*, 1992; Scott *et al.*, 1984). Foraminifera species change up-core into those indicative of warmer and more saline water conditions (e.g., *Adercotryma glomerata* and *Islandiella helenae*).

RADIOCARBON DATES

The establishment of a glacial chronology for western Newfoundland is dependent on radiocarbon dates derived from marine or terrestrial organic material. The inherent uncertainties in ^{14}C activity (Lowe and Walker, 1984) mean that radiocarbon age calculation is reported as a bell-curve distribution, with the date representing the mean determination, and the error bar (±) representing one standard deviation about the mean. A second consideration is the effect of ^{14}C plateaux. These are periods of several hundred years duration of constant radiocarbon age, identified in lacustrine

Table 2. Grain-size characteristics, and number of foraminifera species found in a short core, from the Humber Arm (*see* Shaw *et al.*, 1995)

Depth(cm)	Gravel(%)	Sand(%)	Silt(%)	Clay(%)	Mean(ø)	S.D.(ø)	Species
2-4	0.04	8.58	49.98	42.4	7.7	2.5	20
44-46	0.51	15.34	43.78	40.37	7.3	3.0	13
50-54	11.87	32.24	30.42	25.47	4.8	4.4	20
62-64	0.0	3.61	27.92	68.48	9.3	2.6	9
98-100	0.0	0.54	14.09	85.36	10.2	2.1	3
118-120	0.0	0.99	12.54	86.47	10.3	2.0	4
138-140	0.0	1.07	14.69	84.23	10.2	2.0	3
153-155	0.0	0.62	13.99	85.39	10.3	2.1	3

(Ammann and Lotter, 1989; Lotter, 1991; Lotter *et al.*, 1992) and marine environments (Broecker *et al.*, 1988). They are likely the result of decreased levels of ^{14}C in the atmosphere, possibly correlated with meltwater events in the North Atlantic (Edwards *et al.*, 1993). Radiocarbon plateaux have been identified at 12 800 to 12 600 and 10 000 BP (Ammann and Lotter, *op. cit.*), and at 9500 BP (Becker *et al.*, 1991; Lotter, 1991), and prevent precise dating during the Older Dryas, Bølling, Younger Dryas and Preboreal biozones.

OBJECTIVES

The Humber River basin is an inland basin situated between two possible late Wisconsinan ice-dispersal centres, one on the Long Range Mountains and the other on The Topsails. The existing literature concerning the glacial history for the area is confusing and contradictory, where the individual influences of the two dispersal centres and other possible centres have been poorly defined.

The principal objective of this report is to develop a model that describes both the glacial and postglacial history, and the paleogeography of the landscape in the Humber River basin and surrounding areas. This objective has been accomplished by completing the following tasks:

i) mapping of the distribution of Quaternary sediments, landforms, and features,
ii) description of their geomorphic characteristics, sedimentological and other physical properties, and
iii) recognition and assessment of indicators of ice-flow direction.

The hypotheses developed here are tested against existing descriptions of the late Pleistocene history of the area and Holocene relative sea-level fluctuations.

METHODS

FIELD OBSERVATION

Field work was conducted between 1990 and 1993 and over 650 bedrock and Quaternary sediment exposures were examined. Bedrock exposures were examined for ice-flow indicators, mostly striations. Appendix 1 provides a complete listing of those sites at which ice-flow indicators were identified. Munsell colours presented in this report are moist colours taken from fresh exposures, unless otherwise stated. Fine-grain sediments were collected for microfaunal examination, and for geochemical analysis.

The surface elevation of features were determined using a topographic map or with an altimeter. Determinations from topographic maps with a 10-m-contour interval are considered accurate to ± 5 m, whereas those from altimeter measurements are accurate to ± 2 m. Altimeter measurements were related to mean sea level or lake levels, or to vertical control stations distributed through the field area (Geodetic Survey of Canada, 1978).

LABORATORY ANALYSES

Grain-size analyses were completed in the Geological Survey laboratory in St. John's. Matrix samples (finer than 2 mm/-1ø) were split to provide a 100 to 150 gram sample, and wet sieved through a 4ø (63 µm) sieve. The retained fraction was oven dried and sieved through a nest of 6 stainless steel sieves (-1ø to 4ø) following the standard procedures outlined by Bowles (1978). The finer than 4ø fraction was analyzed using a Coulter Counter Model TAII-L. In addition, standard statistical parameters were derived for each grain-size analysis, including graphic mean and inclusive graphic standard deviation (Folk and Ward, 1957). Data is provided in Appendix 1.

Three-dimensional *clast fabrics* (orientation and plunge) were measured on 25 elongate pebbles having a length:breath ratio of greater than 3:2. Fabric measurements were taken from a small area (<1 m^2), remote from observable contacts, and not adjacent to large boulders. Results were plotted on a stereogram and analyzed using the Stereo™ software package for the Apple Macintosh microcomputer (MacEachran, 1990). Randomness testing shows that for 25 pebble samples S_1 values of greater than 0.46 and S_3 values of less than 0.21 are significantly different from random distributions at the 95 percent confidence level (Anderson and Stephens, 1972; Woodcock and Naylor, 1983)

Boron sediment geochemistry, has been used as an indicator of paleo-environmental conditions during deposition of fine-grain sediments (Shimp *et al.*, 1969; Catto *et al.*, 1981; Mosser, 1983). Boron was analyzed by prompt-gamma neutron activation at the Chemex laboratory in Vancouver. Detection limit was 5 ppm, and standards and controls of known value were incorporated within the data set.

Microfaunal assessment of content of fine-grain sediments were achieved by examination of randomly selected sub-samples of sieved 2ø to 4ø fractions. Samples from Goose Arm, the Lower Humber River valley and the shores of Grand Lake were selected for examination.

Aerial photography was primarily 1:50 000-scale black and white photographs, although areas of thick drift cover were also examined from 1:12 500-scale colour photographs.

Surficial geology maps were initially digitized at 1:50 000 scale, combined with the CARIS GIS and output as colour maps. The enclosed map (Map 2001-42, Open File NFLD/2768) is derived from 1:50 000-scale surficial geology maps (Batterson, 1992, 1994a,b and c).

The *air-photo interpretation* of the surficial geology has been classified using a sediment-landform approach. The classification scheme is similar to that used by Fulton *et al.* (1975), Grant (1973) and Vanderveer (1975), and is suited to areas having physiographic and sediment diversity, plus Quaternary erosional and depositional landforms. The classification has a genetic category that describes the sediment type (e.g., glacial, fluvial, colluvial), and a morphology category that describes the surface expression (e.g., veneer, hummocky, fan). Most mapping units contain more than one genetic and/or morphological type. To accommodate this, units are subdivided by decreasing dominance. For example, Tv/Rc means that the area is dominantly a veneer of till, with a lesser area of bedrock concealed by a mat of vegetation and/or soil. Up to three genetic subdivisions can be shown, with a combination of slashes (/ or //) or hyphens (-) being used to indicate relative proportions. In the map associated with this report (Batterson, 2001) presented, polygons are coloured, based on the dominant sediment type, and thus several polygons are commonly combined and shown as a single colour. The map does not take into account the different texture, colour, or structure that may be found within sediments of the same genetic origin.

SURFICIAL GEOLOGY

MAPPING

Open File NFLD/2768 (Batterson, 2001) summarizes the surficial geology data derived from preliminary air-photo interpretations, followed by field work and subsequent reinterpretation of the study area. Previous mapping of all, or parts, of the Humber River basin (Brookes, 1970a; Grant, 1973, 1989b, 1991; Liverman *et al.*, 1991; Sparkes, 1985, 1987; Vanderveer, 1987) was reviewed following the completion of the air-photo interpretation and field work program. Although there are broad geomorphic similarities between previous mapping and the surficial interpretation presented here, there are also significant differences. For instance, areas described by Grant (1989b) along the eastern shore of Grand Lake as thick till containing moraines, are interpreted here, as glaciolacustrine fan deposits comprising mostly sand and gravel. The lowlands on the northwest shore of Deer Lake that have been mapped as thick till eroded by meltwater channels (*see* Grant, 1989b and Vanderveer, 1987) are here interpreted as fluvial sand and gravel. Similarly, areas mapped as till ridges in the Cormack area by Vanderveer (1987) are interpreted here as till eroded by meltwater channels. Improved access, and the increased exposure has allowed an improved interpretation.

QUATERNARY SEDIMENTOLOGY

Figure 13 is an isopach map for the Humber River basin that used 902 data points, generated from a variety of sources, including water-well data (67), drill-core data from mineral exploration (141), site assessments for buildings and infrastructure development (53), highway construction (9), and field observations by the author (632). A listing of data sources, apart from field sites, is recorded in Appendix 2.

Drilling, as part of commercial mineral exploration programs, was restricted to several small areas. Wigwam Brook in the Upper Humber River valley was the site of exploration for uranium in the late 1970s and early 1980s (Hyde, 1984). Over 80 drillholes were placed within a 10 km^2 area. Similarly, diamond drilling, in support of gold exploration in the Kettle Pond area at the southern end of

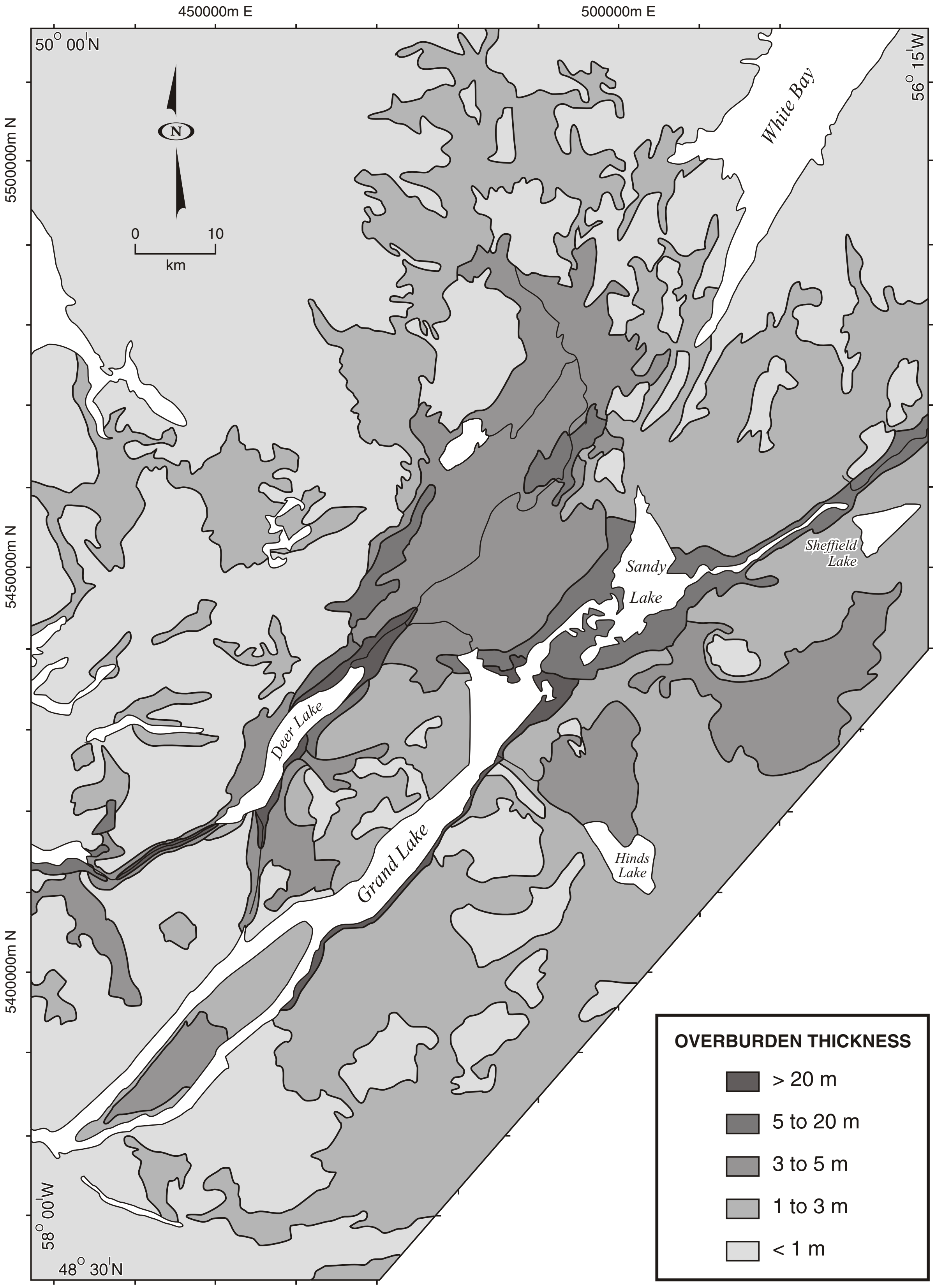


Figure 13. *Isopach map of the Humber River basin.*

Glover Island, provided data from 30 boreholes. Several drillholes were reported in the area north of Sandy Lake and along the east shore of Grand Lake. The latter was the site of the earliest drilling in the Humber River basin. Two drillholes were sunk in the Kelvin Brook area in 1879 in support of coal exploration (Murray and Howley, 1881). The precise location of these drillholes is uncertain because descriptions refer to distances upstream from the lake and the lake shoreline has been changed by the dam at Junction Brook. Further drilling was completed in the same area in 1893 (Murray and Howley, 1918).

Water-well data used in this study (Department of Environment, 1995) provide accurate location and overburden thickness data. Water wells are restricted to communities; therefore, the data are confined to the Humber River valley.

Field observations record exposure thicknesses. Bedrock is commonly not observed in test pits, and consequently most field observations must be considered as minimum depths of overburden. Areas of exposed bedrock were mapped from aerial photographs.

Much of the Humber River basin is characterized by thin overburden (< 5 m). The Long Range Mountains west of the Humber River valley and south of Corner Brook are bedrock dominated having a generally thin (< 1 to 2 m) and discontinuous overburden. Other large areas of bedrock exposure are Birchy Ridge, the western shore of Grand Lake, and parts of The Topsails south of Hinds Brook. Small areas of bedrock exposure are scattered through the basin, mostly restricted to highlands.

The major valleys contain areas of thick Quaternary sediment. The Humber River valley between Reidville and Corner Brook contains thick sedimentary successions. Thicknesses of more than 60 m of sand, silt and clay are reported from Deer Lake airport (Environment Canada, 1980), and in excess of 120 m at Steady Brook (Golder Associates, 1983). Sediment thicknesses greater than 20 m are found at Humber Village (30 m), Little Rapids (30 m), Pasadena (85 m), Pynn's Brook (62 m), St. Judes (76 m) and Reidville (25 m). The consistent sediment thickness found in these drillholes suggests a sediment-filled depression extending from at least as far north as Reidville to the modern coast. Offshore in the Humber Arm, Shaw *et al.* (1995) reported combined water depth and sediment thickness exceeding 115 m. There are no drill-core records from the Humber River gorge, but extrapolation of the data from Steady Brook to Humbermouth suggests the presence of a deep (up to 100 m?), and narrow (less than 200 m wide in places) channel. The extent of the depression north of Reidville is uncertain. Bedrock is exposed in the river bed at Harrimans Steady (about 20 m asl), 8 km upstream of Reidville, with an adjacent 15-m-high sand and gravel terrace. Airphoto mapping suggests sediment thickness is generally less than 10 m upstream of this point.

An area of relatively thick sediment is found in a 150 to 1200 m wide belt along the eastern shore of Grand Lake, extending south from Howley to Connors Brook, and possibly beyond (Figure 1). Lakeshore exposures are up to 30 m high, and Murray and Howley (1918) reported 35 m of overburden near Howley. Similarly, the north shore of Grand Lake also has sediment exposures in excess of 30 m. Sediment thickness below the floor of the modern lake is unknown. The depth of Grand Lake itself is unknown.

Other areas of relatively thick Quaternary sediment are found in the western foothills of Birchy Ridge, the Cormack area, the Old Mans Pond valley, Hinds Brook valley and Birchy Lake valley. The Birchy Ridge foothills have been subject to intensive uranium exploration. Drilling near Wigwam Brook commonly encountered more than 10 m of overburden (Appendix 2). Sediment thicknesses generally decrease both west and east of Wigwam Brook. The Cormack area, on the western edge of the Upper Humber River valley has water-well data showing sediment thickness of at least 10 m, and locally as much as 40 m. Bedrock exposures are more common westward toward the Long Range Mountains, although numerous rock ridges are also found in areas northeast of Cormack. The Old Mans Pond, Hinds Brook and Birchy Lake valley areas contain relatively thick (5 to 10 m) sediment, mainly confined to the valley sides. Bedrock exposures in these areas are rare, commonly exposed during road construction or in stream cuts.

Areas between bedrock-dominated highlands and sediment-filled river valleys have up to 5 m of sediment, an assessment based on the numerous bedrock outcrops exposed through these areas. These intermediate areas characterize 60 to 70 percent of the Humber River basin.

SURFICIAL UNITS

Bedrock (R or Rc)

Exposed bedrock and bedrock concealed by a thin mat of soil and/or vegetation together comprise about 34 percent of the Humber River basin. The largest areas are the Long Range Mountain uplands, west of the Humber River valley; the area west of the South Brook valley and parts of Birchy Ridge, Glide Lake highlands and The Topsails south of Hinds Brook. Areas underlain by Carboniferous strata contain rare bedrock exposures.

In some areas, bedrock structure and morphology control surface relief features (Figure 14). In the Carboniferous Deer Lake basin, Hyde (1982) recognized numerous north–south to northeast–southwest-trending anticlines and synclines within the Humber Falls and Rocky Brook formations in the Humber River valley, and in the Saltwater Cove Formation on Birchy Ridge. The largest of these is the Humber syncline, extending along the axis of the Humber River

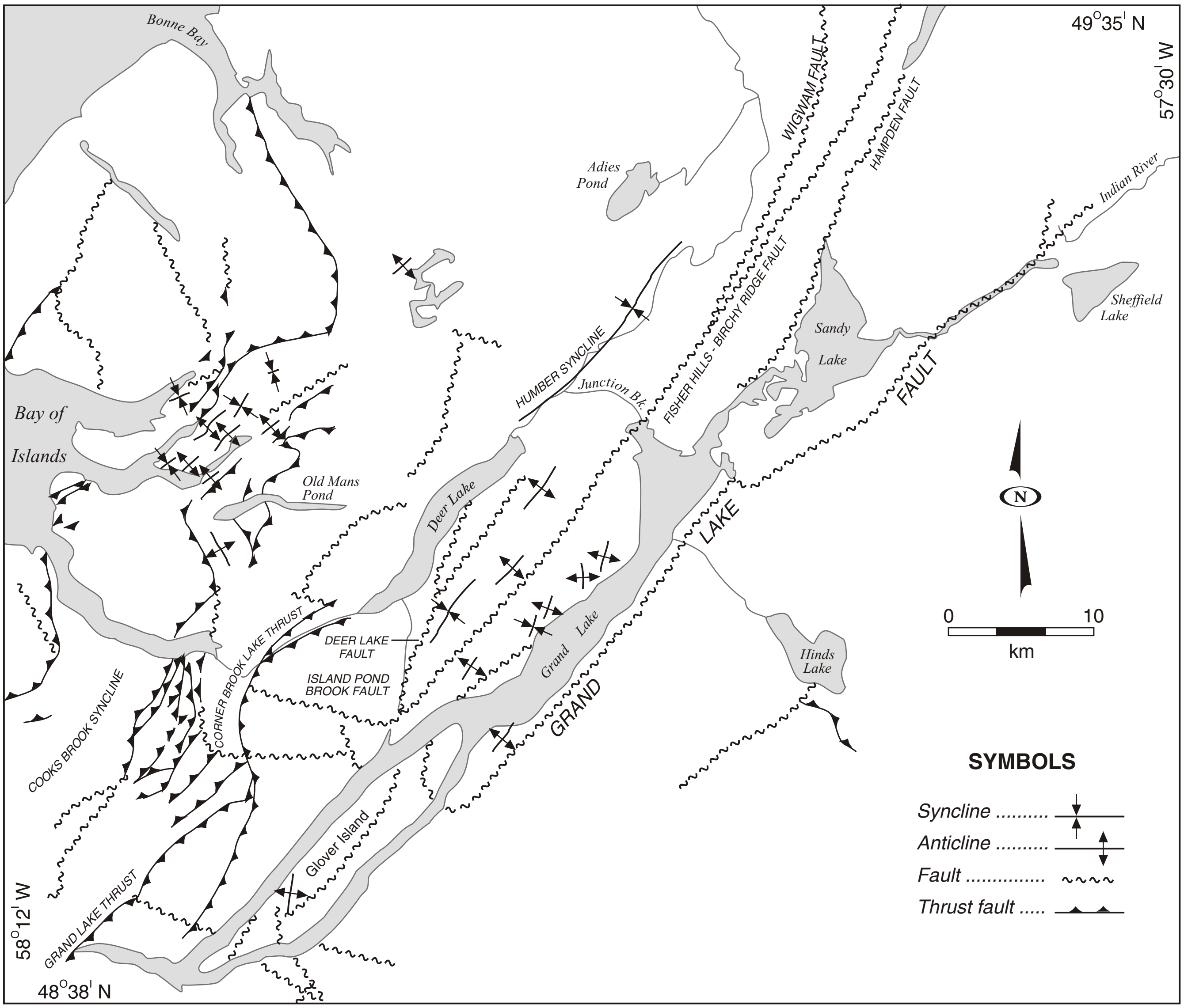


Figure 14. *Bedrock control on surface geology.*

valley from the head of Deer Lake to near Adies Pond. Elongate hills on the Glide Lake highlands are also coincident with the location of isoclinal folds (Hyde, 1982).

The orientation of bedrock ridges and coastal valleys, such as Goose Arm, Penguin Arm and North Arm follow fold or thrust axes within carbonate bedrock (Williams and Cawood, 1989; Knight, 1994). In particular, the Goose Arm, Sugar Loaves, and Raglan Head anticlines, the Middle Arm and Sugar Loaves synclines, and the Goose Arm, Reluctant Head and Alder Steady (Figure 14) thrusts are all aligned roughly northeast–southwest (Knight, 1994). Bedrock ridges in the carbonate rocks south of Corner Brook commonly define the trend of north–south-oriented folds (Knight, 1995). Large areas of The Topsails plateau are underlain by granitoid rocks and lack the fold structures characteristic of the sedimentary rocks to the west. Some valleys are fault guided, notably the Hinds Brook valley (Whalen and Currie, 1988). The prominent erosional remnants found on The Topsails – the Fore, Main, and Mizzen Topsail – are likely bedrock-controlled features. Kerr (1994) suggests that, although there is no lithological contrast with surrounding rocks, the lack of joints oriented perpendicular to glacial flow on the remnants may have impeded erosion.

Diamicton (Tv, Tr, Te, Th)

The Humber River basin contains large areas described as diamicton. In Newfoundland, the revisions of the number of late Wisconsinan glacial advances on the southern Avalon Peninsula are based on stratigraphic re-interpretations (Rogerson and Tucker, 1972; Eyles and Slatt, 1977). Similarly, in the southern St. George's Bay area, Liverman and Bell (1996) demonstrated that sediments previously interpreted as representing glacial advance and readvance, separated by a delta-building phase, could have been entirely deposited within an ice-contact glaciomarine environment.

Areal Distribution

Diamicton is the most common surficial unit across the Humber River basin, comprising 44 percent of the surficial geology (Batterson, 2002). It is found throughout the basin area, with the exception of the main valleys, such as Lower Humber River, Deer Lake, Upper Humber River (only adjacent to the main channel), Grand Lake, Grand Lake–Sandy Lake lowland, Hinds Brook, Goose Brook, Kitty's Brook and Birchy Lake. However, within each of these areas diamicton may be found underlying younger surficial sediments. The diamicton is generally thin (< 3 m). The thickest assemblages, exceeding 10 m, are present in the Upper Humber River valley in areas underlain by soft Carboniferous bedrock. Diamicton thicker than 5 m is also found in the Corner Brook area overlying shale bedrock, in the South Brook valley, and adjacent to Hinds Lake.

Physical Characteristics

Many of the characteristics of diamictons within the field area, especially colour, texture and clast provenance are controlled by the composition of the underlying bedrock.

Grain Size

Ricketts (1993) provided data on 38 diamicton samples from the aggregate-resources inventory (Table 3); these show the diamictons in the Humber River basin are generally coarse, having a matrix content between 25 to 65 percent of total grain size by weight.

Figure 15 shows the grain-size envelope derived from the grain-size analyses of 282 diamicton samples. These show a wide range of curves, reflecting the diverse bedrock geology of the area. The relative proportions of sand (1 to 4ø), silt (4 to 8ø) and clay (finer than 8ø) are plotted on a ternary diagram (Figure 16) which shows that most diamictons in the study area have a sand to silty sand matrix. The mean grain-size proportions are 66 percent sand, 28 percent silt and 6 percent clay. Table 4 provides a summary of the matrix statistics. Diamictons are very poorly to extremely poorly sorted and have a standard deviation (s.d.) ranging between 2.14ø and 5.65ø, with an average of 3.55ø. The low silt–clay component is reflected by the mean grain size that ranges from granule gravel (1.8ø) to medium silt (5.9ø), with an average in the medium sand (1.4ø) fraction. The mean grain-size value of 1.8ø is an artifact of the graphing process that interpolates data into the coarse fractions. Of the 282 diamicton exposures sampled, only 40 had a silt–clay content greater than 50 percent, with only 5 recording silt–clay fractions greater than 70 percent. Of these, three were found in the Humber River valley (Humber Gorge, Pynn's Brook, and Wigwam Brook), and two on the east side of Birchy Ridge. In each case, adjacent diamicton exposures recorded a considerably smaller silt–clay component.

Figure 17 shows the distribution of mean grain sizes overlain on a simplified bedrock geology map. Although the data is concentrated in the central part of the Humber River basin, some general trends are obvious. Those diamictons having a mean grain size finer than 3ø are found overlying

Table 3. Grain-size analysis of diamicton samples from the Humber River basin. Data from the Aggregate Resources Database (Ricketts, 1993)

NTS sheet	Sample	% Gravel (-1ø to >6ø)	% Sand (-1ø to 4ø)	% Silt–Clay (<4ø)
12H03	783464	53.1	37.8	9.1
12H03	783466	52	43.2	4.9
12H03	783463	67.3	28.4	4.3
12H03	784000	64.4	27.7	7.6
12H03	784002	44.5	42.1	13.3
12H03	784003	32.7	46.1	21.2
12H04	783977	40.8	47.7	11.5
12H04	783978	39	55	6
12H04	783979	42.9	47.2	9.9
12H05	770701	61.3	32.1	6.6
12H05	770702	58.4	37.8	3.9
12H05	770708	74	23.5	2.5
12H05	783882	54.3	34.1	11.6
12H05	783884	33.6	50.1	16.3
12H05	783886	45.3	35.1	19.6
12H05	813102	53.3	41.2	5.6
12H06	783936	41.4	42.9	15.7
12H06	783962	48.1	41.4	10.6
12H06	784011	51.9	43.3	4.8
12H06	784013	47.6	30.7	21.7
12H06	784014	78.6	19.9	1.4
12H06	803188	58.9	35.1	5.9
12H06	803343	66	33	1
12H07	782410	50.5	31.7	17.8
12H07	782413	59.5	26.1	14.3
12H07	782415	44.4	43.8	11.8
12H07	782417	62.7	29.1	8.1
12H07	782421	49.5	41.6	8.9
12H07	782423	64.7	32.2	3
12H07	782424	66.6	29.6	3.9
12H07	782436	65.9	22.1	12
12H07	782445	45.3	50.4	4.3
12H07	782451	52.7	39	8.3
12H07	782462	54.6	37.8	7.6
12H07	782464	45.9	43.3	10.8
12H07	782466	59.5	32	8.5
12H07	782467	50.8	42.2	7
12H07	782470	64.2	32	3.8
Mean		53.8	37.1	9.1

Table 4. Mean, range and standard deviation of diamicton matrix samples from the Humber River basin (n=282)

Sediment	Mean %	High %	Low %	S.D.
Sand	66.4	95.8	23	12.1
Silt	28	61.3	4.2	9.9
Clay	5.5	56.4	0	6.8

soft Carboniferous bedrock, or over the limestone terrain directly west of the Carboniferous basin. Some fine-grained diamictons are found overlying schist in the South Brook valley, although areas of schist bedrock both west and east of the valley are overlain by coarse-textured diamictons. Coarser grained diamictons (coarser than 2ø) are found along the shores of Grand Lake, on The Topsails, and overlying gneiss bedrock west of the Upper Humber River basin;

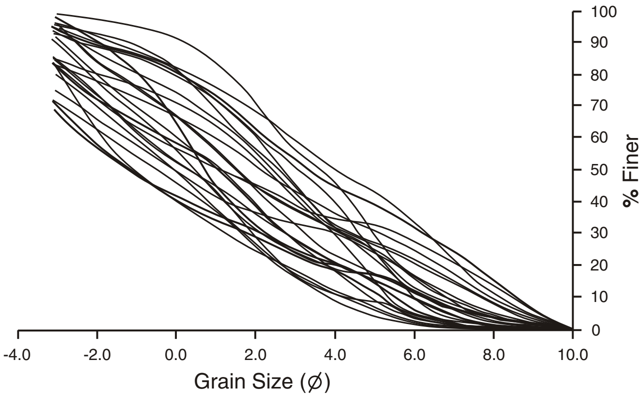


Figure 15. *Grain-size envelope for diamictons within the study area.*

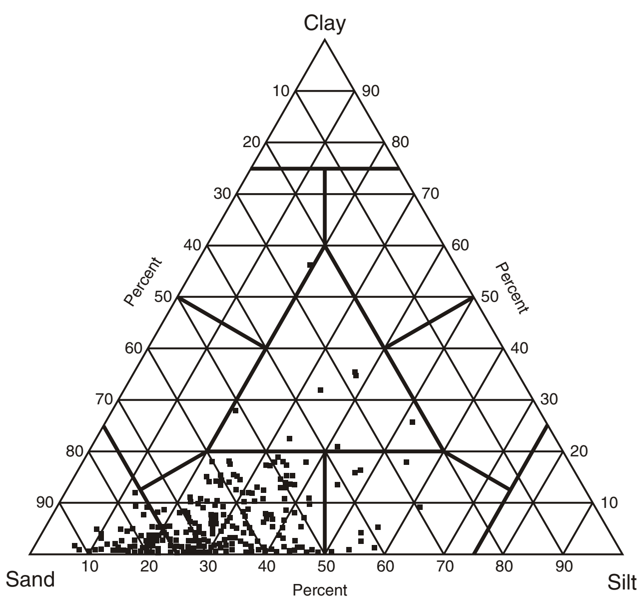


Figure 16. *Ternary diagram of diamicton matrix.*

also, these bedrock types are coarse textured. Although bedrock texture is commonly similar to the texture of overlying diamictons, there are some exceptions. For example, the Upper Humber River basin, west of Birchy Ridge is underlain by soft Carboniferous sandstone and siltstone (Hyde, 1979). However, diamictons in this area are coarser grained and thus are likely not derived from the underlying bedrock. Similarly, Carboniferous sediment underlies the Glide Lake area between Deer Lake and Grand Lake, but is overlain by coarser grained diamictons. In each of these cases, the clast content contains small proportions of the underlying bedrock.

Colour

Diamicton colour is an important field descriptor. In an area, such as the Humber River basin, where there are abrupt lateral changes in bedrock geology the matrix colour is expected to be variable. In an earlier soil mapping survey Kirby (1988) noted a widespread local variation in diamicton colour. Table 5 shows the variations in diamicton colour found during field mapping, in relation to the bedrock geology.

Surficial Geomorphology

Diamicton most commonly occurs as a surface veneer (< 2 m) over bedrock, where the surface expression is controlled by the underlying bedrock structure (Batterson, 2002). In the Upper Humber River valley, diamicton is found as a veneer over northeast–southwest-oriented bedrock ridges. Many of these features have been previously interpreted as till ridges (e.g., Vanderveer, 1981, 1987; Grant, 1989b).

Diamicton ridges are rare but several are found in the Goose Pond area. They are northwest–southeast-oriented linear ridges up to 1500 m long, 400 m wide, and less than 10 m high. A diamicton ridge located in the Hampden River valley, south of Rushy Pond is oriented north–south, about 400 m long, 250 m wide and 20 m high; it has an asymmetric long-profile and is steeper and shorter on the south side. Diamicton thicker than 3 m is found in a south-southwestward- (190°) oriented, 1600-m-long, 400-m-wide and 15 to 20 m high ridge in the Mary Ann Brook valley at the southwest end of Birchy Ridge. The orientation of the ridge is similar to isoclinal bedrock features mapped by Hyde (1979), although no bedrock was found along the ridge.

A group of crag-and-tail hills were mapped in the Long Range Mountains west of Deadwater Brook (Batterson, 2002); they are less than 500 m long, 50 m wide and 10 m high and are oriented west-northwestward (azimuth 280°).

Minor moraine ridges are found at four locations within the study area. In the valley directly south of Blue Grass Brook, a series of small ridges are oriented perpendicular to the valley axis. Similar features are found in the Island Pond Brook valley, east of South Brook. Minor moraine ridges

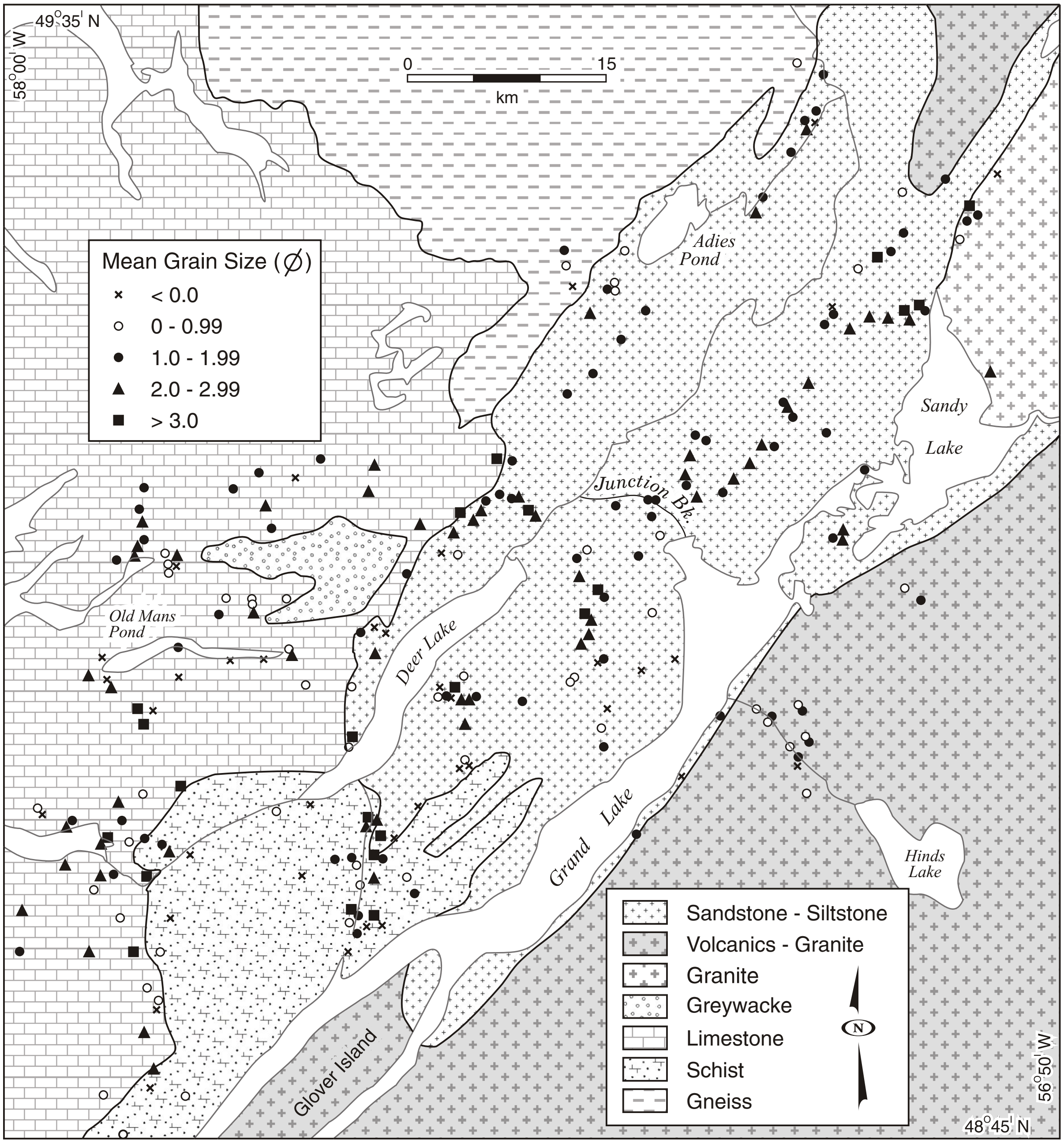


Figure 17. *Mean grain size of diamictons overlain on a simplified bedrock geology map.*

oriented roughly south-southwest to north-northeast are found southwest of Junction Brook, and east–west-oriented moraines are found near Adies Pond. Only the features near Junction Brook were examined in detail. They are 100 to 200 m long, 30 to 50 m wide and less than 3 m high, and are composed of sandy diamicton having a fine sand matrix containing numerous small sand lenses. Clast types are varied and include granites, porphyry, sandstone and siltstone. The granites are similar to those found on The Topsails (units Sm, Oib, Oic, of Whalen and Currie, 1988); the clast fabric is weak (S_1=0.55, S_3=0.17).

Larger moraines are found in the Chain Lakes–Kitty's Brook area of The Topsails and are oriented southeast–northwest, perpendicular to the orientation of the valley, up to 1600 m long and 12 m high and crests up to 300 m apart. Moraines either have roughly symmetrical crossprofiles or are steeper on southwest-facing slopes. Tucker (1974b) showed these moraines to be composed of till, with a preferred clast orientation parallel to the ridge long-axis, and suggested they may have originally been drumlins formed by an ice flow toward White Bay, and subsequently modified by flow down the Chain Lakes valley toward Sheffield

Table 5. Diamicton colours across the Humber River basin (using the Munsell Colour Chart System). All samples are taken from freshly cleaned exposures

Geographic Area	Bedrock Geology	Colour (Moist)	Colour (Dry)
Long Range Mountains, west of upper Humber River valley	Gneiss, granitic gneiss	light greyish brown (10 YR 6/2)	dark greyish brown (10YR 4/2)
Upper Humber River valley	Red, green, grey sandstone, siltstone, conglomerate	dark brown to reddish brown (10YR 3/3 to 5YR 4/3)	light grey to light reddish brown (10YR 7/1 to 5YR 6/3)
Birchy Ridge	Grey sandstone and siltstone	dark greyish brown to very dark greyish brown (2.5Y 4/2 to 10YR 3/2)	grey to light brownish grey (2.5Y 6/0 to 2.5Y 6/2)
East of Birchy Ridge	Grey to red sandstone, siltstone. Granite	dark greyish brown to very dark greyish brown (10YR 4/2 to 2.5Y 3/2)	light grey to light brownish grey (10YR 7/2 to 10YR 6/2)
Hinds Lake area	Granite, granodiorite, rhyolite, gabbro	very dark to dark greyish brown (10YR 3/2 to 10YR 4/2)	light grey to light brownish grey (10YR 7/2 to 10YR 6/2)
Highlands east of Glide Lake	Grey sandstone, siltstone	dark brown to very dark greyish brown (10YR 3/3 to 10YR 3/2)	pale brown to light brownish grey (10YR 6/3 to 10YR 6/2)
West of Glide Lake and Humber River valley	Red to grey sandstone, siltstone, shale	grey–brown (2.5Y 5/2) to dark reddish-brown (5YR 3/3)	very pale-brown (10YR 7/3) to light reddish brown (5YR 6/3)
Highlands west of Deer Lake - sandstone-rich	Red to grey sandstone, siltstone, conglomerate	reddish brown (5YR 4/3)	light reddish brown (7.5YR 6/4)
Highlands west of Deer Lake-limestone-rich	Limestone and dolomite	dark greyish brown (2.5Y 4/2)	light brownish grey (2.5Y 6/2)
Highlands south of Corner Brook	Limestone, dolomite, marble, shale	light olive brown (2.5Y 5/4) to dark yellowish brown (10YR 4/4)	light grey (2.5Y 7/2)
Old Mans Pond area	Limestone, dolomite, shale	olive (5Y 5/3) to dark greyish-brown (2.5Y 4/2)	light yellowish-brown (2.5Y 6/4) to pale yellow (5Y 7/3)

Lake. Tucker (1974b) suggested the ridges were originally formed in the early Wisconsinan, although no supporting data were presented.

Hummocks composed of diamicton are found in the Upper Humber River valley, northeast of Adies Pond; in the Goose Pond to Hinds Lake area; and on the high plateau (~520 m) east of Goose Pond; with smaller areas identified in the Corner Brook Lake valley, and south of Pinchgut Lake.

Hummocks are commonly 50 to 75 m in diameter and up to 10 m high. In the Upper Humber River and Grand Lake areas, these hummocks occur over areas exceeding 5 km^2, whereas those south of Corner Brook are confined to small groups. The hummocks are best seen on aerial photographs where they are surrounded by wetlands. A hummock dissected by a road west of Adies River shows three superimposed diamictons ranging from dark reddish brown (5YR 3/3) to dark yellowish brown (10YR 3/4) to very dark grey (5YR 3/1). The lower two diamictons are broadly similar in texture, having a very poorly sorted (s.d. 3.45ø), silty sand matrix (79 percent sand, 19 percent silt, 2 percent clay). The upper diamicton has a siltier matrix (65.7 percent sand, 34.3 percent silt), and fewer red sandstone clasts than underlying units. A feature near Corner Brook Lake, dissected by a logging road is composed of a very poorly sorted (s.d. 2.99ø), matrix-supported, sandy diamicton (83 percent sand, 14 percent silt, 3 percent clay; mean 0.75ø), containing numerous irregular-shaped sand lenses throughout the unit. Clasts are of local provenance, and up to 2 m diameter and have a poor clast fabric (S_1=0.51, S_3=0.12); hummocks east of Grand Lake were not examined.

Eroded diamicton surfaces are common within the study area. On the west side of the Upper Humber River valley near Cormack, numerous meltwater channels are eroded through the surface diamicton (Plate 1). These channels are commonly 1500 to 3000 m long, although those occupied by modern streams, such as Rocky Brook, Middle Branch, and East Branch, are in excess of 10 km long. Generally, the channels are less than 100 m wide and 5 to 20 m deep and have a steep-sided, flat-bottomed, symmetrical to asymmetrical profile; they are oriented oblique to the valley axis, extending south-southeastward from the basin margins into the valley. The channel gradient varies between 1:40 and 1:95, and no gradient reversals were noted on any of the meltwater channels examined. This suggests that each channel was successively abandoned during northeastward retreat of ice along the Humber River valley. Many paleo-meltwater channels are presently occupied by seasonal streams.

The Hinds Brook valley contains several large steep-sided, flat-bottomed meltwater channels. They are 1100 to 2600 m long, 200 m wide and up to 25 m deep, and are arcuate down-slope and down-valley. Channel gradients are steep, 1:28 to 1:43, with the head of channels becoming successively lower in elevation eastward up the valley, from 390 to 340 m asl. Only the easternmost channel extends to modern Hinds Brook (Plate 2). These channels were ice-marginal or submarginal, formed by a glacier retreating up the Hinds Brook valley toward The Topsails plateau. Numerous small channels are found on the south side of the valley; they are short (< 500 m), steep (1:3) features.

Numerous meltwater channels are found in the northwest slopes of The Topsails plateau, overlooking Sandy Lake. Some of these channels are subparallel to the slope, particularly in the Goose Brook area, between Goose Brook and Kelvin Brook, and east of Goose Pond. These features are interpreted as ice-marginal or submarginal channels formed between ice in the valley and the hillside. Other channels are oriented normal to the slope, and are interpreted as proglacial channels developed from wasting ice on the highlands.

Other areas of eroded diamicton are on the flanks of Birchy Ridge, in the South Brook valley, in the Humber Canal area, in the Glide Lake valley and around Sheffield Lake. Numerous isolated channels are also found scattered throughout the area. In most cases, channels are 200 to 1000 m long and oriented perpendicular to the slope having gradients between 1:10 and 1:100; and indicative of wasting ice farther upslope.

Glaciofluvial (G, Ge, Gh, Gr, Gv)

Glaciofluvial sediments and features are those that are derived from fluvial processes dominated by glacial input, and may occur proximal to a glacier or in meltwater channels some distance from the ice front. Glaciofluvial deposits differ from non-glacial fluvial deposits in that glaciers affect input of sediments and water from fluctuating ice margins, buried ice blocks, and highly seasonal, diurnal and weather dependent variations in discharge (Ashley *et al.,* 1985). In recently glaciated areas, glaciofluvial sediments can be distinguished from non-glacial fluvial sediments by the distribution of sediment, sediment texture, presence of collapse features, and associated structures. Coarse-grained fluvial sediments that are apparently unrelated to modern fluvial activity, e.g., within an ‘underfit' valley or with no modern fluvial source, are likely to be glacially related. Fluvial sediments containing high angled crossbeds (>30°) or faulting may be related to deposition adjacent to an ice mass. Similarly, fluvial sediments showing abrupt lateral and vertical variations in grain size may be related to a glaciofluvial source. A glaciofluvial origin is also indicated by the presence of large boulders within fluvial sediment beyond the competence of any adjacent modern fluvial source.

Distribution

Glaciofluvial deposits comprise about 6 percent of the surficial sediment and are common within the Upper Humber River valley, and in valleys feeding into the Grand Lake and Birchy Lake valleys. Small areas are also found within the Hughes Brook and Goose Arm Brook valleys. Glaciofluvial sediments are relatively rare over the highlands west of Deer Lake, and over the southern extension of the Grenville inlier west of the Upper Humber River valley.

Thickness

There is little data on the thickness of glaciofluvial sediments (Appendix 2). Murray and Howley (1881, 1918) reported at least 29 m of sand and gravel from the Kelvin Brook area near Howley.

In many places, glaciofluvial sediment forms a veneer over diamicton or bedrock. Along the Rocky Brook valley, south of the Cormack Road, glaciofluvial sand and gravel is observed as a 2 m thick veneer over diamicton. Similarly, glaciofluvial veneers were found in the Hinds Brook valley (Appendix 1).

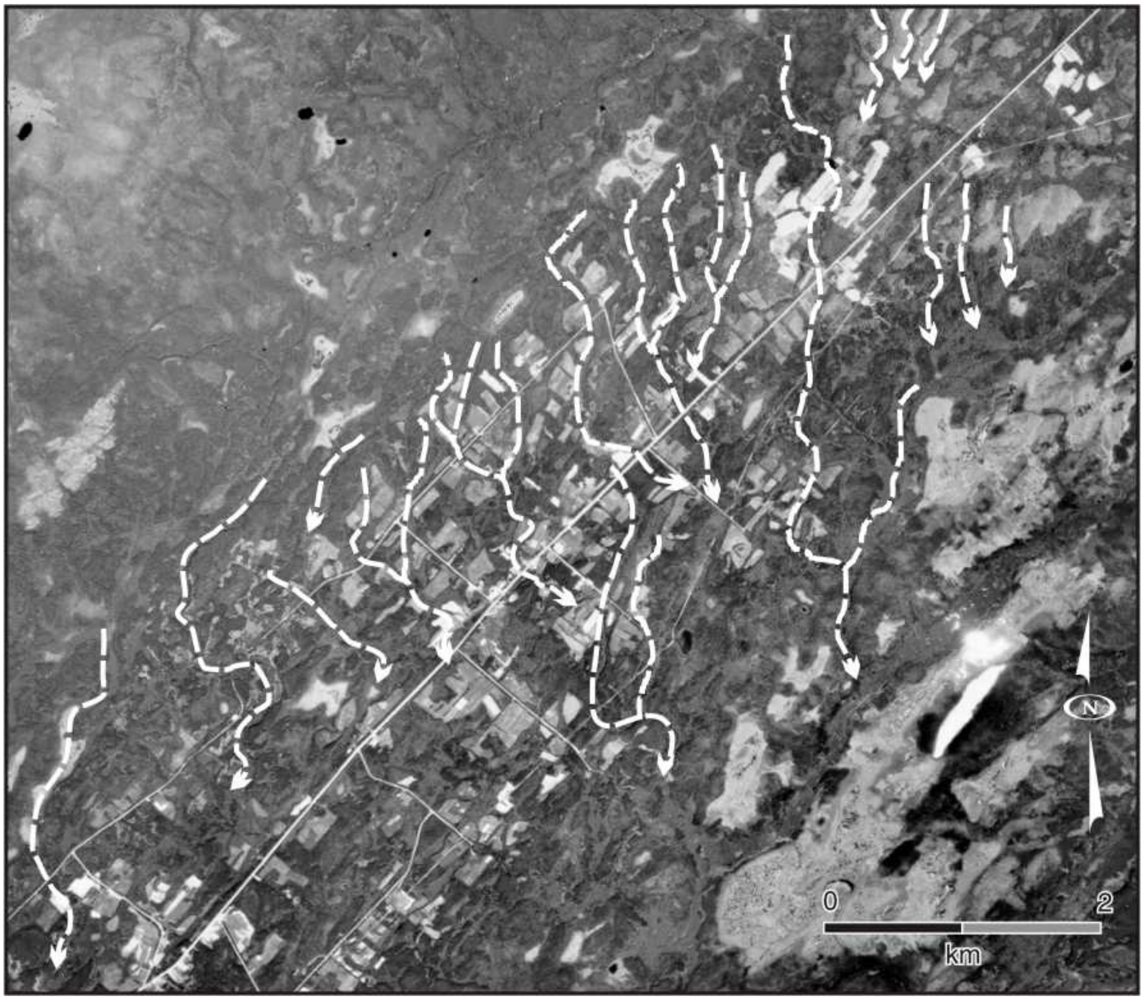


Plate 1. *Vertical aerial photograph of meltwater channels in the Cormack area.*

Grain Size

In the study area glaciofluvial sediments are composed of varying proportions of sand and gravel (~30 to 70 percent gravel, Ricketts, 1993), with less than 5 percent silt–clay. Matrix texture as determined from 64 glaciofluvial samples shows mostly sand matrices (average 91.2 percent) having a mean grain size of 0.17ø (very coarse sand) along with smaller proportions of silt (average 7.6 percent) and clay (average 1.2 percent) (Appendix 1).

Sedimentary Structures

Most glaciofluvial sediments consist of poorly to well-sorted gravel, containing subrounded to rounded clasts up to boulder size in a medium- to coarse-sand matrix. The sediments commonly contain steeply dipping (45 to 60°), trough crossbedded, medium- to coarse-sands and gravels, and numerous randomly distributed, moderately sorted, fine- to coarse-sand lenses; the lenses are irregular shaped and commonly fine upward.

Surficial Geomorphology

Eskers

Eskers are found at several places (Grant, 1989b; Vanderveer and Sparkes, 1982). A prominent ridge, up to 15 m high and 50 m wide, extends 4000 m from Deadwater Brook to Adies Pond (Plate 3). Where exposed, the sediment is gravelly sand (80 percent sand) having a poorly sorted, fine to medium sand matrix; open-work granule gravel and coarse sand lenses are common. Clasts are rounded to subrounded, and are derived from the underlying Carboniferous sandstone bedrock and from gneiss within the Long Range Mountains. The top of the esker is commonly gravel-rich, and is dissected in several places by meltwater channels.

Several smaller esker ridges are found on the north side of Deadwater Brook, each oriented northeast–southwest. Within the Upper Humber River valley, another discontinuous esker ridge is found on the east side of Adies River. Van-

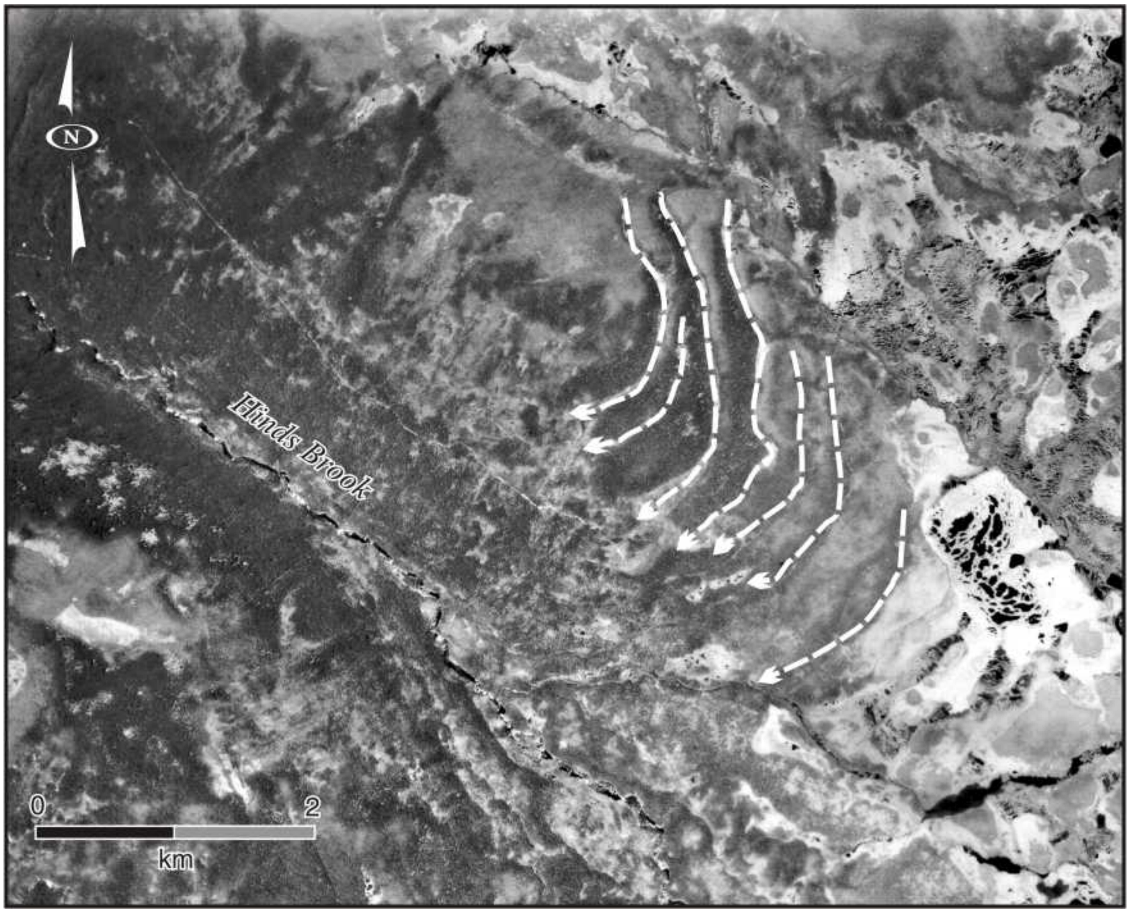


Plate 2. *Vertical aerial photograph of meltwater channels on the north side of Hinds Brook near Hinds Lake.*

derveer and Sparkes (1982) also mapped an esker in the Wigwam Brook area, trending northeast–southwest.

Eskers were also mapped in the Grand Lake basin. The valleys entering Little Grand Lake contain eskers, as does the Red Indian Brook valley farther north. The largest esker extends from near the northwest shore of Hinds Lake, through the unnamed valley south of the Blue Grass Brook valley toward Grand Lake. All these eskers are sinuous, and trend east–west. Similarly, a small esker southeast of Howley is also oriented east–west, although an esker mapped north of Blow Hard Point on the north shore of Grand Lake is oriented northeast–southwest. Grant (1989b) mapped several north–south oriented eskers on the Grand Lake shore at the head of Junction Brook. A number of poorly exposed hummocks composed of sand and gravel were found in this area.

Eskers are not present on the highlands west of Deer Lake or on the Grenville inlier to the west of the Upper Humber River valley.

Kames

A kame was mapped at the head of the Blue Grass Brook valley near Hinds Lake (Site 92100: Appendix 1). It is a hummock about 50 m diameter and about 15 m high. A 5-m-high section was exposed during construction of a logging road, although much of the exposure was obscured by slumping. The sediment is a clast supported, sandy boulder gravel having a medium to coarse sand matrix (10 percent matrix). Clasts are subrounded, ranging from granules to boulders; the largest clast is about 80 cm in diameter. Approximately 40 percent clasts are greater than 5 cm diameter. Clast rock types are dominated by locally derived rhyolite and granite; no bedding was observed.

Some hummocks near Howley are also mapped as kames. They are almost circular to irregular in shape, 20 to 200 m diameter, commonly have a bouldery surface, and are up to 10 m high. They comprise matrix-supported gravelly sand, having a sand matrix containing less than 5 percent silt–clay. A single grain size analysis shows a very poorly sorted sediment (s.d. 2.42ø), 97.5 percent sand and 2.5 percent silt, with a mean grain size of -1.58ø (Site 92016: Appendix 1). Generally, the sediment is structureless, and clast types are dominated by locally derived granites. Clast fabric is moderate, girdle (S_1=0.66, S_3=0.05), showing a preferred clast orientation of about 345°. Glaciofluvial hummocks were also found in groups in the lowlands east of Junction Brook.

Plate 3. *An esker ridge near Adies Pond looking northward.*

the highlands between the Glide Lake and Deer Lake valleys, overlooking the South Brook valley; and The Topsails southwest of Hinds Brook.

Glaciolacustrine and Lacustrine (L, Lt, Lr)

Sediment and geomorphological features produced within bodies of standing water were identified at elevations above the proposed marine limit for the area. Those which are not related to modern lakes have been interpreted as glaciolacustrine. Glaciolacustrine and lacustrine deposits comprise about 1 percent of the study area, and are subdivided into shoreline features (strandlines and deltas), and proximal and distal sediments.

Distribution

Sediment and geomorphological features related to high water levels have been found in the South Brook, Grand Lake, and Birchy Lake valleys and have been mapped as strandlines along the western shore of Grand Lake. They occur as discontinuous, flat benches backed by a steep bluff. Commonly, the bases of the bluffs have a concentration of boulders. The most clearly defined strandlines are on the east side of Thirty-ninth Brook valley, about 350 m inland from Grand Lake (Table 6). Here, three strandlines at 140.5, 153 and 175.5 m asl are separated by 2 well-defined bluffs (Plate 4). Each strandline has been eroded into a diamicton-covered slope, and shows a concentration of granitic boulders (from The Topsails to the east), presumably derived from the slope itself. The uppermost terrace is eroded by a channel, the lower part of which is truncated by a ridge that extends across the entire channel. A test-pit dug in this ridge shows the sediment as a dark greyish brown (2.5Y 4/2), clast-supported (30 percent matrix) sandy gravel having a poorly sorted (s.d. 3.74ø), silty sand matrix (83.1 percent sand, 16.1 percent silt, 0.8 percent clay). The clasts are subrounded to subangular, granule to boulders and, on the basis of its position within the channel and its sediment composition, this feature is interpreted as a beach ridge.

Several other areas showing strandlines were found north of the Thirty-ninth Brook valley (Table 7). North of Wetstone Point, strandlines were noted at 57, 62 and 85 m above lake level, or 143.5, 149 and 171.5 m asl respectively (Sites 93026, 93083: Appendix 1). East of Johnsons Pond, strandlines were found at 49 m and 56 m above lake level (Site 93027: Appendix 1). In each case, and at Thirty-ninth Brook, strandlines were coincident with areas of blown-down trees.

Eroded Glaciofluvial Sediment

Meltwater erosion channels of deposited material are common in valleys containing glaciofluvial sediment, and indicate the rapid variations in discharge, and rapidly shifting stream channels common in proglacial fluvial environments. Channels in these areas are commonly short (less than 100 m), with gradients determined by the valley in which they are found.

Discussion

The areal distribution of glaciofluvial features described here, is generally similar to that mapped by Grant (1989b). The distribution of meltwater channels provides some data on the location of areas of ice wastage. The lack of meltwater channels over the highlands west of Deer Lake, and in the southern part of the Grenville inlier suggests that ice retreated rapidly from these areas, and did not waste *in situ*. In contrast, the numerous meltwater channels on The Topsails plateau show this to be an area of ice disintegration. There is a radial pattern to meltwater channels that extend from The Topsails plateau into the Hinds Brook and Kelvin Brook valleys, and into the Sandy Lake–Grand Lake basins from the 340 m asl plateau between Goose Pond and Hinds Lake. The presence of meltwater channels parallel to the slope between this plateau surface and the higher level at about 520 m asl to the east of Goose Pond, suggests that ice remained on the lower plateau as the slopes to the upper plateau became ice free.

The radial pattern of meltwater channels also suggests that wasting ice covered the southern end of Birchy Ridge;

Table 6. Features associated with higher water levels along Grand Lake, in the lower Thirty-ninth Brook valley. Elevations represent the mean of three separate altimeter readings

Feature	Latitude(°N)	Longitude(°W)	Height(m)	Width(m)	Height above Grand Lake (m)	Elevation (m asl)
Strandline	49°01.3'	57°21.8'			53.5 ± 1.5	140.5
Scarp			6.5 ± 1			
Terrace				approx. 75		
Strandline	49°01.3'	57°21.8'			66 ± 1	153
Scarp			16 ± 1			
Terrace				approx. 150		
Strandline ?	49°01.4'	57°21.8'			88.5	175.5

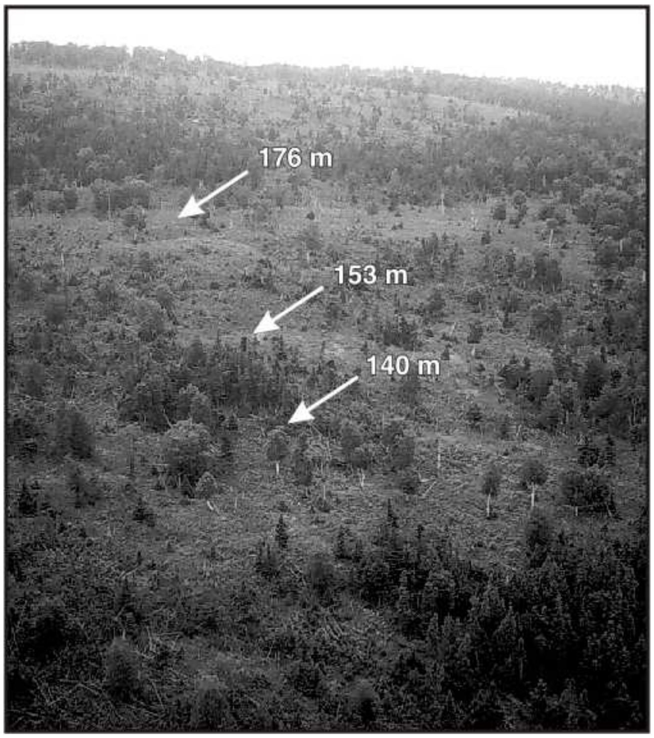


Plate 4. *Aerial photograph showing strandlines along hillside adjacent to the Thirty-ninth Brook valley.*

Strandlines were not found on the east side of Grand Lake. Instead, well-defined, flat-topped deltas were noted (Table 8), at the mouth of Hinds Brook (118 and 157 m asl), north of Grindstone Point (144 m asl), Little Pond Brook (140 m asl), Harrys Brook (144 m and 176 m asl), south of Grand Pond Point (145 m asl), Connors Brook (130 m asl), and at Lewaseechjeech Brook (128 m asl) (Plate 5). Exposures within these features commonly show interbedded sand and gravel dipping toward Grand Lake at 20 to 30°. North of Hinds Brook, deltas are commonly fan-shaped.

The South Brook valley also contains deltas. Two deltas were found at about 135 m asl: one between Whitefish Creek and Carp Creek, and the other south of Salmon Creek. A delta was identified on the west side of the valley south of the transmission line at 145 m asl, and a large flat-topped (delta?) was found on the opposite side of the valley having a surface elevation of 150 m asl. A small delta at 150 m asl, was also identified near Northern Harbour. Each of these features was flat-topped and showed interbedded sand and gravel dipping into the valley.

Sedimentology

Generally, sediments are sandy gravels, and contains about 57 percent gravel, 42 percent sand, and 1 percent silt–clay (mean of 4 samples, Ricketts, personal communication, 1995). Matrix components are generally very poorly sorted (s.d. 2.56ø) sand, (95 percent sand, 5 percent silt and <1 percent clay, mean of 10 samples) having mean grain size of -1ø.

Sections commonly show interbedded fine and medium sand, laterally grading to interbedded sandy gravel and gravely sand. The interbedded sands are well sorted, and planar bedded and the beds extend to over 5 m laterally; pebbles are rare and do not deform surrounding beds. In the deltas adjacent to Grand Lake, beds dip toward the lake at 10 to 15° and the interbedded gravelly sands and sandy gravels have 10 to 50 percent sand matrix. Clasts are commonly subrounded, composed of local rock types, and are less than 5 cm in diameter. Beds dip toward Grand Lake at 20 to 25°.

Grand Lake and Sandy Lake contain several modern lacustrine features. Recurved spits are found at the mouths of Blue Grass Brook and Hinds Brook (Plate 6). The Hinds Brook spit is 800 m long and is anchored to the north shore of Hinds Brook from where it extends 450 m northwest into the lake before curving 350 m toward the north. An area of flats occupies the foreshore northeast of the spit and is composed of sand and gravel. A separate spit is found on the south side of Hinds Brook, although it is smaller (200 m long) than that on the north shore. The Blue Grass Brook spit is smaller, extending 150 m northeastward from the lakeward edge of the delta on the south side of the brook. The spits indicate northward shoreline transport of sediment. Baymouth bars were identified in 4 places: 800 m north of Blue Grass Brook, 900 m south of Alder Brook, at

Table 7. Features associated with higher water levels along the west side of Grand Lake. All measurements are by altimeter, and are accurate to within ± 2 m

Site	Location	Latitude (°N)	Longitude (°W)	Elevation (m asl)	Comments
93014	Thirty-Ninth Brook	49°01.3'	57°21.8'	176	Front of terrace at 170 m. Boulders at base. Diamicton hillside.
93014	Thirty-Ninth Brook	49°01.3'	57°21.8'	153	Front of terrace at 147 m. 17 m bluff behind terrace.
93014	Thirty-Ninth Brook	49°01.3'	57°21.8'	140	Front of terrace 135 m. 7 m bluff behind terrace.
93016	Island Pond Brook	48°56.4'	57°28.7'	137	Front of terrace 137 m. 21 m bluff behind terrace.
93026	Just north of Wetstone Point	49°04.4'	57°16.4'	171.5	14 m bluff behind terrace. Boulders common at base of terrace.
93026	Just north of Wetstone Point	49°04.4'	57°16.4'	143.5	Boulders common at base of terrace. 6 m bluff behind terrace. Front of terrace 139.5 m. Minor slope break at 168 m.
93027	East of Johnsons Pond	49°05.9'	57°16.2'	143	7.5 m bluff behind terrace. Boulders at base of slope.
93027	East of Johnsons Pond	49°05.9'	57°16.2'	136	Possible terrace.
93033	North of Wetstone Point	49°04.8'	57°16.2'	149	7.5 m bluff behind terrace. Boulders at base of slope. Other minor breaks of slope at 126 m, and 179.5 m.

the mouth of Coal Brook, and 600 m west of Blow Hard Point. The latter two locations were not vegetated. A small tombolo was noted at the southern end of Sandy Lake, east of Howley.

All these features have developed since the level of Grand Lake was raised in 1929. The features were formed by wave transport, produced by dominant southwesterly winds that are funneled along the lake. Referring to Grand Lake, Jukes (1842, page 147) commented that ".. there was often a tide in the pond after a high wind. This is no doubt caused by the banking up of the water at one end from the pressure of the wind". Strong southwesterly winds are also common along Deer Lake (Banfield, 1981), oriented the same direction as Grand Lake. Wind velocities along Grand Lake are enhanced by the steep, high valley sidewalls.

Marine (M, Mt, Mf)

Marine sediments are those deposited on, or seaward of, modern or paleo-coastlines and compose about 1 percent of the surface geology. They are subdivided into littoral and sublittoral features and deposits.

Distribution

Areas mapped as marine are found along the modern coast, and in the lower reaches of the Humber River basin below about 60 m asl.

Thickness

Although surface exposures are generally poor, commonly showing less than 3 m of sediment, evidence from drill logs shows that the thickest deposits of sediment in the basin are found in the Lower Humber River valley, occupying a depression up to 100 m below the modern sea level and is filled mostly with silt, clay or sand (Appendix 2).

Surficial Geomorphology

Littoral features include marine terraces and deltas composed of sand and gravel. Brookes (1974) suggested

Table 8. Deltas formed in higher water levels, above marine limit defined for the Humber River basin. Elevations are from altimeter (± 2 m) or topographic maps (± 5 m)

Site	Location	Latitude(°N)	Longitude (°W)	Elevation (m asl)	Comments
93032	Hinds Brook	49°04.8'	57°12.5'	118	Front edge of delta 114 m. Well developed surface.
	Lewaseechjeech Brook	48°38.4'	57°56.7'	128	Topographic map. Unvisited.
	Connors Brook	48°49.4'	57°32.9'	130	Topographic map. Unvisited.
	Grand Lake	49°06.0'	57°11.0'	129-141	Fan deltas. Tops difficult to determine. Poorly exposed.
89006	South Brook	48°58.5'	57°37.2'	134	Abandoned gravel pit. Contains fossil ice wedges.
	2 km south of Grand Pond Point	48°51.9'	57°30.9'	135	Topographic map. Unvisited.
91235	South Brook	48°57.5'	57°37.2'	135	West side of valley. Abandoned gravel pit.
93030	Little Pond Brook	48°57.8'	57°20.3'	140	Delta surface continues on north side of brook.
93017	Harrys Brook	48°54.0'	57°25.6'	144	Delta front at 129 m. 32 m bluff behind delta. Bluff mostly gravels.
93031	North of Grindstone Point	49°00.8'	57°16.5'	144	Front edge of delta at 134 m.
91232	South Brook	48°57.4'	57°36.3'	145	East side of river. Dissected by channels.
91087	Northern Harbour	48°53.9'	57°38.1'	150	Top of 20 m terrace.
93032	Hinds Brook	49°04.8'	57°12.5'	157	Delta surface highly dissected by channels
93017	Harrys Brook	48°54.0'	57°25.6'	176	Lower surface at 144 m.
93111	Birchy Lake	49°16.9'	56°47.3'	210	Possible delta. No internal structure noted.
	Kitty's Brook	49°09.9'	56°51.1'	250	Tucker, 1974b

that the marine limit in the Humber Arm/Bay of Islands area was 49 m asl, based on the elevation of deltas at Corner Brook and Cox's Cove. Marine features have not been recognized above 60 m asl in the western parts of the Humber River basin (*see* Batterson, 1998). Marine deltas and terraces are recognized below 60 m asl in the Humber Arm, Goose Arm, Penguin Arm and Humber River valley areas (Table 9).

At the head of the Humber Arm, raised deltas are found in the Hughes Brook valley (60 m asl), at the head of the Wild Cove valley (50 m asl), and in the Humbermouth area (51 m asl). These features were identified on the basis of morphology, showing flat tops with steep fronts on the seaward side. Internal structure, revealed during gravel extraction operations, showed foreset beds of sand and gravel dipping steeply toward the coast (Plate 7).

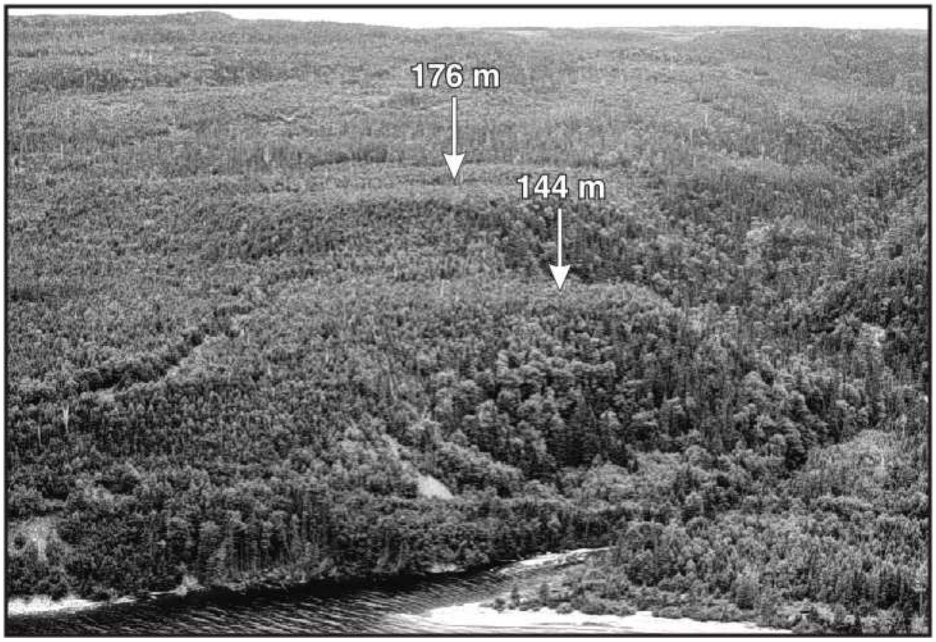


Plate 5. *An aerial oblique view showing a delta at the mouth of Harrys Brook on the east side of Grand Lake.*

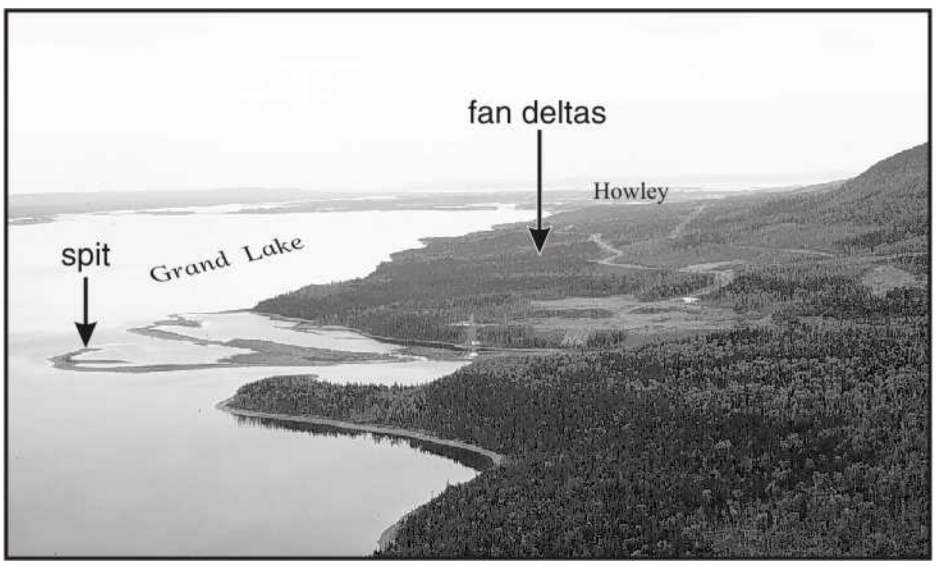


Plate 6. *A recurved spit at the mouth of Hinds Brook.*

width of 7.2 km, 4 km upstream. The delta is flat-topped and has a surface elevation of about 45 ± 2 m (altimeter estimate). It extends on both sides of the Humber River, and is dissected both by the modern river and by a 250 m wide and 30 m deep channel on the south side, part of which is now occupied by Round Pond; the south side of this channel is flanked by bedrock. In 1993 and 1994, road construction revealed interbedded sand and pebbly sand. Medium to fine sand beds were 1 to 3 cm thick, moderately sorted, and ungraded to normally graded. Pebbly sand beds were 5 to 10 cm thick, poorly sorted and have a medium to coarse sand matrix containing granules and pebbles up to 3 cm diameter. Beds dipped about 24° toward 060°, indicating flow down the modern valley and are interpreted as deltaic sediments. The position of the delta at the mouth of Deer Lake, extending across the valley, with a steep upstream face, and foreset beds indicating paleo-flow down the valley, suggests it was an ice-contact deposit. Incision of the delta presumably occurred as the Humber River became re-established in the lower reaches of the valley during periods of isostatic uplift in the early Holocene.

Features interpreted as deltas based on their surface morphology and internal structure are also found at Pasadena (33 m asl) and Pynn's Brook (33 m asl) (Batterson and Vatcher, 1992a), and at Little Rapids (45 m asl), Little Harbour (44 m asl) and Nicholsville (48 m asl). In all cases, these features have steep fronts, but grade upstream to glaciofluvial or fluvial systems.

Brookes (1974) described the delta at Humbermouth as extending on both sides of the Humber Arm. The delta was interpreted as ice-contact based on the presence of ice-contact sediments and lodgement till interbedded with the delta sediments. The position of the feature at the mouth of a long valley, and the extension of the delta on both sides of the Humber Arm suggests that an ice-contact origin is likely.

An ice-contact delta is found at the mouth of Deer Lake, below which the Humber River valley narrows (Plate 8). The valley is only 1.3 km wide at this point, compared to a

Terraces interpreted as marine are found in several locations along, and adjacent to, the coast (Table 10). The terraces are locally continuous over kilometres. Prominent terraces front the Humbermouth delta, extending from Prince Edward Park to Wild Cove at about 33 m asl, and along the lower Hughes Brook valley at about 20 m asl. In Goose Arm, a terrace was identified at 21 m asl. Other terraces were identified adjacent to modern Deer Lake. Terraces at 21 and 22 m asl were found on the south side of Deer Lake near Pynn's Brook, and the north side near Eighth Brook.

Table 9. Deltas formed below proposed marine limit in the Lower Humber River valley. Deltas were identified primarily on the basis of morphology. Internal sedimentary structure was examined where exposed. Additional data is listed in Appendix 1

Site	Location	Latitude (°N)	Longitude (°W)	Elevation (m asl)	Comments
91103	Pynn's Brook	49°05.3'	57°32.5'	33	Abandoned gravel pit.
91069	Pasadena	49°00.6'	57°36.8'	33	Community built on delta surface.
93030	Hughes Brook	48°59.8'	57°53.3'	43	Abandoned gravel pit. Mostly obscured by slumping.
92108	Little Harbour	49°08.3'	57°28.5'	44	Abandoned gravel pit.
93008	Junction Brook	49°13.0'	57°20.1'	45	Abandoned gravel pit. Mostly sloped.
91187	Little Rapids	48°59.7'	57°42.5'	45	Abandoned gravel pit. Fan delta.
91190	West end Deer Lake	49°00.0'	57°42.4'	46	Delta largely removed for highway upgrading.
91144	East of North Brook	49°09.6'	57°31.4'	47	Abandoned gravel pit.
92027	Nicholsville	49°11.5'	57°27.4'	48	Abandoned gravel pit.
91219	Wild Cove	48°58.4'	57°52.1'	50	Reading from top of fan.
	Humbermouth	48°57.8'	57°53.4'	51	Site of Mount Patricia cemetery.
91029	Hughes Brook	49°00.2'	57°52.6'	58	Possible delta in community. Some surface stripping.
91173	Hughes Brook	49°01.8'	57°51.2'	59	Abandoned gravel pit. Some surface stripping.

Terraces at 36 and 40 m asl were also found at Pynn's Brook, and at 28 and 38 m asl near Eighth Brook. Grant (1973, 1989b) also noted these features along Deer Lake.

Sedimentology

Sediments identified as marine are those found below the marine limit of 60 m asl, and/or associated with marine features. They include littoral sand and gravel deposited adjacent to the modern coast as beach-ridges or a discontinuous veneer of sediment over bedrock or diamicton. These are found within a thin belt along the shores of Humber Arm, Penguin Arm, Goose Arm and North Arm. Littoral sediments may also be found in raised marine terraces or deltas.

Rhythmically bedded silt and clay is found within the Humber River valley at elevations below about 50 m asl, commonly capped by a thin veneer of sand and/or gravel. Silt and clay beds are interpreted as sublittoral marine deposits, based on their continuity inland from the coast, paleontology, and association with marine features. They are commonly light reddish brown to reddish brown (5YR 6/3, dry to 5YR 4/3, moist) clayey silt to silty clay. Mean grain size determinations from twenty-one exposures show 11 percent sand (range 0.6 to 35.3 percent), 62 percent silt (range 27.2 to 91.5 percent), and 27 percent clay (range 1.5 to 67.2 percent), having a mean grain size of 6.3ø (fine silt). Spatial textural trends were not evident through the valley.

Fluvial (A)

Fluvial sediments are those deposited postglacially by rivers. Fluvial sediments are defined by their position relative to modern stream channels (i.e., they occur within the modern flood plain), and by their stratigraphic position (e.g.,

Plate 7. *The internal structure of a raised delta at Pynn’s Brook showing foreset and topset bedding.*

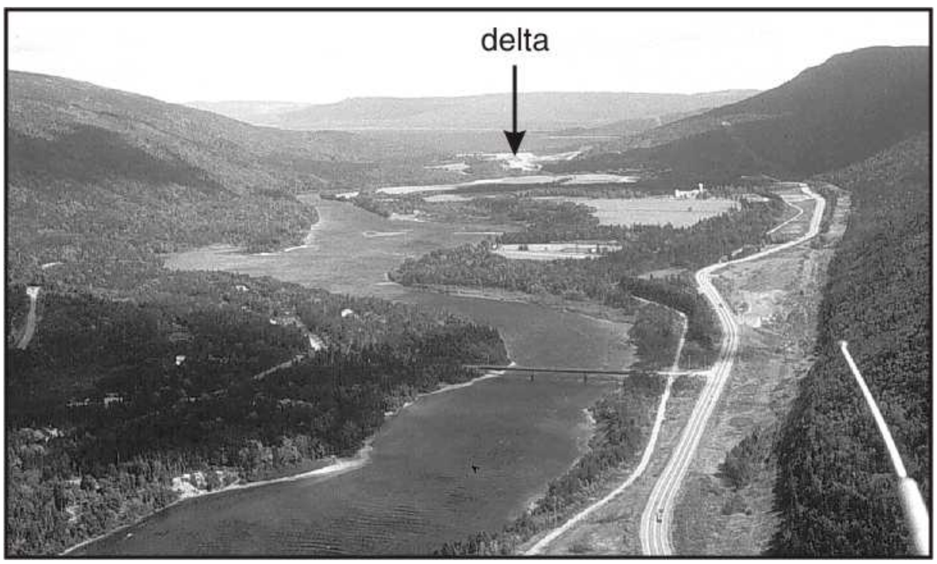


Plate 8. *An aerial oblique view of the Lower Humber River valley looking northeastward.*

occurring above sediments identified as postglacial marine muds). However, the transition from sediments deposited within a fluvial system fed by glacial meltwater, to a system with no glacial input is difficult to define (Ashley *et al.*, 1985). Fluvial sediments comprise about 3 percent of the surficial sediment in the study area, and range from gravel to silt.

Distribution

Postglacial fluvial sediments are common adjacent to the modern Humber River valley, particularly in the lower reaches below Deer Lake. They are also mapped in the Kitty's Brook, Deadwater Brook, Rocky Brook, South Brook, Otter Brook and Hughes Brook valleys, as well as in small valleys scattered across the area.

Thickness

Fluvial sediments are commonly found as a veneer over marine muds, diamicton or bedrock. A drill hole log from Humber Village in the Lower Humber River valley recorded 4 m of sands overlying clay (Appendix 2). Elsewhere across the basin, fluvial sediments are less than 3 m, and commonly less than 1 m thick.

Sedimentology

Fluvial sediments commonly are well sorted fine to coarse sand, to poorly sorted sand and gravel. Near Little Rapids, adjacent to the Humber River, a 3-m exposure shows planar cross-stratified, well-sorted fine to medium sands overlying marine muds. Planar tabular crossbedding indicates deposition by current flow down the modern valley. Some sections show low-amplitude ripples formed under low flow conditions.

In contrast, sections within the Humber River gorge (exposed during road construction in 1991) have a succession of a basal unit of planar bedded, moderately to well sorted, medium to fine sand, having occasional ripples with 10 to 15 cm wavelength and 1.5 cm amplitude. These sands are overlain by 50 to 100 cm of planar-bedded, clast-supported pebble gravel. Clasts are subrounded, of mixed rock types, and have a 2- to 3-mm-surface coating of reddish-brown silty clay. Overlying this gravel unit are poorly sorted, planar-bedded sandy gravels that have a fine to medium sand matrix. The unit commonly contains ovoid, open-work gravel lenses. The sediments within the Humber River gorge suggest increased current flow, probably related to constriction of the valley within the lower reaches. To the east of the Humber River gorge the valley is wide, and gravelly sediments are less common.

Table 10. Location and elevation of terraces found adjacent to the modern coast, or adjacent to modern Deer Lake

Site	Location	Latitude (°N)	Longitude (°W)	Elevation (m asl)	Comments
91019	Wild Cove	48°59.0'	57°54.5'	20	Top of terrace bluff. Approximate elevation.
92161	Goose Arm	49°10.9'	57°51.3'	21	Top terrace bluff.
	Pynn's Brook	49°06.6'	57°31.5'	21	
	North side Deer Lake	49°06.5'	57°35.5'	22	Beach ridge about 2 m high and 12 m wide. Continuous along hillside.
	North side Deer Lake	49°06.5'	57°35.5'	28	Possible shoreline.
91010	Humbermouth	48°58.0'	57°53.7'	33	Prince Edward Park area.
91009	Dawe's Pit	48°57.3'	57°53.2'	35	Top of gravel pit.
	Pynn's Brook	49°06.6'	57°31.5'	36	Top of 14 m bluff.
	South Brook	49°00.9'	57°37.3'	36	Top of 5 m beach ridge.
	North side Deer Lake	49°06.5'	57°35.6'	38	Shoreline. Distinct slope break.
	Pynn's Brook	49°06.5'	57°31.5'	40	Top of 4 m bluff.
	Hughes Brook	49°00.1'	57°52.6'	55	Base of terrace. 11 m bluff behind.
91008	Humbermouth	48°57.0'	57°52.5'	61	Top of terrace.
	Hughes Brook	49°00.1'	57°52.6'	66	Top of terrace.

Surficial Geomorphology

Fluvial terraces are found adjacent to the modern Upper Humber River, and its tributaries, particularly Rocky Brook (Table 11). Lower terraces are commonly continuous for more than 10 km, and are graded similar to the modern channel and have a gradient of about 1:750 in the lower reaches below Harrimans Steady, and about 1:450 between Harrimans Steady and Big Falls. Higher terraces are commonly discontinuous, and no gradient was measured.

Seventeen observations were made on four terrace levels in the Rocky Brook and Upper Humber River near Reidville. They showed terraces at 43.5 ± 0.5 m (2 observations); 36.0 ± 1 m (2 observations), 28.0 ± 2 m (4 observations), and 23.5 ± 0.5 m (7 observations), plus single observations at 10 and 15 m asl. Lower terraces are graded parallel to the modern river, whereas the upper terrace gradient was not measurable.

Organic (O)

The Humber River basin contains large areas of wetland that comprise 9 percent of the surface cover.

Distribution

Wells and Pollett (1983) mapped the distribution of wetland. The Upper Humber River valley contains domed bog and slope fen comprising about 20 percent of the valley floor. Both eccentric and concentric types of domed bog are found that have surface slopes between 1:100 and 1:150, and have concentric pattern of pools. These wetlands are ombrotrophic, compared to the minerotrophic slope fens that are found on 1:20 to 1:4 slopes. Slope fens are found throughout the Deer Lake basin and Lower Humber River valley, being especially common in areas underlain by limestone bedrock. These are less acid and the most nutrient-rich

Table 11. Terraces found adjacent to modern Upper Humber River. All elevations are from altimeter and are accurate to ± 2 m

Location	Latitude (°N)	Longitude (°W)	Elevation (m asl)	Comments
Rocky Brook	49°12.4'	57°26.2	10	5 m terrace bluff. Base of bluff.
Rocky Brook	49°12.4'	57°26.2	15	Top of bluff
Rocky Brook	49°12.4'	57°26.4'	23	14 m terrace bluff. Base of bluff.
Reidville	49°13.4'	57°24.5'	23	Top of bluff
Rocky Brook	49°12.2'	57°26.3'	23	Top of bluff
East Reidville	49°14.6'	57°22.0'	23	Top of bluff
East Reidville	49°14.2'	57°22.4'	24	Top of bluff
Reidville	49°13.6'	57°23.8'	24	Base of bluff. 6 m terrace height.
East Reidville	49°13.8'	57°22.7'	24	Top of bluff.
Reidville	49°13.4'	57°24.0'	27	Top of bluff
Rocky Brook	49°12.8'	57°25.9'	28	Top of bluff (delta?)
Reidville	49°13.6'	57°23.3'	28	Top of bluff. Near Community Park.
Reidville	49°13.9'	57°23.9'	30	Top of bluff. By ball field.
nr Reidville	49°13.1'	57°25.2'	35	Top of bluff
Rocky Brook	49°12.4'	57°26.4'	37	Top of bluff
Rocky Brook	49°12.3'	57°26.6'	43	Top of bluff
Reidville	49°13.7'	57°23.8'	44	Top of bluff

of fens in Newfoundland. Basin bogs are common on The Topsails, characterised by treeless areas of peat having generally flat surfaces and rare pools. In each of these areas, wetlands have developed over an underlying diamicton surface or bedrock. Wetland areas adjacent to the modern coast are commonly underlain by marine sediments. Large areas of organic terrain are uncommon west of the Deer Lake valley and over the Grenville inlier west of the Upper Humber River, although these areas contain small patches of basin and slope bog.

Thickness

Domed bogs are up to 10 m thick, whereas basin bogs and slope fens are generally shallow accumulations, rarely exceeding 2 m thick.

Colluvium (C, Cf, Ca)

Colluvial sediments are deposited at the base of slopes through a combination of falling (free-fall or rolling), sliding or flowing. Colluvium comprises about 2 percent of the surface cover.

Distribution

Colluvial aprons are commonly found along the sidewalls of oversteepened glacial valleys at Old Mans Pond, Hinds Brook, along Goose Arm, Penguin Arm and North Arm, the west shore of Grand Lake, north shore of Glover Island, through the Humber River gorge, and in the Birchy Lake valley.

Colluvial fans form at the mouth of gullies and are fed by landslides and avalanches. Numerous fresh scars show these processes to be active. Historically, landslide and avalanche events have affected the Humber River valley, occasionally with fatal results (Batterson *et al.*, 1995). Small-scale landslides occur in poorly consolidated sediment. Recent examples of such features are in the Humber River gorge (blocking the Trans Canada Highway in 1985), along the banks of West Rocky Brook near Cormack (resulting from stream undercutting in 1993), and along Riverside Drive in Corner Brook (initiated by highway construction in 1994).

Areas of colluvium, from sliding or flowing by solifluction or gelifluction processes, are found in highland areas, although generally not as mappable units. Diamicton covered slopes south of Old Mans Pond likely consist of resedimented diamicton.

Thickness

Colluvial deposits thicken downslope, and are likely up to 5 m thick at the base.

Sediment

Colluvial deposits are commonly coarse-grained gravel and sand. Larger talus slopes show a typical downslope increase in grain size (e.g., Clayton, 1972). No samples were taken from colluvial deposits.

GLACIAL SEDIMENTS AND STRATIGRAPHY

REVIEW AND OBJECTIVES

Surficial mapping of the Humber River basin shows that most of the area is covered by diamicton, and lesser amounts of glaciofluvial, marine and glaciolacustrine sediments; these sediments are exposed in hand-dug test pits, stream and road-cuts and from gravel pits. Most of these exposures are small (less than 1 m), and from these basic observations descriptions were written; sediment characteristics are described in a previous section. Larger exposures are confined to gravel-pit operations (which by their nature are ephemeral features of the landscape), and river sections are produced by undercutting of banks that are areas of slope instability.

Sediment exposures are scattered throughout the basin, rarely exceeding 100 m in lateral extent. Larger sections are preferentially described as they contain better exposures of sediment units, allow examination of lateral extent, and more than one stratigraphic unit commonly is exposed. As a consequence, some areas and/or sediment types may be over-represented. Broad areal sediment distributions were described in the previous section. Examination of over 400 hand-dug and backhoe test pits from across the basin has ensured that no major sediment types were omitted.

Sediment types are discussed in detail, commencing with the glacial deposits. Proglacial and non-glacial (Holocene) sediments are discussed in the section on Deglacial and Postglacial Sediments and Stratigraphy, page 62.

The objective of this section is to describe, and interpret, the larger exposures of glacial sediment found across the basin, discuss the stratigraphy, determine genetic environments, and to correlate, where possible, individual exposures.

GLACIAL SEDIMENTS

Review

Till is a sediment that has been transported and subsequently deposited by, or from, glacier ice, with little or no sorting by water (Dreimanis, 1982). Tills are subdivided into primary tills (ortho-tills) and their re-sedimented secondary products (allo-tills). Primary tills are released directly from glacier ice, whereas secondary tills are remobilized by mass-movement processes or by settling through a water column subsequent to their release from the glacier (Ashley *et al.*, 1985). However, primary and secondary tills are so intimately related that it may be difficult to separate them in any meaningful manner (Dreimanis and Lundqvist, 1984). Drewry (1986) suggested that the term till is no longer applicable as an indicator of genesis, and that it should be replaced by non-genetic terminology, such as ‘diamict'. This term was originally coined by Flint *et al*. (1961), later resurrected by Eyles *et al*. (1983b) and Miall (1983). Environmental interpretation would append a prefix, e.g., glacial diamicton.

Dreimanis and Lundqvist (1984) suggested there are three conditions common to all tills. They consist of glacially transported debris; they have a close spatial relationship to glaciers, either being deposited by, or from, them, and sorting by water is minimal or absent. The usefulness of this proposal lies in the conceptualization of the term till, rather than developing any criteria for the differentiation of till-types or distinguishing tills from other, similar-looking sediments. Sediments that resemble tills are found in other depositional settings, such as sediment-gravity flows in marine or terrestrial environments (*see* Schermerhorn, 1974; Shultz, 1984; Spalletti *et al.*, 1989), diamictons deposited by ice rafting (*see* Ovenshine, 1970; Gravenor *et al.*, 1984; Domack and Lawson, 1985), volcanic environments (*see* Mills, 1984; Car and Ayres, 1991; Smith and Lowe, 1991) and Oberbeck *et al*. (1993) described Precambrian diamicts produced by meteor impacts. This highlights the need for a clear understanding of the genetic and physical differences not only between different glacial environments, but also between glacial and non-glacial diamictons.

Primary (Ortho-) tills

These include lodgement and deformation tills that are produced by actively sliding glaciers; and melt-out till and sublimation till that are deposited in a passive environment.

i) Lodgement till is defined as a sediment, "deposited by plastering of glacial debris from the sliding base of a moving glacier by pressure melting and (or) other mechanical processes." (Dreimanis, 1988, p. 43). Many of the characteristics of these tills have been well described in the literature, including clast ‘traffic jams' (Åmark, 1980), boulder pavements (Humlum, 1981; Hicock, 1991), and faceted and bullet-shaped clasts (Boulton, 1978; Kruger, 1984). The direct contact of ice with the substrate during lodgement produces several features. Sole marks are formed by dragging clasts through a deformable bed (Shaw, 1982). Till fills the resulting grooves producing spoon-shaped ridges at the base of the lodgement till bed. Subglacial deforming beds also exhibit subhorizontal jointing or fissility, characterised by shingled, crosscutting till units. Fissility represents minor failure planes beneath the ice, caused by loading stress (Boulton, 1971; Eyles, 1993), or as a result of brittle deformation (Hart, 1995). The sliding and internal deformation during lodgement means that unlithified intraclasts are absent from lodgement tills, although Kruger (1979) recognized smudges consisting of nearly horizontal bands of substrate material that may be the deformational remnant of intraclasts. There is little water involved in the deposition of lodgement tills and therefore they contain no sorted strata.

ii) The pressure exerted on the substrate by the overriding glacier can, in certain circumstances, drag the till forward as an internally deforming mass. This process was first recognised by Geikie (1863), and the depositional product was named deformation till by Elson (1961). Deformation tills are defined as "... weak rock or unconsolidated sediment that has been detached from its source, the primary sedimentary structures distorted or destroyed, and some foreign material admixed." (Elson, 1989, p.85). Deformation tills consist of remoulded subglacial till that can be incorporated en masse into the overlying ice or bulldozed in front of the ice (e.g., Boulton, 1986). Therefore, they form a continuum with lodgement and melt-out tills, and are here considered to be primary tills.

Finer, impermeable, clast-poor till will be more susceptible to deformation than coarser, permeable, clast-rich till (Boulton and Hindmarsh, 1987; Hart and Boulton, 1991). In the Great Lakes basin, where glaciers have advanced over lacustrine sediments, tills have high silt–clay contents (commonly in excess of 60 percent). Hicock and Dreimanis (1992) identified three layers of a deforming bed, with associated sedimentary structures: an upper ductile layer, a brittle–ductile layer, and a lower brittle layer. These three layers are also referred to as A, B1, B2 (Boulton, 1987), M, Q?, H? (Menzies, 1989) and homogenous, sheared, overturned (Hart and Boulton, 1991) layers. Deformation tills are internally complex, resulting from rapid deposition and each layer may have distinct physical properties. They may contain structures generated by drag of the over-riding ice, such as folds (isoclinal, hook or chevron folds) and boudins (Hart and Boulton, 1991; Hicock and Dreimanis, 1992). They may contain rafts of unlithified substrate, some of which may show evidence of rolling or attenuation (Hicock and Dreimanis, 1992) or they may contain delicate structures formed in pressure shadows protected from the bulk of deformation by the presence of a clast (Hart and Boulton, 1991). Hart (1994) suggests decreasing fabric strength from rigid to deforming bed situations, and between thin deforming beds and thick ones.

iii) In contrast to the active sliding required for the deposition of both lodgement and deformation till, melt-out till "... is deposited by a slow release of glacial debris from ice that is not sliding or deforming internally." (Dreimanis, 1988, p. 45). Melt-out tills form subglacially or supraglacially, and because transport is generally passive, melt-out tills may inherit properties from transport and the melt-out process (Shaw, 1977, 1979, 1982, 1983, 1987; Lawson, 1979; Haldorsen, 1981; Haldorsen and Shaw, 1982; Dreimanis, 1988). During deposition of melt-out tills, large quantities of meltwater are evacuated from the debris-rich basal layers of the ice. Sorted beds are common in melt-out tills, either draping over clasts or truncating against clasts. Some are formed by the scouring of turbulent water in a micro-channel (millimetres thick) beneath a clast as it is let down onto the till surface. Small cavities (millimetres thick) open and close quickly at the base of the ice, into which sorted sediments are deposited (Haldorsen and Shaw, 1982). The sorted sediments may be cross-laminated, plane-bedded or graded (Ashley *et al.*, 1985). Larger cavities (centimetres to tens of centimetres) also exist. Material commonly founders into these cavities from above and the sides producing diapiric structures. The passivity of the melt-out process means that unlithified clasts are preserved in melt-out tills (Shaw, 1982, 1987).

iv) Sublimation tills are restricted to areas such as Antarctica that have long-term extremes of aridity and cold. They are an end-member of supraglacial melt-out till formed by the sublimation of debris-rich ice (Dreimanis, 1988; Shaw, 1988).

Secondary (Allo-) tills

Secondary tills are resedimented primary tills. Some workers suggest that the lack of direct influence of glacial action precludes their inclusion as tills (e.g., Lawson, 1979, 1988), although others (e.g., Dreimanis and Lundqvist, 1984; Dreimanis, 1988) consider this approach too restrictive. The proximity of many of these resedimented deposits to glaciers, and the requirement of a topographic slope is sufficient reason to include resedimented tills as glacigenic. Allo tills are formed by the action of gravity, either by transporting sediment down a slope (the glacier, either subglacially into a cavity, supraglacially or ice-marginally) as glacigenic sediment flows, or by movement through a water column of insufficient depth to cause much sorting (e.g., undermelt diamictons (Gravenor *et al.*, 1984), and subaqueous basal till (Link and Gostin, 1981)).

Hartshorn (1958), Boulton (1968, 1971, 1972), Eyles (1979) and Lawson (1979, 1981, 1988) described glacigenic mass movement deposits (e.g., flow tills, solifluction tills, sediment gravity-flow deposits, ice-slope colluvium, glacigenic mass-flow deposits) that are found in supraglacial or ice-marginal areas, where sediment is concentrated on the surface by ablation and compressive flow, or subglacially within cavities as lee-side tills or cavity fill. Sediment moves down-slope, through a combination of free-fall, collapse, slide, slump or flow processes and commonly produces structureless, internally disturbed, or chaotically structured diamictons.

The most commonly preserved mass-movement diamicton is that produced by sediment gravity flows (Lawson, 1988). These diamictons form a continuum of sediment types depending on water content. With low water contents, cohesive deposits may form in which many of the features of the source material are preserved, whereas at high water contents, shear strength is low and the sediment is supported by flow turbulence, as a turbidity current. Sediment gravity flows can be deposited within standing water, along the grounding line of a tidewater glacier (Powell, 1981, 1984; Molnia, 1983), or in a glaciolacustrine environment (Rust, 1977; Ashley, 1988).

Sediment gravity flows produce a wide range of sedimentary structures that are dependent on the type of flow. The structures found in glacigenic sediment gravity flows are identical to those produced in non-glacial environments, because the process of emplacement is primarily the same. Structures related to sediment movement include boudinage structures, compressional and tensional fractures, load structures in the lower parts of the flow, and a general downslope thickening. Sediment gravity flow deposits may also show crude inverse grading of clast sizes. On a glacier surface this down-slope movement of material is continuous as material is added to slopes by ablation. Flow deposits are commonly buried or partly eroded by subsequent flows, or interstratified with sorted sediments, when flow is into ponds on the ice surface. These processes commonly produce a complex stratigraphy.

Discussion

Till is a variegated sediment and given its diverse genesis (either primary or secondary), it has no single diagnostic characteristic. Grain size in tills is a function of material entrained by the glacier. Some tills are composed mostly of sorted sediments, e.g., Kalix till (Lundqvist, 1977). Some tills are clast-supported and stony, whereas others are clay-rich having few boulders, such as that observed in the deformation tills of southern Ontario (Hicock and Dreimanis, 1992). Some tills are non-stratified, whereas others show either faint lamination (such as the Catfish Creek Till; Dreimanis, 1982) or are stratified (such as the Sveg Till; Shaw, 1979). There may be grain size differences between melt-out and lodgement tills found within an individual section. Less abrasion and greater winnowing by meltwater during melt-out means they are commonly coarser than lodgement tills produced from the same ice. Striated clasts are generally produced by subglacial transport, although Blackwelder (1930) reports randomly distributed striae from clasts in non-glacial debris flows. Kruger (1979) suggested that consistently oriented striations on the upper surfaces of clasts were a characteristic of lodgement tills and are a result of continued abrasion, following deposition. Hicock and Dreimanis (1992) noted the multiple crossing striae on clasts in deformation tills, and suggested this is the result of rotation of the clasts during the deformation process.

Clast fabric is an important characteristic of diamictons. Pioneering work by Holmes (1941) and Harrison (1957) describing clast fabric in tills has been followed by statistical (e.g., Mark, 1974; Woodcock, 1977) and graphical approaches (e.g., Woodcock, 1977; Rappol, 1985; Benn, 1995) designed to characterize clasts fabric within individual depositional environments (e.g., Dowdeswell and Sharp, 1986; Clark and Hansel, 1989; Hicock, 1992; Hart, 1994; Hicock *et al.*, 1996).

Woodcock (1977) developed a descriptive terminology of fabric shape (Figure 18). Two parameters define fabric shape. A strength parameter C, where,

$$C = \ln (S_1/S_2)$$

and which ranges from weak (C less than 1) to strong (C greater than 3). The strength of the fabric increases away from the origin. A shape parameter K, where

$$K = \frac{\ln (S_1/S_2)}{\ln (S_2/S_3)} \quad (\textit{after}\text{ Woodcock, 1977})$$

ranges from cluster (K greater than 1) to girdle (K less than 1). The value K=1 defines the separation of girdle from cluster fabrics.

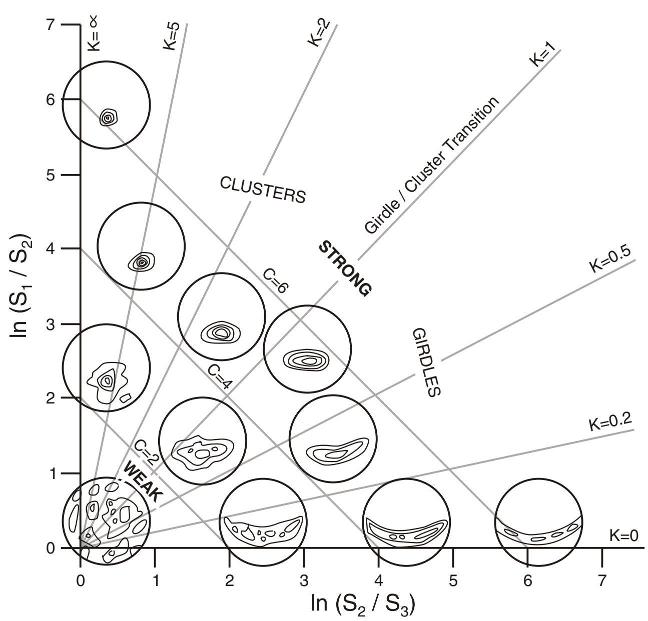


Figure 18. *Description of fabric shape (*after *Woodcock, 1977).*

Data from different genetic environments is presented in Figure 19, as a conventional two-axis plot of S_1 *vs* S_3. The cluster–girdle transition is indicated, and fabric strength increases toward the right. Using the terminology of Woodcock (1977), the diagram shows that melt-out tills have a tendency to have high strength cluster fabrics, although some girdle fabrics are found. Lodgement tills are characterised by moderate to high strength cluster to girdle fabrics.

Deformation tills have a range of fabric shapes from low strength, girdle (S_1=0.53, S_3=0.12), intermediate strength (S_1=0.69, S_3=0.08), to high strength, cluster fabrics (S_1=0.77, S_3=0.06) (Hart, 1994). Glacigenic subaerial sediment gravity flow deposits show commonly moderate to

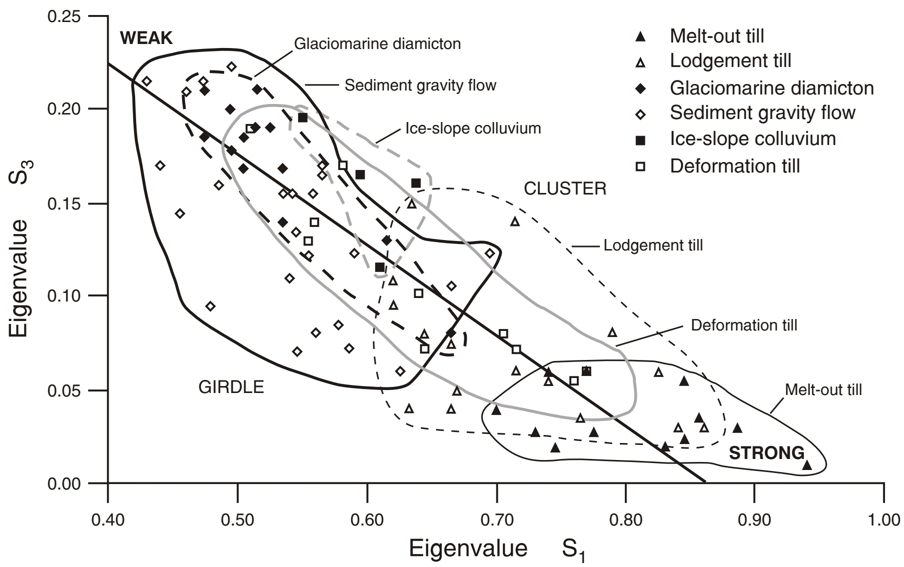


Figure 19. *Plot of S_1 versus S_3 eigenvalues from diamictons deposited in different environments.*

low strength girdle fabrics with some clusters (Lawson, 1979; Dowdeswell and Sharp, 1986; Hart, 1994). Clast fabrics of glacigenic sediment flows resemble those reported from non-glacial debris flows (e.g., Van Loon, 1983; Mills, 1984; Shultz, 1984). Mills (1984) found that clast fabrics from the Mount St. Helens lahar deposits had an upflow dip, and a greater tendency for transverse fabrics and higher dips than glacigenic flows (Lawson, 1979).

Few studies have been completed on the clast fabric of glaciomarine diamictons, formed as either proximal sediment gravity flows, distal underflows, or undermelt diamictons. Undermelt diamictons have generally low strength cluster to girdle fabrics (S_1=0.54, S_3=0.15; Hart and Roberts, 1994), but with a large proportion of clasts dipping at high angles (Dreimanis, 1982; Gravenor *et al.*, 1984; Domack and Lawson, 1985). These high angles result from reorientation of clasts as they drop through the water column.

In both melt-out and lodgement tills, the high frequency of clast interactions at the base of a sliding glacier produces an alignment of elongate clasts, although clast dips may be higher in lodgement tills than in melt-out tills due to settling of clasts in the horizontal plane during melt-out. Clast long axes in melt-out and lodgement tills may be parallel or transverse to the direction of ice movement (cf., Krumbein, 1939; Elson, 1957; Harrison, 1957; Virkkala, 1960; Flint 1961; Boulton, 1970; Mickelson, 1971; Price, 1973; Drake, 1974; Mills, 1977; Lawson, 1979; Shaw, 1982; Rappol, 1985; Dowdeswell and Sharp, 1986; Dreimanis, 1988; Hicock and Dreimanis, 1989). These tills commonly have an S_1 value greater than 0.6 (Figure 19).

Identifying a diamicton as a till can only be accomplished by using multiple criteria that include sedimentological properties, and stratigraphic and regional considerations. The lateral and vertical association with other sediment is commonly a key to assigning diamictons to a glacial or non-glacial environment. Individually, any potential indicator may be unreliable (e.g., grain size, striated clasts, compaction, and stratification). Sedimentary structures and clast fabric, in combination with other parameters, may be used to suggest a genetic origin. Commonly, this is achieved by first eliminating all unlikely processes which allows a resultant most-probable interpretation to be made.

DIAMICTON EXPOSURES–HUMBER RIVER VALLEY

Review

The distribution and general characteristics of diamictons within the Humber River valley were described earlier. Two hundred and seventy-seven diamicton exposures were examined; most of these were small roadside exposures or test pits, from which few sedimentary structures could be seen. Fabric analysis was not attempted in these small pits. Larger exposures were found in scattered aggregate pits, natural river exposures, or were created in backhoe test pits. The diamictons in these exposures were described, clast fabrics recorded, and matrix and pebble contents sampled. Some sections of greater lateral extent, or distinctive character, were described in detail.

Clast Fabric

Clast fabric analysis was conducted on 173 diamicton exposures. Figure 20 presents a clast fabric data plot using the method of Woodcock (1977) and shows that most diamictons have moderate strength, cluster to girdle fabrics. Highly clustered or girdled fabrics (i.e., C > 3) are poorly represented (27 samples). A plot of S_1 (maximum) and S_3 (minimum) eigenvalues is also presented (Figure 21). The

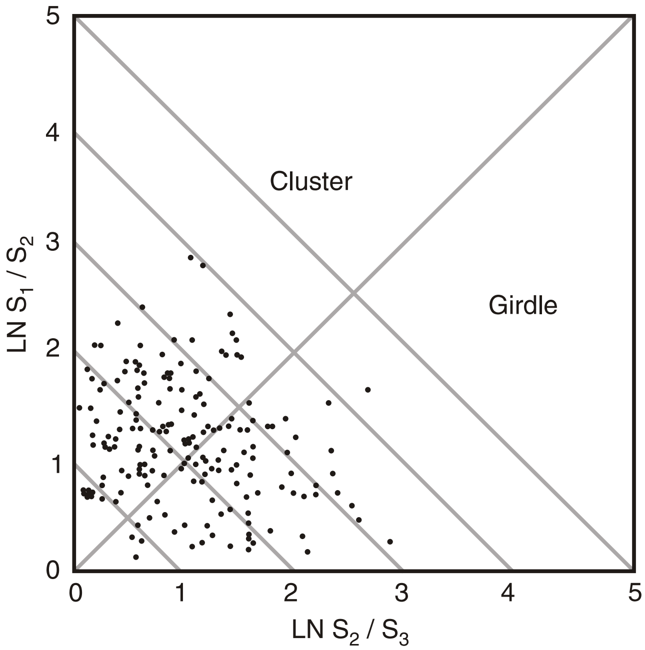


Figure 20. *Shapes of clast fabrics from diamictons within the Humber River basin.*

fabric envelopes generated from the data presented in Figure 19 are superimposed on the Figure 21 diagram and shows that most samples (119) have S_1 values greater than 0.6; of these, 46 are girdle fabrics. Fabrics with S_1 values > 0.8 are clustered, whereas those with an S_1 value < 0.6 are dominantly girdle shapes.

Strongly oriented clast fabrics are not consistent with deposition by sediment gravity flow or glaciomarine processes, and are more compatible with a subglacial depositional environment. This relationship is strengthened by comparison of the preferred trend of clast fabrics with ice-flow patterns derived from independent sources. For those fabrics with an S_1 value > 0.6, 94 of 119 clast fabrics showed similar trends to ice-flow directions. This is discussed more fully in the section on the ice-flow history.

Sedimentary Structures

Most diamictons in the study area are homogenous, and lack obvious sedimentary structures. Of those showing any structural features subhorizontal fissility is the most common. Steeply dipping clastic dykes, and clast pavements are found in rare exposures. Sections at the Pasadena landfill site, near Hinds Lake, and Rocky Brook contain examples of these features.

SECTION DESCRIPTIONS

Diamicton exposures are scattered throughout the Humber River basin, commonly overlying bedrock. Descriptions were noted of sections that were either unusual outcrops or those representative of a group of diamicton exposures with similar characteristics, deposited in a similar depositional environment.

Rocky Brook–An Overconsolidated Basal Till

Rocky Brook, a tributary of the Humber River, has its headwaters in the foothills of the Long Range Mountains north of Deer Lake. The middle reaches of Rocky Brook flow through a gorge cut through Carboniferous sandstone, and the lower 2.5 km is eroded through Quaternary sediment. A small diamicton unit is exposed near the bridge on the Reidville road (Site 92009: Appendix 1).

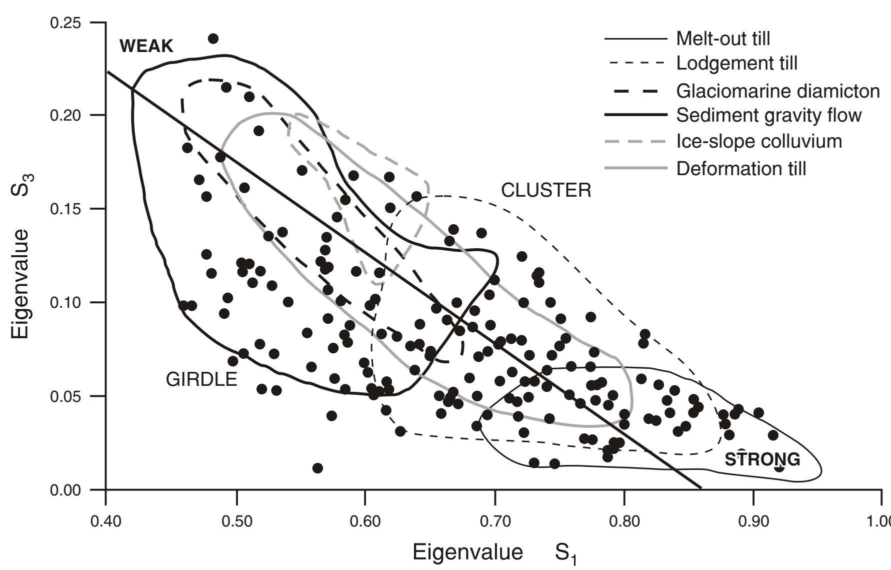


Figure 21. *Plot of S_1 versus S_3 eigenvalues from diamicton fabrics within the Humber River basin.*

Description

The diamicton is exposed on both sides of the river laterally for about 100 m and is best examined when the river is low. The unit extends about 1 m above modal river level, and is of unknown thickness (Figure 22). It is a reddish brown (2.5YR 3/4, moist), compact diamicton. Two samples (sites 924003 and 924004: Appendix 1) have a very poorly sorted (s.d. 2.45ø and 2.14ø), silty–sand matrix, with a mean matrix composition of 63 percent sand, 34 percent silt and 3 percent clay. Clasts are generally granules to cobbles, although rare clasts greater than 50 cm diameter are found. Many of the clasts are striated. One hundred and sixty-seven clasts were identified from the two samples. They are mostly sandstone and siltstone (63 percent) and minor amounts of granite (11 percent), acid volcanic rocks (7 percent), rhyolite (5 percent) and gneiss (3 percent), limestone (2 percent), quartzite–quartz pebbles (4 percent), gabbro (2 percent), quartz pebbles (2 percent), quartz feldspar porphyry (2 percent), and tuff (1 percent) (Figure 22). Two fabrics, one from the north and one from the south side of the river, show strong S_1 (0.72 and 0.70), and low S_3 eigen values (0.04 and 0.08). The preferred trend of both fabrics is 075 (Figure 22). Vanderveer (unpublished data, NLGS database) noted similar results from an exposure on the north bank ($S_1 = 0.74$, $S_3 = 0.05$). The diamicton is very compact, and fractures in sheets parallel with the river bank. A 30-cm-long, single-clast thick, subhorizontal concentration of cobble-shaped clasts was noted about 100 cm below contact with overlying sediment. A thin veneer of reddish brown silt–clay is found in the casts beneath clasts.

The diamicton is overlain above a sharp, planar contact by a 1 to 1.5 m thick, clast-supported boulder gravel. It has a coarse sand to granule gravel matrix, and subrounded to rounded clasts of mixed rock types, up to 1.5 m diameter.

Interpretation

The diamicton is a primary subglacial till. The strongly oriented clast fabrics that have low S_3 eigen values, are typical of subglacial deposition, and plot within the lodgement/melt-out till envelope (Figure 21). The fabric trend is also similar to regional striae directions, showing glacial flow into the Deer Lake basin from the east. Siltstone and sandstone clasts are derived locally from the underlying Rocky Brook Formation, or the adjacent Humber Falls or North Brook formations. Crushing of these rocks produces a reddish brown matrix. Exotic clasts are derived from bedrock located in directions consistent with paleo ice flow. The rhyolite clasts are commonly flow banded, and are likely derived from the Springdale Group, which crops out on The Topsails (Figure 3). Granite (units Oid and Sm) and quartz–feldspar porphyry (Sq) clasts are likely derived from The Topsails (Figure 4). Other rock types such as gabbro, tuff, and acid volcanic rocks also crop out on The Topsails. Limestone clasts may be derived from either the Rocky Brook or North Brook formations both of which have minor limestone constituents (Hyde, 1979). Similarly, the striated clasts suggest a subglacial depositional environment.

The lateral concentration of clasts is a clast pavement. These have been well documented, and are the result of clast collisions due to ploughing on the lodgement surface (Boulton, 1976; Boulton and Paul, 1976; Kruger, 1979; Åmark, 1980; Clark and Hansel, 1989).

The spalling of the till in fractures, parallel with the river banks, are interpreted as structures due to release of pressure. Sladen and Wrigley (1983) consider such fractures to be caused by a reduction in shear strength due to the removal of a supporting load, in this case due to fluvial erosion. Babcock (1977) and Catto (1984) described jointing in Quaternary sediment caused by valley cutting. Similar cases of instability are also associated with cutting exposures through lodgement till during road construction (e.g., Cocksedge, 1983). In Newfoundland, examples of failure due to loss of lateral support have been cited in bedrock along glacially overdeepened fjords (Grant, 1987). An alternative interpretation is that the fractures represent fissility, as described by Muller (1983), Ashley *et al.* (1985) and others. However, fissility is commonly defined by subhorizontal pseudo-beds. As such, the orientation should be similar on both sides of the river, and be aligned approximately normal to the vertical exposure face. The near-vertical trend of the fractures is incompatible with fissility. Fissility should persist throughout the deposit, but the Rocky Brook till becomes less fractured farther into the bank, and this suggests that cracking is ongoing as a result of the loss of lateral support.

The till at Rocky Brook has been previously interpreted as representing ice flow southward down the Humber River valley (Vanderveer and Sparkes, 1982) and was based on the presence of the gabbro clasts, considered to be derived from the Gull Lake intrusive suite, west of White Bay. However, gabbro is also found on The Topsails, within the Rainy Lake complex, Hungry Mountain Complex, Unit Om, and in the Buchans Group (Figure 3). The identification of granite and other rocks compatible with a source on The Topsails suggests westward glacial transport. The previous interpretation is therefore rejected.

The section at Rocky Brook is the only known exposure of this compact diamicton, although Vanderveer and Sparkes (1982) cite unreferenced drill-core data from the Upper Humber River valley to suggest the Rocky Brook till is pre-late Wisconsinan, based on stratigraphic position overlying bedrock. A review of existing data neither supports nor refutes this hypothesis. Pre-late Wisconsinan tills from elsewhere in Atlantic Canada, such as the Halten Till (Stea and Fowler, 1979), McCarron Till (Stea *et al.*, 1985) or Lawrencetown Till (Grant, 1975) were identified on the basis of oxidation, induration or overlying non-glacial sediments containing macro- or micro-fossils dated to pre-late Wisconsinan age. The Rocky Brook till is indurated, but is

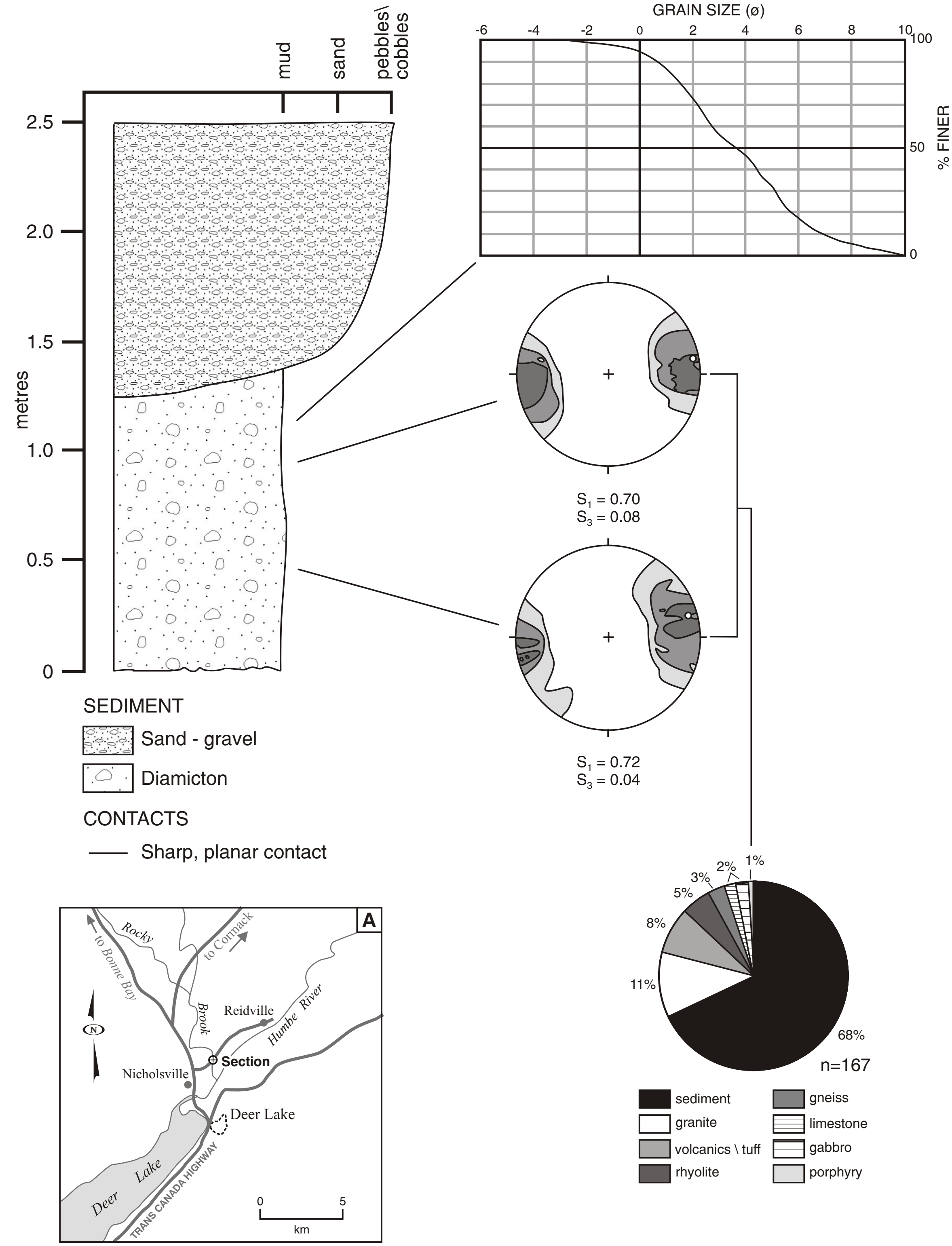


Figure 22. *Stratigraphy of the diamicton exposure at Rocky Brook.*

unoxidised and undated. In the absence of corroborative data the age of this unit remains uncertain.

The boulder gravel overlying the Rocky Brook till is interpreted as a postglacial fluvial deposit, as indicated by the stratigraphic position of the unit, the position of the unit adjacent to the modern stream channel, and its sedimentological characteristics. The section is directly downstream of a confined gorge, at a point where the stream likely lost its competence to transport large clasts. Clasts of similar dimension are observed during periods of low flow in the base of the modern stream channel.

Hinds Lake Dam–Subglacial Melt-Out Till

The area adjacent to the Hinds Lake dam is generally sediment covered (1 to 5 m) having rare bedrock exposures (Bajzak *et al.,* 1977; Whalen and Currie, 1988). It has numerous small borrow pits developed during dam construction. Regional ice-flow indicators show an early flow to the northwest (320°), followed by a later southwest (250°) flow (*see* Section on Ice-flow History, page 103).

Description

A borrow pit (Site 92195: Appendix 1) on the north side of the Hinds Brook valley exposes two sandy diamictons (Figure 23). The basal diamicton has a minimum thickness of 1.5 m. It is matrix supported (40 percent matrix), very dark greyish brown (10YR 3/2, moist) to light brownish grey (10YR 6/2, dry) having a very poorly sorted (s.d. 3.03ø) silty sand (73.7 percent sand, 26.2 percent silt, 0.1 percent clay) matrix and a mean grain size of 1.17ø. Clasts are subrounded to subangular, granules to boulders and are up to 40 cm diameter. Commonly, they are covered by silt on their upper surfaces and have clean lower surfaces. Rock types include gabbro (30 percent), granite (25 percent), basalt (19 percent), porphyry (14 percent), and minor amphibolite (3 percent), rhyolite (3 percent), and acid volcanic (2 percent). The diamicton has a weak girdle fabric (S_1=0.53, S_3=0.05). Distributed throughout the unit are small (8 cm lateral extent and 2 cm thick), roughly subhorizontal lenses composed of moderately to well sorted, fine to medium sand, and containing rare trough crossbedding. Sorted medium to fine sand lenses also occur beneath clasts, the lateral dimensions of which are restricted to clast dimensions.

The lower basal diamicton is overlain by 1.5 m of a very dark brownish grey (10YR 3/2, moist) to pale brown (10YR 6/3, dry) sediment (Figure 23) having a very poorly sorted (s.d. 3.07ø) sand matrix (82.3 percent sand, 17.7 percent silt), and a mean grain size of 0.06ø. Clasts are subangular to subrounded, granules to boulders, up to 60 cm diameter. Clast types are dominated by gabbro (42 percent), granite (26 percent), and porphyritic rhyolite (15 percent), and minor basalt (8 percent), acid volcanic (7 percent) and rhyolite clasts (1 percent). Aphanitic clasts were commonly striated. The unit has a strong, clustered clast fabric (S_1=0.88, S_3=0.03), with a preferred clast orientation towards 238°. Most clasts have low dips (15/25 < 10°). The unit contains numerous convexo–concave and planar lenses of moderately to well-sorted fine to medium sand. The lenses are crudely bedded, although not graded. Some lenses are found on the north to west side of larger clasts, and have a ribbon-shaped plan view. The maximum thickness of these lenses was controlled by the clast diameter, and they pinched out laterally, towards the west. These lenses contained crudely bedded sand, with beds dipping 10 to 15° northward to westward. Other small (less than 2 cm lateral extent by 1 cm thick), subhorizontal, planar lenses of moderately to well sorted, fine to medium sand were found throughout the diamicton. Sorted lenses are found beneath clasts, the dimensions of which are controlled by the clast dimensions.

Interpretation

The upper unit has a strong clast fabric, showing a well oriented, clustered distribution that is typical of subglacial deposition (Lawson, 1979; Dowdeswell and Sharp, 1986; Dreimanis, 1988; Hart, 1994) and is interpreted as a subglacial melt-out till. The preferred trend of clasts is similar to the regional ice flow derived from striations, i.e., southwestward. The low dip of many clasts is also common in melt-out, produced by the settling of clasts in the horizontal plane during melt-out (Lawson, 1979; Shaw, 1983; Ashley *et al.,* 1985; Dreimanis, 1988).

The lower basal diamicton till has a weaker clast fabric, and may have been deposited by debris flow.

The clast rock types are consistently transported from the northeast. The site is underlain by Hinds Brook granite (Figure 4), from which most of the granite clasts are derived. Other granitic clasts originate from the Topsails intrusive suite (units Sp and Sm). The gabbro may be from the Hungry Mountain Complex, and the basalt from the Springdale Group both of which are found in outcrop adjacent to the site. Porphyritic rhyolite is found within the Topsails intrusive suite (Unit Sq) and the Springdale Group. All of these clast types indicate southwestward glacial transport.

The sediment lenses indicate sorting by water, and the crude bedding suggests current flow. This is supported by the low silt–clay content, produced by winnowing of fines. These sedimentary features are uncommon in lodgement tills. Shaw (1982, 1983) and Haldorsen and Shaw (1982) described sorted layers, formed as the result of meltwater flow during the melt-out process. Intra-till sorted lenses are distributed throughout the exposure. Lenses form in subglacial cavities that are abandoned relatively quickly (Haldorsen and Shaw, 1982). Convex–concave sorted lenses are scour features caused by turbulent flow beneath clasts suspended from overlying melting ice (Shaw, 1983). These lenses extend beyond the confines of the overlying clast and are commonly part of laterally continuous beds and are syndepositional features. The presence of a thin layer of silt

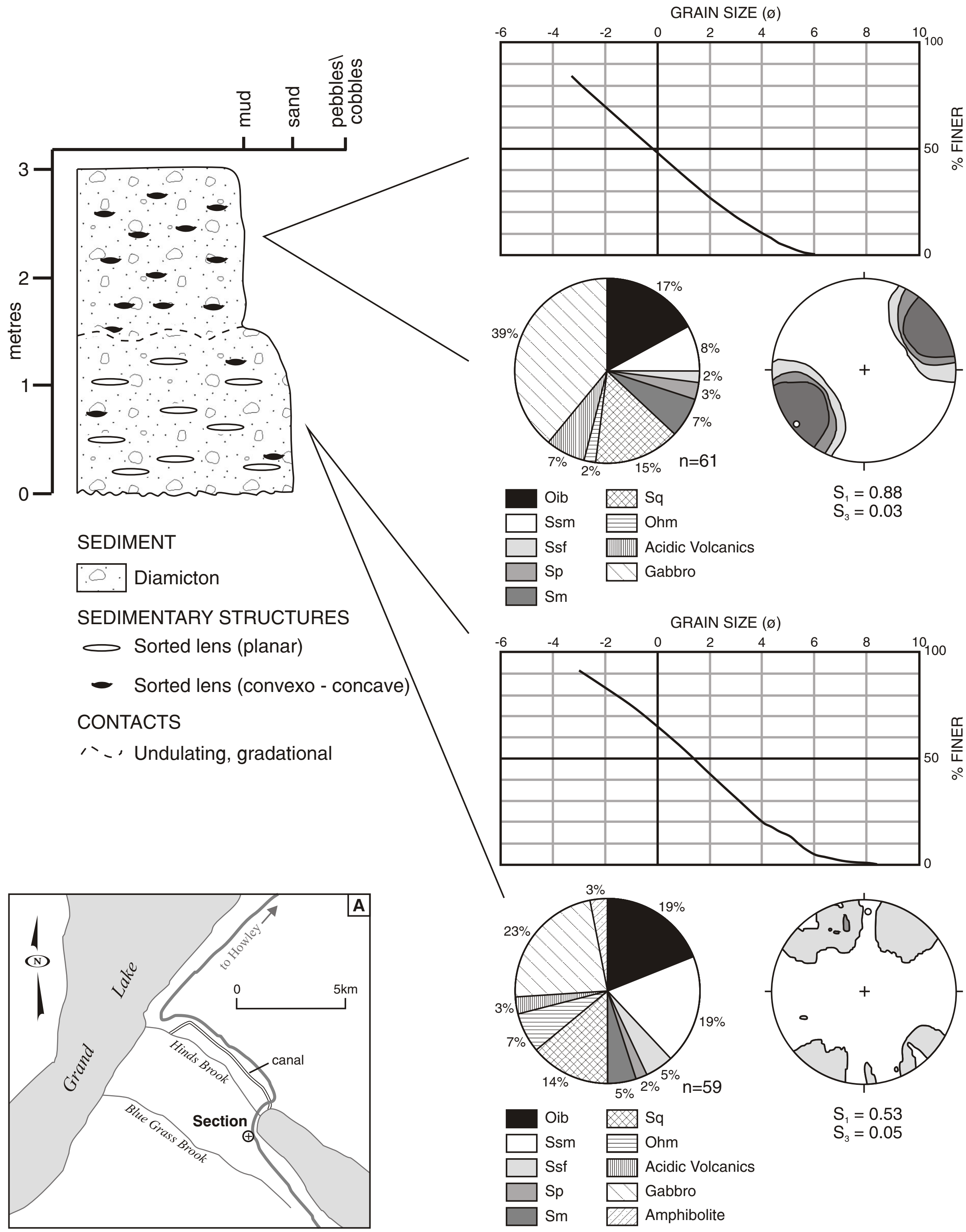


Figure 23. *Stratigraphy of a diamicton exposure near Hinds Lake dam.*

covering the upper surfaces of clasts and a thin sorted, medium to coarse sand to granule layer beneath has not been described by previous authors. It is a common feature of many diamictons in Newfoundland. Several hypotheses exist for the formation of these sorted layers:

i) They represent boulder scours formed in melt-out tills (Shaw, 1983). This is rejected because boulder scour features commonly are traced laterally beyond the confines of the clast. The sorting described above is confined to beneath the clast. Similarly, this model does not explain the upper silt layer.

ii) They are ‘perched clasts' and represent fines protected from erosion (by water?) by the overlying clasts (Fisher, 1989). This is rejected because the features described above are not associated with other features suggestive of erosion. Moreover, it does not explain the silt cap.

iii) They are secondary features resultant from the downward percolation of water during the Holocene. This suggestion is neither accepted nor rejected. No evidence was found, such as organic matter, that would demonstrate downward percolation of water. This may suggest that this process occurred shortly after deposition, where there was little or no surface soil profile. The silty sand matrix common in diamictons suggests downward migration of water may be continuous, and may result in minor turbulence beneath clasts, and deposition above.

Sedimentary features exposed near the Hinds Brook dam are compatible with deposition in a subglacial environment through a process of melt-out. Other potential genetic environments, such as sediment gravity flow are rejected. The strong clast fabric is not compatible with sediment gravity flow deposits, that commonly have weak cluster to girdle fabrics (Figure 19). The low dip angles of clasts are uncommon in sediment gravity flows (cf. Lawson, 1979). Similarly, the presence of numerous sorted lenses does not support a debris flow origin for the sediment.

Goose Arm–Two Diamictons

This site is located along the Goose Arm road north of Old Mans Pond (Site 92147: Appendix 1). It is a 4-m roadside exposure, cut into a hillside and has a slope of about 17°. The exposure consists of a ~2-m diamicton layer dominated by limestone clasts, overlain by a 2 m sandy diamicton containing numerous exotic clasts (Figure 24).

Description

The lower diamicton is compact, greyish brown (2.5Y 5/2, moist) to light grey (10YR 7/2, dry), is very poorly sorted (s.d. 3.45ø), and has a silty sand matrix (61 percent sand, 39 percent silt, and <1 percent clay, with a mean grain size of 0.3ø. The unit is generally structureless, although medium to coarse sand lenses and rare open worked granule gravel lenses are found beneath clasts, within dimensions equal to the diameter of the clast. The upper surfaces of clasts are commonly silt covered, and lower surfaces are clean. The diamicton is clast-rich (~70 percent clasts); clasts are angular to subangular, granules to boulders. Limestone (89 percent), minor quartzite (9 percent) and dolomite (2 percent) dominate the clast rock types. Clast fabric is of moderate strength (S_1=0.68, S_3=0.09) and clustered, with a preferred clast orientation trending northwest.

Overlying a sharp, undulating contact is a loose, dark greyish brown (2.5Y 4/2, moist) to light brownish grey (10YR 6/2, dry) diamicton. The matrix is a very poorly sorted (s.d. 3.42ø), silty sand composed of 74 percent sand, 24 percent silt, 2 percent clay having a mean grain size of 0.07ø. The unit is structureless, except for small lenses of medium to coarse sand and occasional open worked granule gravel lenses found beneath clasts. The texture of the lens coarsens in relation to increasing clast size. Larger lenses are normally graded. Clasts are subangular, granule to boulder sized, with the largest clast about 50 cm diameter. Quartzite (32 percent) and limestone (22 percent) are the major clast rock types, but they also include granite (10 percent), sandstone (10 percent), rhyolite (6 percent), gabbro (4 percent), porphyry (4 percent), shale (4 percent), and siltstone (2 percent). Clast fabric is weak, with a girdle distribution (S_1=0.56; S_3=0.07). Clasts are rarely striated.

Interpretation

The lower diamicton is interpreted as a basal till. This is indicated by the moderately oriented clustered fabric, and the preferred clast orientation parallel to regional ice-flow directions determined from ice striations. The till, which is locally derived, has angular clasts from the underlying Port au Port Group (Knight, 1994), and is similar to the ‘immature' tills described by Croot and Sims (1996).

The overlying diamicton is interpreted as a secondary till, possibly deposited by debris flow. It contains more far-travelled material than the lower diamicton. The granite clasts are derived from the Topsail intrusive suite (units Sp, Sm and Sq). The rhyolites are flow-banded and have their source in the Springdale Group, and gabbro is found within several rock units on The Topsails (Hungry Mountain Complex, Rainy Lake group, Unit Om). Transport from The Topsails has crossed the Deer Lake basin, incorporating red Carboniferous sandstone clasts. Derivation of clasts from The Topsails and the Deer Lake basin is consistent with the regional trend of ice movement, and thus the sediment is glacial in origin. The weak, girdle fabric does not support primary subglacial deposition (cf. Lawson, 1979; Dowdeswell and Sharp, 1986; Dreimanis, 1988), and the sandy matrix (75 percent sand) suggests winnowing of silt and clay. Winnowing is also suggested by the sorted layers beneath larger clasts, and may have been post-depositional

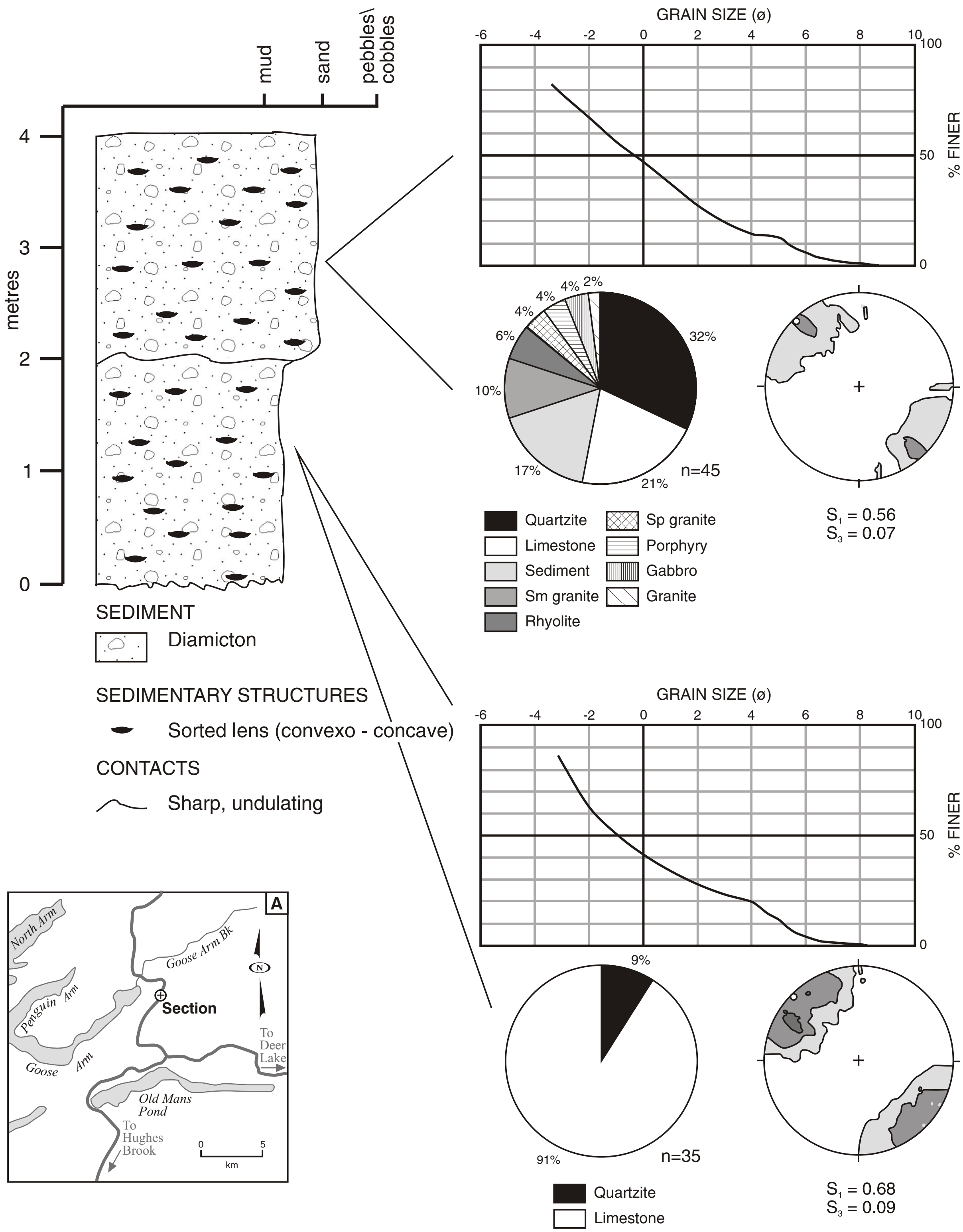


Figure 24. *Stratigraphy of a roadside exposure near Goose Arm.*

or syn-depositional. The unit is located on a hillside, and thus downslope movement under the influence of gravity is likely.

Pasadena Dump–Complex Stratigraphy

On the west side of the road to Northern Harbour along the west side of the South Brook valley, 100 m north of the gate to the Pasadena dump is an abandoned gravel pit (Plate 9). The pit face is oriented north–south, and is about 8 m high, including the lower colluviated 2 m. The upper surface of the pit face is about 65 m asl. The face exposes three diamicton units (Site 91094: Appendix 1) (Figure 25).

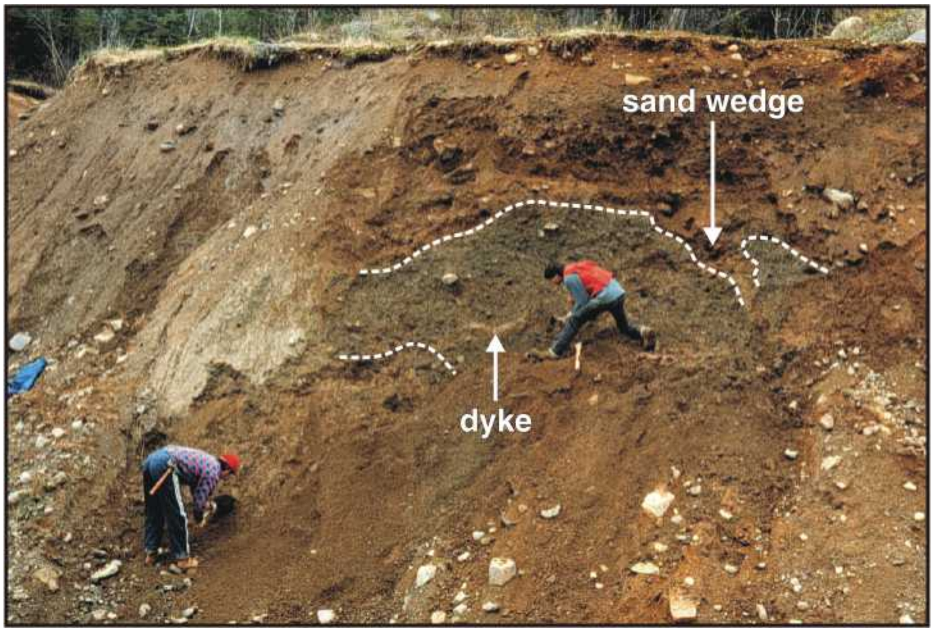


Plate 9. *A general view of the Pasadena dump exposure.*

Description

The lower unit is at least 2 m thick. It is a dark brownish grey (10YR 4/2, moist) to pink (7.5YR 7/4, dry), matrix-supported (~ 60 percent matrix) diamicton. It has an extremely poorly sorted (s.d. 4.24ø), silty sand matrix having a mean matrix composition of 66 percent sand, 25 percent silt and 9 percent clay (n=2). The unit is compact and structureless. Clasts are subrounded to subangular, generally granule to cobble size and the largest clast size is 30 cm in diameter. Clast surfaces commonly are striated. Two clast fabrics were completed on the diamicton from similar stratigraphic positions, separated laterally by 10 m. Both are strongly oriented and weakly clustered (Table 12), with the direction of preferred orientation toward the northwest. Sandstone (48 percent), siltstone (26 percent), and arkose (15 percent) dominated clast rock types. Minor amounts of schist (5 percent), quartz feldspar porphyry (3 percent), and conglomerate (3 percent) are also present.

Overlying an undulating, gradational contact extending 60 cm vertically is a middle, 1.5 m-thick diamicton unit, which is olive grey (5Y 4/2, moist) to light brownish grey (2.5Y 6/2, dry), compact, and matrix-supported (~60 percent matrix) (Figure 25). It has an extremely poorly sorted (s.d. 4.08ø), silty sand matrix, composed of 57 percent sand, 31 percent silt and 12 percent clay. Clasts are subangular to subrounded, mostly granule to cobble size, have rare boulders up to 50 cm diameter and are commonly striated. Two fabrics about 6 m apart, at the same stratigraphic position within the middle of the unit, are strongly oriented and clustered (Table 12). Preferred clast trends are northwest and southeast. Clast types are mostly sandstone (50 percent), siltstone (26 percent) and arkose (16 percent), as well as minor granite (3 percent), quartz pebbles (3 percent), gabbro (1 percent), and quartz feldspar porphyry (1 percent).

The upper olive grey diamicton unit contains three irregular shaped sheets of sand extending downward through the unit (Plate 10). The uppermost sheet is a thin (0.2 to 1.5 cm), contorted unit and is composed mostly of moderately sorted fine sand. Thicker parts of the layer contain interbeds of fine and very fine sand oriented parallel to the sheet margins. The sheet is about 1.5 m long, and dips at about 5 to 50° toward the south. The middle sheet is about 150 cm long and subparallel to the upper sheet, separated by 5 to 20 cm of olive-grey diamicton. This middle sheet is 1.0 to 3.5 cm thick, contorted, and composed of poorly sorted medium to coarse sand with occasional granule to pebble clasts. Interbeds of moderately sorted sand were noted. The lower sheet is the least extensive, being about 50 cm long, with similar internal structure to the upper sheet. All three sheets were traced into the exposure for greater than 20 cm.

The three sheets extend downward from a 10 to 15 cm thick, laterally discontinuous, north–south trending wedge of sandy gravel exposed along the contact between the olive grey diamicton and an overlying reddish brown diamicton. The sandy gravel has a poorly sorted fine to medium sand matrix, subrounded to rounded, granule to pebble clasts of mixed rock types. The sandy gravel bed is generally structureless, although small, convexo–planar lenses of well-sorted coarse sand to open-worked granule to pebble gravel were noted beneath some clasts. The northern extension of the sandy gravel penetrates the olive grey diamicton as a wedge-shaped feature, extending downward about 1 m (Plate 11). The wedge varies in inclination from 5 to 80 , with steeper angles at depth. Upper and lower contacts of the wedge are sharp.

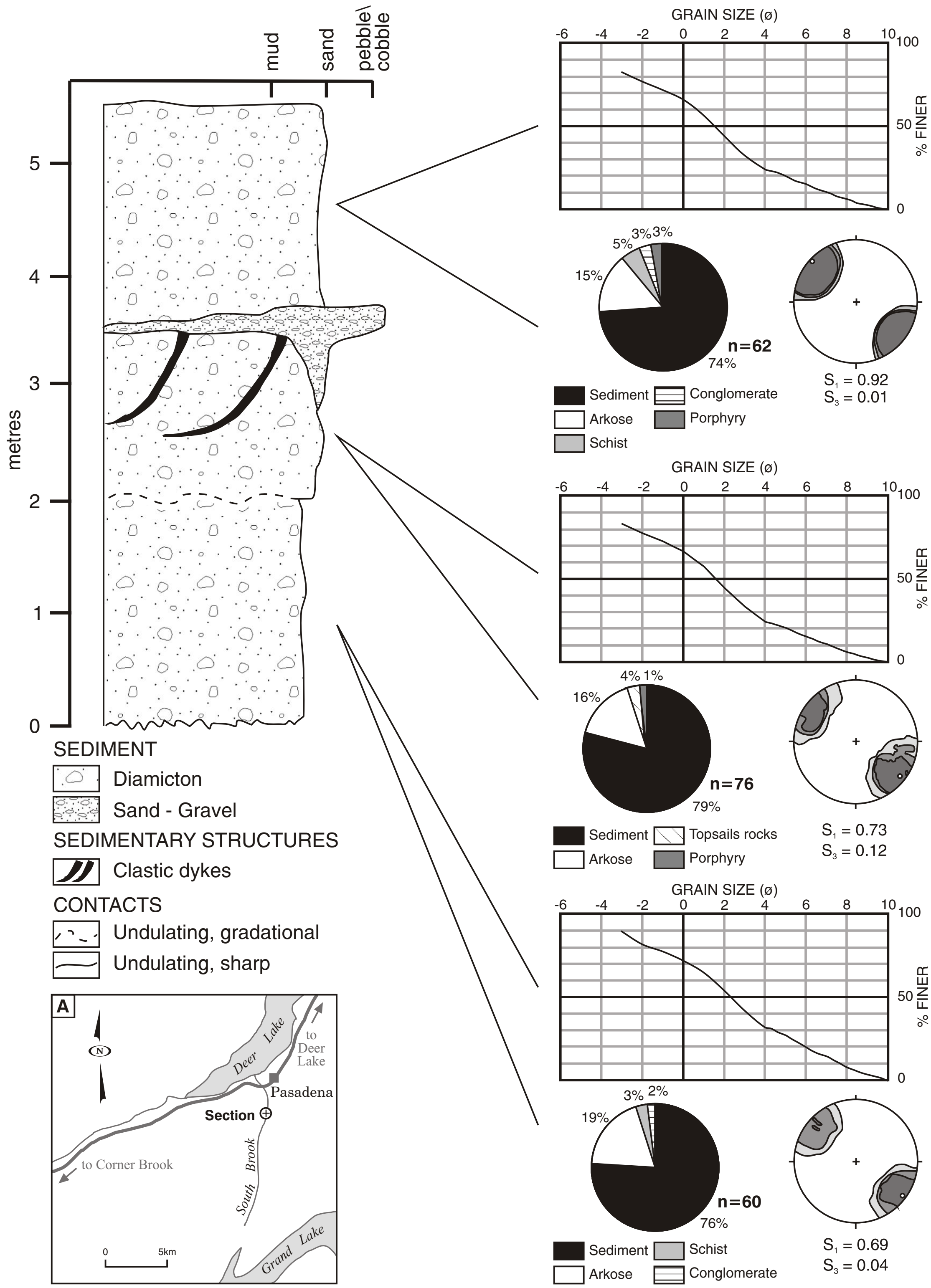


Figure 25. *Stratigraphy of a diamicton exposure near Pasadena dump.*

Table 12. Statistics of clast fabrics from the Pasadena dump site section

Unit		S_1	S_3	Trend (°)	Plunge (°)
Lower	1.	0.79	0.1	303	14.2
	2.	0.92	0	308	2.3
Middle	1.	0.73	0.12	126	7.6
	2.	0.72	0.13	316	2.3
Upper	1.	0.75	0.1	84	4.2
	2.	0.74	0.1	310	6.5
	3.	0.69	0	129	1.9

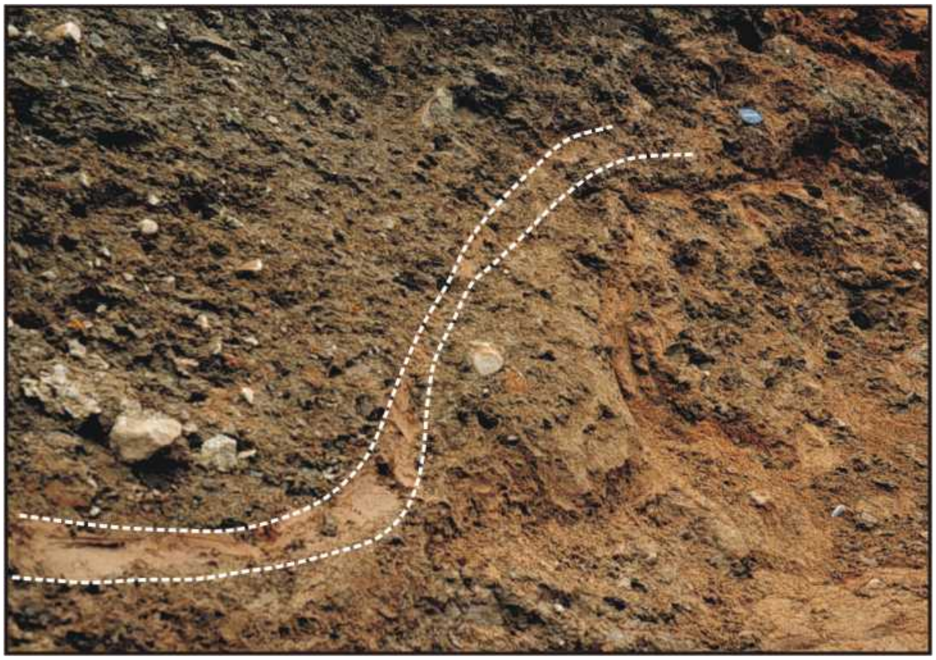


Plate 10. *A clastic dyke within the middle till unit at the Pasadena dump exposure.*

The upper olive grey diamicton and the adjacent sandy gravel unit are overlain by a 2 m-thick compact, matrix-rich (60 percent) diamicton, along a generally sharp, undulating contact (Figure 25). The diamicton is dark reddish brown (5YR 4/2, moist) to light brown (7.5YR 6/4, dry). It has an extremely poorly sorted (s.d. 4.05ø), silty sand matrix composed of 59 percent sand, 30 percent silt, and 11 percent clay. Clasts, which are commonly striated, are subangular to subrounded and mostly granules to cobbles and the occasional boulders. Three fabrics were determined in the unit, spread laterally over about 10 m at the same stratigraphic position. Two of the fabrics are strongly oriented and clustered (Table 12); preferred trends are 129° and 084°, respectively. A third fabric shows a strongly oriented, girdle distribution having a preferred clast orientation of 309°. Clast rock types are mostly sandstone (50 percent), siltstone (27 percent) and arkose (19 percent) and minor schist (3 percent) and conglomerate (1 percent).

Interpretation

The lower, dark brownish grey diamicton is interpreted as a primary, subglacial till. This interpretation is supported by the striated clasts, and very strong and clustered clast fabrics. Preferred clast fabric trend is southeast–northwest, parallel to the regional striation patterns (see section on Ice-flow History). Sandstone, arkosic sandstone, siltstone and conglomerate all occur within the North Brook Formation of the Deer Lake Group that underlie the Pasadena area, or from the Anguille Group that crops out to the east. Quartz-feldspar porphyry is found in the Topsails intrusive suite (Unit Sq), which crops out west of Hinds Lake, and the schist clasts are likely derived from the Mount Musgrave Group that outcrops on the hills, west of the South Brook valley. The unit is homogenous. No sorted lenses, boudins or other sedimentary structures were noted that would suggest either the presence of significant quantities of water during the sedimentation process, or that the till was resedimented.

The middle, olive grey diamicton also is interpreted as a primary, subglacial till. The gradational contact suggests depositional continuity with the underlying unit, and the presence of striated clasts and strongly oriented and clustered fabrics support a subglacial origin. Preferred clast trend is southeast–northwest, parallel to the northwestward regional striation pattern. The distribution of clast types suggest a northwestward ice flow direction. Sandstone, siltstone and arkose clasts are found in the North Brook Formation that underlies the dump site area. Granites are fine grained and pink, similar to rock types found in the Topsails intrusive suite (units Sg and Sm). This suite also contains quartz-feldspar porphyry (Unit Sq). All of these units outcrop southwest of Hinds Lake. The gabbro clast is also likely from The Topsails, exposed in outcrop up-ice of the dumpsite in the Rainy Lake complex (Figure 4).

The sand sheets are interpreted as clastic dykes. They have been described from many areas (Larsen and Mangerud, 1992; Dreimanis and Rappol, 1997), including Nova Scotia (Mörner, 1973; Dredge and Grant, 1987). They include wedges injected downward from a basal till layer, either composed of till (e.g., Dreimanis, 1969, 1992; Åmark, 1986), or sorted sediment (e.g., Kruger, 1938; Åmark, 1986; Larsen and Mangerud, 1992; Dreimanis and Rappol, 1997). The clastic dykes at the Pasadena dumpsite penetrate till,

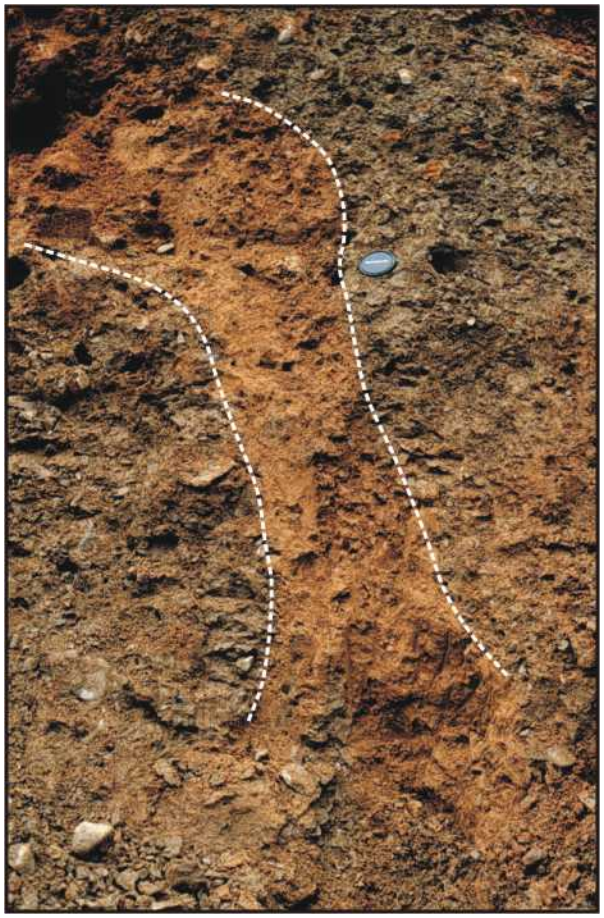


Plate 11. *A wedge of sand and gravel between the middle and upper till units at the Pasadena dump exposure.*

and therefore, are subglacially derived. The dykes comprise sorted and laminated sediments, with laminae parallel to the dyke walls. They extend downward from a unit of sandy gravel that must also have been subglacially deposited. The parallelism of laminae to dyke walls suggests incremental dyke development, (Larsen and Mangerud, 1992). A wide crack that developed rapidly would likely produce horizontal lamination. Therefore, the propagation of the cracks is by hydraulic splitting generated by high hydraulic pressures. Such pressures are required to explain the near horizontal angle of the dyke at depth that would otherwise have closed due to the weight of the overlying sediment.

These dykes are similar to glaciotectonic tension fractures produced by glacier drag (Hicock and Dreimanis, 1985), and from the Northern Superior till, north of Lake Superior, Ontario (Hicock and Dreimanis, 1992). The thermal condition of the till into which the dykes were injected is difficult to determine, however, Dreimanis (1969, 1992) suggested that till wedges require a frozen substrate to facilitate cracking. Humlum (1978) argued that a frozen substrate would promote basal freezing and consequently glacial flow by internal deformation, rather than basal sliding required to promote glacial drag. At the Pasadena dumpsite, the presence of contorted segments within the dykes, and the irregular course of the dykes suggests the surrounding diamicton was unfrozen during dyke formation.

The upper, reddish brown diamicton is interpreted as a primary subglacial till. It contains striated clasts, and well oriented, girdle to cluster, clast fabrics, that generally have a preferred trend parallel to regional striation patterns. Clast rock types are of local provenance, derived from the underlying North Brook Formation.

Discussion

All three diamicton units have broadly similar characteristics. Grain size distributions for the lower, middle and upper tills are similar, and show values of 34 percent, 43 percent and 42 percent silt–clay, respectively. Clast fabrics are strong and clustered. The lowest diamicton possesses the strongest fabrics, ($S_1 \leq 0.75$ and $S_3 > 0.05$). The middle and upper diamictons have similar fabrics (S_1=0.72 to 0.75, and S_3=0.09 to 0.13). Generally, clast dips are low within each diamicton unit. The preferred trend of clasts within each of the diamicton units is similar for 6 of 7 fabrics (~130 or 310°), and similar to regional paleo ice-flow directions determined from bedrock striations. The regional trend shows ice moving west to northwest from a source on The Topsails, crossing the South Brook valley. Clast–provenance data is consistent with this interpretation because mostly granite or quartz–feldspar porphyry exotic clasts were found in outcrop on The Topsails.

All three diamictons are interpreted as basal tills. There is no evidence for an ice-free period, suggesting that glacial cover was continuous during deposition. The contact between the lower and middle tills is gradational. There were no non-glacial or oxidized sediments noted between the middle and upper tills. Sandy gravel exposed along the contact is formed subglacially, as indicated by the clastic dykes that propagate downward from it into the middle till. The upper till represents a different depositional phase, being in sharp contact with the middle diamicton and devoid of sedimentary structures indicative of glaciotectonism. Preferred clast orientation and clast provenance show that the upper diamicton was deposited by ice from the same source as that which deposited the middle diamicton.

The clastic dykes found within the middle diamicton have not been reported from elsewhere in Newfoundland. Within the Humber River basin, similar sorted and laminated dykes are found north of Deer Lake (Site 91163: Appendix 1). At this site a lower reddish brown to light brown (5YR 5/3, moist to 7.5YR 6/4, dry), very poorly sorted (s.d. 3.58ø), silty sand (49 percent sand, 36 percent silt, 15 percent clay) diamicton possessing a strong fabric (S_1=0.69; S_3=0.07) is overlain along an undulating, sharp and dipping contact by a reddish brown (2.5YR 4/4, moist) to light red (2.5YR 6/6, dry), very poorly sorted (s.d. 3.91ø), silty sand (48 percent sand, 34 percent silt, 18 percent clay) diamicton having a strong fabric (S_1=0.74; S_3=0.07). The lower unit contains steeply dipping (~ 38°), thin (2 to 3 mm wide) dykes, containing reddish brown silt–clay, commonly with very thin (< 1 mm) medium sand partings.

Evidence for glaciotectonism also comes from an exposure (Site 91104: Appendix 1) about 400 m upslope and 20 m higher than the Pasadena dumpsite section. This section, termed Pasadena dump II, exposes about 2 m of diamicton, overlain by 1 to 2.5 m interbedded fine sand, medium sand and diamicton, further overlain by 0.8 m diamicton (Figure 26). The upper and lower diamictons are similar, although the upper is slightly coarser (43 percent silt–clay) compared to the lower (32 percent silt–clay). Both are extremely poorly sorted, structureless, brown to dark brown (10YR 5/3, moist to 7.5YR 4/4, moist) diamictons. Clast fabric data from each unit are strong and clustered. Clasts are striated, and provenance is dominantly local; the exotic clasts found in the diamicton have been derived from The Topsails (quartz–feldspar porphyry from the Topsails intrusive suite, and rhyolite from the Springdale Group). The upper diamicton contains fewer exotic clasts than the lower diamicton. The two diamictons differ in the degree of compaction (lower more compact than upper), and the direction of preferred clast orientation (262° and 253° in the lower diamicton (similar to the Pasadena dump tills), and 297° in the upper diamicton).

Both diamictons are interpreted as primary subglacial tills, based on structure, striated clasts, and strong clast fabrics with preferred clast trend parallel to regional flow direction. The upper till is correlative with the upper till at the Pasadena dumpsite. The lower till at Pasadena dump II and the lowest diamicton at the Pasadena dumpsite are also correlative.

The middle unit separating the two diamictons units is a very compact, crudely horizontally stratified pebbly sand, with diamicton lenses. The unit includes interbeds of moderately sorted medium to very fine sand. The diamicton forms a series of plano-convex lenses about 10 cm wide and 3 cm high. The lenses appear internally structureless, have a fine sand matrix and granule to pebble sized clasts.

Numerous thin (0.1 cm), discontinuous, medium to fine sand strata or inclusions, up to about 30 cm long occur throughout the middle unit. In places, these inclusions truncate existing bedding. Twenty-five of these strata were excavated across the section, and their plunge and orientation measured (Plate 12). They commonly are steeply dipping (mean 49°) toward 180°. They are also found in the lower till, although not in the upper one. In places, the middle unit contains clastic dykes, commonly deformed and faulted (Plate 13). The dykes consist of medium to coarse sand, and dip steeply (20 to 60°) westward. Larger dykes contain laminated silt and sand, with laminations parallel to the wedge margins. One of these extends into the lower diamicton unit.

A clast of unconsolidated sediment was found at the contact between the middle unit and the upper till. It was about 100 cm wide by 75 cm high, and was composed of loose, mica-rich, interbedded, moderately to well sorted, medium and fine sand. The bedding dips about 10° eastward, and is more distinct towards the top of the clast, whereas lower down it is more structureless (Plate 14).

This middle unit is interpreted as having been deposited in a small ice marginal or subglacial pond. The size of the pond may be estimated from the extent of the unit, which appears to pinch out to the south, and from the lack of similar exposures elsewhere. The pond is ice proximal, suggested by the lack of silt–clay, and the diamicton lenses (e.g., Lawson, 1982; Ashley *et al.*, 1985; Eyles and Eyles, 1992). These are interpreted as having been deposited by debris flow. The geographic location of the deposit on a valley side-wall suggests an ice marginal origin.

The pond has been overridden by ice. This is indicated by the clastic dykes and fine sand inclusions or plates contained within the pond sediments, interpreted as glaciotectonic features. Dreimanis (1992), Hicock and Dreimanis (1985, 1992), Broster *et al*. (1979), Broster (1991) and Derbyshire and Jones (1980) described similar features. These glaciotectonic features are interpreted as tension gashes or cracks, produced in fine-grained sediments as a result of glacial overriding. Whether the upper diamicton represents a readvance of regional significance, or reflects a local reactivation of the ice front cannot be determined from this section alone.

To be preserved, a loose, sand intraclast must have been frozen during emplacement, and remained frozen during glacial overriding and the development of the glaciotectonic features. The upper part of the sand intraclast was eroded during overriding (Plate 14). The sorted and laminated clastic dykes suggest the surrounding sediment was not frozen. Therefore, it is possible that the sand intraclast was introduced to the pond as a frozen block derived from the subglacial drainage system. The sharp contacts with surrounding sediment, and dip angle of beds within the intraclast unrelated to bedding within the pond sediment support this conclusion. Glacial overriding occurred subsequently, leading to the development of tension cracking and clastic dykes. Boulton (1970, 1979), Hooke (1970), Shaw (1982), Harris and Bothamley (1984), Menzies (1990a, b), and Hicock and Dreimanis (1992) described sand intraclasts within deformed or glaciotectonised sediment.

Pynn's Brook Valley–Evidence for Glacial Readvance (?)

This exposure was within a backhoe test pit (Site 91239: NTS 12H04; UTM reference 461920E 5436600N) in the Pynn's Brook valley, north of Pasadena (Plate 15). The surface is flat (elevation 161 m asl), sloping gently toward Pynn's Brook to the east.

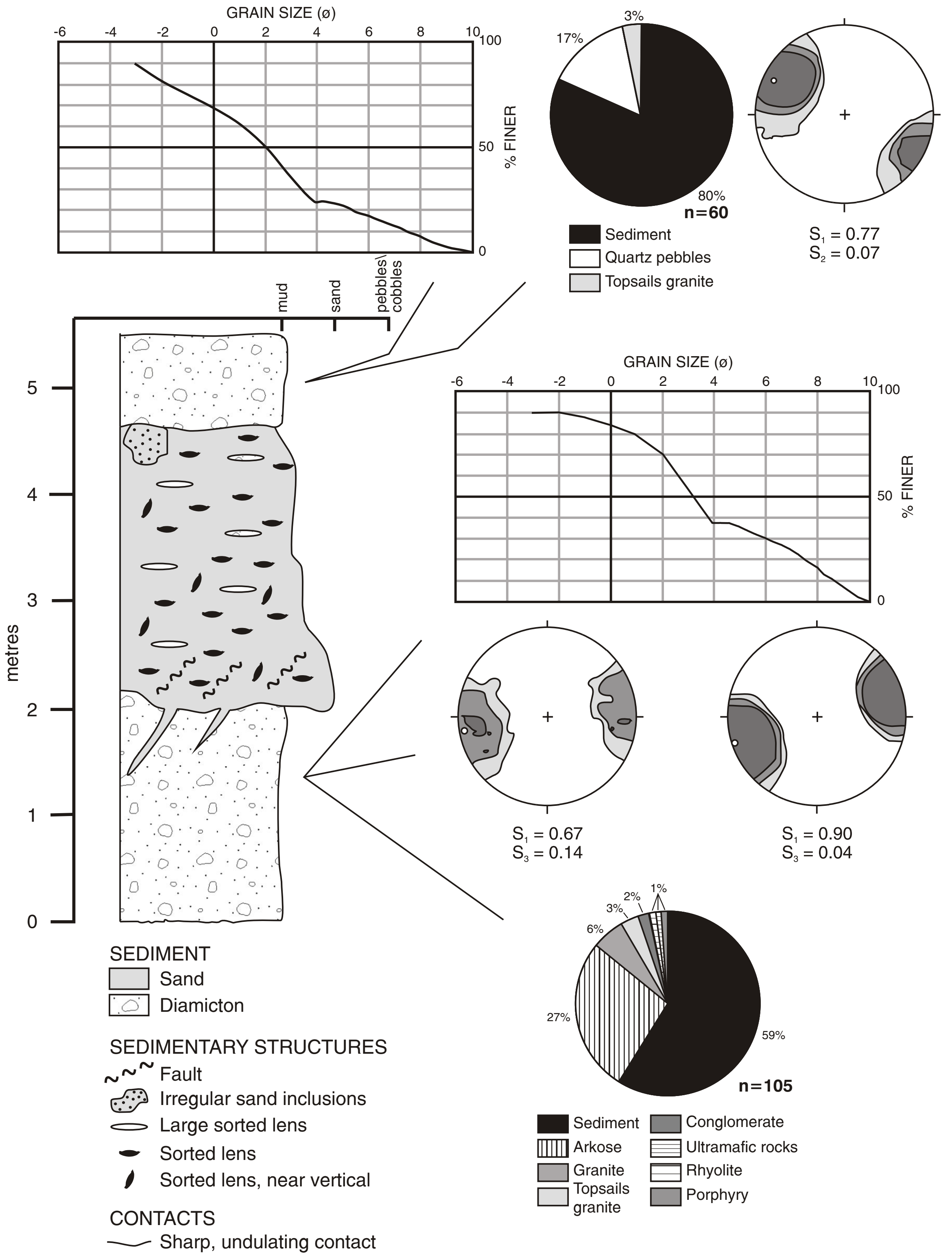


Figure 26. *Stratigraphy of the Pasadena dump II exposure.*

Plate 12. *Sand strata found within compact proglacial lacustrine sands in the Pasadena dump II exposure.*

Description

The section exposed about 3 m of sediment with a lateral extent of about 2 m (Figure 27). At the base was at least 1 m of light yellowish brown (10YR 6/4, moist) to pale brown (10YR 6/3, dry) silty sand, moderately sorted (s.d. 1.4ø), comprising 70 percent sand, 28 percent silt and 2 percent clay, and having a mean grain size of 3.9ø; the unit is graded normal. The unit contains rare pebble to cobble-sized clasts (<5 percent) and is crudely stratified and contains planar, subhorizontal, moderately sorted medium and coarse sand lenses, up to 15 cm long and 3 cm thick. Individual sand beds commonly are convoluted, showing simple folds in beds up to 20 cm in length. No current flow structures were noted.

This is overlain by a 1 cm thick, moderate to well sorted, structureless, very fine sand to silt unit, above a sharp, wavy contact (Plate 16) which is in turn overlain by a 20-cm thick, cohesionless and structureless pebbly sand along a sharp, irregular contact. This 20-cm-thick unit has a fine to medium sand matrix, and subangular to subrounded granule to pebble clasts and contains a steeply dipping (about 60°), 3 to 4 cm long lens extending from the underlying sand bed about 15 cm into the pebbly sand. The lens contains a structureless, moderately sorted, fine to medium sand.

Overlying a gradational contact is 2 m of diamicton, which is generally homogenous, apart from a 15-cm thick, structureless, subhorizontal, laterally continuous, planar, fine to medium sand lens, toward the base of the unit. The sand lens has sharp upper and lower contacts. The diamicton is reddish (5YR 4/3, moist) to light brown (7.5YR 6/4, dry), very poorly sorted (s.d. 4.58ø), sand–silt–clay matrix (28 percent sand, 37 percent silt and 35 percent clay) having a mean grain size of 4.48ø. Clasts are subangular to subrounded, commonly striated, and randomly distributed through the unit. The upper surfaces of clasts commonly are covered by a thin (< 1 mm) veneer of silt to fine sand, whereas lower surfaces are clean. Clast rock types include sandstone (56 percent), schist (28 percent), siltstone (10 percent), gneiss (2 percent), quartz pebble (2 percent) and granite (2 percent). Clast fabric is moderately strong and clustered (S_1=0.67, S_3=0.13) having a preferred clast orientation of 018° and a modal plunge of 17°.

Interpretation

The lower sands are a subaqueous deposit, as indicated by their texture and sorting, and the low proportions of coarse clasts. Waning flow or decreasing sediment supply explains the normally graded bed. Lack of current is suggested by the high (30 percent) silt–clay content in the matrix. The lenses suggest discrete current flow events, but whether these represent turbidity currents or channeled flows is not clear. The disturbance of the beds are a result of dewatering produced by loading. This is indicated by the overlying wavy contact with the pebbly sand, and the steeply dipping sand lens (flame structure?) in the overlying pebbly sand bed. Loading is provided by the overlying diamicton. Development of dewatering structures suggest rapid emplacement of the overlying diamicton onto a saturated lower unit, inducing the gravitational instability to induce upward movement of the underlying fines. The sands were deposited in a proglacial or subglacial environment. Sand beds have been described from both these environments (e.g., Rust and Romanelli, 1975; Cheel and Rust, 1982; Ashley, 1988; Miall, 1992).

The overlying diamicton is a basal till as indicated by the lack of sedimentary structures, striated clasts and moderate strength clustered fabric. Clast provenance is local. The site is underlain by metasediments of the Mount Musgrave Group, and sandstone, siltstone and quartz pebbles are likely derived from the adjacent Anguille Group. The gneiss and granite clasts are exotics. The granite is porphyritic and may have its source in The Topsails (Unit Ssya?). Gneiss is also found on The Topsails, and in the Long Range Mountains. It could not be determined whether the glacier that deposited the diamicton had its source in the Long Range Mountain or The Topsails. Clast fabric is moderately strong and clustered, and falls within the basal till envelope on Figure 19. Preferred clast orientation is north–south, perpendi-

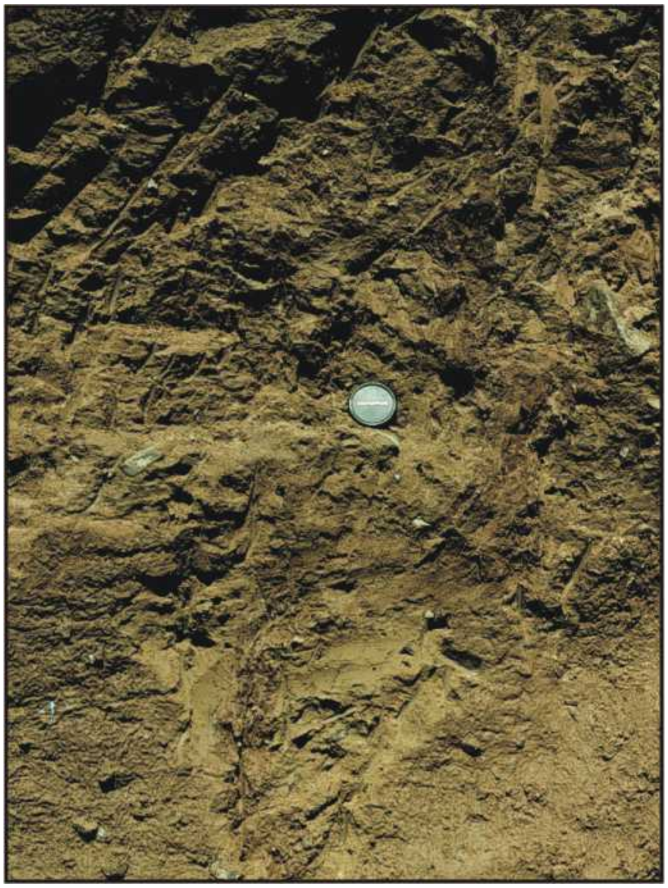


Plate 13. *Small, steeply dipping subparallel normal faults within a compact sand bed in the Pasadena dump II exposure.*

cular to the trend of the valley in which the section was exposed.

A possible alternative interpretation is deposition as a sediment gravity flow. The diamicton is generally structureless, loads the underlying sediment (cf. Leeder, 1982), and contains steeply dipping clasts (8 of 25 > 30°), similar to characteristics noted by Lawson (1979). However, clast fabric in sediment gravity flow deposits is commonly weak (S_1 < 0.6) and girdled (Dowdeswell *et al.*, 1985; Dowdeswell and Sharp, 1986; Hart and Roberts, 1994). The diamicton exhibits a moderately strong, clustered fabric, is ungraded, shows no evidence for a traction carpet in the lower parts of the bed, and has a flat surface morphology. All of these factors are not typical of a sediment gravity flow deposit.

The presence of a basal till overlying sand suggests a reactivation of the ice margin, if the sand was deposited proglacially. The source and extent of this reactivation is unclear. The Pynn's Brook exposure may be related to a poorly exposed section located further to the east in the Pynn's Brook valley (Site 92172: Appendix 1), that shows a thin diamicton bed overlying interbedded sand and diamicton. Southward-oriented striae noted in the Glide Lake area have a similar trend to the clast fabric orientation at Pynn's Brook (*see* Section on Ice-flow History, page 103). The Pasadena dump II section exposes a similar stratigraphy. The surface diamicton at that section is locally derived, generally structureless, and has a similar clast fabric strength (although different preferred orientation) to the till at Pynn's Brook. These two tills are tentatively correlated with each other.

The anomalously high clay content in the diamicton (35 percent) is likely the result of overridding of a fine-grained source by ice; the source of these fines could not be determined. Other diamictons in the area have clay contents of less than 10 percent. The clast component in the diamicton (mostly sandstone and schist) suggests the clay is not from a bedrock source. Similarly, the diamicton is located well above muds found in the Deer Lake basin to the west, and no fine grained sediments were located in the Pynn's Brook valley from which the clay may have been derived.

Plate 14. *An unconsolidated clast along the contact between the compact sand bed and the upper diamicton unit in the Pasadena dump II exposure.*

The Pynn's Brook exposure may represent depositional evidence for a local reactivation of the ice front during deglaciation. The presence of till overlying sand and gravel is unusual in Newfoundland, but the limited lateral extent of the exposure means that an alternative hypothesis of all sediment being deposited in a subglacial environment could not be dismissed. The suggestion that the

Plate 15. *General view of the Pynn's Brook valley exposure.*

exposure shows evidence for a readvance therefore remains tentative. Excavation of more extensive exposures in the Pynn's Brook valley is required to test the validity of alternative hypotheses.

DIAMICTON IN THE HUMBER RIVER BASIN: DISCUSSION

The diamicton exposed at Rocky Brook is a primary subglacial till deposited by lodgement; this exposure is anomalous. The high compaction and fracturing of the diamicton due to pressure release is not found elsewhere across the Humber River basin. Other characteristics of deposition by lodgement occur rarely in diamicton exposures. A section near Hinds Lake dam (Site 92089: Appendix 1) shows an intra-till clast pavement, similar to the one noted at Rocky Brook. The pavement is composed of flat-lying clasts up to 15 cm diameter forming a well-defined 10 cm-thick layer (Plate 17), separating a lower dark greyish brown (10YR 4/2, moist), diamicton with a strong clast fabric (S_1=0.79; S_3=0.03) from an upper dark brown (7.5YR 4/2, moist) diamicton, with a moderate clast fabric (S_1=0.60, S_3=0.07), and containing numerous irregular-shaped, subhorizontal sand lenses.

Other small diamicton exposures that showed characteristics found in lodgement tills, including fissility, having strong fabrics with preferred orientation parallel to regional ice flow, and striae on clasts parallel to clast long axis were found at six sites (Table 13) (Sites 91048, 91104, 91215, 92093, 92207, 92212, and 93133: Appendix 1).

The diamictons exposed at the Hinds Brook dam site are primary basal tills deposited by melt-out. Diamicton exposures showing similar characteristics (lenses of sorted sediment, possibly with strong fabrics, and clast orientation parallel to regional ice flow) are found at 39 sites across the study area (Sites 91062, 91064, 91193, 91222, 91231, 91242, 92018, 92024, 92033, 92035, 93036, 92039, 92089, 92096, 92111, 92112, 92154, 92178, 92196, 92202, 92205, 92207, 92209, 92213, 92214, 92215, 92216, 92217, 92224, 93011, 93019, 93020, 93174, 93093, 93097, 93103, 93105, 93122, and 93134: Appendix 1).

The diamictons exposed near Goose Arm are a basal till of local origin, overlain by a secondary till deposited by sediment gravity flow. Diamictons with characteristics consistent with subglacial deposition and short transport (e.g., dominated by local clasts, possibly with strong clast fabric with a preferred clast orientation parallel to regional ice flow) were found at 39 sites across the study area (Sites 91006, 91027, 91033, 91036, 91044, 91053, 91060, 91061, 91063, 91078, 91080, 91081, 91098, 91120, 91141, 91142, 91187, 91191, 91198, 91205, 92059, 92060, 92061, 92062, 92063, 92145, 92152, 92199, 92200, 92203, 92208, 92210, 92220, 92221, 93068, 93095, 93102, 93136, 93137, 93140: Appendix 1).

Diamictons with similar properties to those found in the upper till at Goose Arm are commonly found along, or at the base of, steep slopes. Some had preferred clast trends and dips parallel to local slopes. Examples are found in the Lower Humber River (Sites 91150 and 91184), Old Mans Pond (Site 91208), South Brook (Sites 91090 and 91097), Wild Cove (Site 91222) and Grand Lake (Site 93015) valleys, and along the foothills of Birchy Ridge (Sites 93054 and 93104) and the Long Range Mountains (Site 93044).

Other exposures showed thin (5 to 30 cm) diamicton beds interbedded with sand and gravel. The diamictons are interpreted as sediment gravity flow deposits. Examples of these are found in the South Brook (Site 91087) and Wild Cove (Site 91220) valleys, and near Cloudy Pond (Site 92124).

The diamictons at the Pasadena dumpsite and Pasadena dump II site are primary tills deposited in a subglacial depositional environment. The lowest till shows evidence of lodgement, the middle till contains glaciotectonic sedimentary structures, and the upper till is a primary basal till. Other exposures showing evidence for glaciotectonism are rare across the basin. High proportions of silt and clay (commonly greater than 60 percent; Hicock and Dreimanis, 1992) are required to produce the necessary porewater pres-

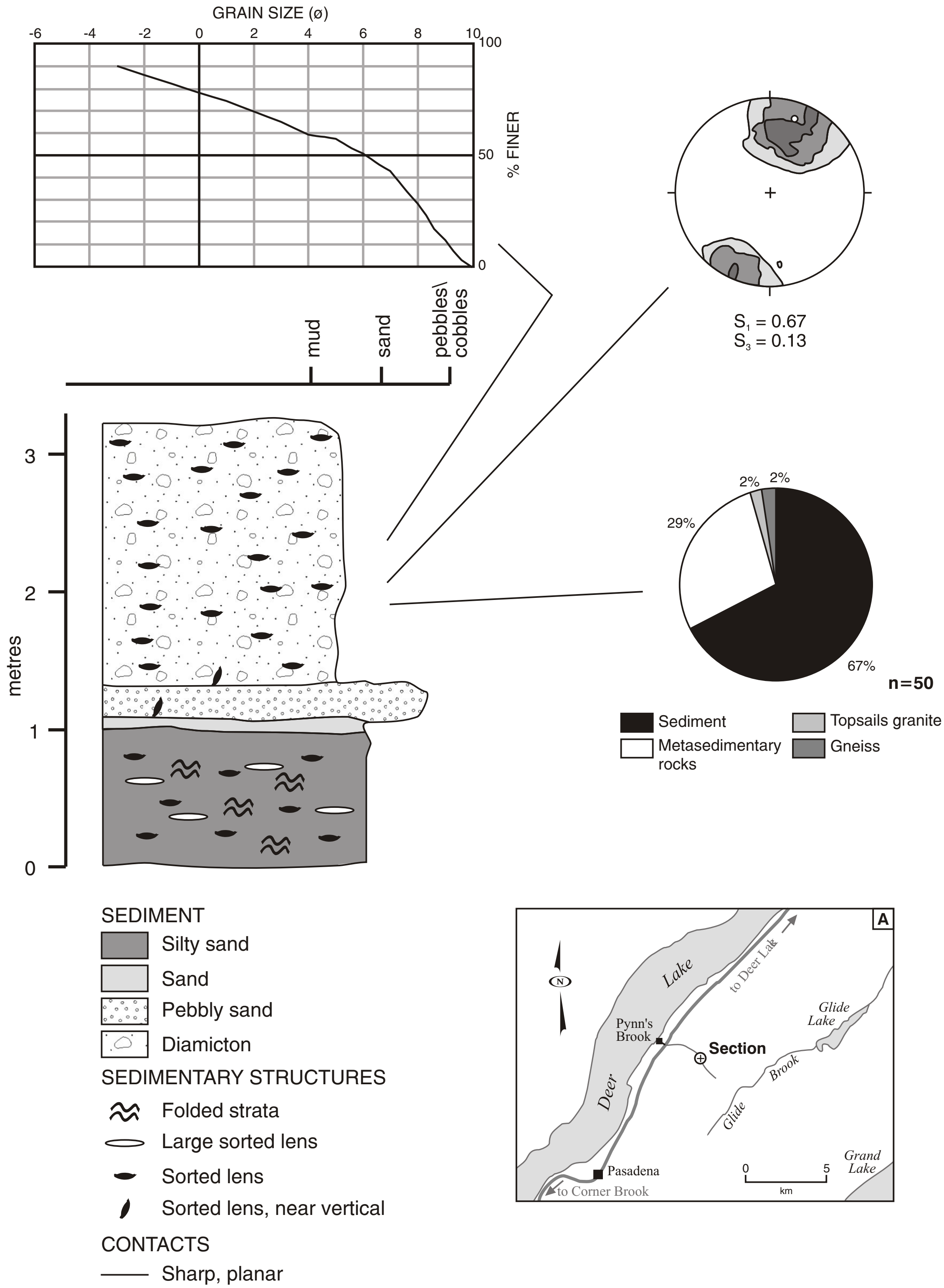


Figure 27. *Stratigraphy of an exposure in the Pynn’s Brook valley.*

Plate 16. *Sand overlain by structureless very fine sand to silt, pebbly sand and diamicton in the Pynn's Brook valley exposure.*

sures required to induce deformation. These high proportions of fines are rarely found in tills in the Humber River valley.

The diamicton exposed at Pynn's Brook is correlated with the upper Pasadena dump II till, and together these provide tentative evidence for a local readvance. Areas showing evidence of reactivation of an ice-margin are rare in the Humber River valley. Apart from those at Pynn's Brook and Pasadena dump, a diamicton near Birchy Lake in the Upper Humber River valley (Site 93088: Appendix 1) shows 60 cm diamicton overlying a minimum of 15 cm sand–silt. The lower 5 to 10 cm of the diamicton contains rip-up intraclasts composed of sand–silt, derived from the underlying unit. This section is poorly exposed, but may represent evidence of local readvance.

Diamictons are generally either primary tills (ortho-till) or their derivatives (allo-till). Diamictons commonly have a silt sand matrix, and contain exotic clasts whose source in bedrock is found up-ice. Striated clasts are common.

Of the 249 diamicton exposures examined, over half (136) were small, with little vertical or lateral continuity from which genesis could be determined. There was little between-site variability in grain-size and clast provenance for closely spaced exposures. Clast provenance (i.e., erratics transported parallel to flow direction indicated by striae) or clast fabric (with a preferred orientation parallel to glacier flow) suggested glacial transport. Whether these diamictons were primary tills or their secondary derivatives (e.g., sediment gravity flows) could not be ascertained. The preservation of structures within diamictons is dependent on numerous factors, including local slope angle and aspect, sediment texture and water content, post-depositional weathering and modification by frost processes. Therefore, it is reasonable that diamictons showing features of primary glacial deposition should be juxtaposed with their secondary derivatives. It may therefore be assumed that diamictons, unless shown otherwise, were initially of glacial origin.

Plate 17. *An intra-till clast pavement within a diamicton unit from an exposure near Hinds Lake dam.*

Single unit diamicton exposures were described from 136 locations throughout the study area. Of these, 32 exhibited well-oriented clast fabrics (S1 > 0.6) (Sites 91014, 91050, 91176, 91177, 91225, 91227, 91233, 91236, 92007, 92020, 92021, 92056, 92058, 92097, 92163, 92171, 92191, 92193, 92197, 92206, 92218, 92222,

Table 13. Characteristics of lodgement tills found across the Humber River basin.

Site	Compaction	Fabric	S_1	S_3	Parallel to Ice Flow	Fissile	Overlain by	Other Structures	Striated Clasts
92093	high	strong	0.78	0.06	yes	yes	sand-gravel	none	yes
92207	mod	mod	0.64	0.09	yes	yes	diamicton	none	no
92212	mod	strong	0.84	0.04	no	yes	surface	none	yes
93133	high	mod	0.62	0.08	yes	yes	surface	none	yes. // to flow
91048	high	mod	0.64	0.08	yes	yes	surface	none	yes
91215	high	mod	0.68	0.10	yes	no	sand-gravel	none	yes. // to flow

93028, 93035, 93045, 93067, 93077, 93080, 93087, 93101, 93127, 93128, and 93147: Appendix 1). These sediments may indicate subglacial deposition. A further 23 exposures showed weak clast fabrics (Sites 91040, 91066, 91223, 91241, 92178, 92201, 92211, 92219, 93082, 93125, 93126, 93129, 93130, 93131, 93132, 93135, 93138, 93139, 93144, 93145, 93146, 93150, and 93151: Appendix 1). These exposures may indicate secondary mobilization. Fabrics at the remaining 81 sites were not recorded.

DEGLACIAL AND POSTGLACIAL SEDIMENTS AND STRATIGRAPHY

The Humber River basin contains two areas of continuous, or near-continuous, cover of subaqueous sediments; these are in the Lower Humber River valley, and along the east shore of Grand Lake. In addition, valleys contain coarse grained sand and gravel deposits that are commonly exposed in small borrow pits or roadside ditches. These exposures were described and discussed in the section on Surficial Geology, page 20.

The Lower Humber River valley between the community of Deer Lake and the Humber River gorge has a thick (up to 100 m near Steady Brook) cover of sand, silt and clay near. Generally, this area is at, or below, the marine limit of 50 m (Brookes, 1974). Exposures are poor, apart from that at the head of Deer Lake, where sections at Rocky Brook and North Brook are described in detail (Figure 28); a small, fossil-bearing section in the Humber River gorge is also discussed. Adjacent to the modern coast, three exposures are described that provide the range of depositional environments encountered. A section at Dawe's Pit is located near the mouth of the Humber River at the head of the Humber Arm. A series of small exposures along the south side of the Wild Cove valley, a parabolic valley 4 km north of the Humber River, shows sediment containing marine shells within fan-shaped features. In the Hughes Brook valley that enters the Humber Arm, north of Wild Cove, a section in a gravel pit provides evidence for deltaic sedimentation.

A more-or-less continuous series of exposures are found for at least 20 km along the eastern shore of Grand Lake (Figure 28). These are well above the proposed marine limit. Most are obscured by slumping, although descriptions are presented from exposures near Grindstone Point, Little Pond Brook and Alder Brook.

DEER LAKE AREA

Rocky Brook

This is a 19.5-m thick cut-bank river exposure (Site 93034: Appendix 1) in the side of a terrace (surface elevation 28 m asl), located 400 m downstream of the bridge on the Reidville road (Plate 18). A diamicton exposure located near the bridge was discussed in the section, Glacial Sediments and Stratigraphy, page 40.

The following is divided into 12 units (Figure 29), but the bottom 2.0 to 2.5 m directly above the river is obscured.

Unit 1: Laminated Silt–Clay Rhythmites and Clay

Description

This basal unit extends 14 cm above the base of the exposed face. It has rhythmically laminated silt, clayey silt and clay, in a 3-couplet system. The couplets are between 0.3 to 0.5 cm thick, consisting of a thicker horizontally stratified grey silt grading upward into a thinner reddish brown clayey silt. The couplets are commonly overlain by 0.8 to 1.2 cm clay, containing 2 to 5 silt laminae. These laminae are planar horizontal, about 0.5 mm thick and have sharp upper and lower contacts. All laminae are laterally continuous for more than 2 m. A rhythmic sequence consisting of eight couplets was noted. Upper and lower contacts of individual rhythmites are sharp and individual rhythmites showed no current structures.

The exception to the rhythmic bedding of the unit is a 1 mm-thick, structureless, roughly horizontal, silt lamina

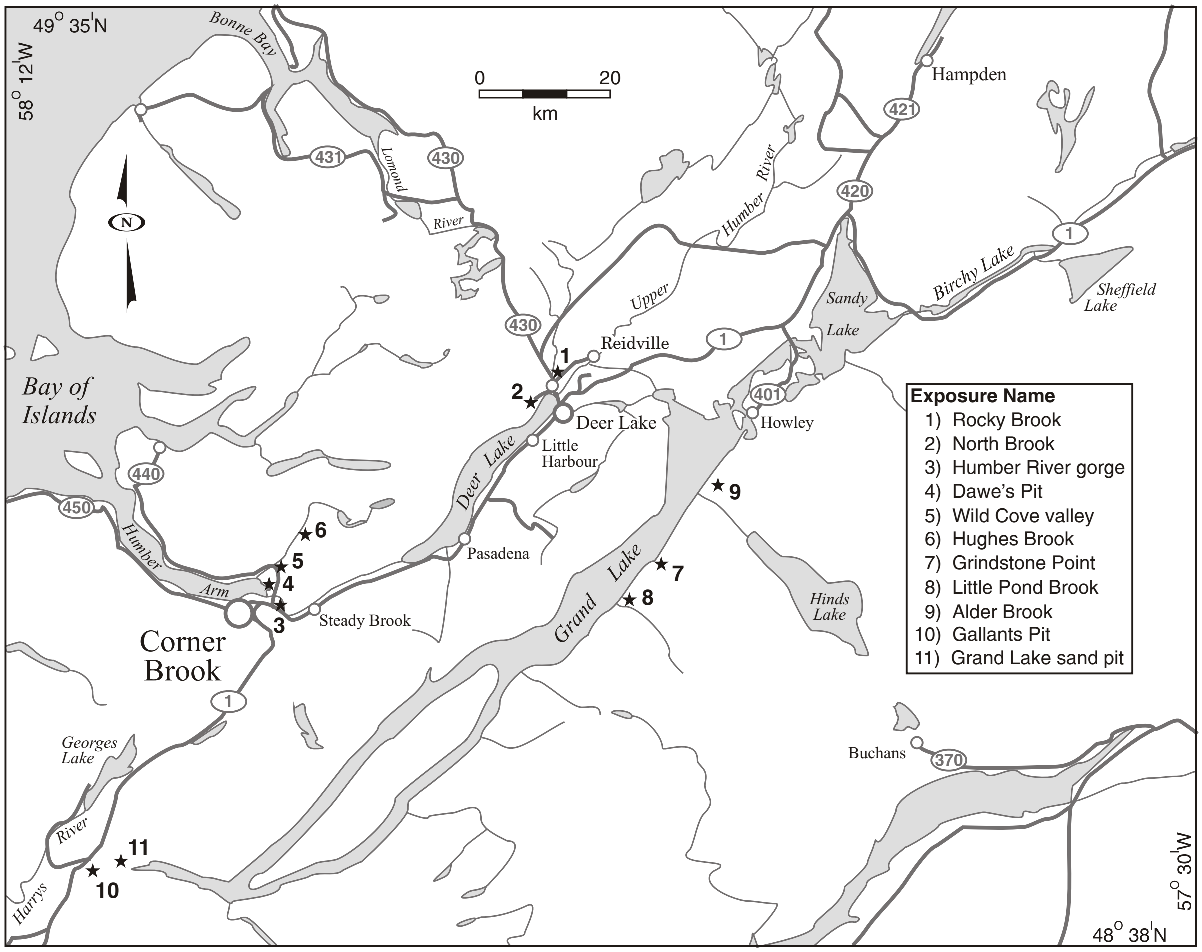


Figure 28. *Location of exposures described in text.*

found 8 cm from the base that pinches and swells across the exposure.

Interpretation

This unit was deposited in standing water where the silt–clay couplets may either represent cyclical seasonal deposition (i.e., varves) or rapid deposition by underflow and overflow–interflow, but having no temporal implication.

The silt–clay beds found in the Rocky Brook section are normally graded containing rare clay–silt interbeds; no lebenspurren were found in any of the couplets. Thus, the couplets likely represent deposition from a combination of current flow and suspension settling. Similar sediments have been well described from modern lacustrine environments (e.g., Gilbert, 1975, Gilbert and Shaw, 1981; Ashley, 1988), and marine (e.g., Miall, 1983; Mackiewicz *et al.*, 1984; Powell, 1990), and applied to the interpretation of ancient fine-grained sediments (e.g., Ashley, 1975; Catto *et al.*, 1981; Catto, 1987; Liverman, 1991). Deposition by suspension settling is suggested by the lack of current-flow structures, and graded bedding. Within individual beds, the relative proportion of silt and clay is similar, suggesting that the sediment was transported to the site at the same time, probably by turbidity currents (Ashley *et al.*, 1985). The overlying clay beds are interpreted to have been deposited by suspension settling from overflow or interflow, except for the interbedded thin silts that likely result from underflows. Gravenor and Coyle (1985) and earlier, Shaw and Archer (1978) reported similar structures. The sharp lower contacts show depositional hiatuses and/or erosion.

Unit 2: Interbedded Draped Sand Ripples and Silt–Clay

Description

This unit is 11 cm thick, and comprises flat–lying beds laterally continuous for greater than 2 m. The base of this unit is marked by a ripple bed. This single bed contains asymmetric, erosional stoss (Ashley, 1975) ripples, with a wavelength (l) of 15 cm, amplitude (H) of 1 cm, and ripple index (R.I.) of 15, composed of planar cross-stratified, very fine sand. The ripples indicate flow toward 150°. The rip-

Plate 18. *View of part of the Rocky Brook exposure.*

ples are draped by 0.1 cm clay, over a sharp contact. Two other ripple beds are found within this unit, at 20 and 23 cm elevation above the base of the exposed section; the upper ripple bed marking the top of the unit. Both are similar to the lowest ripples in dimensions (λ=14 cm, H=1, R.I.=14; l=13 cm, H=1.5, R.I.=9), and interpreted flow direction (150°). The lower ripples are draped by three silt–clayey silt–clay couplets, similar to those of Unit 1. The higher ripples are draped by a 0.5 cm clay layer. Between the lowest and middle draped ripple beds are two, normally graded beds of fine sand to silt, 0.6 and 2.3 cm thick, separated by a sharp, horizontal contact. They are overlain by 1.5 cm, normally graded, fine to very fine sand, 0.2 cm clay, and 1.0 cm normally graded very fine sand to silt. A 1.2-cm-thick, fine to very fine sand bed with 1 mm, discontinuous, planar, horizontal clay laminae separates the middle and upper draped ripple beds.

Interpretation

Erosional stoss ripples indicate unidirectional flow (in this case toward 150°), where ripple migration dominates over deposition (*see* Jopling and Walker, 1968; Ashley, 1975). The draped laminations observed here were formed from deposition by suspension settling as the current flow waned or ceased (*see* Allen, 1984). This association of ripples and draped laminations is interpreted as the result of density underflow (cf. Jopling and Walker, 1968; Ashley, 1972; Gustavson *et al.*, 1975; Shaw, 1975; Clemmensen and Houmark-Nielsen, 1981; Allen, 1993). Planar beds are interpreted as having been deposited from underflow–interflows, as in Unit 1.

Unit 3: Interbedded Silt–Clay Rhythmites and Clay

Description

This unit is 33 cm thick. The base of the unit is defined by a clay drape over the upper ripple bed in Unit 2. A 1-cm-thick, silt–clayey silt–clay couplet lies above a sharp contact. There is a gradational contact between silt and clayey silt. A sharp upper contact separates the clayey silt from an overlying 0.7-cm-thick clay stratum containing three 0.5 mm-thick silt layers, each with sharp upper contacts. This is overlain by 1 cm-thick bed of well sorted, structureless fine sand, overlying a sharp contact and is itself overlain by a rhythmically bedded sequence, of seven silt–clayey silt–clay couplets, overlain by a planar bedded, structureless fine sand–silt bed. This is overlain by two rhythmically beds composed of three and four silt–clayey silt–clay couplets separated by a 6 cm-thick, normally graded, fine to very fine sand bed that contains discontinuous clay rip-up intraclasts (Plate 19). The four couplet bed is overlain by structureless fine sand to silt, grading upwards into clay containing thin (<0.2 mm) planar silt laminae. The top of the unit is a 1.8-cm-thick rhythmite bed containing three silt–clay couplets.

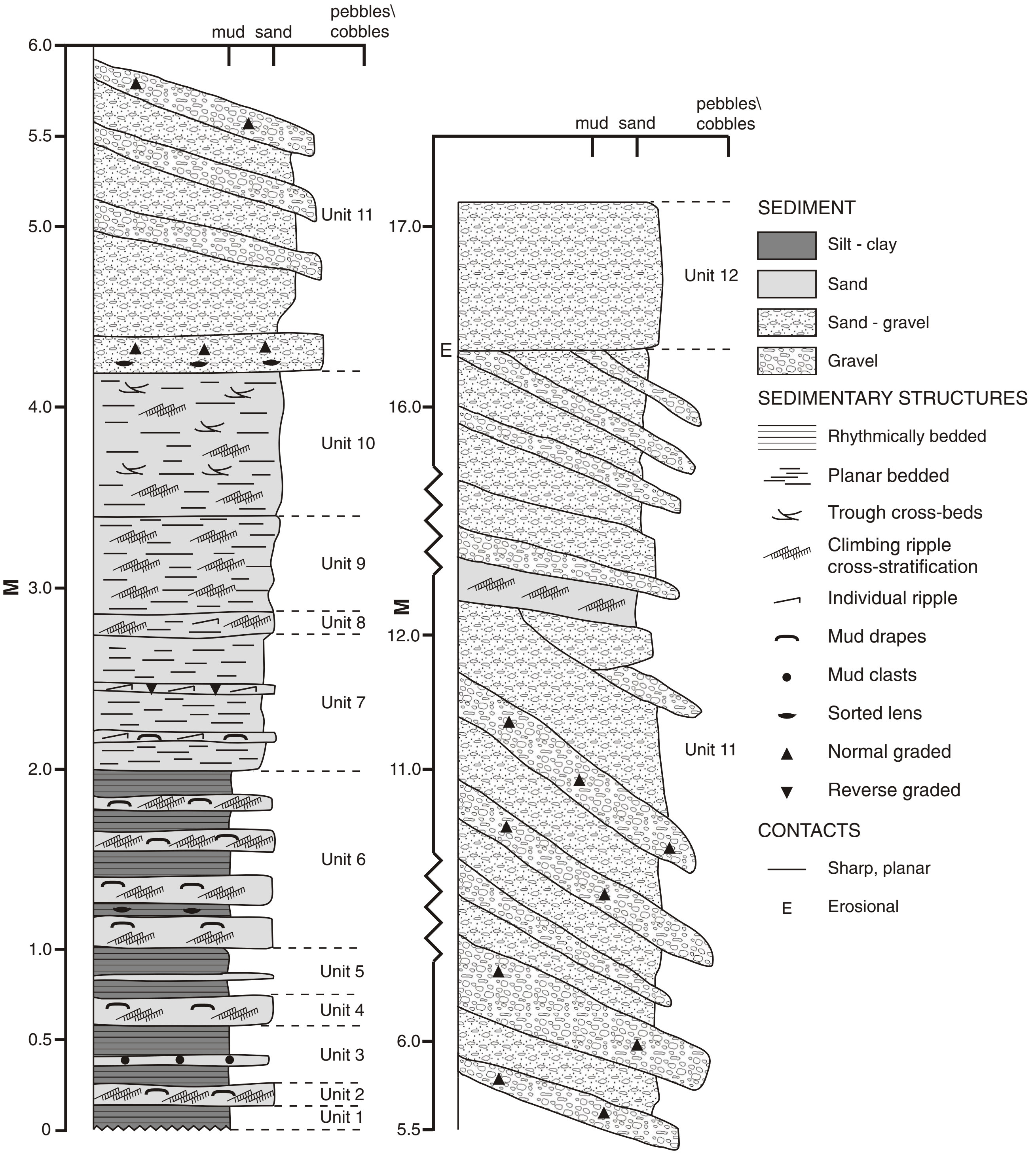


Figure 29. *Stratigraphy of an exposure at Rocky Brook.*

Interpretation

The silt–clayey silt–clay rhythmites were deposited in standing water by underflow either with or without overflow–interflow (similar to Unit 1). The discontinuous clay bed at the base of the unit and the clay intraclasts suggest erosion by the flow that deposited the bed. The intraclasts are interpreted as rip-up clasts, (based on their shape and their random distribution within the bed) that were eroded and deposited by turbidity currents.

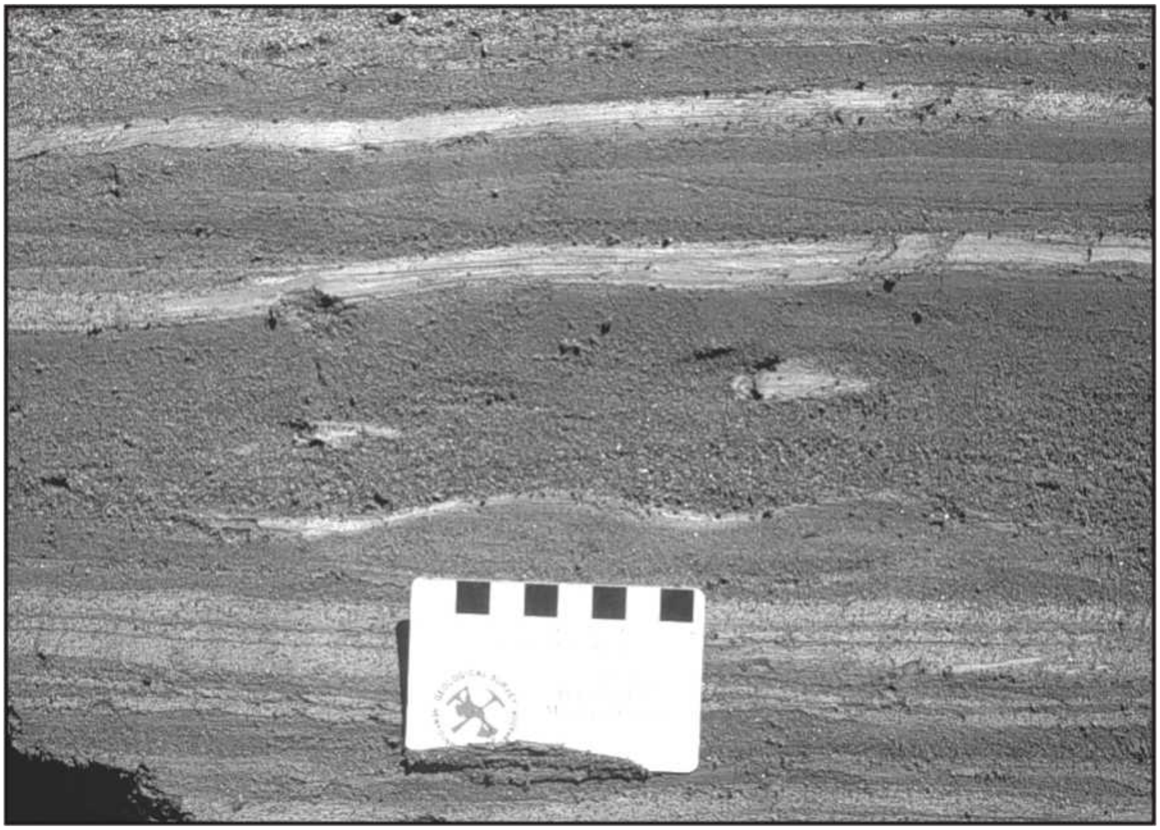


Plate 19. *Rip-up intraclasts from Unit 3 of the Rocky Brook exposure.*

Unit 4: Interbedded Draped Sand Ripples and Sand–Silt Beds

Description

This unit is 15 cm thick. The base is a 2.0 to 4.5 cm thick bed of rippled sand that are asymmetric, erosional stoss (λ=15 cm, H=3, R.I.=5) planar cross-laminated, fine sand ripples, showing flow toward 150°. They are draped by 0.1 cm clay, 2.0 cm normally graded very fine sand to silt, and 4.5 cm normally graded fine to very fine sand, all with sharp contacts. The draped strata are overlain by 4.5 cm fine to very fine sand, with a planar, horizontal upper contact, overlain by 1.5 cm silty clay containing three small (0.5 to 3 mm thick) silt laminae that have sharp upper and lower contacts. A second rippled sand bed occurs at 70 cm, and these are asymmetric, erosional stoss (λ=12 cm, H=1.2, R.I.=10), planar cross-laminated, fine to very fine sand ripples, showing flow toward 150°. The ripples are draped by five, 0.1 to 0.5 cm thick strata of alternating clay, and silt to very fine sand, each with sharp contacts.

Interpretation

The sand ripples were probably deposited by turbid underflow current, in a manner, similar to that described in Unit 2. The overlying draped sequences were deposited by suspension settling during waning current flow or following flow cessation.

Unit 5: Rhythmically Bedded Silt–Clay and Interbedded Sand

Description

This unit is 24 cm thick. The base is a 2-cm-thick bed of rhythmically stratified silt–clayey silt–clay couplets, similar to those described previously. This bed contains five couplets. Overlying a sharp, horizontal contact is 1.5 cm of planar, interbedded fine and very fine sand. This sequence is repeated in three overlying rhythmite and sand beds. Rhythmite beds are 2 to 3.8 cm thick and contain 5, 5 and 12 couplets respectively. The sand beds are 0.8 to 1.5 cm thick. The top of the unit is a 2-cm-thick rhythmite bed containing 4 silt–clay couplets. Unit 5 contains a total of 31 couplets.

Interpretation

This unit is similar to units 1 and 3, and is also interpreted as having been deposited within standing water by suspension settling. Individual rhythmites represent separate surge events; interbedded sand laminae are deposited by underflow currents.

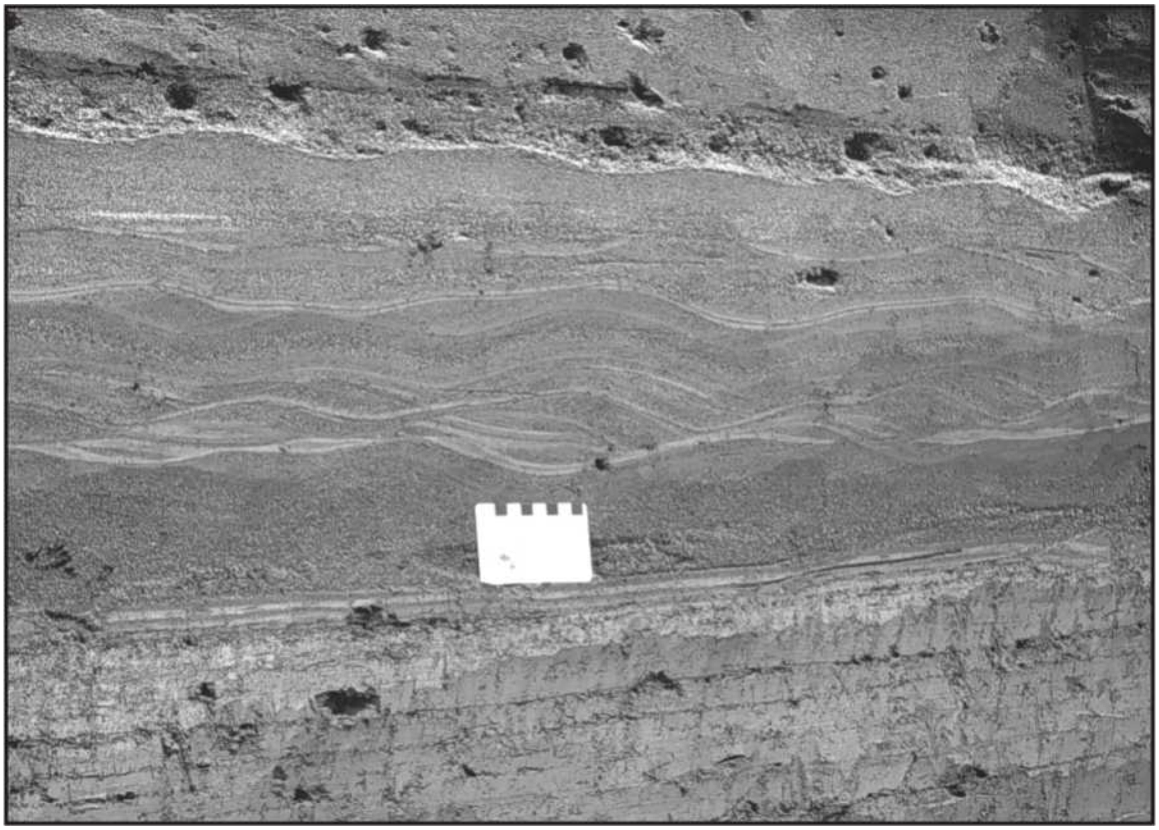


Plate 20. *Cross-stratified climbing ripples at the base of Unit 6 in the Rocky Brook exposure.*

Unit 6: Draped Rippled Sands and Silt–Clay Rhythmites

Description

This unit is 99 cm thick. At the base is a 12-cm-thick bed of asymmetric, erosional stoss ripples (λ=13 cm, H=1.5 cm, R.I.=9) containing fine sand planar cross-laminae, indicating flow toward 150° (Plate 20). Individual ripples are draped by 0.5 to 5 cm silt–clay; the silt–clay is commonly rhythmically bedded. Four ripple sets are draped by silt–clay. A 10-cm-thick bed of planar bedded silt–clay rhythmites lies above the draped ripples, containing 20 couplets. This bed also includes a biconvex lens (16 by 1.5 cm thick) of structureless fine sand. Overlying a sharp contact is 0.2 cm grey clay. The sequence of ripples draped by silt–clay is repeated four times through the unit, with the texture of the drape sediment coarsening upward from clay to very fine sand. Similarly, the rhythmites, which coarsen upward from silt–clay to very fine sand–silt, are found within seven beds, containing 6 to 36 couplets. Unit 6 contains 129 couplets. Interbeds between the rhythmite beds or between ripple and rhythmite beds also coarsen upwards from clay near the base to fine sand at the top. These strata are up to 4 cm thick, although they are commonly much thinner (average 0.5 cm). Strata are ungraded to normally graded.

Interpretation

The thickness of ripple strata is greater in this unit compared to that observed in later units. However, they are interpreted in a manner similar to those in units 2 and 4, and formed by turbidity currents. The larger ripple size is a function of increased current velocity. The ripples are commonly draped by silty clay, and deposited by suspension settling as current flow dropped.

Unit 7: Planar-Bedded Sand and Silt

Description

This unit is 73 cm thick. At the base is a 5-cm-thick rhythmically bedded sequence of seven normally graded, very fine sand to silt laminae with sand commonly thicker than silt. Sharp, horizontal contacts separate the laminae. This unit also contains a small amount (<1 percent) of coarse sand, randomly distributed as individual grains throughout the unit.

There are 24 rhythmic beds, containing 1 to 12 normally graded laminae, for a total of 59 laminae within the unit. The rhythmite beds are commonly separated by 0.1 to 2.2 cm thick planar, horizontal beds of normally graded to ungraded, well sorted, very fine to fine sand having sharp upper and lower contacts.

The exceptions are three beds of rippled sand. The lowest, at 218 cm (above the base of the exposure) contains asymmetric, erosional stoss ripples (λ=14 cm, H=2 cm, R.I.=7), and erosional remnants, composed of structureless, well sorted, medium sand. They are draped by 1.5 to 5.5 cm interbedded fine sand, very fine sand and silt. These beds are truncated by an erosional surface dipping down about 20 northwestward, overlain by a normally graded medium to fine sand bed, thickening northwestward. The middle ripple bed at 244 cm, contains asymmetric, erosional stoss ripples (λ=11 cm, H=0.7 cm, R.I.=16), comprising inversely graded, well sorted, fine sand and showing flow toward 180°. No drapes were noted on these ripples. The upper ripple bed, at 246 cm, also contains asymmetric, erosional stoss ripples (λ=5 cm, H=0.5 cm, R.I.=10), composed of well sorted, fine to very fine sand and showing flow toward 180°. No drapes were noted on these ripples.

Interpretation

This unit is similar to units 1, 3 and 5. The absence of clay and the relative increase in fine sand within the couplets suggests this unit was deposited closer to the sediment source than the underlying units. The ripples are of the same type and depositional environment to those described previously.

This unit contains small proportions of medium to coarse sand, especially within silt layers. The sand is randomly distributed, suggesting it may be unrelated to current flow. The origin of the sand could be as an overflow from an injection of sediment above the thermocline, or as a high density turbidity current in a manner similar to that described by Lowe (1982). Alternatively, the sand could originate as windblown sand onto seasonal ice. Given the depositional environment within a recently deglaciated setting, and the presence of a near-continuous ice cover on modern Deer Lake during the winter, the latter is considered the simpler and more elegant interpretation.

Unit 8: Rippled Sands and Sand–Silt-Graded Beds

Description

This unit is 11 cm thick, and contains 4 thin (0.5 to 1.3 cm) rippled sand strata. The ripples are all asymmetric (λ=4-11 cm, H=0.3 to 1.3 cm, R.I.=8 to 17), erosional stoss planar cross-laminated, fine sand ripples indicating flow toward 180 to 190°. The rippled strata are separated by 1.2 to 3.5 cm thick beds consisting of between five and seven normally graded, fine sand to silt laminae, separated by sharp, horizontal contacts. In all cases, the sand thickness is approximately equal to the silt. Contacts between ripple beds and normally graded sand–silt are sharp, and undulating. This unit contains 17 graded beds.

Interpretation

These sediments are similar to those described in units 2, 4 and 6. The rippled beds and rhythmites are composed of coarser sediment than those found in the lower beds and therefore, this unit is likely more proximal to the sediment supply than the lower ones.

Unit 9: Planar-Bedded Sand to Silt

Description

This unit is 56 cm thick and consists of 91 beds and laminae of normally graded sand and silt. Each stratum contains ~ 1 percent coarse sand, distributed randomly. In examples observed, the silt component was thicker than the sand. Toward the top of the unit, the sand–silt rhythmites are interbedded with planar horizontal, normally graded 0.2 to 0.4 cm thick fine sand containing minor coarse sand lamina and having sharp upper and lower contacts.

A single, 0.6-cm-thick interbed of rippled sand bed occurs at 303 cm elevation. It contains asymmetric (λ=9 cm, H=0.6 cm, R.I.=15), erosional stoss ripples comprising normally graded fine sand. Ripple dimensions vary laterally. A pebble (1 cm diameter) is found within a coarse sand–granule and fine sand bed at 310 cm. The clast lies on a sharp, horizontal contact, but is draped by a thin (1 mm), silt bed.

Interpretation

The graded sand to silt beds are interpreted to have been deposited by suspension settling from underflow with or without overflow–interflow, in a manner similar to that described for Unit 7. The unit is clay-poor, suggesting either that inflowing sediment lacked a clay component, or that the sediment was deposited relatively close to the source and the clay component has been carried by suspension into more distal parts of the basin. The preferred interpretation is the latter, given the presence of clay-rich units within the section. Randomly distributed coarse sand is interpreted as rain-out of windblown sand from seasonal ice.

The pebble clast draped by silt is a drop stone deposited from floating ice or have been rafted on glacier ice. Support for former interpretation is the coarse sand found within the unit. The clast may have rolled onto seasonal ice and subsequently dropped through the water column. No other clasts or carapace structures were noted within the finer grained components of the section, and no sediments deposited in an ice-proximal environment were exposed.

Unit 10: Rippled Sands

Description

This unit is 81 cm thick. The base of the unit is a 17 cm-thick bed of asymmetric, sinuous out of phase, planar cross-laminated, fine- to medium-sand, erosional stoss ripples. These are relatively large ripples (λ=25 cm, H=3 cm, R.I.=8), and are draped by 0.2 to 0.4 cm thick beds of fine sand. The ripples show flow toward 190°. The ripple sequence is truncated along a sharp, horizontal surface by a 3-cm-thick, planar, medium to fine sand bed, containing fine

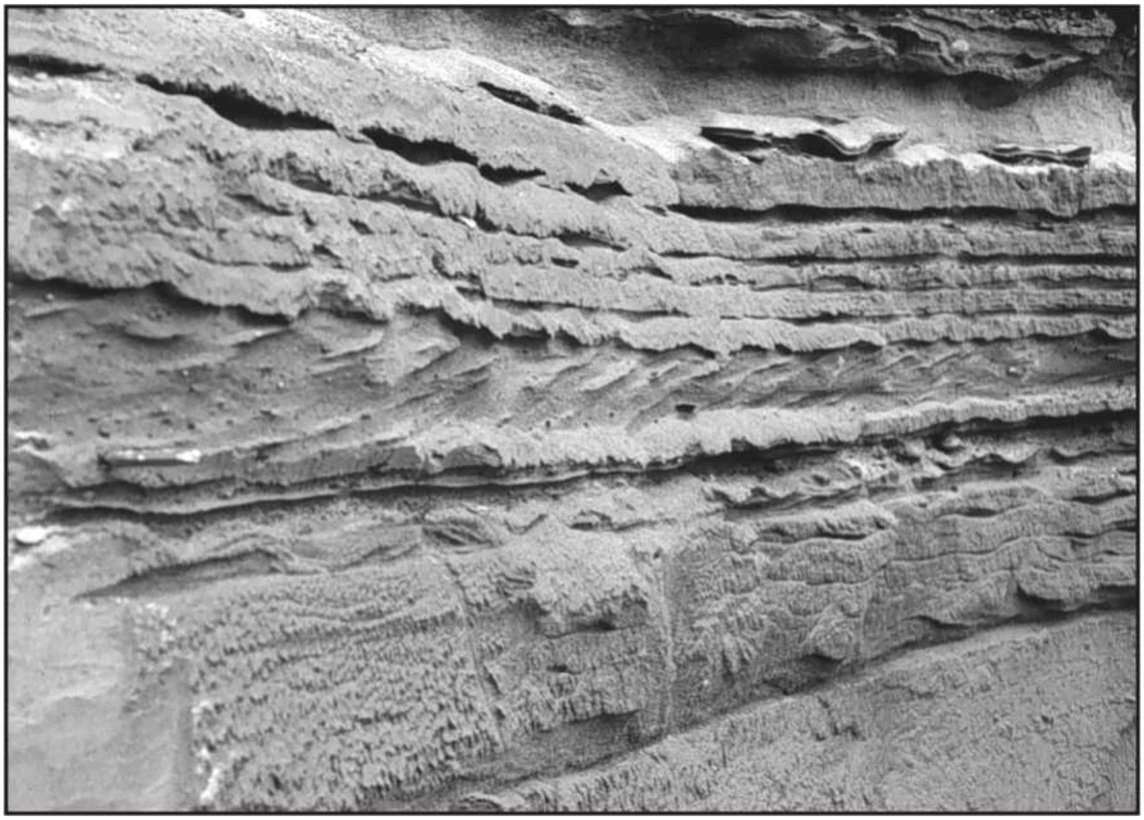


Plate 21. *Trough crossbeds and climbing ripples within a cohesionless, well-sorted fine sand bed (Unit 10).*

sand partings, overlain by a 28-cm-thick bed of trough crossbedded and rippled sand. The ripples are erosional stoss, out of phase, asymmetric (λ=8 cm, H=1 cm, R.I.=8), moderately to well sorted, fine to medium sand, showing flow toward 190°. Planar tabular crossbeds occur within a loose, well-sorted fine sand bed and indicate flow toward about 200° (Plate 21). The crossbed are overlain by well defined 9 to 15 cm thick rippled sand, with sharp lower contacts. Ripple beds are separated by normally graded, sand to pebbly sand beds 3 to 15 cm thick. Individual beds within this unit dip about 9° toward 087°.

Interpretation

The erosional stoss ripples show current flow. Sediment grain size indicates that flow velocities are between 20 and 80 cm sec^{-1} (Harms *et al.*, 1982). The silt to fine sand drapes indicate deposition by suspension settling produced as the current flow waned.

Planar sand beds are interpreted as upper flow plane beds. The alternative interpretation of deposition under lower flow conditions is considered unlikely because of the erosional contact, and the presence of fine sand partings, the result of grain sorting in the bedload (Kuenen and Migliorini, 1950; Moss, 1963; Kuenen, 1966a, 1966b). The grain size (medium sand) suggests flow velocities of 100 to 200 cm sec^{-1} (Harms *et al.*, 1982). The erosional stoss ripples, draped lamination, and upper flow plane beds show Unit 10 was deposited under a variable flow regime.

Unit 11: Interbedded Gravel and Sandy Gravel

Description

This unit is 1.22-m thick, and is composed mostly of loose gravels on steeply dipping surfaces. The base of the unit is a 17-cm thick, normally graded, gravelly sand having a fine to medium sand matrix. Coarse sand lenses occur beneath granule to cobble clasts of mixed rock types. Many of the clasts (~5 cm diameter) dip at 5 to 10° toward 065°. This bed grades upward to medium to fine interbedded sand.

The bottom 350 cm and the upper 400 cm of this unit are sand dominated. The middle part of the unit is gravelly sand, sandy gravel, and gravel beds. Gravelly sand beds are 15 to 43 cm thick, normally graded and matrix supported. Sandy gravel beds are 26 to 38 cm thick, commonly normally graded to ungraded, clast-supported with a medium to coarse sand matrix. They also contain interbeds of open work granule gravel to pebble gravel. Much of the unit is dominated by 13 to 59 cm thick beds of granule–pebble to pebble–cobble gravel. These beds are clast-supported, commonly with an open-work structure, and contain less than 10 percent sand matrix. Planar interbeds of sandy gravel are common. Clasts are of mixed rock types and are dominated by locally derived sandstone and siltstone, and but also include granite, gabbro and volcanic clasts; clasts commonly dip parallel to the bedding.

Toward the base of the unit (416 to 672 cm above section base), beds dip about 24° (range 15 to 28°) toward 070° (range 060 to 080°). Above 672 cm from the section base, beds are more steeply dipping (mean 28.5°, range 20 to 36°) toward about 118° (range 105 to 130°).

Toward the top of the unit, the regular pattern of dipping beds is truncated by a trough-shaped bed having a lateral extent of about 90 cm. It contains normally graded, trough cross-stratified, gravelly sand and interbedded sand. This bed is truncated above a sharp, undulating contact by a well-sorted bed of crossbedded and rippled sand. These are out of phase, asymmetric (λ=10 cm, H=2 cm, R.I.=5), cross-laminated, fine sand ripples. The ripples indicate flow toward about 150°. The sand bed is inclined about 20° toward 108°. The sand bed is overlain by interbedded gravelly sands and open work gravel where individual beds are 8 to 20 cm thick, and all dip about 26° toward 120°.

Interpretation

This unit consists of foreset beds deposited within a delta produced by stream inflow into a basin. Fining upward sandy gravel and gravel beds were produced during periods of sedimentation from either high discharge events or changes in the location of distributary channels on the delta surface (Ashley *et al.*, 1985). Individual beds dip at about 24 to 36°, the latter close to the angle of repose for gravel. These high angles indicate gravity-driven transport and deposition. The range of bed orientations from about 070° toward the base to 118° toward the top, reflects changing locations of delta-front channels. Clast long-axis is commonly parallel to dip-direction of the depositional slope. This is produced by grain avalanching and has been observed and modeled in crossbedded gravels (e.g., Wadell, 1936; Johansson, 1963; Allen, 1984; Ashley *et al.*, 1985).

The trough-shaped bed containing trough cross-stratified sand is a channel scour formed by a distributary channel on the delta surface. The asymmetric ripples overlying the channel are depositional stoss or stoss preserved (Ashley *et al.*, 1982). They are slightly asymptotic climbing ripples having an angle of climb of about 55°. They show unidirectional flow towards about 150°, under low flow velocities (< 40 cm sec^{-1}) (Harms *et al.*, 1982), where suspension sedimentation dominates over ripple migration rates. This interpretation is compatible with formation within a deltaic environment.

Unit 12: Planar-Bedded Sandy Gravel

Description

This unit is 80-cm thick, and is mostly within the zone of pedogenic modification. It truncates underlying beds over a sharp, horizontal contact, and consists of a sandy gravel, has a sand matrix, and gravel to cobble clasts of mixed rock types. Clasts are commonly subrounded, and either imbricate toward about 300° (i.e., dipping up the modern stream), or are flat-lying.

Interpretation

This unit is poorly exposed and is mostly within the soil profile. The coarse texture, moderate sorting, subrounded clasts commonly imbricate toward ~320° (i.e., parallel with the modern river) or flat-lying suggests this is a fluvial deposit.

Summary

The section at Rocky Brook exposes a variety of sediments deposited within a subaqueous environment. The compact diamicton found 400 m upstream of the section was not exposed here. Rhythmically bedded silt–clay and sand–silt in the lower part of the unit show sedimentation within a basin, in an environment distal from sediment supply. The Rocky Brook section contains over 300 silt–clay to sand–silt couplets. Generally, the couplets coarsen and thicken upward, reflecting increasing sedimentation rates and proximity to sediment supply (cf. the work of Gilbert (1975), Smith (1978), and Smith *et al.* (1982) in modern environments). The couplets show characteristics of rapid deposition rather than those of seasonally controlled varves. Thus, they do not provide a reliable measure of the length of time taken to deposit the delta at Rocky Brook. Interbeds of rippled sands are from periodic surge currents. The upper part of the section shows deposition within a delta, proximal to stream input. This is a 'Gilbert-type' delta (Gilbert, 1980; Leeder, 1982; Miall, 1984; Ashley *et al.*, 1985; Edwards, 1986) characteristic of sediment-laden streams entering fresh or brackish water (Miall, 1984). The general upward coarsening of sediment suggests that this delta was prograding.

The ripples were produced by turbidity currents. The drapes and rhythmic bedding were deposited by suspension settling under slack water conditions. Some beds were structureless and poorly sorted. Others show better sorting and normal grading.

The paleogeography of the Rocky Brook area suggests that the delta formed at the head of a fjord, encompassing modern Deer Lake. The surface elevation shows that the water levels were at least 29 m above present. It is not an ice-proximal delta. This is suggested by the lack of diamicton interbeds from either sediment gravity flows or iceberg dumping, and the absence of load structures (cf. Gustavson *et al.*, 1975; Eyles *et al.*, 1987; Ashley, 1988). Seasonal ice deposited rare dropstones. The delta was ice-distal, formed by water flowing from the north-northwest (i.e., similar to the flow direction of modern Rocky Brook), with headwaters in the Long Range Mountains west of Adies Pond.

North Brook

This is a natural cut, bank exposure, 300 m upstream from the mouth of North Brook, where it enters Deer Lake (Site 93157: Appendix 1). The section has a surface elevation of about 20 m asl. North Brook flows south in a channel incised through a generally flat surface that gently slopes toward Deer Lake. The source of the brook is in a small basin in the eastern foothills of the Long Range Mountains west of Deer Lake. The North Brook section consists of about 635 cm of Quaternary sediment overlying bedrock and having a succession of 91 cm gravelly sand, overlain by 109 cm of rippled and crossbedded sands, 315 cm rhythmically bedded silt–clay with sand interbeds, and 120 cm pebbly sand (Figure 30).

Unit 1: Gravelly Sand

Description

This unit is 91 cm thick, and overlies grey to green siltstone and mudstone of the Rocky Brook Formation (Hyde, 1982) that extends about 6 m above the level of North Brook. The gravelly sand is structureless, dark brown (7.5YR 3/4, moist) to brown (7.5YR 5/2, dry), poorly sorted (s.d. 1.4ø) and has a mean grain size of -0.7ø. Silt and clay forms less than 2 percent and is mostly found on the upper surfaces of clasts. Clasts are subrounded to subangular granules to boulders, and are mostly sandstone and mudstone. Clasts appear randomly distributed throughout the unit and have no obvious clast fabric. Local clast-supported zones containing open-work granule gravels are common beneath clasts.

Interpretation

This is a matrix-supported, polymodal, unstratified, ungraded sediment having an unordered clast fabric. It has no horizontal bedding, clast imbrication or crossbedding that may suggest a fluvial or glaciofluvial origin; clasts are not striated and no percussion marks were noted. The unit may be a sediment gravity flow deposit (*see* Middleton and Hampton, 1973; Lowe, 1982; Middleton, 1993), a slurry flow or hyperconcentrated flow. Similar type deposits are found in a range of environments, including alluvial (e.g., Miall, 1977, 1978; Rust, 1978; Rust and Koster, 1984), glacial (e.g., Eyles *et al.*, 1983b, 1988; Goldthwait and Matsch, 1988), glaciomarine (e.g., Powell and Molnia, 1989; Lønne, 1995) and glaciolacustrine (e.g., Rust, 1977; Eyles *et al.*, 1987; Ashley, 1988).

Unit 2: Rippled and Crossbedded Sands

Description

This unit is 109 cm thick. The base of the unit is marked by a sharp, planar contact overlain by 54 cm of rippled and crossbedded sand. The ripples are erosional stoss, climbing ripples (angle of climb ~5°). Ripples are out of phase, asymmetric (λ=13 cm, H=2 cm, R.I.=6.5) and formed in moderately to well-sorted fine sands and show flow toward 150°. Coarse sand to granule laminae, commonly 1 to 3 grains thick, are found along the bounding surfaces. Small (< 0.5

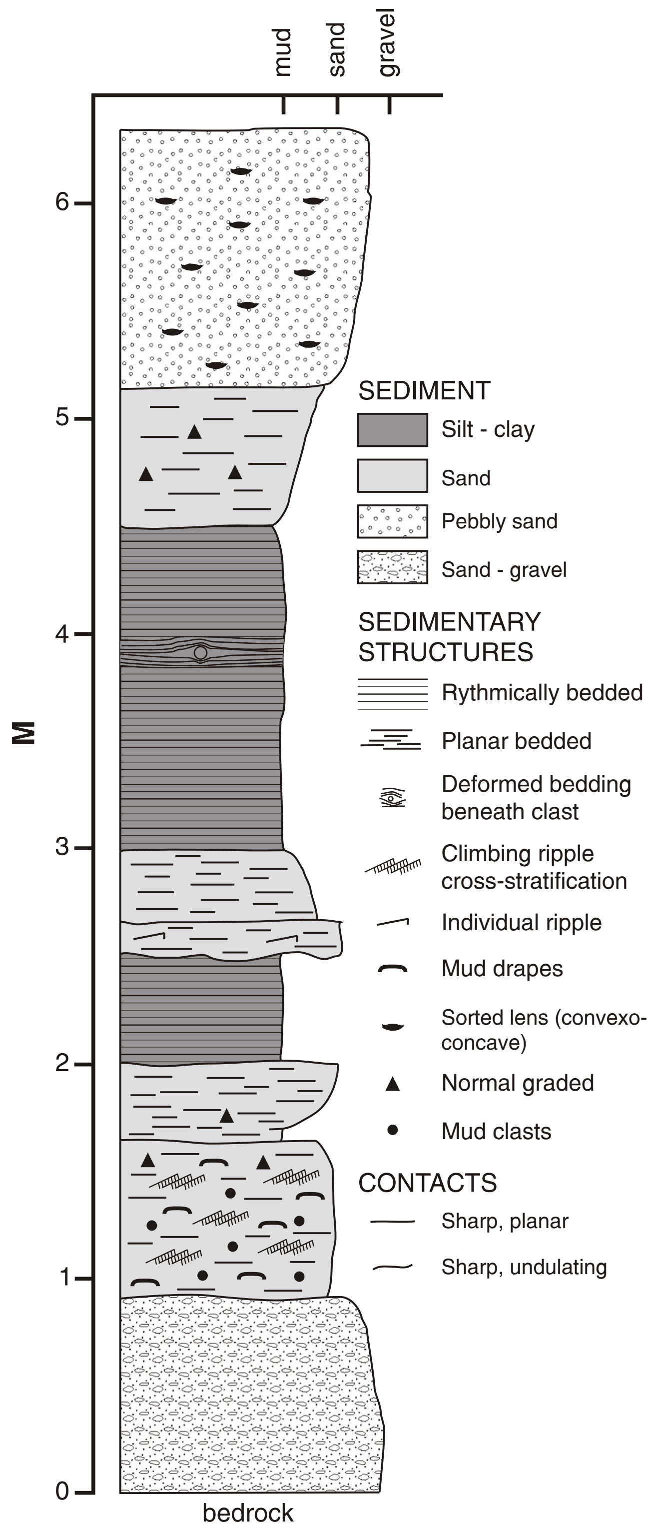


Figure 30. *Stratigraphy of an exposure at North Brook.*

cm diameter), rounded to subrounded silt intraclasts are found randomly distributed through the ripple beds. Some ripples are draped by a thin (<1 mm), very fine sand layer, although most show no draping.

Ripples are draped by 0.1 to 0.3 cm thick, moderately to well sorted, normally graded, medium sand laminae, that are thicker in the ripple troughs and thinner over the crests. This bed grades up into a 1-cm-thick bed containing four normally graded fine sand to very fine sand laminae that have sharp, planar contacts. The rhythmites are overlain along a sharp, planar contact by a 38-cm-thick bed of horizontally laminated fine sand. Small (<1 cm), subrounded to rounded, randomly distributed, silty clay intraclasts occur in the lower 13 cm of this bed. This bed is overlain along a sharp, undulating contact by 2 cm bed of planar tabular crossbedded, fine to medium sand indicating flow toward 140°.

The top of the unit consists of 16-cm-thick planar bedded sands. Individual strata are moderately to well sorted, normally graded, 0.2 to 2.5 cm thick and have sharp, horizontal upper contacts. Rare, clay intraclasts (<1 cm diameter) are found within the coarser sand beds, but not in the finer ones.

This unit fines upward from medium to fine sand at the base, to fine sand to silt at the top. Individual beds are ungraded to normally graded, and show flow structures indicating flow toward between 140 to 150°.

Interpretation

The ripples indicate unidirectional current flow, where the migration rates exceeded deposition. Current flow was discontinuous, as suggested by draping of the very fine sand to silt, deposited by suspension settling. Some draped beds are normally graded or show rhythmic bedding. This unit is a density underflow deposit. Similar draped laminations over fine sand current ripples have been reported elsewhere (*see* Jopling and Walker, 1968; Ashley, 1972; Gustavson *et al.*, 1975; Shaw, 1975; Clemmensen and Houmark-Nielsen, 1981). The most common environment, where this association occurs, is mid-delta (Ashley *et al.*, 1985), where underflow velocities are less than about 60 cm sec^{-1} (Harms *et al.*, 1982).

Unit 3: Rhythmically Bedded Silt, Clay and Sand

Description

This unit is 315 cm thick and is dominated by a repetitive sequence of laminated sand–silt–silty clay–clay beds.

The lower 100 cm of the unit is a fining upward sequence of rhythmically bedded fine to very fine sand, and silt couplets. Individual couplets are defined by normally graded sand laminae, 0.1 to 1.0 cm thick, overlain by silty sand to silt across sharp planar contacts. This lower unit contains a total of 68 couplets. The rhythmic bedding is interrupted by 12 ungraded to normally graded, 0.2 to 3.7 cm thick, moderately to well sorted, fine to medium sand (mean 1.3ø) strata. These beds have sharp, planar to undulating contacts. Rare sand beds contain small (<1 cm diameter), randomly distributed, slightly elongate clay intraclasts that have no preferred orientation.

The middle part of the unit, between 301 and 451 cm above the base of the section, contains rhythmically bedded silt–silty clay–clay couplets (average 32 percent silt and 68 percent clay). Individual couplets comprise normally graded silt to silty clay (0.1 to 1.2 cm thick), overlain by clay (0.1 to 1.5 cm thick) across a sharp, planar contact. Clay is commonly thicker than silt. Couplet thickness decreases from about 1 cm at the base to 0.3 cm toward the top. This part of the unit contains 237 couplets.

Above the silt–silty clay rhythmites is 49 cm of laminated silty sand–silt and medium sand. The silty sand–silt laminae are about 1 cm thick, normally graded with silt thicker than silty sand. A total of 38 rhythmites were counted. The rhythmic bedding is interrupted by six, 0.4 to 3.0 cm thick, ungraded to normally graded sand (mean ~1.3ø) beds overlying silty sand–silt rhythmites across sharp, planar contacts.

There are two exceptions to the sequence of rhythmic bedding and sand beds. The first is a 16-cm-thick bed of rippled and planar tabular crossbedded sands, exposed 253 cm above the base of the section. Ripples were typically out of phase, erosional stoss, asymmetric (λ=10 cm, H=1 cm, R.I.=10) indicating flow toward about 240°. The ripples are commonly draped by very fine sand to silt. The second exception is deformation of silt–clay beds beneath a pebble located at 397 cm from the base of the unit. The pebble deforms 2 couplets (~1 cm thick) and two couplets are draped over the pebble surface.

Interpretation

This unit was mostly deposited by suspension settling in standing water. The high clay content (~68 percent) suggests at least 2 months for settling to occur (e.g., Ashley *et al.*, 1985), although chemical flocculation (e.g., Chase, 1979) or biogenic pellitization (e.g., Smith and Syvitski, 1982) provide mechanisms to increase clay settling rates by up to two orders of magnitude. Individual couplets are normally graded, and the similar relative proportions of clay and silt within individual couplets suggest sediment was introduced by turbid underflow currents (with or without overflow–interflow). The couplets show characteristics of rapid deposition, similar to those at Rocky Brook, rather than those of seasonally controlled varves. The rhythmites thin upward from 1 cm to 0.3 cm, indicating either decreasing sediment input, increasing frequency of turbidity current events and/or increasing proximity to sediment input. Coarsening upward suggests increasing proximity to sediment supply.

Indicators of current flow are rare, confined to several thin beds, where ripples indicate unidirectional flow toward about 240°; contacts with underlying beds are sharp. These sand beds were deposited by underflow currents. Sayles (1919), Caldenius (1932), Agterberg and Banerjee (1969), Shaw *et al.* (1978), and Shaw and Archer (1978) described similar deposits within rhythmically bedded silt and clay, and variously interpreted as turbidity currents generated by either cyclonic activity (Sayles, 1919), catastrophic drainage of glacial lakes (Caldenius, 1932) or failure of an adjacent delta front (Shaw *et al.*, 1978). Given the location of the exposure on a basin margin, adjacent to paleo-deltas, the latter is the preferred interpretation. Failure at the basin margin is common in proglacial subaqueous environments (Ashley, 1988).

Some of the silt beds include small clay intraclasts. These are rip-up clasts derived from underlying beds, deposited by turbid underflow currents, similar to features described by Shaw (1977) and Shaw *et al*. (1978) from lakes in British Columbia.

The pebble clast embedded within the silt–clay rhythmites is interpreted as a dropstone. The bending of stratum beneath and above the clast are formed by release of a clast from floating ice (Thomas and Connell, 1985).

Unit 4: Pebbly Sand

Description

This unit is about 120 cm thick, of which 80 cm has been modified by pedogenesis. The unit is a loose, matrix supported, pebble gravel. The matrix is dark reddish brown (5YR 3/3, moist) to light reddish brown (5YR 6/4, dry), medium to coarse sand (silt–clay less than 1 percent). Clasts are subrounded to rounded granules to pebbles (up to 6 cm diameter). Clast rock types are mixed and contain granite, porphyry, sandstone, and conglomerate. The unit is generally structureless, except for coarse sand lenses and rare openwork granule gravel lenses beneath clasts.

Interpretation

This unit is poorly exposed. The general absence of silt and clay, and the crude stratification suggests deposition by current flow. The gently inclined surface slope toward modern Deer Lake suggests it is part of a graded fluvial system.

Summary

The lowest sandy gravel unit is a hyperconcentrated gravity flow deposit, although whether from within a glaciofluvial, glaciomarine or glacial depositional environment is not clear. The lack of striated clasts does not support, or reject, a glacial hypothesis, although the overlying rippled sand beds makes a glacial origin unlikely. The sands, overlying silt and clay, and the topmost pebbly sand may be found in a marine or lacustrine environment that deepened and subsequently shallowed. This is supported by the gradual fining of sediment within Unit 3 from fine sand at the base to silt–clay in the middle part of the unit, which reflects gradual deepening of the water. Turbidity current activity is indicated by rippled beds and clay rip-ups. Bed inclination shows a source to the east or northeast, consistent with the orientation of ripples, and flow down the modern Humber River valley. Thickness of silt–clay couplets reaches a maximum in the middle of Unit 3, above which couplets thin and become increasingly coarse. This shows deposition within a shallowing water body. The unit is capped by a fluvial deposit.

Discussion

Apart from the delta at Rocky Brook others have been identified at Nicholsville (surface elevation 48 m asl), near Little Harbour (44 m asl), and near Junction Brook (45 m asl). Each delta exposes steeply dipping interbedded sand and gravel (foreset beds). Deltas are graded to fluvial systems upstream, and are likely not ice-proximal. Fine-grained sediment are found at the base of the Rocky Brook and Nicholsville deltas, and on the flanks of the Junction Brook delta and were deposited distal from the basin margins on the basin floor. The sediments at North Brook, that are similar to those in the lower part of the Rocky Brook section, were deposited in similar distal basin environments. Also, similar fine-grained muds are found north along the modern Humber River to Harrimans Steady (20 m asl). Water level thus was at least 45 m above modern elevation at the time of formation of the deltas. Figure 31 provides an interpretive sketch of the head of Deer Lake during the time of delta formation.

Coarsening-upward sediment sequences show increasing proximity to source. Rapid isostatic rebound of the coast in the early Holocene and the consequent fall in relative sea level (*see* Section on Sea-level History, page 121) is the cause. The fall of sea level below the top of the delta resulted in fluvial sedimentation (topset beds) having flow directions similar to those of modern streams. Continued sea level fall led to grading of fluvial systems and incision of the delta at this site. The lack of deltaic deposits adjacent to the modern Humber River suggests reworking of deltas in this area, down to the muds.

HEAD OF HUMBER ARM

Humber River Gorge

This is a small exposure excavated during highway reconstruction on the west side of the Humber River gorge in 1991 (Site 91138: Appendix 1). The site is within a north–south oriented fissure into limestone of the Reluctant Head Formation (Williams and Cawood, 1989), along a water-eroded west–east trending valley sidewall.

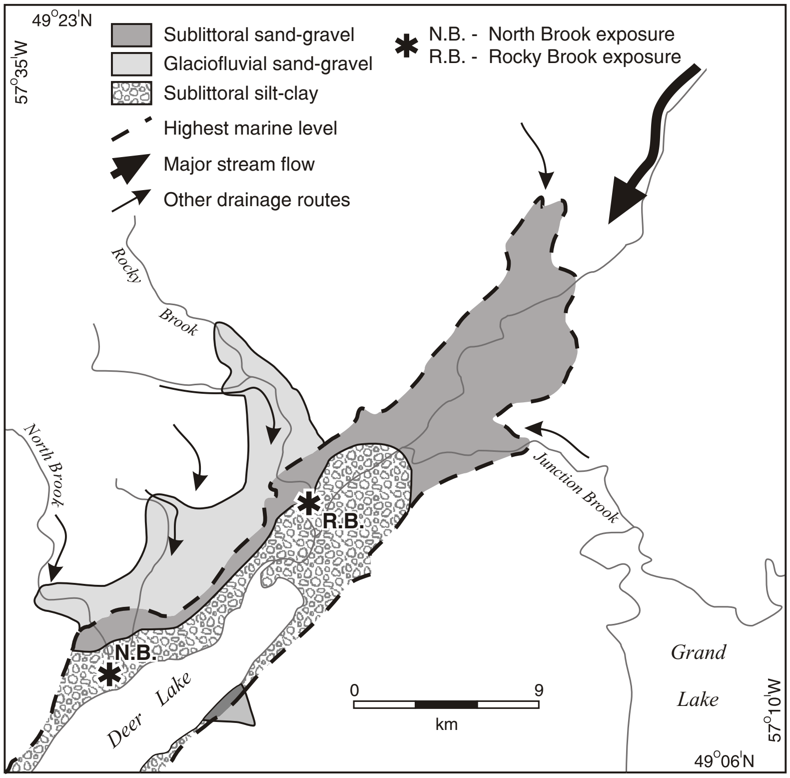


Figure 31. *The paleogeography of the northern part of the Deer Lake valley.*

Description

Diamicton is smeared against the valley sidewall, and is up to 2 m thick. The matrix is a reddish brown (5YR 4/3, moist), matrix-supported (60 percent matrix), clayey silt. Soft-sediment deformation structures including regular folding, and disrupted and faulted (?) bedding are common; the matrix contains numerous marine shell fragments. These are mostly *Balanus hameri* (plates up to 7 cm long) and a few whole *Hiatella arctica* shells. Clasts are angular carbonate fragments, granule to cobble size and the largest clast is 40 cm in diameter. Larger clast sizes are common and are scattered throughout the unit.

Interpretation

The fine grained matrix containing marine fossils has been deposited in the deeper continental margin. *Balanus hameri* has ecological preferences of deep water (20 to 300 m), salinity in excess of 33‰, and mean annual sea surface temperatures of 3 to 15° C (Pilsbury, 1916; Nilsson-Cantell, 1978; Dyke *et al.*, 1996). The preferred salinity is similar to the modern values in the Humber Arm (Shaw *et al.*, 1995). The reddish brown matrix is likely derived from the red Carboniferous sediments within the Deer Lake basin, upstream of this site. The clast component is entirely local and angular clasts suggest no fluvial transport. The sediment also contains no exotic clasts that would be expected if the deposit was fluvially derived and are therefore derived from the adjacent slopes, by mass movement. The introduction of clasts into the muds initiated the soft sediment deformation structures seen in the sediment.

The *Balanus hameri* shell fragments were dated at 12 220 ± 90 BP (TO-2885) (*see* Table 17). This date suggests that the Humber River gorge was ice free before 12 200 BP.

Dawe's Pit

This is a gravel pit on the north side of the Humber River (Plate 22), just across the bridge connecting the north shore highway (Route 440) to Riverside Drive in Corner Brook (Site 93037: Appendix 1). The top of the pit is flat having a surface elevation of 30 m asl, and a surface area of about 0.4 km^2.

Much of the pit face is slumped but exposed areas show a sand dominated north face (Figure 32), and a gravelly west face, capped by a silt–clay layer (Figure 33).

North Face

The basal 3 m of the North Face is obscured by slumping (Figure 32). The lower unit has a minimum thickness of 25 cm, consisting of well sorted, interbedded fine–medium and medium–coarse sand; it has asymmetric (λ=17 cm, H=3 cm, R.I.=5.5), out of phase, erosional stoss, climbing ripples showing flow toward 330°, and trough crossbeds. The ripples are draped across a sharp, undulating contact by a 7-cm-thick layer of interbedded very fine sand and silt and rare pebble clasts. The clasts do not disturb the underlying beds.

The drapes are overlain above a sharp, undulating contact by 20 cm fine and very fine sand. The sands contain out of phase, erosional stoss, trough crosslaminated, climbing ripples. This rippled sand bed is truncated across a sharp, undulating contact by 68 cm of well sorted, large-scale planar tabular crossbedded, medium to coarse sand. Crossbeds dip about 20° toward 340° at the base of the unit, and toward 300° at the top. Individual sets of crossbeds are up to 28 cm thick. The unit contains rare pebble clasts that are aligned parallel to the crossbeds.

The crossbeds are truncated by a sharp, planar horizontal contact, overlain by 170 cm of clast-supported (80 percent clasts) sandy pebble gravel having a coarse sand matrix. The unit contains ten, laterally continuous beds of granule gravel to pebble gravel, each 1 to 3 cm thick. These beds have sharp, planar, horizontal upper and lower contacts, are ungraded, and contain subangular to subrounded clasts of mixed rock types.

Plate 22. *Aerial view of Dawe's Pit, located on the north side of the Humber River near Corner Brook.*

A 142-cm-thick unit of well sorted, mica-rich medium sand lies above a sharp, planar contact. The unit is generally structureless, although poorly defined planar tabular crossbeds showing flow toward about 300° are present near the top of the unit. The unit contains two lenses, both oriented north–south. The lower lens is 35 cm above the base of the unit. It is 8 cm thick, has a sharp, planar lower contact and is composed of interbedded very fine sand and silt. The lens dips 12° toward 280°. The upper lens is 130 cm above the base of the unit, has a lateral extent of about 8 m, and consists of normally graded very fine sand and silt beds. Contacts between beds are sharp. The lens is overlain by 50 cm of planar interbedded fine sands.

A 28-cm-thick rippled sand bed lies above a sharp, planar horizontal contact. Ripples toward the base are out of phase, asymmetric (λ=12 cm, H=1.5 cm, R.I.=8), erosional stoss, planar laminated. In the central part of the unit are 4 cm erosional stoss, trough crosslaminated, medium to fine sand, climbing ripples. At the top, are asymmetric (λ=20 cm, H=2.5 cm, R.I.=8) trough crosslaminated ripples. Flow directions indicated by ripple asymmetry are 0° at the base of the bed, 340° in the middle, and 160° at the top of the bed. The drapes are overlain by 4-cm-thick bed of erosional stoss, climbing ripples and trough crosslaminations showing flow toward about 340°. All the ripples within this unit are draped by 0.3 to 1.0 cm thick, very fine sand–silt normally graded strata having similar characteristics to those described in the previous paragraph.

The next unit is 45-cm thick and comprises planar interbedded, fine and fine–medium sand, across a gradational, undulating contact, which in turn is overlain by 120 cm sandy gravel across a sharp, planar horizontal contact. This unit is clast supported (80 percent clasts) having a coarse sand matrix. It contains 2 to 5 cm thick interbeds of coarse sand–granule gravel, and sandy pebble gravel containing clasts up to 5 cm diameter, that are commonly imbricate (dip at 10 to 14° toward 220 to 250°).

The sandy gravel bed is overlain across a sharp, planar horizontal contact by 20 cm thick layer of interbedded sands and pebbly sands where with individual beds dip at ~8° toward 240°, and subsequently across a sharp, planar contact by 50 cm of sandy pebble gravel.

The North Face is capped by about 1 m of loose, interbedded pebble gravel and sandy gravel. Individual beds are up to 6 cm thick, normally graded and dip ~10° toward 050°, into the face.

Several of the beds in the lower 5 m of the North Face are truncated by normal faults (Plate 23). Seven fault planes were noted having angles of dip commonly 60 to 80°, with 5 to 25 cm offsets. Faults are spaced 30 to 100 cm apart.

West Face

Figure 33 is a stratigraphic log of the West Face of Dawe's Pit. The lower part of the West Face is similar to the exposures of the north face (*see* Figure 32). The lower slope is a unit of interbedded, medium to medium–coarse sand, and coarse sand to granule gravel having sharp planar contacts. Individual beds dip at about 20° toward ~270 to ~360°, but there is no clear pattern distinguished. Beds are commonly truncated laterally by normal faults with up to 3 cm offsets. This unit is in sharp, planar contact with an overlying gravel unit.

The gravel unit is approximately 5-m thick, crudely stratified, and clast supported (15 percent coarse sand matrix). Clasts mostly are subrounded, granules to boulders (up to 50 cm diameter), dominated by micaceous schist, and less common basalt, rhyolite and sandstone clasts. Generally, clasts dip into the face, at ~20° toward 360° (Plate 24). Open work granule–gravel to pebble–gravel lenses are common throughout the unit. A veneer (1 mm) of reddish brown silty clay covers the upper surfaces of clasts within the upper 50 cm of the unit.

The gravel unit is overlain across a sharp contact, by interbedded sand, silt and clay, that occupy a trough-shaped depression which pinches out laterally toward the east (Plate 25); the central part of the trough is 550-cm thick. At the base of the trough is a 3 cm dark reddish brown (5YR 3/3,

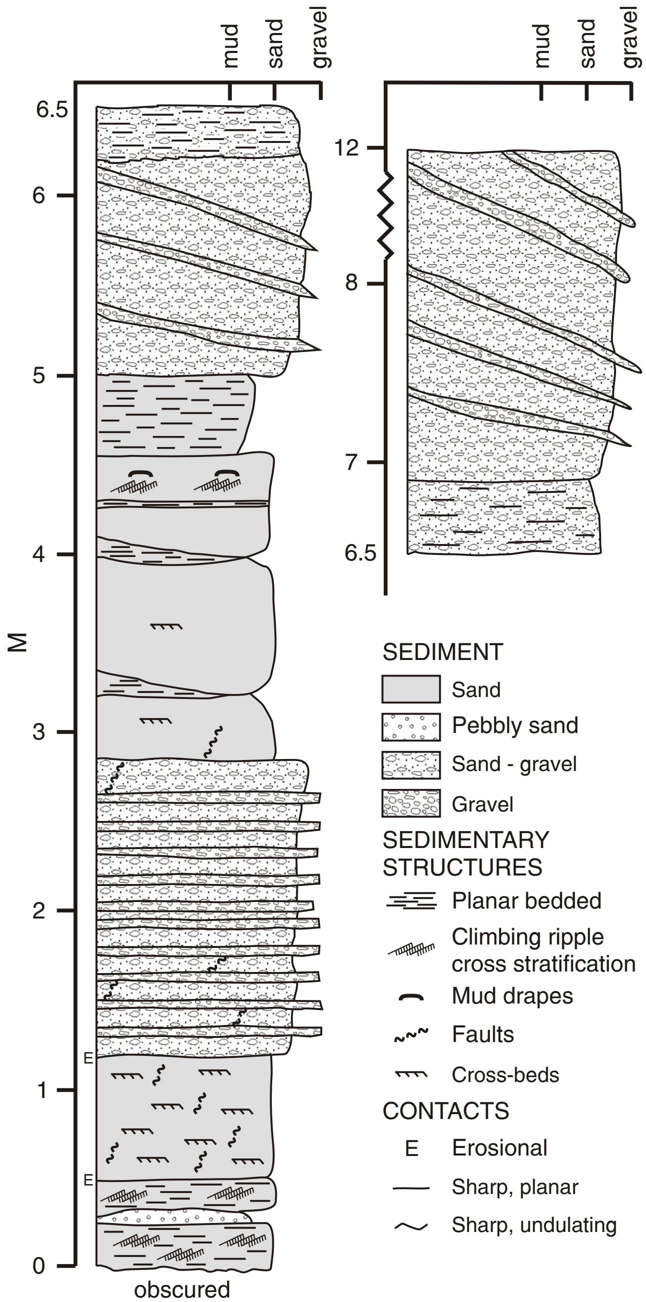


Figure 32. *Stratigraphy of the northern exposure at Dawe's pit.*

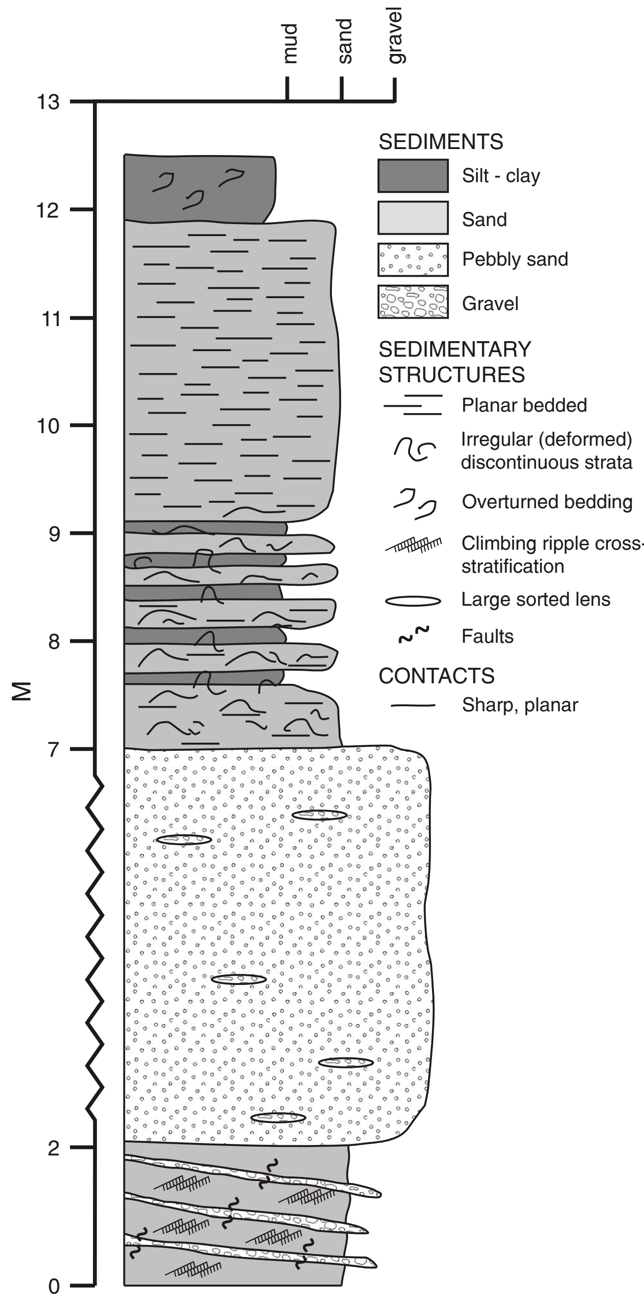


Figure 33. *Stratigraphy of the western exposure at Dawe's pit.*

moist) to reddish brown (5YR 5/3, dry), structureless silty clay layer. It is overlain by a 25 cm thick planar bedded, well sorted, coarse and medium sand layer. Nine similar sequences of silty clay (or clayey silt) overlain by sand are present vertically through the unit. The silty clay (or clayey silt) beds are structureless, and 0.2 to 15 cm thick, increasing in thickness upward. Lower contacts are sharp and planar. The sand beds are thicker (7 to 43 cm), and have sharp, irregular lower contacts. Both, the sand and silt–clay beds, dip steeply into the face (~40° toward 330°). This is overlain by about 260 cm interbedded fine and very fine sands, commonly having sharp irregular contacts between beds. The North Face is capped by about a 60 cm reddish brown, structureless, silty clay layer.

The general stratigraphy within the sand, silt and clay unit is commonly deformed, particularly in the lower part. The eastern margins of the unit dip steeply (60 to 80°), and some beds are deformed into recumbent folds. In other places, the bedding has been completely deformed. Soft-sediment deformation features include regular folding, and chaotic and dislocated beds associated with diapiric struc-

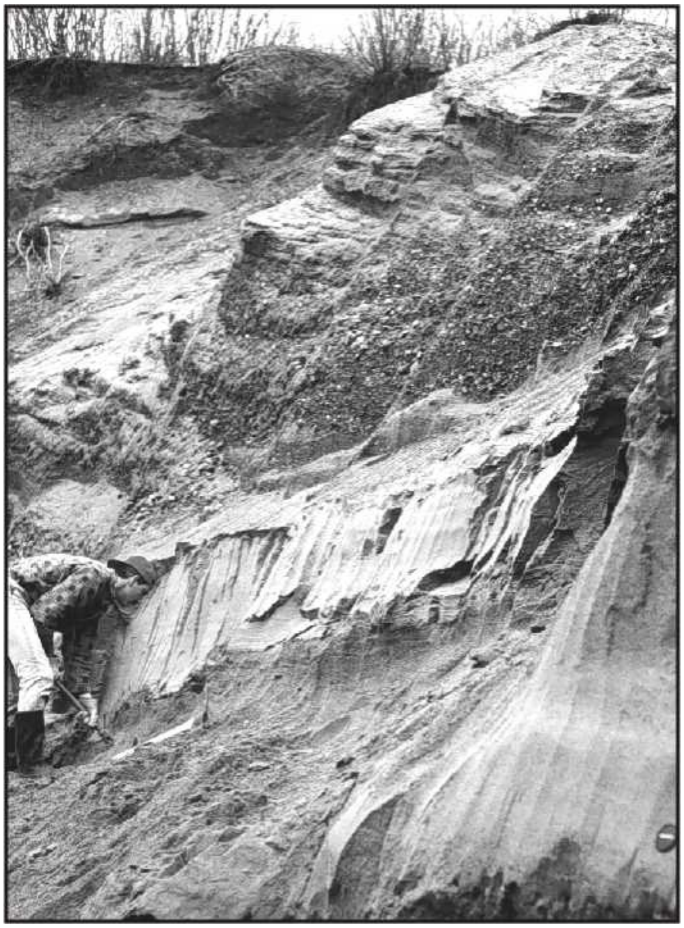


Plate 23. *Synsedimentary normal faults within sand–gravel on the north face of Dawe's pit.*

tures. Generally, the folds are regular and symmetrical, have a wavelength of less than 4 cm and an amplitude of less than 2 cm. Chaotic bedding and associated diapiric structures are common. Individual beds are contorted or disrupted. Diapirs are 4 to 8 cm high, and almost symmetrical about the vertical axis.

Interpretation

The rippled sand, sandy gravel and interbedded silt that dominate the West Face and that are found also at the base of the North Face are fluvial sands and gravels. The erosional stoss, climbing ripples were deposited by unidirectional current flow, in which ripple migration is greater than deposition. Ripples are generally small and fine grained, suggesting deposition by relatively low flows (~ 50 cm sec^{-1}) (Harms *et al.*, 1982). Flow directions are variable, ranging from 160 to 360°. Similar flow structures have been reported from sand flats on the South Saskatchewan River (Cant and Walker, 1976), and meandering streams (Jackson, 1976; Leeder, 1982), in which current velocities were low, but variable. Waning flow conditions are interpreted to have formed the thicker beds of cross bedded sands overlying rippled sand beds (Harms *et al.*, 1982). Draped silt and silty clay beds overlying the ripples record periods of flow cessation. The dominantly sand unit coarsens upward. This sequence is either the result of increasing proximity to a sediment source or hydrological changes that produce greater runoff and sedimentation rates.

The clast-supported gravel unit on the West Face is interpreted as a fluvial sediment. The matrix component is low (less than 10 percent) and composed of coarse sand. The sediment was deposited by an energetic flow that kept sand in suspension. The imbricate clasts, dipping into the bank, show that flow was from the north. Crudely bedded gravels that have imbricate clasts are deposited in longitudinal bars (e.g., Smith, 1974; Hein and Walker, 1977; Miall, 1977, 1978). Clast imbrication is not found in debris flow sediments (e.g., Rust and Koster, 1984). The unit is laterally discontinuous, suggesting channeled flow. The reddish brown silt–clay caps on the clasts in the upper 50 cm of the unit may result from postdepositional translocation of fines from overlying beds.

The silt–clay, and interbedded sand unit at the top of the West Face lies within a channel-shaped depression in the underlying fluvial gravels. The sediment contain no macrofossils and were not sampled for microfauna. Its position above fluvial gravels on an isostatically rebounding coast suggests the muds are not marine and, therefore, have been deposited in an abandoned channel by suspension settling. Similar fine-grained deposits associated with coarse gravels have been described from gravely braided systems (Smith, 1974; Hein and Walker, 1977, 1982; Rust and Koster, 1984). The soft-sediment deformation structures are loading features and the small regular folds are convolute laminations. Similar structures were described from fine-grained turbidites (e.g., Kuenen, 1953; Bouma, 1962; Allen, 1984), or their fluvial–deltaic analogs (e.g., Picard and High, 1973; Allen, 1984). The chaotic bedding and associated diapiric features are interpreted as load casts. They are typical of fluvial and deltaic deposits (e.g., Potter and Pettijohn, 1963; Collinson and Thompson, 1982; Allen, 1984), and turbidites (e.g., Kuenen, 1953; Bouma, 1962; Allen, 1984). These features are formed by rapid deposition of a relatively dense unit onto a saturated, finer substrate. The deformed silt–clay bed is overlain by up to 200 cm fine and very fine sand, showing poorly defined ripples. The grain size and the presence of current structures suggest the sand was deposited by a current flow of relatively low velocity that deposited the sands quickly over the sand–mud beds causing the unit to dewater and leading to the soft-sediment deformation structures observed. The structureless, undeformed silty clay bed overlying the sands shows that sedimentation by suspension settling continued following introduction of the sand bed.

Much of the section shows normal faults, indicating tensional stresses, with up to 25 cm offsets. Fault planes are sharp and have small throws. Sediment thickness is consistent on both sides of faults. The upper part of the section is

Plate 24. *Gravel bed showing clasts dipping northward on the west face of Dawe's Pit.*

commonly not faulted. These factors show the faults are synsedimentary, probably produced by slumping.

Section Interpretation

Dawe's Pit lies on the north side of the modern Humber River, within a small bedrock controlled embayment immediately downstream of the Humber River gorge. Although sediments exposed within the pit exhibits considerable lateral and vertical variability, they are all interpreted to have been deposited within a fluvial environment. Fluvial sediments indicate current flow opposite to the general flow of the modern Humber River. This may be explained by current flow into a back channel in which Dawe's Pit is now situated (Figure 34). This interpretation is supported by the variable direction of current flow as indicated by the ripples, the local geomorphology in relation to river flow, and the abundant sediment supply provided by the Humber River. Current eddying explains ripples showing flow oblique or opposite to the modern Humber River. Draped ripples and planar bedded gravels with similar characteristics to those found in Dawe's Pit are found within the Humber gorge, and were exposed during highway construction in 1991 (Sites 91138, 91139: Appendix 1).

The alternative explanation is that flow was from an adjacent small valley north of the pit area, but which is bedrock floored and has a small watershed situated on the bedrock-dominated highlands overlooking the Humber River valley. The lack of an obvious sediment supply suggests this is an unlikely source for the sediment in Dawe's Pit.

The section is a fluvial sediment, deposited by current flow into a bedrock controlled back channel on the north side of the modern Humber River. Flow strength was low, and stopped during certain periods, perhaps during winter freeze up. The section generally coarsens upward, reflecting increasing proximity to a sediment source, possibly caused by a fall in relative sea level. The alternative explanation is increased discharge and sedimentation rates produced by changing climate, but this could not be corroborated (cf. Macpherson, 1981). The gravel-rich bed is a longitudinal bar formed across the mouth of the embayment, possibly as river depths decreased during sea-level lowering. A channel on the north side of this bar isolated from the main channel became an area of standing water. Sedimentation in this pond was by suspension. Introduction of sand into the pond (during a flood?) produced the loading structures observed. Sediment was eroded and a terrace formed during sea level fall and the continual grading of the Humber River to reduced base levels. The removal of the lateral support produced the collapse features noted throughout much of the sand unit.

In a discussion of a radiocarbon date (12 700 ± 300 BP (GSC-4272) from *Macoma balthica*) found at 15 m above present sea level in muds adjacent to the Dawe's Pit. Grant (*in* Blake, 1987, p. 6) previously described the exposure as a "kettled glacier-marginal kame delta". Such a delta would have formed in a glaciomarine environment, similar to features and sediments described by Cheel and Rust (1982). The present interpretation based on additional data, of the depositional environment at Dawe's Pit does not support this interpretation. Fluvial deposition is indicated by the presence of low velocity current ripples, and the moderately sorted, imbricate gravels show no evidence (e.g., striations) of glacial transport. Dawe's Pit does not contain oversized, angular boulders, debris flow deposits, and rapid lateral and vertical changes in grain size that may be expected from an ice-proximal environment. Although the Humber River discharge was probably dominated by glacial meltwater, ice was not proximal to this site during deposition of the sediments found in Dawe's Pit.

Wild Cove Valley

The Wild Cove valley is an east–west oriented, flat-bottomed valley, the mouth of which is about 5 km north of the mouth of the Humber River. Wild Cove is separated from

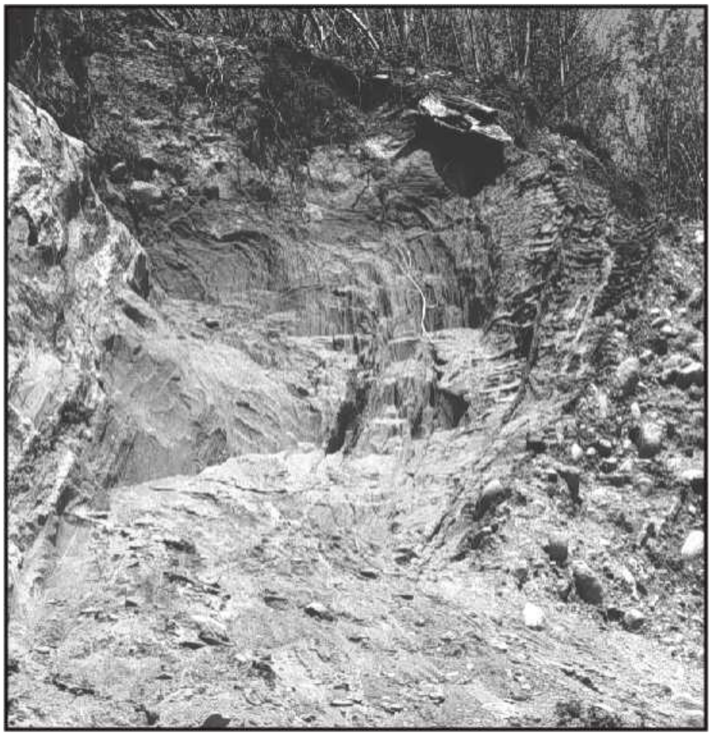


Plate 25. *Steeply dipping silt–clay within the upper part of the west face of Dawe's Pit. These were produced by loading from a rapidly sedimented sand bed that overlies the silt–clay.*

the Humber River valley near Steady Brook by a col at an elevation of 90 m. Up to 16 m of marine clay cover the valley floor (Ricketts, 1987). A small delta having surface elevation of 50 m asl (Site 91219: Appendix 1) is found at the head of Wild Cove. The north side has active scree slopes. In contrast, the south side of the valley shows fan-shaped features (Plate 26), extending 1.8 km from the southwestern end of the valley to near the delta. Sediment within the fans are poorly exposed. Two sections are described. One is a small borrow pit that contains marine macrofossils. The other is a backhoe pit.

Borrow Pit

A small borrow pit at the southwestern end of the valley (Site 91025: Appendix 1) exposes 2 m of sediment, over a lateral extent of about 4 m, at the base of a steep slope. A section log is shown in Figure 35. Top of the pit is about 33 m asl.

At the base of the pit is at least 20 cm of sandy gravel. The gravel unit is matrix-supported having a moderately sorted medium to coarse sand matrix. Clasts are subangular to subrounded, granules to cobbles of mixed rock types. Some clasts are striated. The unit contains subhorizontal lenses of open-work pebble gravel, dipping northward into the valley.

This unit is overlain across a sharp undulating contact by a 30 to 60 cm thick bed of pebbly sand, composed of moderately sorted fine sand, that pinches out east and west. Medium- and coarse-sand fractions are largely absent. Pebbles are subangular to angular, and of mixed rock types. The matrix contains marine shells, mostly *Mya truncata*, but also *Mya arenaria* and *Macoma calcarea*. Specimens are commonly whole valves, although not in growth position. The fossiliferous pebbly sand bed is draped by a 10 to 30 cm thick moderately sorted, normally graded, coarse to medium sand (80 percent) and granule gravel (20 percent) bed.

This unit is overlain, across a sharp, undulating contact, by about 60 cm of pebbly sand. It is composed of about 60 percent, structureless, moderately sorted fine sand; medium and coarse sand is poorly represented. Subangular to subrounded pebbles of mixed rock type account for 40 percent of the unit. The bed contains convexo–concave and convexo–planar lenses, 40 cm to greater than 80 cm lateral extent, and 10 to 20 cm thick. The lenses are structureless, and contain medium to coarse sand and pebbles. The lenses pinch out east and west, and are continuous into the exposure, dipping 15 upslope. The lateral extent of this unit is unknown.

Backhoe Pit

A backhoe pit was excavated in the fan about 1100 m east of the borrow pit along a narrow gravel road about 20 m above the valley floor (Site 91220: Appendix 1). It exposed a vertical section of 3.8 m containing interbedded diamicton, sand and silt–clay (Figure 36). The pit was 3 m wide, and had a surface elevation of about 41 m.

The bottom unit is a >150-cm-thick bed of loose, brown (10YR 4/3, moist), poorly sorted (s.d 1.9ø), fine sand (mean 2.8ø). The unit is generally structureless, apart from planar, very fine sand laminae, and a single subhorizontal, irregular-shaped lens, 5 cm wide by 1 cm high, pinching out east and west, containing structureless very fine sand; the lens dips about 6 downslope. The unit is overlain by 100 cm diamicton across a sharp, wavy contact. The diamicton is dark yellowish brown (10YR 4/4, moist), matrix supported, and structureless, and has an extremely poorly sorted (s.d. 4.2ø), silty sand matrix (mean 0.8ø). Clasts are subangular to subrounded, granules to boulders up to 70 cm diameter, of mixed rock types. Larger clasts are concentrated toward the top of the unit. Clasts have a strong, slightly clustered fabric (S_1=0.76, S_3=0.05) and a preferred clast orientation toward 325° (i.e., downslope).

The diamicton is overlain by 10 cm of brown (7.5YR 5/4, moist), laminated, very poorly sorted (s.d. 2.8ø), silt (mean 5.6ø) across a sharp, wavy contact, that grades upward into 5 cm of laminated fine sand. Laminae are up to 0.5 cm thick, and ungraded, and contain 10 to 20 percent

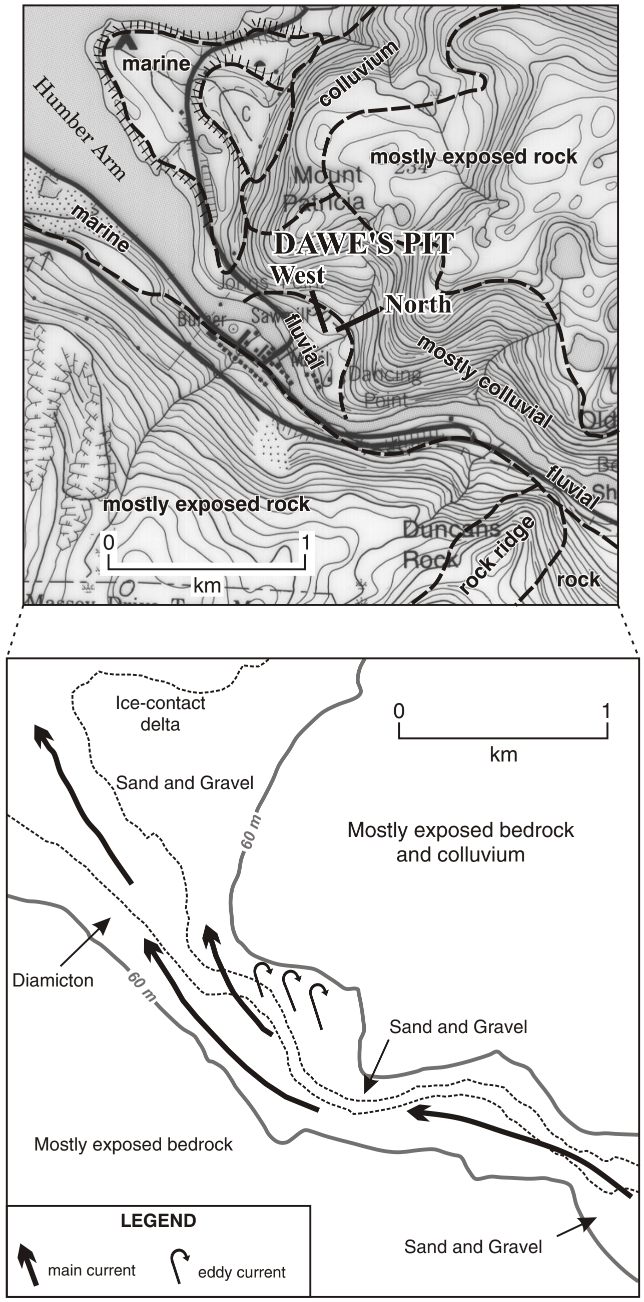


Figure 34. *Paleogeography of the Dawe's Pit area.*

(increasing upward), randomly distributed coarse sand and granule gravel. Individual laminae in both the silt and sand units are commonly contorted (mostly regular folds) and, rarely, disrupted.

The sand bed is overlain by 100 cm of diamicton across a wavy (loaded?), gradational contact. It is matrix-supported (mostly fine sand), with less than 10 percent silt–clay. Subhorizontal lenses of structureless sandy gravel having a medium to coarse sand matrix, and subangular to subrounded granule to cobble clasts, are common throughout the unit. Clasts are preferentially oriented downslope. The diamicton extends to the top of the backhoe pit.

Interpretation

Sediment exposed in the backhoe pit are interpreted as having been deposited in an ice-proximal subaqueous fan environment. The basal sand bed is waterlain, deposited on an inclined surface, likely by sediment gravity flow (grain flow?). The overlying diamicton bed is interpreted as a hyperconcentrated sediment gravity flow deposit. This is supported by deposition on an inclined surface, poor sorting, and the preferred concentration of larger clasts toward the top of the bed (cf. Lowe, 1982; Lønne, 1995; Benn, 1996). The clast fabric is strong for many sediment gravity flow deposits (cf. Lawson, 1979; Dowdeswell and Sharp, 1986), but the preferred clast orientation downslope is most likely due to flow controlled by slope, rather than being related to the westward glacial flow (*see* Section on Ice-flow History, page 103). The diamicton is both underlain and overlain by fine-grained sediment interpreted to have been deposited in a subaqueous environment.

The uppermost diamicton is also interpreted as a gravity flow deposit. The gravity flow that deposited the diamicton was likely subaqueous. Diamictons located on inclined surfaces, and interbedded with sand, silt and/or gravel, are interpreted as sediment gravity flow deposits, possibly formed in an ice-contact environment (Powell, 1981, 1983; Lawson, 1988; Lønne, 1995). The deformation of sand and silt beds is interpreted as having been produced as a result of loading by the diamicton bed that overlies it.

The sediments described above and the gentle dip to inclined beds suggests deposition as an ice-contact subaqueous fan. The glacier terminated in the sea, rather than on land. This would preserve diamicton beds that otherwise would likely have been reworked by surface streams. Apart from the section at the western end of the fan, the sediments examined along the southern wall of the Wild Cove valley do not contain marine macro-fossils. This may be the result of the high sedimentation rates likely in ice-proximal subaqueous environments (e.g., Powell, 1991; Syvitski *et al.*, 1996), and the consequent unsuitable habitats.

Sediment genesis in the borrow pit section is difficult to determine due to the poor exposure. The basal sand and gravel unit contains striated clasts, suggesting a glacial source. The inclined granule–gravel lenses indicate concentrated flow down slope. The fossiliferous sand unit is marine. Macro-fossil species are pelecypods of a pioneer assemblage (Dyke *et al.*, 1996), that prefers shallow arctic waters. A single *Mya truncata* shell was radiocarbon dated at 12 450 ± 90 years BP (TO-2884), and provides a minimum date for deglaciation of this site (*see* Table 17). Pebbly sand beds are interpreted as sediment deposited in a channelised subaqueous fan environment (cf. Hein and Walker, 1982; Walker, 1978, 1984). This interpretation is supported

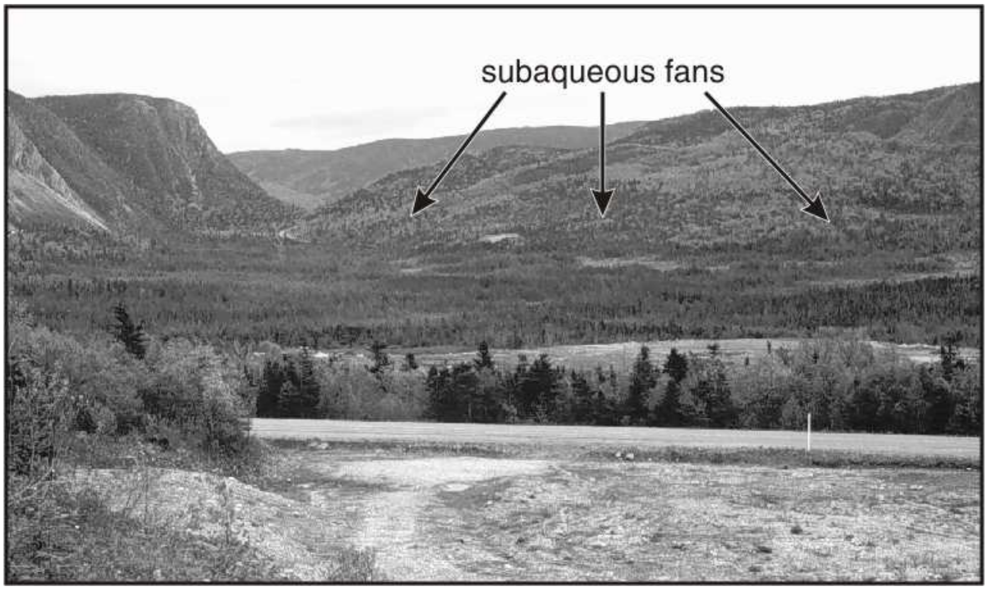


Plate 26. *View of the Wild Cove valley showing subaqueous fans (right side).*

by the close association with marine deposits, and the gravelly sand lenses found within the unit.

The angle of inclined beds toward the valley, the lack of features (e.g., faults) showing collapse, and the flat bottomed valley underlain by up to 16 m of silt–clay, suggests that the sediment source was the highlands to the south, rather than ice-marginal sedimentation from a glacier occupying the Wild Cove valley.

Hughes Brook Pit

Hughes Brook has its source in Hughes Lake, located 6 km west of Deer Lake; it flows west, through Long Pond and Balls Pond, before turning south and flowing through a broad valley until it enters Humber Arm east of Irishtown. The upper reaches are bedrock dominated; the middle contains sand and gravel; and the lower reaches are incised through a narrow rock gorge about 400 m long characterized by numerous pot holes, downstream of which are marine muds at the outlet. Much of the lower reaches lie below the previously suggested marine limit of 49 m asl (Brookes, 1974). The valley contains several abandoned sand and gravel pits, but most are slumped or sloped with no good exposure. One active pit exists, operated by North Star Cement of Corner Brook, hereafter called the Hughes Brook pit.

The Hughes Brook pit is located on the east side of the valley about 5 km upstream from Humber Arm (Site 91173: Appendix 1). The surface of the pit is 61 ± 2 m asl (altimeter estimate). The pit is at the mouth of a small, narrow valley that extends 7 km eastward. Descriptions were made from two fresh faces, on the south and west sides of the pit.

South Face

The South Face shows a 5.5-m-thick exposure of sand and pebbly to gravelly sand extending from the pit floor to modern surface. An unknown quantity of material has been removed, but likely less than 3 m, based on a comparison to the adjacent, unmodified slopes. Beds are commonly laterally continuous for at least 2 m, except where otherwise noted (Figure 37).

The base shows more than 20 cm gravelly sand, interbedded with coarse sand–granule gravel. The gravelly sand is loose, poorly sorted (< 2 percent silt–clay), and has subangular to subrounded, granule to cobble clasts (up to 8 cm diameter), composed of mixed rock types. Coarse sand–granule gravel interbeds are 1 to 1.5 cm thick and have sharp, planar upper and lower contacts. Granule gravel interbeds are commonly open-work.

This is overlain by 15 cm of planar stratified, sand and silt–silty clay. At the base of the unit is a 2-cm-thick bed of normally graded clayey silt, with the silt component thicker than clay, overlain by rhythmically bedded fine to very fine sand. Individual couplets are normally graded, 0.1 to 1.8 cm-thick. Couplets have gradational internal contacts (i.e., fine to very fine sand), and sharp between-couplet (i.e., very fine to fine sand) contacts. Thirteen couplets were counted within this bed.

This unit is overlain by 16 cm of interbedded fine, and medium to fine sand. Individual beds are 0.2 to 1.0 cm thick, normally graded, and dip at ~20° toward 160°. Planar tabular crossbeds are common and show flow toward 210°. The unit also contains a small (3.5 cm wide and 1 cm high) channel containing normally graded coarse sand. The channel lies on an inclined surface that dips at ~20° toward 220°.

The crossbedded sands are overlain by 9 cm of interbedded coarse sand and granule gravel across a planar gradational contact, 10 cm of interbedded fine and medium–fine sand, and 44 cm of gravelly sand and interbedded coarse sand and granule gravel, each with gradational planar lower contacts.

These beds are overlain by a 10-cm-thick bed of draped rippled sand. They are mostly erosional stoss, asymmetric, trough crossbedded, climbing ripples (λ=13 cm, H=1.8 cm, R.I.=7) showing flow toward 080° (Plate 27). Some deposi-

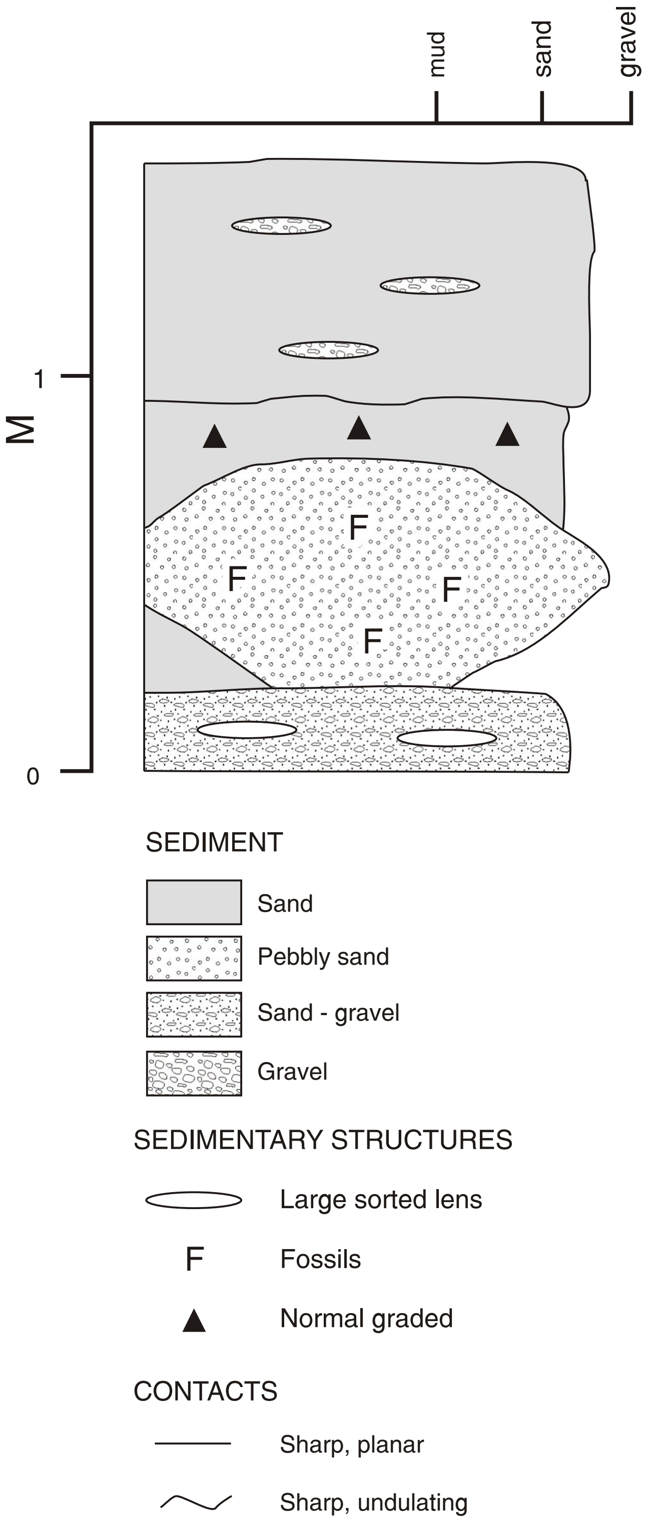


Figure 35. *Stratigraphy of an exposure in the Wild Cove valley.*

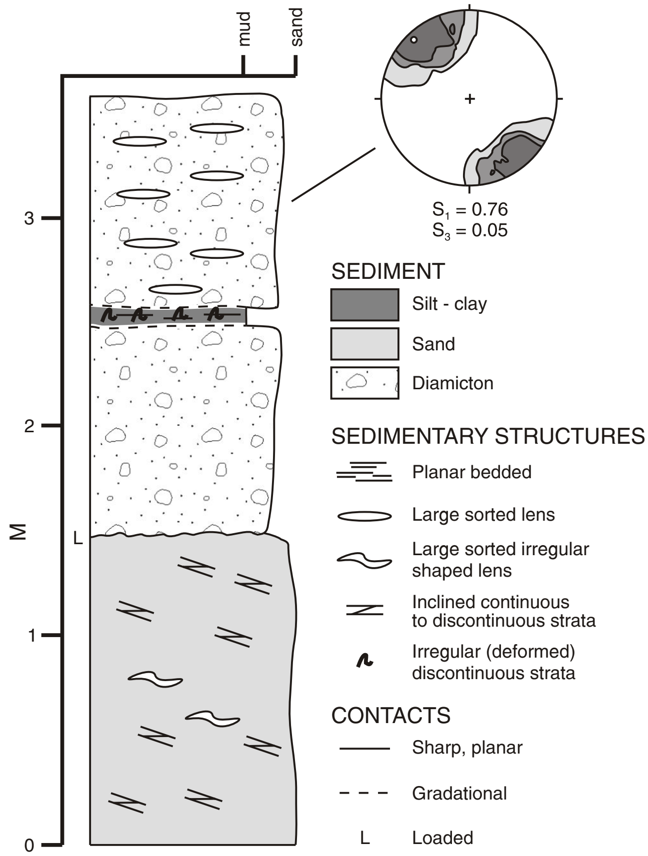


Figure 36. *Stratigraphy of a backhoe-pit exposure in the Wild Cove valley.*

tional stoss ripples of similar size were also found. Angle of climb is low (~6°). The ripples are draped by 0.1 to 0.3 cm thick very fine sand to silt laminae. In places, the draped laminae are contorted or, rarely, discontinuous (Plate 28).

The remaining 420 cm of the south face is a repetitive sequence of rippled and crossbedded sands, planar bedded fine to coarse sands, and gravelly sands, occasionally containing interbedded coarse sands and granule gravel.

Ripples commonly are composed of moderately to well-sorted fine to medium sand, showing little vertical variation in texture. Where measurable, ripples are generally small (λ = 5.5 to 16.5 cm), have low to moderate ripple indices (7 to 11), and erosional stoss to depositional stoss climbing ripples; angle of climb varies between 2 and 16°. The ripples are oriented between 20 and 270°, being more variable toward the base of the section (65 to 250°), and consistently toward 240 to 270° in the upper part. Upper parts of ripples are commonly eroded.

Rippled beds are commonly draped by silt to very fine sand. Drapes are 0.1 to 2.0 cm thick, rarely thicker over ripple troughs, and are either ungraded silt to very fine sand laminae, or normally graded, rhythmically bedded silt and fine sand. Rippled beds commonly truncate underlying sediment. Draped laminae commonly show soft sediment deformation structures, including flame structures, chaotic and discontinuous strata. The flame structures are commonly less than 1 cm high, and angled down flow (Plate 29). Discontinuous beds, where found, are commonly on the downflow side of ripple peaks.

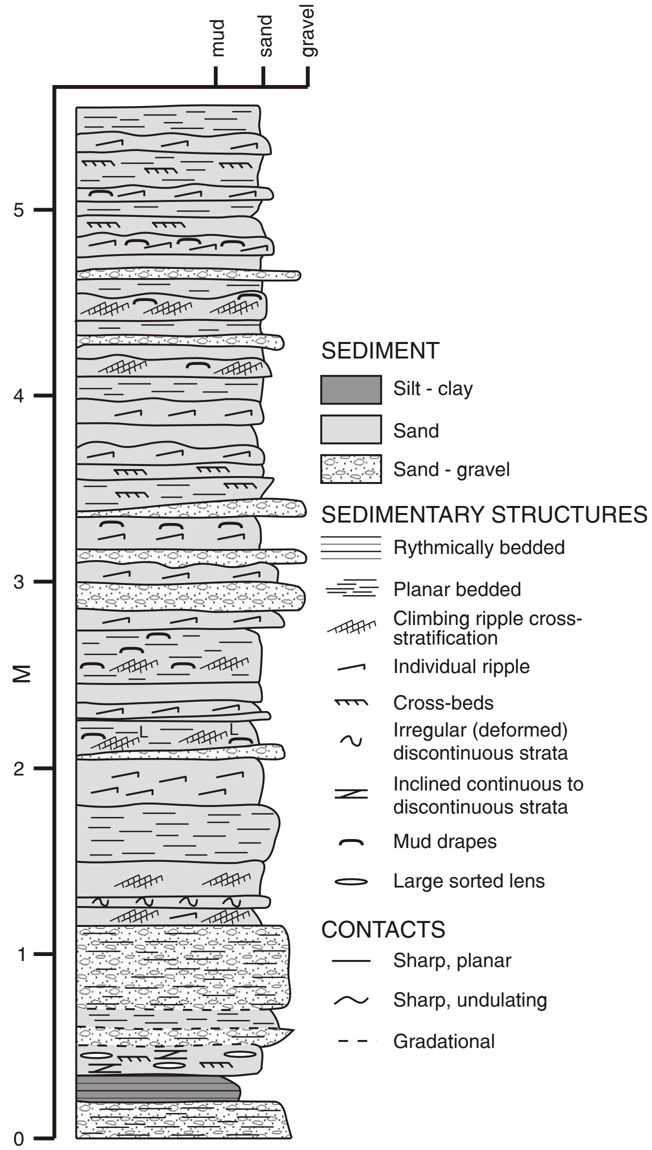


Figure 37. *Stratigraphy of the southern exposure in the Hughes Brook pit.*

Pebbly sand beds are moderately sorted, have a fine to medium sand matrix, and have granule to pebble clasts randomly distributed throughout the unit. They are normally graded to ungraded, 1 cm to 13 cm thick, and are generally confined to the upper 300 cm of the section. Basal contacts are sharp, planar and horizontal. Gravelly sand beds are thicker (up to 44 cm), coarser than the pebbly sand beds, and commonly contain interbeds of normally to ungraded, well-sorted coarse sand and open-work granule gravel.

Planar laminated fine to coarse sands have sharp, flat contacts between laminae. Individual sand strata are 0.1 to 0.5 cm thick, and coarser laminae are 0.2 to 1.0 cm thick. Units of laminated sands commonly coarsen upward.

Individual beds are inclined. The degree and direction of inclination varies between 16° toward 160° at the base of the section, to 8° toward 230° in the middle, and 8 to 14° toward 225° at the top. Although most beds are planar and can be traced laterally for greater than 1 m, some beds are truncated along sharp contacts by troughs.

West Face

The west face is a 600-cm-thick exposure, the bottom 150 cm of which is obscured. The rest consists of interbedded pebbly sands, pebbly gravels and sands (Figure 38).

The lowest exposed unit is at least 45 cm of gravelly sand, composed of moderately sorted fine to medium sand (60 percent matrix) with less than 5 percent silt–clay. Clasts are subrounded and up to 1.5 cm diameter. The unit contains 0.5- to 1.0-cm-thick laterally continuous (> 2 m) interbeds of normally graded, granule gravel to coarse sand having sharp, planar lower contacts. Granule gravel beds are commonly open-work. Individual beds within this unit are inclined at ~24° toward 340°. This unit is overlain by 18 cm of moderately sorted, normally graded, coarse sand to granule gravel, containing 80 percent sand, across a sharp, planar contact. A sharp, undulating contact separates this bed from 3 to 15 cm of normally graded, open-work granule gravel to pebble gravel bed that occupies a northward-thickening trough. This bed is overlain across a sharp, planar contact by a 97-cm-thick planar laminated sand, containing 1 to 10 cm thick (mostly 1 to 3 cm) interbeds of pebbly sand. Individual beds are inclined at 25° toward 340°.

A 77-cm-thick sand unit is next and is composed of 0.2- to 1-cm-thick interbedded fine, medium and coarse sand. Individual beds are ungraded having sharp, planar lower contacts. Bed inclination varies from 25° toward 310° at the base of the unit, to 18° toward 325° in the middle, to 26° toward 315° at the top. One 30-cm-thick sand bed showed poorly defined ripples having no clear asymmetry. Individual beds are commonly continuous laterally for greater than 2 m, although a lens truncates some beds. The lens is convexo–planar having sharp lower contacts, 11 cm wide and 1.5 cm high, and containing a core of granule gravel, flanked by moderately sorted coarse sand.

The sands are overlain across a sharp, planar contact by 30 cm structureless pebbly sand, with ~90 percent matrix and rare clasts up to 3 cm diameter. This bed is overlain across a sharp, planar contact by 18 cm clast supported, normally graded sandy gravel, with ~10 percent matrix and granule to pebble clasts up to 3 cm diameter. The uppermost inclined bed is a 60-cm-thick bed of open work, normally graded, interbedded pebble gravel to granule gravel; lower contacts are sharp and planar. The unit contains subrounded clasts, of mixed rock types up to 5 cm diameter. Clasts are commonly oriented parallel to bedding. Beds are inclined at 25° toward 240°.

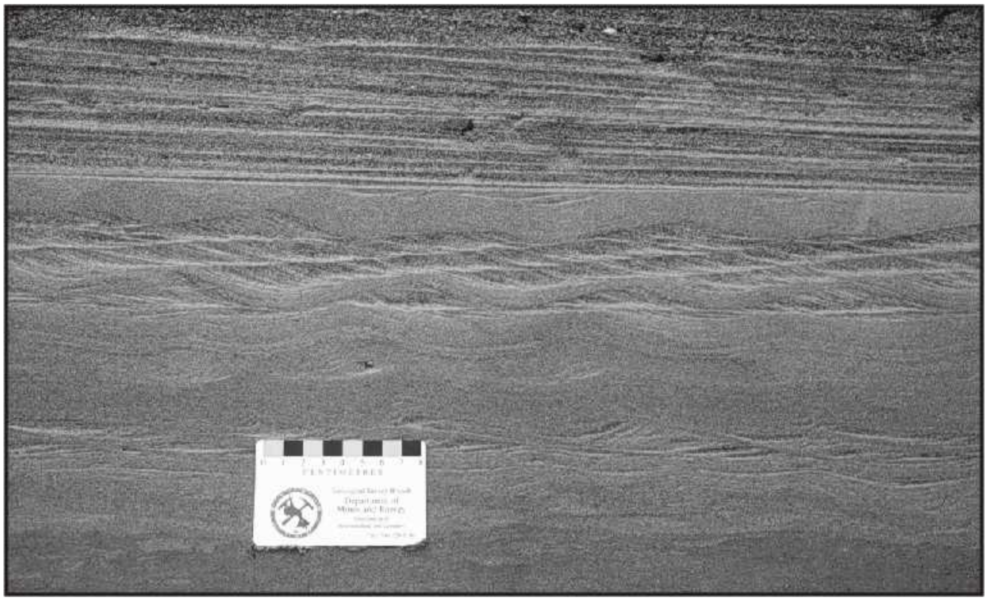


Plate 27. *Climbing ripple cross-stratification having a low angle of climb, found within sand beds, on the south side of the Hughes Brook pit.*

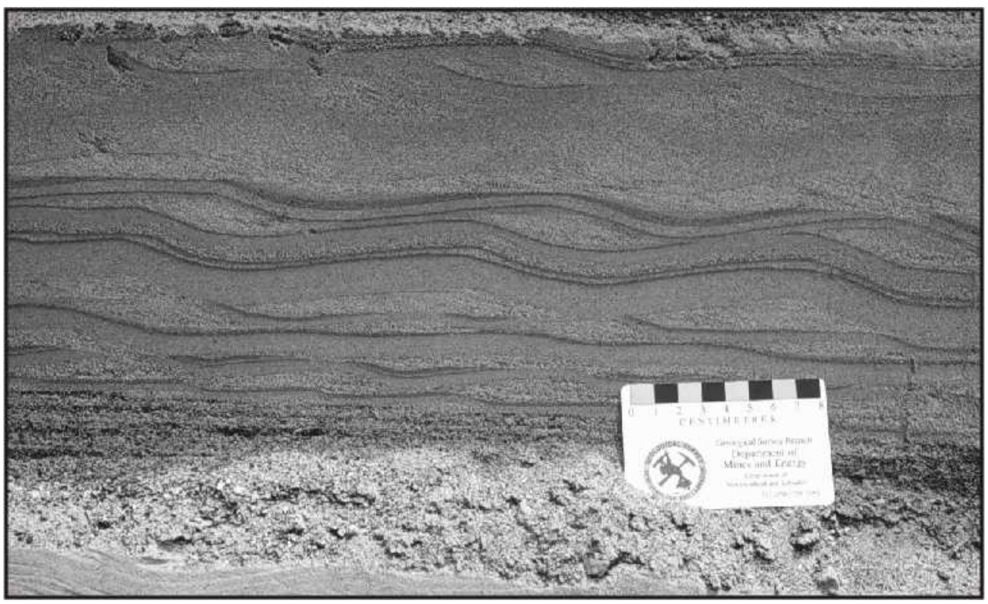


Plate 28. *Draped ripples within sand beds on the south side of the Hughes Brook pit.*

The inclined beds are truncated along a sharp, planar horizontal contact by 65 cm sandy gravel. This uppermost unit lies entirely within the soil profile. It is structureless with a sand matrix, and subrounded, mixed rock type, granule to boulder clasts up to 13 cm diameter of mixed rock types. Clast long axis is commonly oriented toward ~255° (i.e., perpendicular to the axis of Hughes Brook), with generally flat dips. The surface has been grubbed-off during pit development. From the presence of roots in the soil profile, it is estimated that 10 to 20 cm of soil has been removed from this site.

Interpretation – South Face

The sand dominated units that compose the entire south face are interpreted as mid-delta foresets deposited by a combination of current flow and sediment gravity flow.

Climbing-ripple crosslaminae (Ashley *et al.*, 1982) are common throughout the section. Ripple texture and angle of climb indicates fluctuating rates of ripple migration relative to ripple aggradation rates. Climbing-ripples have been described from fluvial and delta environments (e.g., Sorby, 1859; Jopling and Walker, 1968; Picard and High, 1973; Allen, 1984; Ashley *et al.*, 1985). Draped laminations (Gustavson *et al.*, 1975) are produced during waning flows. Ashley *et al.* (1982) demonstrated experimentally the production of drapes from suspension settling over inactive ripples. This mechanism was postulated by Allen (1963), McKee (1965), Gustavson *et al.* (1975) and Hunter (1977), although challenged by Jopling and Walker (1968) and Banerjee (1977) who preferred formation during periods of low flow (less than 10 cm sec^{-1}). The presence of climbing ripples, commonly capped by draped laminae, and deposited on inclined beds is typical a mid-delta depositional environment where sediment and current are supplied by underflows. Most drapes were supplied by overflow–interflow, because they produce beds of constant thickness over the ripples. Some however, show thinning over ripple crests and thickening over ripple troughs. These drapes were likely deposited by underflow (e.g., Ashley *et al.*, 1985).

Normally graded to ungraded, pebble sand beds, deposited on slopes inclined about 8 to 16° also indicate a delta depositional environment. These pebbly sand beds are interpreted as sediment gravity flows, from either grain flow or high density turbidity currents (cf. Kuenen, 1950, 1966a,

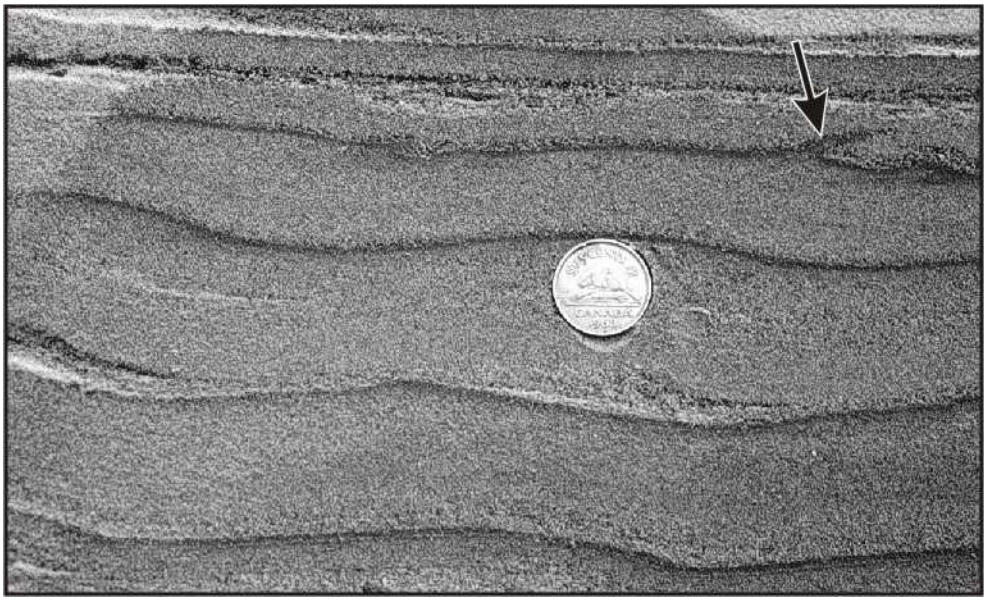


Plate 29. *Flame structures (indicated by arrow) on ripple surface from sand beds on the south side of the Hughes Brook pit.*

b; Kuenen and Migliorini, 1950; Middleton, 1967; Lowe, 1976, 1982; Collinson and Thompson, 1982; Leeder, 1982; Allen, 1984). Sediment was deposited rapidly, supported by the presence of soft sediment deformation structures (Collinson and Thompson, 1982; Allen, 1984).

Interpretation – West Face

The interbedded sands and gravels that characterize the west face are interpreted as upper delta foresets of a typical 'Gilbert-style' delta (e.g., Gilbert, 1890; Nemec and Steel, 1988; Colella and Prior, 1990; Prior and Bornhold, 1990; Postma, 1995). Individual beds are inclined at between 18 and 26°, and were deposited mostly by grainflow and avalanching down the delta front. Clast long axes oriented parallel to bedding support this interpretation. The sharp contact between beds shows episodic accretion of sediment. Although most beds are planar, some are truncated by channels that interpreted as ephemeral distributary channels on the delta. Channel migration explains the presence of sandy interbeds in the delta foresets. Progradation of the delta into the valley, and changes in sediment entry points also are shown by changes in the direction of bed inclination, from ~315° in the central part of the face, to 240° near the top.

The uppermost sandy gravel unit is interpreted as a fluvial sediment, and thus forms delta topsets. It truncates the underlying inclined beds along an erosional contact. The unit contains no noted current flow indicators. Given the location of the delta extending into the Hughes Brook valley, it is likely the sandy gravel unit was deposited by current flow down Hughes Brook.

Discussion

The sediments exposed in Hughes Brook pit were deposited in a delta produced by a stream entering the Hughes Brook valley on the east side. Only the western and southern parts of the delta currently are exposed, showing steeply dipping, gravelly upper foresets and sandy mid-foresets, respectively.

The surface morphology of the delta, grading upstream into fluvial sediment, shows this was not an ice-contact delta. Sediment deposited in standing water adjacent to a glacier commonly consist of coarse-grained outwash deposited close to ice at the mouths of ice tunnels (Rust and Romanelli, 1975; Ashley *et al.*, 1985). Sedimentation rates are commonly high in these environments and syndepositional collapse features, as the result of the melting out of buried ice blocks, are common. Similarly, ice-proximal lacustrine or marine sediment commonly exhibits abrupt lateral and vertical changes in texture, as the result of constant shifting in the point sources of sediment input. Diamicton, deposited from debris flows, is also common (e.g., Lawson, 1982). Finer grained sediments commonly contain dropstones deposited from floating ice (e.g., Thomas and Connell, 1985). None of these features are found in the Hughes Brook delta, suggesting it was distal to the ice front. However, sediment and water discharge to the delta was likely controlled by melting ice on the highlands. The modern stream occupies a small, narrow valley with discharge buffered by three lakes within the drainage basin. Following emergence of the delta during postglacial isostatic rebound, the stream was unable to incise through the delta and was instead deflected northward around it.

The delta was formed by fluvial discharge into a lake or the ocean. Formation in a lake would require the presence downstream of an ice dam in order to impound standing water. Eventual draining of a proglacial lake in this area should have produced drainage channels. The existence of the delta only 5 km upstream of the modern coast suggests that formation adjacent to a higher postglacial sea is more likely.

The existence of a marine delta in the Hughes Brook valley having a surface elevation of 61 ± 2 m asl requires a revision of the postglacial sea-level history for the area.

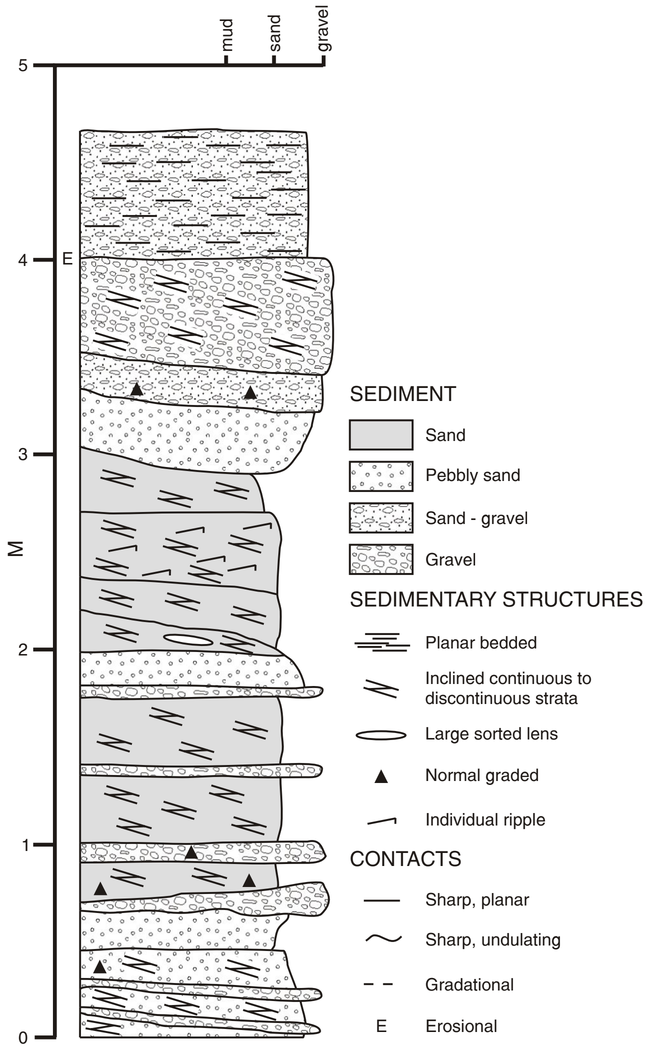


Figure 38. *Stratigraphy of the western exposure in the Hughes Brook pit.*

Brookes (1974) suggested a marine limit at the head of the Humber Arm of approximately 49 m asl, based on the presence of a delta at Humbermouth extending on both sides of the modern Humber River. Other supporting evidence for a higher marine limit is fragmentary. A possible delta (Site 91029: Appendix 1) exists near the community of Hughes Brook at 58 m asl, and a terrace (Site 91008: Appendix 1) at ~61 m asl was identified at Humbermouth, on the opposite side of the Humber River from Dawe's Pit.

Discussion: Deposition at the Head of the Humber Arm

During deglaciation, inflow of water and sediment into the head of the Humber Arm was from three major valleys, Humber River, Wild Cove, and Hughes Brook. Each shows evidence of fluvial, marine, or ice proximal sedimentation.

Primary basal till, found at the head of the Humber Arm (Site 94003: Appendix 1), has a reddish brown matrix and red sandstone clasts both of which were derived from the Carboniferous rocks of the Deer Lake basin. Clast fabric is strong (S_1=0.78, S_3=0.06) having a preferred clast orientation, showing flow from the adjacent Humber River gorge. A diamicton exposure on the north shore of Wild Cove (Site 91014: Appendix 1) shows a dark greyish brown, primary basal till with a strong clast fabric (S_1=0.64, S_3=0.16) and a preferred clast orientation indicating flow from the Wild Cove valley. These two sites indicate that ice entered the Humber Arm via the Wild Cove and Humber River valleys. Only small exposures of till were present in the Hughes Brook valley.

During deglaciation, melting ice in the gorge produced a large ice contact delta at Humbermouth (Brookes, 1974), shown by a flat-topped, steep sided feature composed of sand and gravel having a surface elevation of 50 m asl. Sedimentary evidence for this is fragmentary, following aggregate extraction on the south side of the Humber Arm, and the development of a cemetery on the north side. The internal structure of a feature in the Hughes Brook valley having surface elevation of about 60 m asl, was identified as a delta formed by fluvial input from a tributary valley. Similarly, a delta was identified at the head of Wild Cove (50 m asl), also on the basis of internal structure (Site 91221: Appendix 1).

The elevation of delta surfaces suggests the marine limit was about 60 m asl. A postglacial sea flooded the lower reaches of Hughes Brook valley, and the Wild Cove valley, as shown by the delta, subaqueous fans, and the thick silt–clay covering the valley floor. Marine shells found near Steady Brook record marine inundation through the Humber River gorge. Radiocarbon dating of marine shells in the Humber River gorge and Wild Cove provides a minimum date for marine inundation and thus deglaciation at 12 200 – 12 500 BP.

The re-establishment of subsequent fluvial sedimentation in the lower reaches of the Humber River valley is shown by sediments in Dawe's Pit, the Humber River gorge, and in the Hughes Brook valley. There has been minimal fluvial sedimentation in the Wild Cove valley following deglaciation, and the emergence of the valley floor from below sea level.

SECTIONS ON THE SHORES OF GRAND LAKE

Grand Lake is the largest lake in insular Newfoundland, with a surface elevation of ~82 m. The shoreline shows a distinct contrast between the west and east shore. The west shoreline is bedrock dominated, with scattered outcrops of sand and gravel mostly found at the mouths of small tributaries. In contrast, the east shore is dominated by Quaternary sediment, and rare bedrock outcrops.

On the east shore, Quaternary sediment occupies a narrow belt, between 500 and 1300 m wide, increasing in width northward. Coastal bluffs (Plate 30) are commonly separat-

Plate 30. *Lakeshore exposure of a fan-delta on the east shore of Grand Lake.*

ed from the lake by a gently sloping, cobble to boulder beach, up to 50 m wide. Prevailing winds are parallel to the lake orientation, whereas the effects of those oriented across the lake are minimal because of protection from the surrounding highlands. Hence, there is little evidence of wave impact on the bluffs and sediment exposure is poor and obscured by slope failure.

Thirty-two lakeshore exposures, which had little vertical or lateral continuity, were examined between Harrys Brook and Howley. Descriptions are presented from three of the best exposed sections, representing the range of sediment types found.

Grindstone Point Section

The section is located about 1200 m north along the coast from Grindstone Point (Site 93009: Appendix 1). The section is about 22 m high, is poorly exposed and has 4 m of Carboniferous red sandstone (of the Little Brook Formation) at its base (Figure 39). The overlying 12 m is obscured, and only the upper 6 m of the section is well exposed. The lowest exposed unit is at least 1 m of moderately sorted (s.d. 0.8ø), planar bedded fine, medium and coarse sand (1 percent granules, 98 percent sand and 1 percent silt; mean grain size 2ø). Beds are 1 to 5 cm thick, normally graded to ungraded, and laterally continuous for more than 2 m; contacts between beds are commonly sharp. The unit contains rare subrounded cobble clasts that disturb the bedding (Plate 31). Sand beds characteristically are compressed beneath the clasts. Clasts comprise less than 10 percent of the unit and range from granule to cobble size, and are randomly distributed throughout.

The sand unit is overlain across a sharp, planar contact by 4.5 cm of laminated clayey silt and coarse sand. Clayey silt laminae are about 0.5 cm thick, normally graded to ungraded and have sharp lower contacts. The coarse sand beds are 1 cm thick, well sorted, normally graded and have sharp lower contacts. There are three clayey silt laminae separated by three sand beds. A single channel sample for this unit showed a very poorly sorted sediment composed of 60 percent sand, 37 percent silt and 3 percent clay and having a mean grain size of 3.6ø.

The sand–silt bed is overlain across a sharp, planar horizontal contact by a 4-m-thick unit of sandy gravel. The unit is crudely stratified and matrix supported. The matrix is medium to coarse sand, supporting subrounded, granule to boulder clasts up to 30 cm diameter and numerous cobbles. Rhyolite (36 percent), basalt (24 percent), granite (17 percent), tuff (14 percent), minor porphyry (7 percent) and gneiss (2 percent) comprise the rock types. Clasts long axes are commonly oriented downslope (at ~ 20° toward 290°).

A 45-cm-thick unit of interbedded granule gravel and pebbly sand overlies a gradational, planar contact. Five granule gravel beds were noted, each greater than 2 m in lateral extent. They were 1 to 2 cm thick, moderately to well sorted, normally graded (2 beds) to ungraded (3 beds) having undulating, gradational lower contacts. Pebbly sand beds are 5 to 12 cm thick, have a poorly sorted, fine to medium sand matrix and contain granule to pebble clasts up to 1.5 cm diameter. Four pebbly sand beds were found, all of which were internally structureless.

A 12-cm-thick bed of structureless, well-sorted coarse sand overlies a sharp, planar contact, which is overlain by a 50-cm-thick bed of pedogenically modified sandy gravel, that forms the top of the section.

Sediments exposed in the Grindstone Point section are interpreted to have been deposited in standing water, by a combination of sediment gravity flow and suspension settling.

The basal planar-laminated sand, and overlying laminated clayey silt and coarse sand are interpreted as having been deposited by sediment gravity flow, likely grain flow

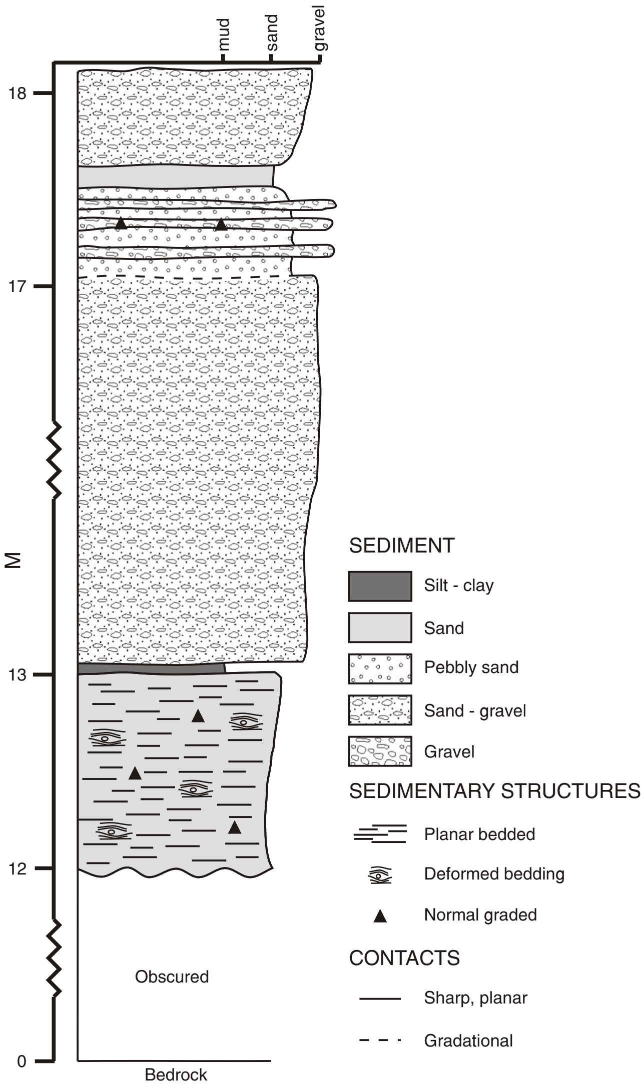


Figure 39. *Stratigraphy of an exposure at Grindstone Point, Grand Lake.*

(cf. Lowe, 1976, 1982). The thin, commonly ungraded beds that have sharp lower contacts, suggest deposition from separate pulses of sedimentation. Periods of no flow are indicated by the normally graded beds of clayey silt, deposited by suspension settling. Sedimentation was thus initially by grain flow and subsequently by suspension settling. There are no flow structures suggesting current flow, and the bed geometry and lateral continuity of individual strata are compatible with deposition by turbidity current (Bouma, 1962; Ashley, 1988). The downfolding of sand beds beneath cobble-size clasts that deform underlying sands–silts are interpreted as dropstones. Thomas and Connell (1985) described similar features, indicating that ice was in contact with at least part of the basin.

The crudely stratified sandy gravel bed is also interpreted as having been deposited in standing water. Clasts lying conformable to bedding are interpreted as clasts moved by grain avalanching down bedding plane surfaces (Wadell, 1936; Johansson, 1963; Allen, 1984; Ashley *et al.*, 1985). The sediment source was from the adjacent hills. Clast assemblages from Grindstone Point show mostly rhyolite, basalt derived from the Springdale Group (units Ssf and Ssm), porphyry and some granite from the Topsails intrusive suite (units Sq and Sm), and coarse grained pink granite from the Hinds Brook granite (Whalen and Currie, 1988), all of which are found on the hills above Grindstone Point. The sediments exposed at the Grindstone Point section were all deposited in standing water. The laminated sand–clayey silt, lack of current flow structures indicating fluvial transport, and the presence of dropstones all support this conclusion.

Little Pond Brook Section

This is the southernmost section examined and is located about 2.1 km south of the mouth of Little Pond Brook (Site 93013: Appendix 1). It is 16-m-thick, but only small areas are well exposed (Figure 39). The lower 4.5 m of the section is obscured, above which is a small area with 100 cm lateral and 142 cm vertical extent. It is composed of interbedded sand, gravelly sand, and diamicton. The central 8 m of the section is obscured, although small exposures show gravelly sands and interbedded fine, medium and coarse sand beds. The upper 2 m is composed of rhythmically bedded sand–silt couplets, overlain by sandy gravel and gravelly sand.

The base of the lower 4.5 m is a planar stratified sand having a minimum thickness of 30 cm. Individual strata are moderately sorted, fine, medium and coarse sand, 0.5 to 1.0 cm thick, ungraded and contain rare pebble clasts; the lower contacts are sharp and commonly undulating.

A 15-cm-thick bed of structureless, medium to coarse grained sand having rare granule to pebble clasts, up to 1.5 cm diameter, overlies a sharp, undulating contact. This unit is overlain across a sharp, undulating contact by a 10-cm-thick bed of reverse graded, pebbly sand, with a medium to coarse sand matrix (85 percent matrix), and granule to pebble clasts up to 2 cm diameter.

A 7-cm-thick, structureless, fine to medium sand bed overlies the pebbly sand bed above a sharp, undulating contact, and this in turn is overlain by a diamicton. The lower contact is sharp and undulating, and the diamicton bed is up to 25 cm thick, pinching out to the north, over 100 cm. The diamicton is structureless, has a fine sand to silt matrix, and has subangular to subrounded, granule to cobble clasts, up to 20 cm diameter. A 40-cm-thick bed of structureless gravelly sand having a medium to coarse sand matrix, and subangular to subrounded, granule to cobble clasts up to 20 cm diameter, overlies a sharp, undulating contact. This unit is overlain by 5 cm diamicton across a sharp, undulating contact. The diamicton has similar characteristics to that underlying the gravelly sand. This is overlain by 5 cm structureless, poorly sorted, dark reddish brown (5YR 3/4, moist) to

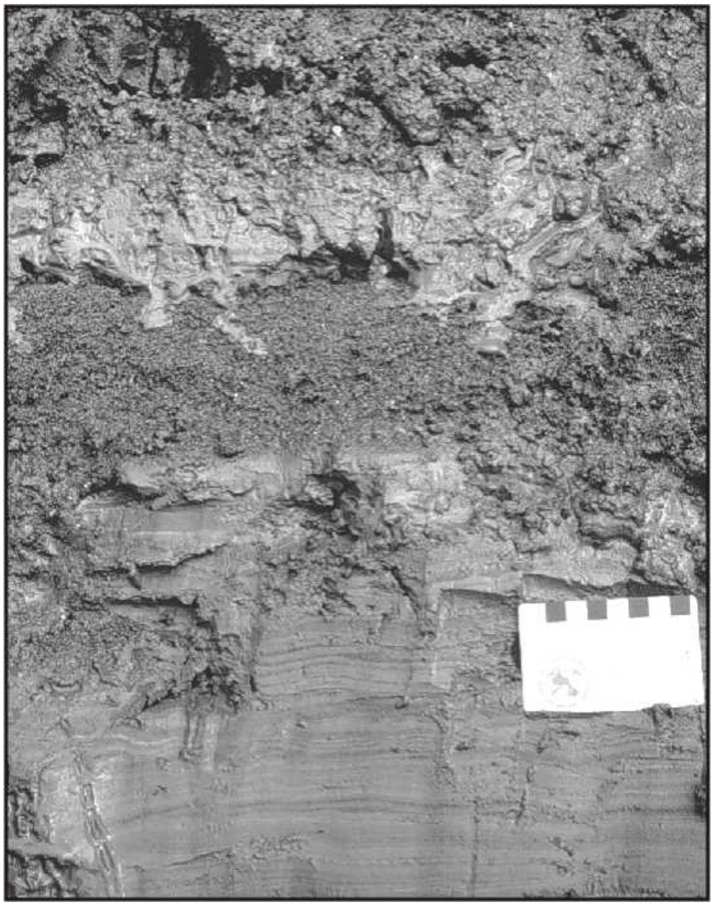


Plate 31. *Sand–silt rhythmites within the Little Pond Brook section on Grand Lake.*

pinkish grey (7.5YR 7/2, dry), silt (8 percent sand, 90 percent silt, 2 percent clay) having a mean grain size of 5.6ø (Sample 934014: Appendix 1).

The upper 200 cm, including the soil profile is well exposed (Figures 39 and 40). A stratigraphic log shows a basal unit of at least 10 cm of structureless gravelly sand, with a medium to coarse sand matrix, and granule to cobble clasts. It is overlain by a 3-cm-thick bed of well sorted, structureless coarse sand, and 6-cm rhythmically bedded very fine sand to silt, both across sharp, planar contacts. The rhythmites consist of planar and horizontal, 0.1 to 0.5 cm thick, normally graded laminae; there are twelve rhythmites. Contacts between laminae are sharp, and planar. Grain size analysis from a small area covering several silt–sand laminae shows a sediment composed of 24 percent sand and 76 percent silt and having a mean grain size of 4.9ø (Sample 934015: Appendix 1). Individual beds are inclined at about 10° toward 320°.

The silt–sand unit grades upward into 5 cm layer of structureless medium sand. This is overlain across a sharp contact, by a 5 cm layer of clast-supported sandy gravel having a coarse sand matrix, and granule to pebble clasts. Open work granule gravel lenses occur throughout the unit. These beds are overlain by a sequence of very fine sand–silt rhythmites (5 cm), structureless medium to coarse sand (7 cm), and sand–silt rhythmites (25 cm) (Plate 32). These beds dip at ~10° toward 320°. They are overlain by 20 cm structureless, matrix-supported, sandy gravel having a medium to coarse sand matrix, and granule to pebble clasts up to 5 cm diameter.

The section is capped by a 120-cm-thick bed of structureless gravelly sand having a medium sand matrix, and granule to cobble clasts up to 20 cm diameter; much of this unit is within the soil profile.

The sediment exposed within the Little Pond Brook section have been deposited within standing water and have characteristics similar to those exposed at Grindstone Point. Planar laminated sands at the bottom of the exposure are sediment gravity flow deposits, likely from grain flow (cf. Lowe, 1976, 1982). The thin, commonly ungraded beds that have sharp lower contacts, suggest separate pulses of sedimentation. Reverse graded beds are formed by grain avalanching on slip faces (Bagnold, 1954; Allen, 1984), and are typical of grain flows (Lowe, 1976; Walker, 1984). Rhythmically bedded sand–silt are interpreted to have been deposited by suspension settling from overflow–interflow. There are no structures to indicate current flow within the exposure. Matrix-supported sandy gravel beds found in the upper part of the section suggest deposition by sediment gravity flow rather than current flow.

The Little Pond Brook section contains diamicton beds that are thin, structureless and pinch out laterally above sharp contacts. They are commonly overlain by sands or sandy gravels. Their characteristics and stratigraphic relationships indicate that these diamicton beds are derived from debris flows within a subaqueous depositional environment. Diamictons interbedded with sand and gravel, with bedding dipping towards modern Grand Lake suggests deposition within an ice-proximal, deltaic (?) environment (e.g., Middleton and Hampton, 1976; Lawson, 1982; Ashley *et al.*, 1985; Nemec and Steel, 1988; Colella and Prior, 1990; Lønne, 1995).

Alder Brook Section

Several sand and gravel pits alongside the road leading to the hydroelectric station at Hinds Lake have been graded and no exposures remain. A section located 800 m south of Alder Brook bridge (Site 93018: Appendix 1) has a surface elevation of 125 m asl. The pit has a 15 m face, most of which is obscured by slumping; however, the upper 3.6 m was cleared for examination (Figure 41). Most of the exposure comprises a monotonous sequence of planar laminated sand, capped by 80 cm of planar bedded pebbly sand, and sand and gravel.

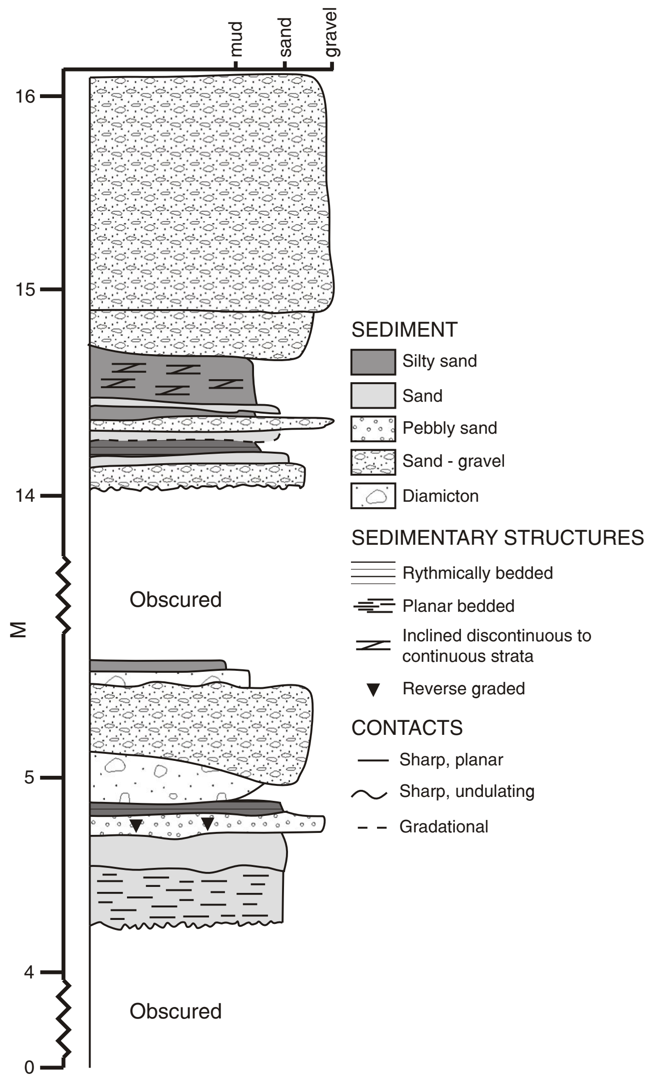


Figure 40. *Stratigraphy of an exposure at Little Pond Brook, Grand Lake.*

The lower 2.8 m of the exposure is mostly planar laminated to planar bedded sand. All beds are laterally continuous over 2 m, and are inclined at 10 to 14° toward ~275° (i.e., downslope). Textural analysis shows a moderately sorted (s.d. 0.8ø) sediment, composed of 99.8 percent sand and 0.2 percent silt–clay having a mean grain size of 2.3ø (Sample 934020: Appendix 1). There are 161 strata within the exposure, most separated by sharp, undulating contacts. Individual beds are ungraded to normally graded medium–fine, medium–coarse and coarse sand. Coarse sand strata are thin (0.2 to 0.5 cm) and confined to the lower 20 cm of the exposure. Medium to coarse sand strata range in thickness from 0.2 to 4.0 cm, are mostly ungraded or normally graded, although rare reverse graded strata were noted low in the exposure. The medium to coarse sand strata are interstratified with 0.2 to 5.5 cm thick, ungraded, medium to fine sand layers. The only interruption to the rhythmic stratification is a 5-cm-thick bed of ungraded pebbly sand containing subrounded pebble to granule clasts up to 4 cm diameter, found near the base of the exposure. It overlies a medium to fine sand lamina across a sharp (erosional?) contact.

The upper 80 cm of the exposure is planar bedded sand, pebbly sand and gravelly sand. Coarser beds are normally graded, and contain subangular to subrounded, granule to cobble clasts up to 12 cm diameter. Clasts are commonly flat lying and have axes conformable to bedding. Coarse sand accumulations are commonly found on the upslope side of clasts. Clast rock types include granite, rhyolite, basalt, gabbro, sandstone, and shale.

The Alder Brook section contains the least variety of sediment types of the three sections described, being largely composed of a repetitive sequence of inclined beds of planar laminated sand. These sands represent grain flows into a body of standing water. Laminae are thin, commonly ungraded and have sharp lower contacts, suggesting they were deposited incrementally. Reverse graded beds are formed by grain flow avalanching on slip faces (Bagnold, 1954; Allen, 1984). The uniformity of grain size and bed geometry suggest that rates of sediment input were consistent during deposition of this sediment. Dropstones and diamicton beds deposited by debris flow found at the Grindstone Point and Little Pond Brook exposures, indicated ice-proximal conditions. During deposition of sediments at Alder Brook, these dropstone and diamictons were not found here, and that may indicate that ice was distal to this part of the basin. Individual beds dip toward modern Grand Lake, although angles of dip are low (10 to 14°). The grain size, lateral continuity of beds, and the low angle bedding indicates deposition in a subaqueous fan.

Generally, the sequence coarsens upwards, suggesting either increased sediment input or increasing proximity to sediment source. The source of the sandy gravel was likely from the overlying hillside. Clasts in the upper part of the Alder Brook section are derived from the Topsails intrusive suite (units Sp, Sm and Sq), the Springdale Group (units Ssf and Ssm), the Hungry Mountain Complex and Ordovician granites (units Oib and Oic), all of which are found on the hills above the section.

Discussion

During a period of higher water level in the Grand Lake basin, sediment exposed within the Grindstone Point, Little Pond Brook and Alder Brook sections were deposited in a subaqueous environment. Sediments at Little Pond Brook and Alder Brook were deposited in a fan or deltaic environment. There is evidence for ice-distal and ice-proximal sedimentation in these exposures.

Although the Grindstone Point, Little Pond Brook and Alder Brook exposures are best along the shoreline, twenty-nine other sections were examined (Figure 42). Many of the characteristics described from these three sections are found in sediments from the other exposures.

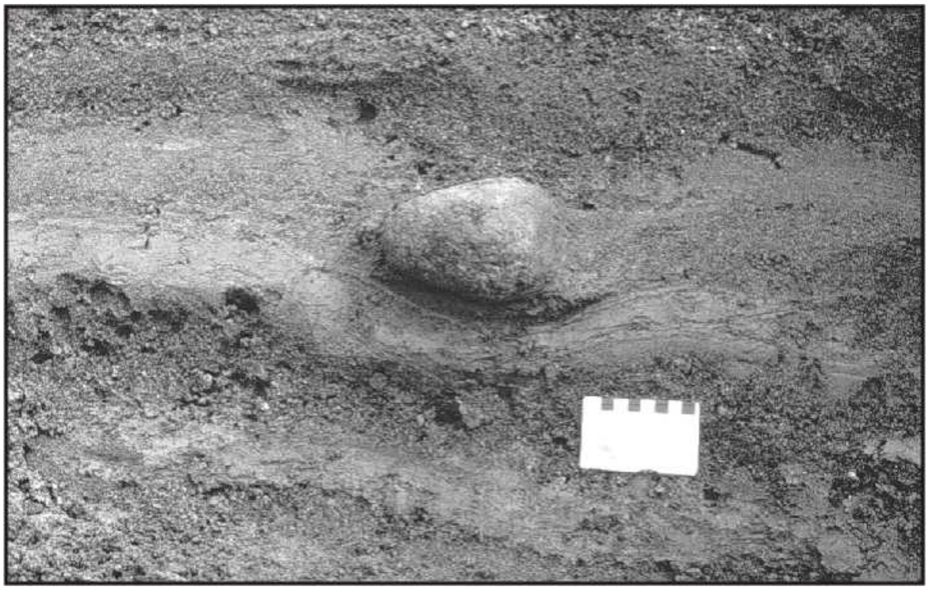


Plate 32. *Deformed bedding beneath dropstone in the Grindstone Point section. This is evidence for a proglacial lake in the Grand Lake valley.*

Silt and Clay

Rhythmically bedded silt and sand, with or without clay are found at five locations, apart from the Little Pond Brook section.

On the west side of Grand Lake, just north of Thirty-fifth Brook (Site 93015: Appendix 1) (Figure 42), 3 m of interbedded fine sand, silt and clay is found. The sediment is dark reddish brown (5YR 3/3, moist) to reddish grey (5YR 5/2, dry), with one channel sample showing sediment composed of 35 percent sand, 40 percent silt and 25 percent clay, having a mean grain size of 5.4ø. Beds are highly contorted, and overlain across a sharp, undulating contact by gravelly sand. This relationship suggests rapid deposition of the gravelly sand onto a saturated substrate producing soft sediment deformation structures.

On the north shore of Grand Lake, 1.3 km east of Blow Hard Point (Figure 42), a 7-cm-thick bed of contorted silt and clay is found at an elevation of 89 m (Site 92022: Appendix 1). Soft sediment deformation features include isoclinal folds and flame structures, and were induced by deposition of an overlying crossbedded sand unit.

On the east side of Grand Lake, 2.2 km north of Hinds Brook (Figure 42), a 6-m exposure of mostly rippled sands was found on the west side of a well defined, northwest-trending, sand-dominated ridge. The section contains a unit of at least 20-cm thick planar laminated silt and clay having interbeds of sand, extending up to 91 m (Site 92078: Appendix 1). The unit shows 16 normally graded silt–clay laminae; sharp, planar contacts separate individual laminae. Silt beds, within this unit are commonly overlain unconformably by 0.5 to 3.0 cm thick, well sorted, ungraded fine sand strata having sharp lower contacts. The unit is overlain by a bed containing draped ripples. The silt and clay rhythmites were deposited by suspension settling in standing water, and the fine sand interbeds represent grain flows or high density turbidity currents.

About 500 m north of the Hinds Brook dam gate on the east side of the road (Figure 42), an 8-m-high section shows poorly exposed fine sand and silt (Site 93085: Appendix 1). These sediments are exposed in a unit containing laminated fine and very fine sand, and silt. A single channel sample showed a grain size distribution of 12 percent sand, 86 percent silt and 2 percent clay, have a mean grain size of 5.4ø. Laminae are highly contorted and deformed by subrounded, pebble to boulder clasts up to 60 cm diameter. The laminae were deposited by suspension settling in a body of standing water and the clasts were deposited following the laminae, likely by sediment gravity flow. This is suggested by the absence of medium to coarse sand in the matrix and the distortion of beds beneath clasts. The deformed unit is overlain by planar laminated medium, fine and very fine sands.

About 1.4 km north of Blue Grass Brook (Figure 42), a 20 m exposure of mostly interbedded sandy gravel and gravelly sand, contains a 90-cm-thick bed of planar laminated silt and clayey silt (30 cm) grading up into fine sand and silt (60 cm) (Site 92178: Appendix 1). The bed extends up to 99 m. Individual silt–clay lamina are normally graded and separated by sharp, planar contacts. Silt is thicker than clay, and commonly contains 2- to 4-mm thick, planar, well sorted, ungraded fine sand strata. The unit was deposited mostly by suspension settling and by periodic rapid sedimentation of sand from traction current.

Gravel and Sand

Loose gravel and sand beds are found in 24 of 29 sections examined. Units of sand–gravel commonly comprise the bulk of the sections and consist of loose, crudely stratified, normal to ungraded beds of sandy gravel, gravelly sand, open-work granule to pebble gravel and sand. Lenses of stratified sediment are common throughout. Sand strata are commonly thin (< 10 cm), well sorted, ungraded and have sharp upper and lower contacts. All beds are laterally continuous across the sections for at least 2 m. Interbedded sands and sand–gravel commonly are inclined, dipping at between 8 and 26°, and dips consistently increase with stratigraphic elevation. The direction of dip is always toward Grand Lake (Figure 42), suggesting the sediment was derived from the adjacent hills.

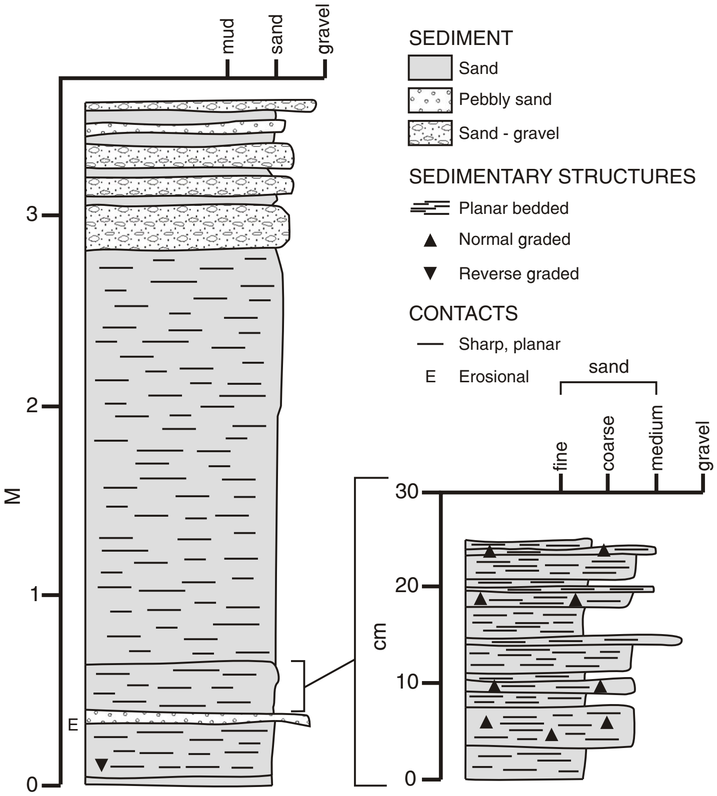


Figure 41. *Stratigraphy of an exposure at Alder Brook, Grand Lake.*

Pebble samples from 22 exposures on the east side of Grand Lake indicate that the clasts are mostly of rock types found on The Topsails, generally with sources directly upslope of the sections. Carboniferous sandstone and siltstone are relatively rare, despite the fact they underlie most of the sections along Grand Lake. Carboniferous clasts are generally confined to exposures on the north and west shore (e.g., Sites 93014, 93023, 93025: Figure 42), or are associated with deposits at the mouths of the larger streams south of Hinds Brook (e.g., Sites 93013, 93016: Figure 42).

Diamicton

Diamicton beds are found in six sections (Sites 92178, 93008, 93011, 93015, 93028, 93029: Figure 42), all on the east side of Grand Lake. Table 14 lists the characteristics of these diamictons. Where diamictons are found interbedded with fine grained sediments or sand–gravel, they have weak, girdle clast fabrics, and irregular-shaped sand lenses, and are debris flow deposits, similar to those at Little Pond Brook. This interpretation applies to diamictons exposed at Sites 93011 and 93015. The close association of diamicton beds with fine-grained sediments or other waterlain deposits suggests they were deposited in a subaqueous environment.

The diamicton exposed at the base of the section at Site 93028 is a debris flow deposit. Clast fabric is moderate, slightly girdled (S_1=0.61, S_3=0.12) having a preferred clast orientation toward 302°. Ice flow in this direction would have crossed gabbro and diorite of the Rainy Lake Complex (Unit Sorl of Whalen and Currie, 1988; Figure 4). The diamicton contains none of these rock types. Instead, clast rock types suggest derivation from the northeast. The presence of large proportions of rhyolite and porphyry suggests a source in the Springdale Group (Unit Ssf) and Topsails intrusive suite (Unit Sqa) respectively. Based on the fabric strength, the clast provenance unrelated to preferred clast orientation, and geographic location of the unit at the base of a steep hill, the diamicton at Site 93028 was deposited by debris flow.

Crossbedded and Rippled Sand

Crossbedded and rippled sands are found in six sections (Sites 92012, 92078, 92082, 92194, 93022, 93024: Figure 42). Most of these are on the north shore of Grand Lake. Current flow indicators are toward the Junction Brook area (Sites 93022, 93024, 92194), toward Howley (Site 92012), or toward Sandy Lake (Site 92082). On the east side of Grand Lake, 2.8 km north of Hinds Brook, a 2-m section shows fine to medium sand, asymmetric erosional stoss climbing ripples (λ=12 cm, H=1.5 cm, R.I.=8) indicating northward current flow. The ripples are commonly draped by 0.5 to 1.0 mm reddish brown silty clay. The draped ripples showing deposition from suspension settling during waning flow conditions, and is consistent with deposition within standing water. Draped ripples are also found at Site 92194, on the north shore.

Discussion: Grand Lake Sections

The sediment and features along the east side of Grand Lake are the result of deltaic or fan-delta deposition. Sand and gravel beds dipping into the lake, rhythmically bedded sand, silt and clay, draped ripples, and geomorphology all support this interpretation. An alternative hypothesis is deposition as paraglacial alluvial fans, similar to those described by Ryder (1971a, b), and Church and Ryder (1972). Rhythmically bedded sediment dropstones and diamicton lenses are not typical of alluvial fans. Alluvial fans should grade toward their source areas, and therefore the elevation of the tops of alluvial fans within a valley should vary. Along Grand Lake, the elevations of the surfaces are consistent.

Flat-topped features at the mouths of larger streams, such as Harrys Brook, Little Pond Brook, Hinds Brook, and brooks north of Grindstone Point having surface elevations of 118 to 157 m asl, are interpreted as deltas (*see* Section on Methodology, page 19). These features show higher water levels in Grand Lake. The development of flat-topped surfaces in these deltas presumably reflects more stable streams, rather than the more episodic discharge onto the fan-shaped deltas north of Hinds Brook.

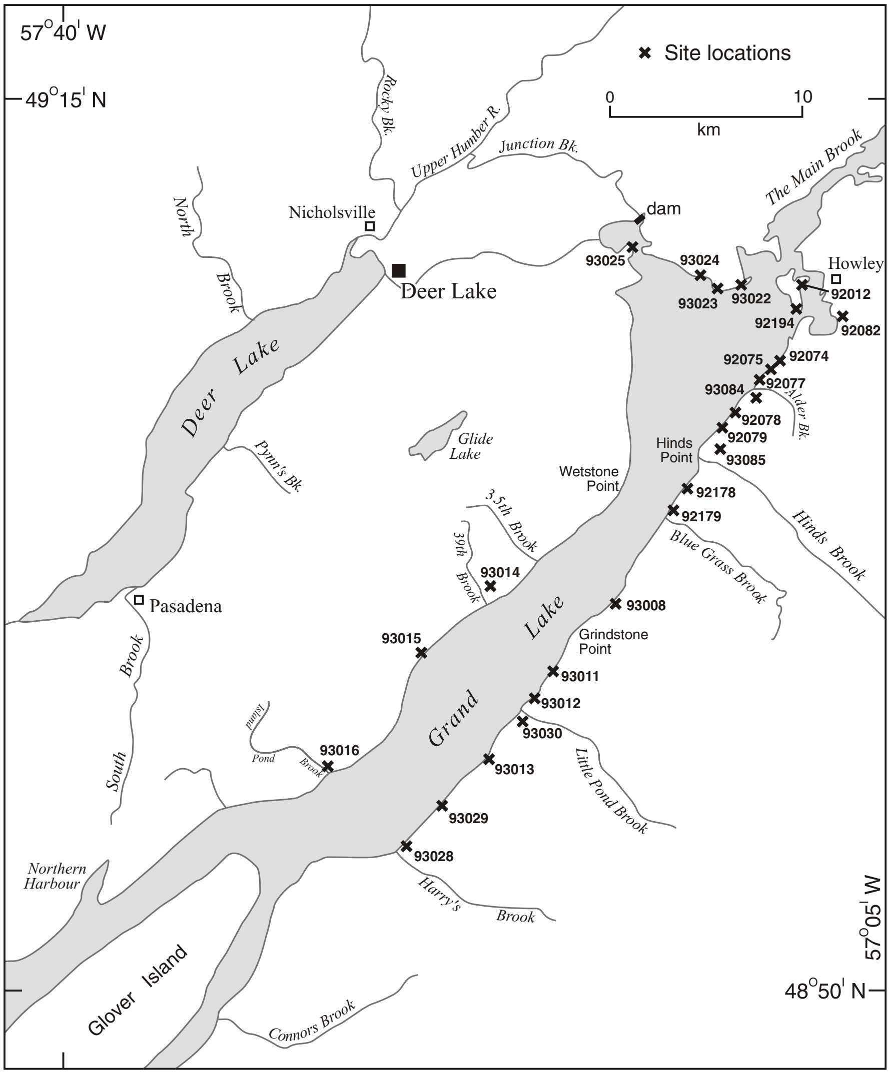


Figure 42. *Location of sections described around the shores of Grand Lake.*

Rhythmically bedded sand, silt and clay, and draped ripples are evidence for standing water conditions at a higher elevation than the surface of modern Grand Lake. These deposits are found up to an elevation of 99 m, 12 m above the modern lake surface. Individually, each of these deposits could be explained as being developed in ice-marginal lake. The distribution of sites containing silt and clay, on the west, east and north sides of the lake, all have similar elevations. Deposition within a single body of standing water is thus considered more likely.

The elevation of sediments and features indicating standing water shows water surfaces up to 150 to 160 m. This is substantially above the marine limit of 60 m defined for the Humber Arm in this report, and elevations derived from other coastal areas of western Newfoundland (e.g., Liverman, 1994; Grant, 1980; Brookes *et al.*, 1985). The deltaic deposition was therefore associated with a lacustrine environment. The orientation of inclined beds, the distribution of sediment and clast provenance show that the major source of sediment and water discharge was from The Topsails, to the east of the basin. The area north of Hinds Brook is characterized by fan deltas. This area has only steep-walled, small tributary valleys entering the modern lake where the source of water and sediment was wasting ice on the adjacent hilltops that entered the valley as poorly confined flows. South of Hinds Brook, several larger streams enter Grand Lake. Meltwater and sediment discharge in these areas was along well defined channels. At the mouths of these streams flat-topped deltas are found having surface elevations, from north to south, of 157 and 118 m asl (Hinds Brook), 144 m asl (Harry's Brook, brook north of Grindstone Point), 140 m asl (Little Pond Brook), 135 m asl (south Grand Pond Point), 130 m asl (Connors Brook), and 128 m asl (Lewaseechjeech Brook). The progressive decrease in elevation from north to south is likely related to postglacial isostatic rebound (*see* Section on Sea-level History, page 121). These flat-topped features were established during a period of relative stability in lake development.

Sediment Exposed at the Southwestern End of Grand Lake

Several small sections were cursorily examined at the southwestern end of Grand Lake. These sections are outside the Humber River basin, but were examined because of their relevance to discussions of the Quaternary history of the basin.

Gallants Pit

A large, abandoned gravel pit is located opposite the junction between the TCH and the road to Gallants (Site 91137: Appendix 1). The pit is located on the south side of a channel extending from modern Grand Lake toward the Harrys River valley. Surface elevation of the pit is about 45 m.

The sediments are more than 15 m thick, poorly exposed and consist of weakly stratified gravel, sandy gravel and sand beds. The gravel and sandy gravel beds are 20 to 30 cm thick, normally graded or ungraded, commonly clast-supported, and contain subrounded, commonly granite and felsic volcanic granules to boulders; clasts show no obvious imbrication. The gravel beds are open-work, normally graded and planar bedded; the beds have sharp, subhorizontal

Table 14. Characteristics of diamictons exposed along the eastern shore of Grand Lake

Site	Thick (cm)	Colour	Sand (%)	Silt (%)	Clay (%)	Sorting (ø)	S_1	S_3	Trend (°)	Structures	Overlain by	Underlain by
92178	50+	7.5YR 4/2	71	27	2	3.5				None	Silt–clay rhythmites	?
93008	30	5YR 3/3	73	15	12	2.4				None	Sandy gravel	Sandy gravel
93011	120+	5YR 3/3	76	21	3	3.1	0.6	0.1	47	Irregular planar lenses	Sandy gravel	?
93011	110	7.5YR 4/2	86	11	3	4	0.5	0.2	25	Irregular planar lenses	Surface	Sandy gravel
93015	100	7.5YR 3/4	64	26	10	4.6				None	Surface	Gravelly sand
93028	600	10YR 3/4	70	23	7	4	0.6	0.1	302	Irregular planar lenses	Surface	?
93029	400	5YR 3/4	75	21	4	3.4				None	Sandy gravel	?

upper and lower contacts. The sand beds are 10 to 30 cm thick, laterally discontinuous, planar bedded, well-sorted medium to coarse grained and some trough crossbedded sands showed flow directions between 230 and 110°.

The sediments found within the Gallants Pit were deposited by current flow. Beds are commonly well sorted gravels to sands, indicating variable current flow velocities, including high-energy flows. Trough crossbedding indicates variable flow directions. The combination of roughly horizontally stratified gravels, horizontally stratified sands and planar cross-stratified sands suggests deposition in an ice-proximal braided stream environment (*see* Rust, 1975, 1978 and Miall, 1978). No collapse structures, faulting, diamicton beds, or high-angle crossbeds were noted that would suggest that the deposits formed in an ice-contact environment.

Grand Lake Road Sand Pit

An abandoned pit within the Grand Lake Brook valley is located about 1.1 km from Grand Lake on the north side of a gravel road leading to the TCH (Site 91136: Appendix 1); the pit exposes about 15 m of sediment and has a surface elevation of about 120 m.

Sediment exposed within the pit is mostly moderately sorted medium sand. Clasts are rare, and where found are pebble to cobble size. Sands are rippled and mostly erosional stoss. Of the ripples noted, wavelength was 15 to 25 cm, and ripple height 4 to 7 cm, producing a ripple index of 4 to 6. Planar, tabular crossbeds and trough crossbeds are common across the pit. Paleo-flow indicators showed westward flow (~ 230 to 270°), i.e., away from Grand Lake.

The sediment in the Grand Lake Road Sand Pit was deposited by unidirectional current flow, as suggested by the rippled and crossbedded sands. Erosional stoss ripples indicate migration rates greater than deposition, and grain size and ripple dimensions suggest low current flow velocity (less than 50 cm sec^{-1}) (Harms *et al.*, 1982).

The pit is within a channel, and the lack of clasts in an area elsewhere dominated by sand and gravel (cf. Gallants Pit) suggests deposition within a lake. Alternative depositional environments are subglacial or proglacial. Deposition within a subglacial tunnel would likely include coarse material and diamicton, either incorporated by falling from the tunnel roof, or as debris flows from the channel sides (e.g., Cheel and Rust, 1982; Eyles and Eyles, 1992). A proglacial depositional environment is characterized by highly variable water and sediment discharge, commonly producing a gravel-dominated braided stream system (e.g., Miall, 1992). Non-marine, sand-dominated systems are rare in Newfoundland. Sommerville (1997) describes similar sediments in the Terra Nova River area, eastern Newfoundland and also interprets them as having been deposited in a proglacial lacustrine environment.

Discussion: Fine-grained Sediment in the Humber River Basin

Other exposures of fine grained, rhythmically bedded sediments in the Deer Lake–Grand Lake basin were described by other workers. Normally graded, rhythmically stratified, sand to silt laminae occur at the base of the Gillard's Lake section at 140 m asl, overlain by 30 m of interbedded silt, sand and diamicton (Liverman and St. Croix, 1989b). These sediments were interpreted as glaciolacustrine, deposited in a pro-delta setting and overlain by debris flow diamictons, originating from an ice margin abutting standing water.

Lundqvist (1965) described a section at an elevation of 104 m asl in a canal cut across the Indian Brook–Birchy Lake drainage divide. Liverman and St. Croix (1989b) re-examined the section and found a metre of rhythmically laminated sand, silt and clay, overlain by sands and gravels. This sequence indicates that a brief period of glaciolacustrine rhythmite sedimentation was followed by progradation of deltaic sediments. Clast supported, horizontally bedded gravel exposed at the surface is probably fluvial, and was deposited following lake drainage.

Elsewhere, throughout the Humber River basin, below approximately 25 m asl elevation, diamicton and glaciofluvial sediment is overlain by reddish-brown, rhythmically bedded silt and clay. These extend from Harrimans Steady near Reidville to the modern coast, the exception being the Humber River gorge, where only small pockets of rhythmites are found. Silt–clay thickness exceeds 60 m at Deer Lake airport (Environment Canada, 1980), and over 100 m near Steady Brook (Golder Associates, 1983), but more commonly is 10 m or less (Appendix 2). The sedimentology of these fine-grained sediments has been described from exposures at Rocky Brook and North Brook.

Fine-grained deposits found below published estimates of marine limit may be marine and those above lacustrine or glaciolacustrine. To test this hypothesis, other indicators of depositional environment, including macro- or microfauna, and geochemistry are considered.

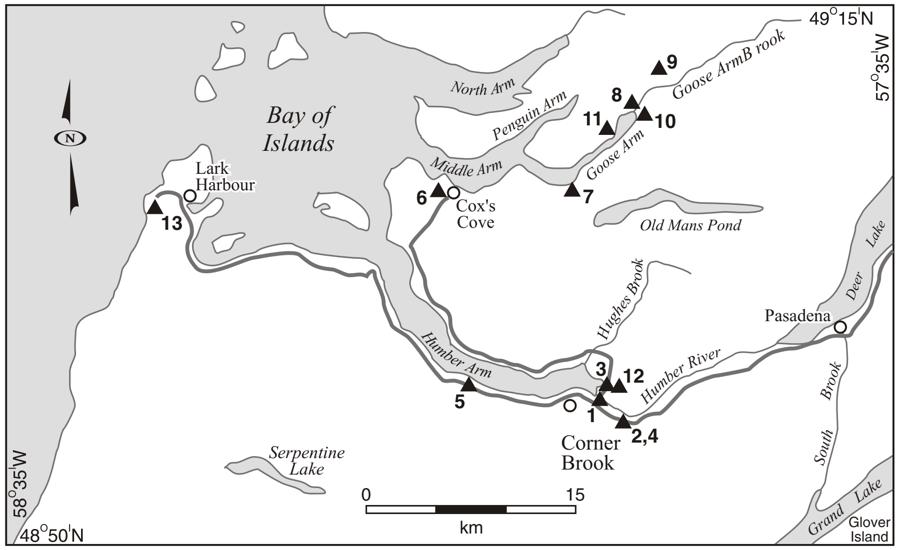


Figure 43. *Location of marine shells found in the Humber River basin area.*

Macrofauna

Marine shells found within fine-grained sediments above modern sea level are restricted to areas around the modern coastline (Figure 43; Table 15). The table is restricted to reported data, and in some cases may not include the range of species found at a particular site. Radiocarbon dates on shells from these locations range between 10 600 and 13 100 BP.

Dyke *et al*. (1996) have reviewed the distribution of collected marine shells in Canada, and in comparison to modern distributions have characterised fossil assemblages as they relate to water temperature and, to a lesser degree, salinity. The restricted Arctic assemblage contains few species (*Hiatella arctica, Mya truncata, Portlandia arctica*). No known modern assemblage contains only these species, and thus it may characterize ice-proximal marine environments. The diverse Arctic assemblage contains a range of gastropods (e.g., *Natica clausa*), pelecypods (e.g., *Mya truncata, Macoma calcarea, Hiatella arctica, Nuculana penula*) and Cirrepeds (e.g., *Balanus balanus*). These species are found in environments that have 7 to 12 months sea-ice cover, shallow water (less than 100 m) and summer temperatures up to 5°C. This assemblage reflects the arrival of species shortly after deglaciation. The arctic assemblage having boreal species characterizes a zone of slightly warmer water temperatures, allowing survival of more thermophilous species (e.g., *Chlamys islandicus, Macoma balthica, Mytilus edulis, Balanus hameri*). Other assemblages (the boreal assemblage with Arctic species, boreal assemblage, and Virginian assemblage) reflect increasingly temperate conditions. Marine fossils may thus provide an indicator of paleo-environmental conditions at the time of their death.

Marine pelecypods, gastropods and cirrepeds found in the Humber River basin are cold water species (Lubinsky, 1980; Dyke *et al.,* 1996). Arctic and diverse Arctic assemblages are found in Goose Arm and Wild Cove. These are typical of recently deglaciated marine environments. The remainder of the area is generally characterised by an Arctic assemblage containing boreal species. These areas may have experienced increased seasonal warming, but not sufficient to prevent arctic species from dominating (Dyke *et al.,* 1996). Marine shells in the inner part of the Humber Arm, including the Humber River gorge are part of this assemblage. *Macoma balthica* shells identified at Dancing Point also fall into the Arctic assemblage containing boreal species. Grant (*in* Blake, 1987) interprets the sediment in which the shells are found to be part of an ice proximal delta (*see* discussion of Dawe's Pit, page 74). Macoma balthica are dominantly a boreal species, although they are commonly found in arctic assemblages (Dyke *et al.,* 1996). They are not associated with modern, ice-proximal, glaciomarine environments (Rodrigues, 1992), and are not found within the Arctic or diverse arctic assemblages. The presence of marine shells having boreal affinities at the head of Humber Arm indicates the infusion of warmer water into the area following deglaciation.

The presence of marine shells in fine-grained sediment adjacent to the modern coast, and within the Humber River gorge shows these areas were inundated by the sea shortly after deglaciation. Fine-grained sediment examined in sections at Pasadena, Reidville, and Steady Brook, and in numerous small roadside exposures between Corner Brook and Deer Lake, did not contain macrofossils.

Microfauna

Fourteen silt–clay samples from coastal exposures, the Deer Lake valley, and the shores of Grand Lake were exam-

Table 15. Marine macrofauna species found in the coastal areas of the Humber River basin, and their modern habitats (using the terminology of Dyke *et al.*, 1996)

#	Location	NTS	Latitude (°N)	Longitude (°W)	Elev. (m asl)	Reference	Species	Fossil assemblage
1	Dancing Point	12A13	48°57'	57°53'	15	GSC Paper 87-7	*Macoma balthica*	Arctic-dominated with boreal (?)
2	Humber River gorge	12A13	48°56.9'	57°50.8'	13	This report	*Balanus hameri* *Hiatella arctica*	Arctic-dominated with boreal
3	Wild Cove	12A13	48°58.3'	57°52.7'	18	Batterson *et al.*,1993	*Mya truncata* *Hiatella arctica* *Mya arenaria* *Macoma calcarea*	Diverse Arctic
4	Humber River gorge	12A13	48°56.9'	57°50.8	13	Batterson *et al.*, 1993	*Balanus hameri* *Hiatella arctica*	Arctic-dominated with boreal
5	Curling	12A13	48°57.5'	57°59.3'	10	This report	*Mya truncata* *Hiatella arctica* *Chlamys islandica*	Arctic-dominated with boreal
7	Goose Arm	12H04	49°07.4'	57°56'	6	GSC Paper 89-7	*Mya truncata*	Arctic (?)
8	Goose Arm	12H04	49°11.0'	57°51.8'	28.5	This report	*Hiatella arctica* *Chlamys islandica* *Macoma balthica* *Mya arenaria*	Arctic-dominated with boreal
9	Goose Arm	12H04	49°12.7'	57°49.7'	50	Batterson *et al.*, 1993	*Mya truncata*	Arctic (?)
10	Goose Arm	12H04	49°10.9'	57°51.5'	7	This report	*Nuculana penula* *Mya truncata* *Balanus crenatus* *Natica clausa*	Diverse Arctic
11	Goose Arm	12H04	49°10.1'	57°53.1'	26.5	This report	*Mya truncata*	Arctic (?)
12	Wild Cove	12A13	48°58.3'	57°53.3'	7	This report	*Macoma calcarea* *Hiatella arctica* *Balanus crenatus* *Nuculana* sp. *Chlamys islandica* *Natica clausa*	Arctic-dominated with boreal
13	Little Port	12G01	49°06.7'	58°24.8'	37	Brookes, 1974	*Mytilus edulis*	Arctic-dominated with boreal

ined for foraminifera. Samples were wet sieved through a 4ø sieve, and the coarser than 4ø fraction dry sieved through a nest of 1ø, 2ø and 3ø sieves. Residue was examined using a binocular microscope.

Samples from fine-grained sediments adjacent to the modern coast commonly contained the foraminifera *Elphidium excavatum* and rare specimens of *Cassidulina reniforme*. These species tolerate a wide range of temperature and salinity conditions, and are commonly the first species to occur following deglaciation (Knudsen, 1971; Scott *et al.*, 1984; Vilks *et al.*, 1989). Thus, they represent a glaciomarine environment, having temperature and salinity about 0°C and between 25 and 30‰, respectively. This assemblage has been identified from marine muds in the Humber Arm (Shaw *et al.*, 1995), and elsewhere in eastern Canada (e.g., Vilks *et al.*, 1989; Maclean *et al.*, 1992; Rodriques *et al.*, 1993).

Samples of fine-grained sediment collected from along the shores of Deer Lake and Grand Lake showed no well preserved foraminifera. Near Reidville (Site 91010: Appendix 1), and at Eastern Brook, near Pasadena (Site 91192: Appendix 1) samples contained agglutinated grains, possibly agglutinated foraminifera (C.P.G. Pereira, Department of Mines and Energy, personal communication, 1996). These are silt-sized grains agglutinated using extraneous material (Moore, 1964) having chitin-like endoskeletons. Species identification is hampered by the agglutinates, and no surface pattern was noted to aid classification. Scott *et al.* (1980) and Murray (1991) state that agglutinated foraminifera dominate brackish water and estuarine environments in Atlantic Canada.

Fine-grained sediment samples examined from elsewhere in the basin were devoid of microfauna. The lack of microfauna (or microflora, such as diatoms) does not imply the area was not inundated by the sea. Low to zero concentrations of microfauna were reported in other areas of eastern Canada, including the Gulf of St. Lawrence (Rodriques *et al.*, 1993), in the Hudson Strait (Vilks *et al.*, 1989), and at the base of a core from the Humber Arm (Shaw *et al.*, 1995). Rodriques *et al.* (1993) attributed these low values to high meltwater discharge. It is also likely that turbidity within the water column hampered the influx of microfauna (or microflora), and in areas undergoing isostatic rebound, such as the Lower Humber River valley, the length of time available for migration was short. The Humber River gorge provides a narrow entrance to the Deer Lake basin through which modern discharge rates are high (mean annual discharge about 120 m^3/s) (Department of Environment and Lands, 1992). This would also prove a severe impediment to the in-migration of marine microfauna. Paleontological evidence in muds inland of the Humber River gorge is scarce, and does not conclusively demonstrate a marine or lacustrine depositional environment.

Geochemistry

Geochemistry is used as a paleo-environmental indicator of conditions during deposition of silt–clay, particularly boron chemistry. Sea water contains about 4.6 ppm boron (mostly as hydroxides), compared to 0.1 ppm for freshwater environments (Goodarzi and Swaine, 1994). Boron may be incorporated rapidly into clay minerals (particularly illite), initially by surface absorption and more slowly by replacement of alumina in tetrahedral sites of the mica structure. The concentration of boron within fine-grained sediments is proportional to the salinity of the solution in which it was deposited, and thus boron may be used as an indicator of paleosalinity (Lerman, 1966; Shimp *et al.*, 1969; Couch, 1971; Catto *et al.*, 1981; Mosser, 1983; Goodarzi and Swaine, 1994). Harder (1970) suggested that boron content is dependent on a number of factors including; the boron content of water in the zone of sedimentation, distance travelled by particles during sedimentation, water temperature and grain size. Shimp *et al.* (1969) and Levinson and Ludwick (1966) considered that the boron content increases with decreasing grain size. Catto *et al.* (1981) suggested that boron values greater than 105 ppm are likely representative of marine environments, whereas those below 30 ppm are freshwater. Similar ranges were reported by Goodarzi and Swaine (1994) in coals.

Boron analysis was completed on silt–clay samples from sediments independently identified to represent raised marine and freshwater environments (Table 16). Methods of analysis were described in the Review of the Humber River Basin. The marine samples were from coastal locations in the Humber Arm and near Springdale, and commonly the sediments sampled contained marine mollusca. Freshwater sediments were from raised proglacial or ice-marginal lakes in the Tulks River valley in central Newfoundland, the Birchy Lake valley at a site previously described by Lundqvist (1965), and from the South Brook valley near Pasadena. In each case, sediments were from well above marine limit for the local area. Samples from within the Humber Valley were all from elevations below 50 m asl.

Data shows that freshwater muds had values from 36 to 73 ppm boron (mean 56 ppm). The boron content of muds from areas of known marine or glaciomarine sedimentation in the Cook's Brook and Wild Cove areas ranged in value from 57 to 70 ppm (mean 63 ppm), indicative of brackish water conditions. Samples from the Indian Brook valley, east of Birchy Lake, record a wide range of boron values from 28 to 131 ppm (mean 54 ppm). Scott *et al.* (1991) suggested samples from this area were deposited in a brackish water environment based on their vanadium content. Samples from below the marine limit in the Humber River valley ranged from 50 to 78 ppm (mean 64 ppm). These data may also be interpreted to indicate brackish water conditions.

There is considerable overlap in the range of values between different sedimentary environments, and conclusions concerning the paleo-salinity of muds within the Humber River valley must be considered speculative. Samples of fine grained sediment from the area are clay-rich (mean 43.4 percent, n=10), and clay mineralogy is dominated by illite (Vanderveer, 1977), both suitable for boron absorption. However, the fine-grained sediments in the Lower Humber River valley were deposited in an area of rapid isostatic adjustment that was a major conduit for glacial meltwater, and they may thus contain a large component of non-clay minerals in the clay-sized fraction. There may also have been insufficient time for the absorption of boron to occur,

Table 16. Boron geochemistry of fine-grained sediment in the Lower Humber River valley, compared with other areas in west–central Newfoundland. Samples originally collected by B. Sparkes, D. Liverman and F. Kirby (all Newfoundland Geological Survey) were analysed for boron and data released below

Geographic Location	Source	Environment	Value (ppm)
Tulks Valley	Sparkes, unpublished data	freshwater	45
Birchy Lake	Liverman, unpublished data	freshwater	49
Birchy Lake	Liverman, unpublished data	freshwater	36
South Brook	Kirby, unpublished data	freshwater	63
South Brook	Kirby, unpublished data	freshwater	68
South Brook	Kirby, unpublished data	freshwater	73
South Brook	Kirby, unpublished data	freshwater	57
Indian Brook 1 – base	Liverman, unpublished data	glaciomarine	31
Indian Brook 1	Liverman, unpublished data	glaciomarine	104
Indian Brook 1	Liverman, unpublished data	glaciomarine	28
Indian Brook 1	Liverman, unpublished data	glaciomarine	29
Indian Brook 1 – top	Liverman, unpublished data	glaciomarine	32
Indian Brook 2 – middle	Liverman, unpublished data	glaciomarine	63
Indian Brook 2 – all	Liverman, unpublished data	glaciomarine	131
Indian Brook 2 – top	Liverman, unpublished data	glaciomarine	34
Indian Brook 3	Liverman, unpublished data	glaciomarine	30
Cook's Brook	This report	marine	57
Cook's Brook	This report	marine	64
Cook's Brook	This report	marine	70
Wild Cove	This report	marine	58
Near Steady Brook	This report	unknown	54
North Brook	This report	unknown	72
North Brook	This report	unknown	57
North Brook	This report	unknown	59
North Brook	This report	unknown	57
North Brook	This report	unknown	68
North Brook	This report	unknown	78
North Brook	This report	unknown	64
Rocky Brook	Liverman, unpublished data	unknown	69
Rocky Brook	Liverman, unpublished data	unknown	50
Rocky Brook	Liverman, unpublished data	unknown	76
Rocky Brook	Liverman, unpublished data	unknown	67
Rocky Brook	Liverman, unpublished data	unknown	73
Reidville	This report	unknown	50
Reidville	This report	unknown	60

or that the large input of turbid freshwater served to dilute sea water and produce local differences in geochemistry (Scott *et al.*, 1991). Samples containing marine macrofossils were deposited in fjord environments, rather than open coastal waters. They may also represent brackish water conditions.

The boron content of fine grained sediments from within the Humber River valley suggests brackish water conditions (Catto *et al.*, 1982; Goodarzi and Swaine, 1994). Values are above those of sediment deposited in known freshwater environments from the Tulks valley and Birchy Lake valley, although not above those from South Brook. Boron values from marine or glaciomarine sediments found in the Indian Brook valley contain low concentrations that may be explained by turbid freshwater input.

Conclusion

The depositional environment of fine-grained sediment in the Lower Humber River valley remains speculative. The sediments contained no macrofossils, no microflora, and only rare foraminifera, and had no distinct geochemical signature. A sedimentological examination showed the sediments were deposited in standing water. Isotopic examination has been used to differentiate freshwater from marine environments in Israel (Reinhardt, 1996), but analyses were of the molluscs, foraminifera and ostracods, rather than the sediment in which they were found. Analyses of strontium isotopes within sediment deposited in a brackish marine environment may be hampered by a $^{87}Sr/^{86}Sr$ ratio that may be similar to that in sea water, because there may not be enough dissolved Sr in the glacial meltwater to overcome the high concentration of Sr in sea water (E. Reinhardt, Dalhousie University, personal communication, 1996).

The problem of differentiating sediments deposited in a marine from freshwater environment is not unique to the Humber River basin. Similar problems have been encountered in environments where fine-grained sediments found below published marine limits are devoid of macro- or microfauna, e.g., Alaska (D. Barclay, University of Buffalo, personal communication, 1996), Cape Cod, Massachusetts (Winkler, 1994), and Puget Sound, Washington State (R. Thorson, University of Connecticut, personal communication, 1997).

Stratigraphic Relationships

Exposures of Quaternary sediment scattered across the Humber River basin, described here and previously, contain sediment deposited in a range of environments, including subglacial, proglacial, glaciolacustrine, glaciomarine and fluvial. Correlation between sections is unclear, because of the wide spatial distribution, the consequent inability to demonstrate facies changes, and the poorly defined temporal relationships. Sediments from the same depositional environment are time transgressive. Therefore, stratigraphic relationships are on the basis of similar depositional environments where stratigraphic succession provides a relative dating control.

Stratigraphic relationships are presented as a series of five transect lines oriented almost parallel with, and perpendicular to, the basin axis (Figure 44a). Descriptions of individual sections found along these up-valley and across-valley profiles, are used to construct an idealized chronostratigraphic column for the area (Figure 44b). It represents the sediment types and depositional environments found across the basin, rather than inferring regional correlation of individual sections. The complete package illustrated is a chronological representation and does not exist at any individual site.

Transect A–B extends along the axis of the Humber River valley from Corner Brook to north of Cormack. It includes exposures described at Humbermouth, Hughes Brook, the Lower Humber River valley, and from the head of Deer Lake, as well as general descriptions from the Deer Lake basin and the Cormack area. The lowest Quaternary stratigraphic unit in the Humber River valley is a diamicton. Diamicton is found in each of the areas described, but only exposed as the surface sediment in the Upper Humber River valley, away from the modern Humber River. The diamictons vary considerably along the valley; the colour and grain size of the matrices are strongly influenced by local bedrock. Diamictons interpreted as primary tills are commonly found adjacent to those with characteristics suggesting remobilisation. This suggests a similar initial genesis as subglacial deposits for most of the diamictons. Exposures that showed more than one diamicton (e.g., Pasadena dump) commonly had similar textural, clast fabric and clast provenance characteristics, and were interpreted as having been deposited during the same event. Preferred clast orientations are generally parallel to the last ice-flow direction derived from striation data, and clasts within diamictons have their source in bedrock up-ice of the sections in which they were found.

Diamicton exposures representing deposition during more than one glacial advance have not been recognised. Apart from the diamicton exposed at Rocky Brook, all diamictons are interpreted to have been deposited during the late Wisconsinan. The basal diamicton at Rocky Brook was described as a pre-late Wisconsinan till (Vanderveer and Sparkes, 1982). Although some characteristics of this sediment (e.g., compaction) are different than those of other diamictons in the Humber River valley, clast provenance, texture and fabric data are similar to other adjacent diamictons. No datable material was found below the single exposure of this diamicton. The suggestion by Vanderveer and Sparkes (1982) that this diamicton represented the oldest till in the Upper Humber River valley could not be confirmed by an examination of drill core data. Without dating and stratigraphic corroboration from elsewhere in the basin, the Rocky Brook till cannot be definitely assigned a pre-Wisconsinan age.

Ice contact sand and gravel is found within deltas in the Humbermouth area, originally described by Brookes (1974). Exposures are fragmentary as a result of removal for aggregate and community development. Ice contact sediment is also found at the mouth of Deer Lake, within the delta that impounds the lake.

Rhythmically bedded silt and clay dominate the lower part of the Humber River valley below about 30 m asl, except for the Humbermouth–Humber River gorge area where current velocity precluded deposition. These sediments have been described from the Rocky Brook and North Brook sections, and are also exposed in sections at Coal Brook, Reidville, Pasadena, Humber village, Goose Arm, and Wild Cove. In drill core from Steady Brook, sediment thickness was over 100 m. Silt–clay deposits are found as far north as Harrimans Steady on the Humber River (elevation 30 m asl). These were deposited in standing water by a combination of underflow currents and suspension settling. Micropaleontological evidence shows foraminifera in sediments adjacent to the modern coast (at Wild Cove and Goose Arm), and agglutinated silt grains in fine grained sediments inland. Boron geochemistry, used as a potential means of distinguishing deposition within fresh or saline water was inconclusive. The similarity in physical characteristics, the continuity in sediment outcrop and the similarity in elevation suggests that all rhythmically bedded silt–clay sediments in the Humber River valley below about

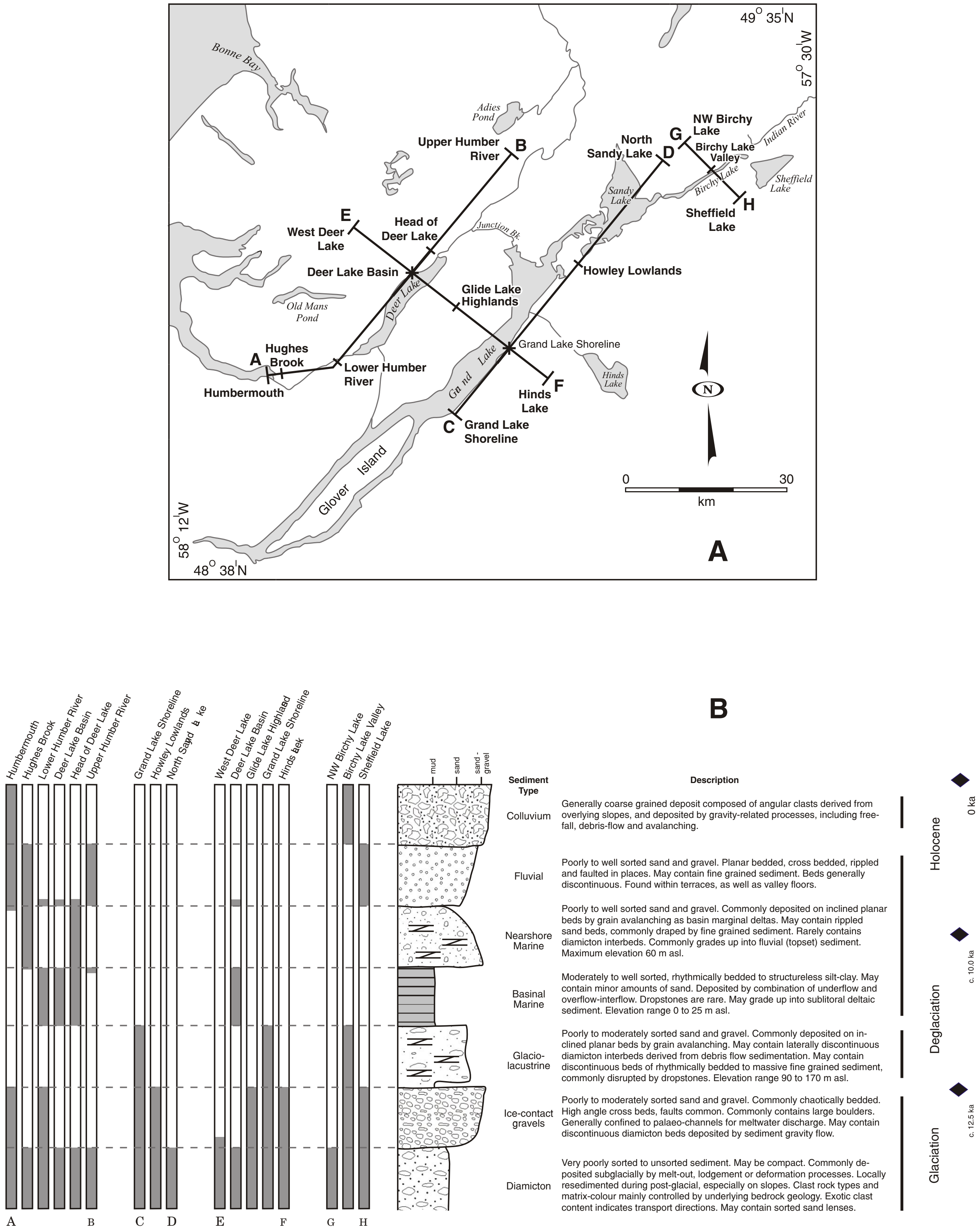


Figure 44. *Stratigraphic relationships across the Humber River basin.*

30 m asl, were deposited in a marine environment. Differences in couplet thickness and grain size are a function of proximity to sediment source, and local sediment and water input. During periods of higher relative sea level, it is likely that sea water inland of the narrow Humber River gorge would have a large freshwater component due to wasting ice in the Humber River watershed. Modern discharge is about 3.8 billion cubic metres of water annually (Department of Environment and Lands, 1992), which likely would have been increased by glacier meltwater contributions during deglaciation, assuming a similar drainage basin area.

Mid-delta and upper-delta foresets lie above silt–clay rhythmites at Rocky Brook. These sediments were deposited in a prograding deltaic environment. A similar stratigraphy is found at Hughes Brook. Delta foresets in the Humber River valley are found at Nicholsville, Little Harbour, Pynn's Brook, Pasadena, Little Rapids and Corner Brook. Deltas at similar elevations in an area experiencing isostatic adjustment suggests that all deltas are roughly the same age. Deltas are absent from the Upper Humber River valley area above Cormack.

Fluvial topset gravels cap deltas at Rocky Brook, North Brook and Hughes Brook. Similar deposits are also found overlying all other deltas in the Humber River valley. Fluvial sediments were deposited in postglacial drainage systems, that have reworked previously deposited sediment. Fluvial systems were re-established in the lower parts of the Humber River valley as sea level fell due to isostatic rebound. Fluvial sediments are therefore found in river terraces, up to elevations of about 30 m asl. These deposits are found at Dawe's Pit, in the Humber River gorge, and on the Humber River flood plain. Contrasts in grain size reflect changes in local flow conditions, where increased current flow through the Humber River gorge deposits gravel-dominated sediment, in contrast to the sandy (commonly cross-bedded) sediments found along the lower reaches of the Humber River.

Colluvium is restricted to the base of steep slopes, and is common within the Humber River gorge, and along North Arm, Penguin Arm and Goose Arm. The lack of vegetation on talus slopes indicates modern activity.

Transect C–D extends along the axis of the Grand Lake valley to the Sandy Lake lowland (Figure 43a, b). Diamicton interpreted as subglacial till is found in some locations at the base of exposures, along the shores of Grand Lake, and is the surface sediment along the north shore of Sandy Lake. Where strong fabrics are found, they either indicate flow northward out of the Sandy Lake basin or westward across Grand Lake. Clast types indicate ice flow and dispersal from The Topsails. The east and north shores of Grand Lake are dominated by sand and gravel. These sediments are interpreted as glaciolacustrine deltas or fan deltas, formed in a proglacial lake with an ice dam located at the northern end of the Grand Lake basin and where water and sediment is derived from melting ice on The Topsails. Proximity of glacier ice is indicated in some exposures by the presence of diamicton beds or dropstones in fine-grained sediment. Glaciolacustrine sediments extend to the surface, and were not reworked, thus suggesting the lake drained rapidly. Deltas at the mouths of modern permanent streams were however incised following lake drainage. Sections along the north shore show lacustrine sediments at the base of some exposures, but they mostly contain glaciofluvial gravels. Current-flow structures commonly indicate flow toward Junction Brook. Near Howley, sandy gravel exposed in hummocky terrain is interpreted to represent areas of ice stagnation. The presence of large boulders (commonly greater than 1 m diameter) near the surface of some hummocks, suggests subglacial deposition.

Transect E–F shows a stratigraphy reflecting variations in elevation across the basin. The lowest Quaternary stratigraphic unit is diamicton. Physical characteristics of these diamictons reflect underlying bedrock geology. Thicker diamictons commonly are found in areas underlain by softer, easily eroded rock types. The highlands west of Deer Lake has diamicton exposed at the surface. Diamictons are common between Deer Lake and Grand Lake, and on The Topsails east of Grand Lake. In the Pynn's Brook area, sediment interpreted as subglacial till is found overlying sand and gravel. This sequence is interpreted as indicating local readvance of ice down the Pynn's Brook valley into Deer Lake. This stratigraphy is unique across the Humber River basin. In the Hinds Brook valley, diamictons are commonly overlain by glaciofluvial sand and gravel derived from meltwater draining through the valley toward Grand Lake.

The intervening Deer Lake and Grand Lake valleys contain thick sediments deposited in standing water. The higher Grand Lake basin contains glaciolacustrine sand and gravel, commonly indicating ice-proximal conditions. The Deer Lake basin contains fine-grained sediment deposited in a postglacial sea that inundated the Deer Lake basin to beyond Reidville. The Grand Lake and Deer Lake valleys are connected by the Junction Brook valley, through which the lake drained.

Transect G–H is across the Birchy Lake valley from the northwest to the Sheffield Lake area. Diamicton is the lowest Quaternary stratigraphic unit. It is generally thin (less than 3 m), with characteristics largely determined by the underlying bedrock geology, and was commonly formed in a subglacial depositional environment. It forms the surface sediment on the highlands on either side of the Birchy Lake valley. Diamicton is overlain by glaciofluvial sand and gravel in some exposures in the Sheffield Lake area, and more commonly, in the Birchy Lake valley; some of these gravels were deposited in an ice-contact depositional environment (Grant, 1989b). Glaciofluvial sediments grade upward into deltaic sediments, indicating the presence of standing water in the Birchy Lake valley. This was noted in the Gillard's Lake valley (Liverman and St. Croix, 1989b), on the watershed with the Indian Brook valley (Lundqvist, 1965), and in the Voyins Brook valley south of Mount Sykes. Much of the

Birchy Lake valley, particularly the steeper central and southern portions is flanked by colluvium derived from the easily weathered slopes above; some slopes are still active. Many of the small peninsulas marking the southern shore of Birchy Lake are debris-flow fans and are partially drowned, following dam construction at Junction Brook.

Periglacial Evidence

Periglacial features do not occur as mappable units, but are significant for reconstruction of the postglacial history of a region. Structures, interpreted as indicative of former permafrost, are found within the South Brook valley.

In a borrow pit, adjacent to the Pasadena incinerator (Site 91105: Appendix 1), a distinct wedge structure was formed within a poorly sorted sand and gravel unit, about 1 m below the present surface, at about 109 m. The wedge was 20 cm wide, 150 cm deep, and contained vertically aligned clasts on its margins surrounding a structureless sand and gravel fill. It truncated primary bedding planes that dip eastward into the South Brook valley.

In a small gravel pit (Site 89006: Appendix 1) at 135 m asl another two wedge-structures are found. The pit exposes deltaic sediments in the form of interbedded, moderately sorted sand and gravel foresets dipping 10 to 20° into the South Brook valley. Each structure consists of a narrow wedge, 30 to 40 cm wide at the top, and 50 to 60 cm deep. The wedges are filled with poorly sorted sand and gravel. Clast sizes in the gravel are mostly granules, but with some pebbles in the upper part of the wedge; the wedge truncates the beds. Where beds meet the wedge in its upper part, they are gently deformed to dip parallel to the wedge sides. The structure is overlain by poorly sorted pebble to cobble gravel.

Interpretation

The wedges in the South Brook valley are interpreted as epigenetic ice wedge casts. Epigenetic features are those that form subsequent to deposition of surrounding sediment, as opposed to syngenetic features whose development accompanies the accumulation of sediment (French, 1976; Washburn, 1980). Syngenetic ice wedges are found on alluvial plains in northern Siberia, and are commonly large features, up to 3 to 4 m wide and 5 to 10 m deep, generated during extended periods of permafrost conditions. Epigenetic features are commonly smaller, and may form during relatively brief periods of permafrost conditions. An epigenetic origin for the South Brook wedges is suggested by the truncation in bedding in confining sediment, the coarse sediment that has been interpreted to have formed over a brief time period, and the palynological record from the southern end of the South Brook valley (Thane Anderson, personal communication, 1996) that suggests a brief period of tundra conditions before the development of forest vegetation.

The development of ice wedges in coarse grained sediment requires that drainage be impeded. This may be achieved by a higher water table or by the presence of a frozen layer. The delta surfaces on which the ice wedges are found are not directly dated, but were developed in a proglacial lake formed during deglaciation.

Discussion

Ice wedges generally require a mean annual air temperature of -6 to -8 °C (Péwé, 1966), although slightly higher temperatures as a result of micro-climatic effects has been suggested by Washburn (1980). Mackay (1990, 1992), working in the Mackenzie Delta, shows that although ice wedges develop at the temperatures indicated by Péwé (1966), they may develop at higher temperatures, up to -1 °C in some areas.

Fossil ice-wedge casts have been described from several other areas adjacent to the study area. Liverman and St. Croix (1989b) describe ice-wedge casts from the Indian River valley. They are found at about 65 m, within cross-bedded and crosslaminated sands and gravels, interpreted as part of a delta sequence. Liverman and St. Croix (1989b) argue the delta was formed in a proglacial lake because current flow indicators within the section show westward flow, opposite to modern drainage. Brookes (1971) described structures from St. David's on St. George's Bay, and from York Harbour in the Bay of Islands (Figure 1). In both areas, the ice wedges were found within bedded gravels, and the wedge casts were filled with chaotic sand and gravel. They formed sometime after deglaciation, at about 12 600 BP (Brookes, 1974). Corney (1993) described ice-wedge casts in beach ridges close to modern sea level in the Fox Island River area. Forbes *et al.* (1993) produce a well-constrained sea level curve for St. George's / Port au Port Bays, that date these features as younger than 12 000 BP.

Liverman *et al.* (2000) examined evidence for fossil ice wedge casts across Newfoundland, and suggested that most, especially those below marine limit, may have formed during the Younger Dryas cooling event, between 11 000 and 10 400 BP. The features found in the South Brook valley are poorly dated. It is possible that they were formed during the Younger Dryas, but it is equally possible they formed shortly after deglaciation of the area.

ICE-FLOW HISTORY

INTRODUCTION

Determination of patterns of ice flow is a critical component of any reconstruction of the Quaternary history in glaciated areas. Previous reconstructions of the ice flow in the Humber River basin by Rogerson (1979), Vanderveer and Sparkes (1982), and Batterson and Taylor (1990) have been contradictory. The objective of this section is to examine the evidence used to construct these hypotheses and produce one that best explains the paleo-ice-flow data, including striations and other erosional features, clast fabrics, and clast-dispersal patterns, and distribution of glacial landforms. Glacial landforms and clast fabrics from diamictons have been considered previously, but will be discussed in the summary of this section.

ICE-FLOW INDICATORS

Striations and Other Erosional Features

Erosional forms on bedrock surfaces cover a wide range of dimensions from large scale (e.g., fjords) to micro forms (e.g., striations). The larger features indicate regional ice-flow trends and are commonly formed over successive glacial periods. The orientation of these features is generally controlled by bedrock structure (e.g., faults). Small-scale wear marks on bedrock outcrops indicate local flow at that site.

The most common small-scale features are striations. These are produced subglacially by abrasion of material held within the base of the ice on the underlying bedrock surface. Striations are shallow (< 1 to 2 mm), parallel scratches commonly centimetres long. Longer and deeper parallel indentations are termed grooves.

Striations are commonly of three types: those that increase in depth and width to a point from which width and depth decrease to end as it began; those that increase in depth and width from the up-ice end to the deepest point when it suddenly ends; and those which start abruptly and gradually taper off (Chamberlin, 1888; Iverson, 1991). Striation morphology is determined by the relative hardness of the tool and bedrock surface, the angle of the clast tip, and the overlying load (Sugden and John, 1976; Drewry, 1986). Most striations indicate a trend of ice-flow movement, from which no direction can be determined. Ice-flow direction is commonly inferred from bedrock surface morphology (e.g., miniature stoss-and-lee features) or from rat-tail features. These are formed when a resistant part of the bedrock surface (e.g., a quartz crystal) protects softer rock immediately down-ice producing a tapering ramp of bedrock (Plate 33). Some rock surfaces are unlikely to preserve small erosional features such as striations. In particular, coarse-grained rocks commonly do not host fine striations, and erosional indicators are restricted to bedrock stossing and grooves. Many of the rock types within the study are coarse-grained (coarse sandstone, gneiss, schist, and most of the granites) on which striations are poorly preserved. Finer grained rocks, such as limestone, commonly have well-striated surfaces when first exposed, but weather quickly, removing the finer striations. In the study area, outcrops exposed along roads constructed for logging that have been abandoned for more than 10 years have few or no preserved striations. New logging roads commonly have numerous striated outcrops. The development of a vegetation mat over bedrock outcrops and the effects of acidic precipitation on soluble bedrock surfaces, such as limestone, result in striation erosion.

Striations reflect ice flow at a given site. Subglacial topography locally may be important in determining ice-flow direction, particularly during periods of glacial build-up and decline. Areas with considerable relief may show divergence of flow around topographic highs (e.g., Lawson, 1996). Ice flow initially may be topographically controlled, but as ice thickens flow is less influenced by subglacial topography and more by ice-sheet topography (Denton and Hughes, 1981; Eyles, 1983). Individual rock outcrops therefore may show more than one set of striations, that reflect changes in the location of glacial dispersal centres, rather than separate glacial advances, especially if there is no evidence for intervening ice free periods (e.g., weathered surfaces). Determinations of regional ice-flow directions should be from numerous striation measurements (e.g., Virkkala, 1960; Embleton and King, 1968; Flint, 1971; Kleman, 1990; Kleman and Borgström, 1996).

Other non-glacial processes produce features morphologically similar to glacial striations. These include mudflows (Sharpe, 1938; Scott, 1988), avalanches (Dyson, 1937), river, sea or lake ice (Washburn, 1947; Clayton *et al.*, 1965), and movement along faults (Engelder, 1974; Hancock and Barka, 1987; Power and Tullis, 1989). In the Falkland Islands, ‘striations' are produced by penguins sliding down coastal rock surfaces (Splettstoesser, 1985). In the study area, most striation sites occur away from steep slopes, above the limits of present or paleo lakes, or removed from faults, and makes a glacial origin most likely. Individual sites may have been influenced by debris flows. Analysis of the regional pattern of striations allows potential influences from debris flow activity to be recognised and discounted.

Bedrock outcrops were systematically examined for ice-flow indicators. Numerous striations were commonly found on a single outcrop, showing a wide divergence of orientations. Similar patterns are found at the margins of modern glaciers where ice-flow direction is evident (Iverson, 1991). Striations commonly showed minor (< 5 to 10°) divergence, and mean directions were determined from 10

Plate 33. *Rat-tail from quartz crystal in shale bedrock, near Corner Brook.*

to 20 observations. Individual striations that are at variance with the general trend were ignored. Ice-flow direction was determined from rat-tail features and stoss-and-lee forms on the bedrock surface. More than one set of striations commonly were found on a single bedrock outcrop. The relative age of striations was determined from crosscutting relationships, and lee-side preservation (Lundqvist, 1990; Figure 45). A regional pattern was recognised where a similar age relationship could be determined over a number of outcrops within an area.

Across the study area, 250 striated bedrock outcrops were examined, 43 of which recorded more than one ice-flow direction (Batterson and Vatcher, 1992b; Batterson, 1994d, e, f). This data was supplemented by ice-flow records from published sources (e.g., Neale and Nash, 1963; Brookes, 1974; Alley, 1975; Alley and Slatt, 1975; Vanderveer, 1981, 1987; Grant, 1989b, 1991; Liverman *et al.*, 1990, 1991; Taylor and Vatcher, 1993; Taylor, 1994), as compiled in the Newfoundland striation database (Taylor *et al.*, 1994). More than 800 observations have been recorded in the study area.

All the striations examined appeared fresh, and where more than one set of striations was found on an outcrop there were no visible contrasts in the degree of weathering. Therefore, it is likely that all the striations were formed during the late Wisconsinan.

Clast Fabrics

Clast fabric data has been discussed in the Glacial Sediment and Stratigraphy Section. Diamictons with S_1 values > 0.6, and which contain other characteristics of glacigenic origin (e.g., striated clasts), were, more than likely, primary tills. Clast long axes in these depositional environments are commonly parallel or transverse to ice flow as determined from independent sources.

Of the 173 clast fabrics completed in the Humber River basin, 119 had S_1 values greater than 0.6. Clast fabric trends derived from strongly oriented clast fabrics are commonly similar to striation directions on adjacent bedrock surfaces. This has been well demonstrated in the area around Corner Brook (Figure 46). The striation record shows two regional flows, an early southward flow from a source in the Long Range Mountains (?), and a later event toward the Humber Arm from a source in The Topsails. Clast-fabric data shows similar patterns. Most preferred clast orientations in strong fabrics (i.e., $S_1 > 0.6$) show flow toward Humber Arm. However, several sites show southward paleo-flow. A site near Summerside (Site 91036: Appendix 1) shows two diamicton units, the lower of which has a preferred clast orientation toward 206° (S_1=0.78), and the upper has a preferred clast orientation toward 295° (S_1=0.78). A similar relationship between striations and preferred clast orientation in tills is found in other areas.

Clast Dispersal

The ice-flow pattern from striations indicates the general direction of glacial flow, but not concerning the specific location of ice-dispersal centres. Rock types, identified from glacial diamictons, that possess a distinctive character and a defined bedrock source can help determine both direction and distance of glacial transport.

Clast samples were collected from 351 diamicton sites across the Humber River basin. Samples consisted of at least 50 pebbles and/or cobbles, and included all those clasts distinct from the underlying bedrock geology, determined from bedrock geology maps. Exotic clasts commonly represented far less than 1 percent of the clast component, but were sampled because of their potential significance to dispersal patterns. Therefore, the exotics will probably be over-represented in clast percentages. Clasts were cleaned and the presence of striations or other surface features were noted. Clast rock-type was identified, commonly by comparison to a reference collection of bedrock samples. A total of 19 723 clasts were identified and assigned to 31 different categories (Appendix 3). Groupings were into rock type (e.g., limestone, sandstone, gneiss, rhyolite), or, where possible, specific rock groups or formations (e.g., Sp granite, Hungry Mountain Complex, Hinds Brook granite, Whalen and Currie, 1988). An outline of the bedrock geology of the study area is provided in the Introduction Section, page 1 and includes a discussion of distinctive rock types and their characteristics (Table 1). The dispersal patterns of distinctive rock types is described below.

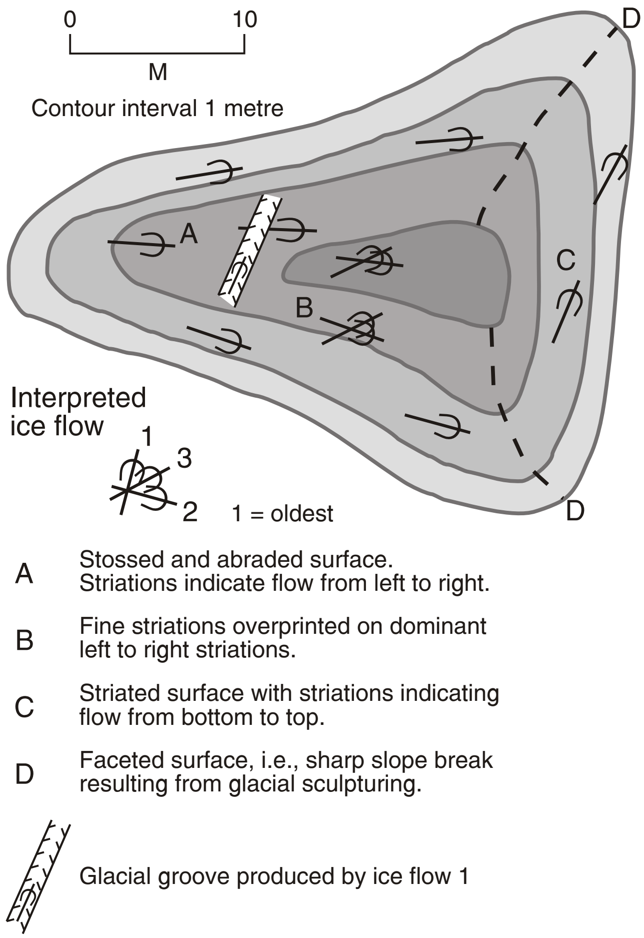


Figure 45. *Schematic diagram showing striation types and their relationship to ice-flow directions.*

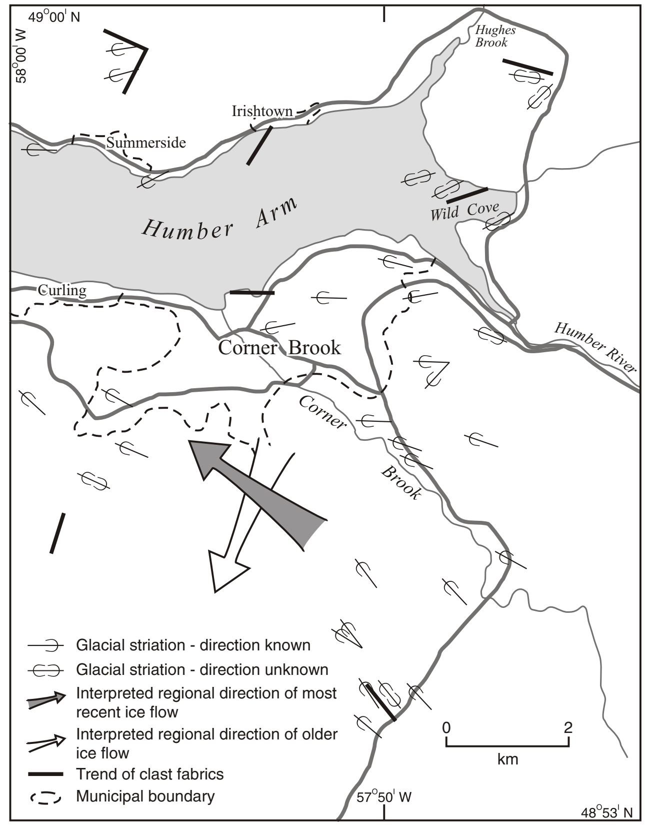


Figure 46. *Comparison of striation and preferred clast-fabric orientation data for the area around Corner Brook.*

Carboniferous Rocks

The central Humber River basin is underlain by Carboniferous sandstone and siltstone. The Humber Falls, Rocky Brook, and North Brook formations of the Deer Lake Group are dominantly red (Hyde 1979, 1984) and they underlie the Upper Humber River basin, and the area around the northern end of Glover Island. The Little Brook Formation and the Howley Formation in the Grand Lake–Sandy Lake basins are red and grey sediment, whereas the Anguille Group that underlies Birchy Ridge is grey. Red sediment in western Newfoundland is restricted to the Deer Lake basin. Carboniferous rocks also contain fluvial conglomerates, found within the North Brook Formation that outcrops on the western, northern and southern margins, and the Humber Falls Formation found in the central part of the basin. The North Brook conglomerate contains arkosic sandstone clasts, and felsic volcanic clasts dominate the Humber Falls Formation conglomerate.

Sandstone–siltstone clasts are dispersed across much of the Humber River basin (Figure 47). They are found along the shores of Grand Lake, between Grand Lake and Deer Lake, on the highlands north and south of Old Mans Pond, on Birchy Ridge and in the Upper Humber River valley. Red sandstone clasts were also noted near the summit of North Arm Mountain, at 640 m but generally are not found on the Long Range Mountains west of Adies Pond, except at one site in the foothills. Similarly, they are rare on The Topsails, with only three sites recording sandstone clasts. A site near Hinds Lake showed two arkosic sandstone clasts. Apart from within Carboniferous sediments, arkose is reported within the Springdale Group (Whalen and Currie, 1988), which outcrops in the vicinity and is therefore a potential source of these clasts. A second site in the Hinds Brook valley recorded a single brown sandstone clast for which there is no obvious source, and a third site near Goose Pond recorded four red sandstone clasts. This latter site is about 2 km east and 100 m higher than the nearest sandstone outcrop, in an area with only a few striations showing ice flow toward the northeast, from the Grand Lake basin toward Sheffield Lake. A flow in that direction would not have crossed Carboniferous bedrock. Other clasts found at the site are generally consistent with northeast transport, including units Oib (granite–granodiorite), OHma (gabbro) clasts, and Oic (granite) clasts, the latter of which only crop out to the southwest. However, the sample also includes Unit Sp (granite) clasts which are found on the highlands 2.5 km to

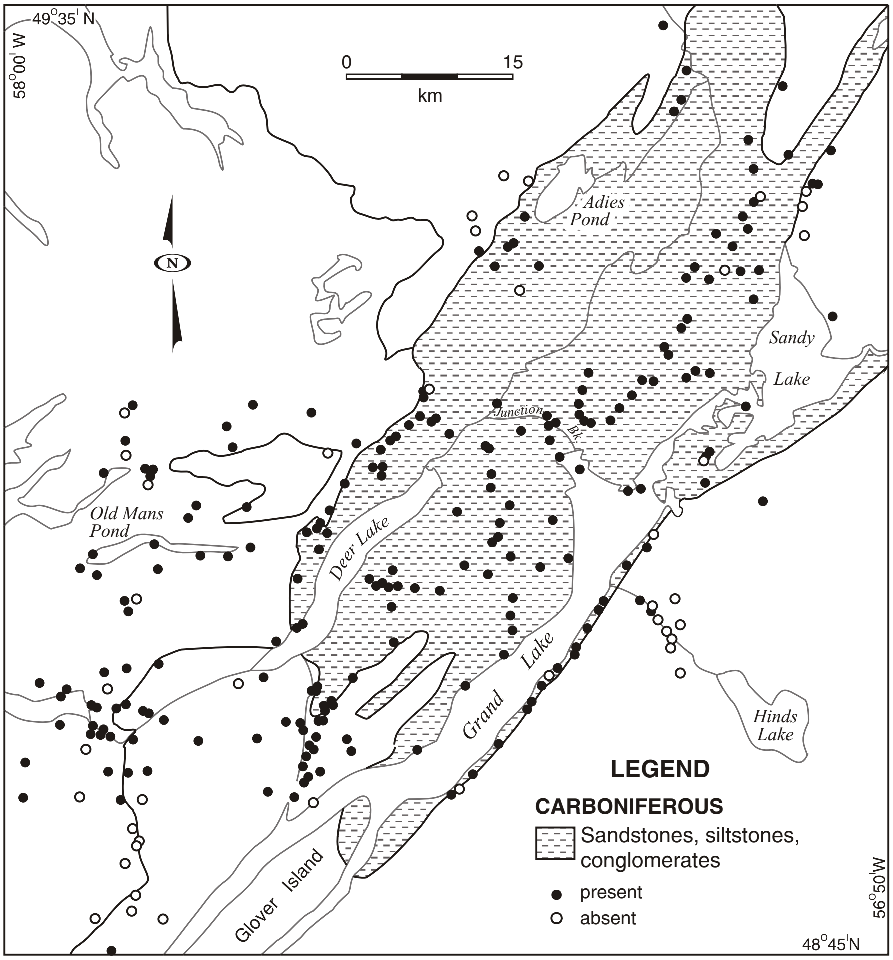


Figure 47. *Dispersal of Carboniferous clasts across the Humber River basin.*

No Unit Sp or Sq clasts were found in the Upper Humber River basin west of Birchy Ridge or over the Long Range Mountains west of Adies Pond. Examination of boulder piles created during land clearing in the Cormack area (Plate 34) showed that clasts derived from The Topsails were present as far north as the Middle East Branch valley. Only Grenville gneiss and Carboniferous sandstone clasts were found north of this area.

Gabbro

Gabbro bedrock is found on The Topsails within the Rainy Lake Complex (Unit SOrl), Hungry Mountain Complex (Unit OHma), Unit Om, or the Buchans Group (Unit Ob) (Whalen and Currie, 1988). Gabbro also crops out within the Gull Lake intrusive suite west of White Bay (Saunders and Smyth, 1990).

Gabbro clasts are dispersed across the central and southern parts of the Humber River basin (Figure 50). They are found south of Birchy Ridge, in the Old Mans Pond area, and south of Corner Brook. They were not found on Birchy Ridge, within the Upper Humber River valley, or around Sandy Lake.

the east, and which are inconsistent with northeast transport. The transport history for clasts at this site remains unresolved.

Topsails Intrusive Suite

The Topsails intrusive suite is confined to the east side of Grand Lake (Whalen and Currie, 1988). Clasts from distinctive bedrock units within it such as Sp granite (Figure 48) and Sq porphyry (Figure 49) have similar dispersal patterns. Clasts from these areas are found dispersed in the highland between the Grand Lake and Deer Lake valleys, and over the limestone bedrock west and southwest of Deer Lake. In the north, Sp granite and Sq porphyry clasts are recorded on Birchy Ridge and in the Sandy Lake lowlands to the east. Other granitic rock types from The Topsails, but with less distinctive characteristics, such as Unit Sm granites (white to red, fine- to medium-grained, equigranular granites), and Unit Sg granite (white to pink, medium-to coarse-grained, biotite–amphibole granite) showed similar patterns to dispersal of the Sp and Sq units. Clasts from bedrock Unit Ssy (mostly quartz syenite) rarely were found, possibly because of the small areas of surface exposure.

Red Flow-Banded Rhyolite

Red, flow-banded rhyolite crops out in the Springdale Group on the central and northern parts of The Topsails (Whalen and Currie, 1988), and within the Sops Arm Group, west of White Bay (Smyth and Schillereff, 1982).

Red, flow-banded rhyolite clasts are found across much of the area (Figure 51), including Birchy Ridge and in the Sandy Lake basin to the east, but are absent from the Upper Humber River basin.

Limestone

Limestone bedrock underlies a large area of western Newfoundland, from St. George's Bay extending north along the Great Northern Peninsula (Williams and Cawood, 1989; Colman-Sadd *et al.*, 1990). Within the Humber River basin, limestone underlies the western part, as far east as Deer Lake. Limestone is found within numerous bedrock units in the study area, including the St. George and Port au Port groups, Reluctant Head Formation, and Old Mans Pond Group. It also composes small parts of the Humber Falls and Rocky Brook formations in the Deer Lake basin. Limestone from each of these areas is visually similar, and therefore

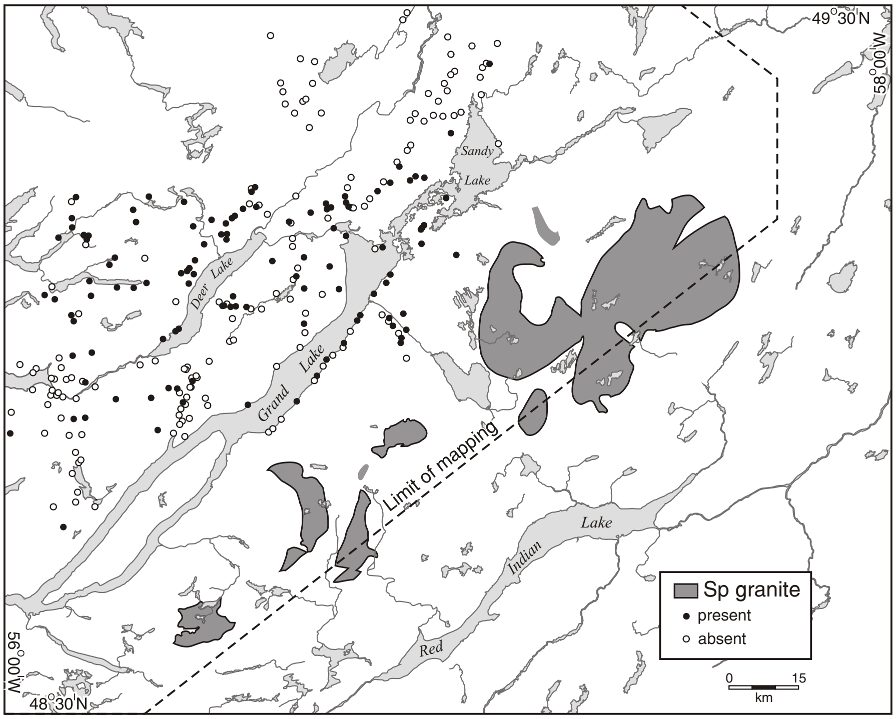


Figure 48. *Dispersal of Sp (Topsails) granite clasts across the Humber River basin.*

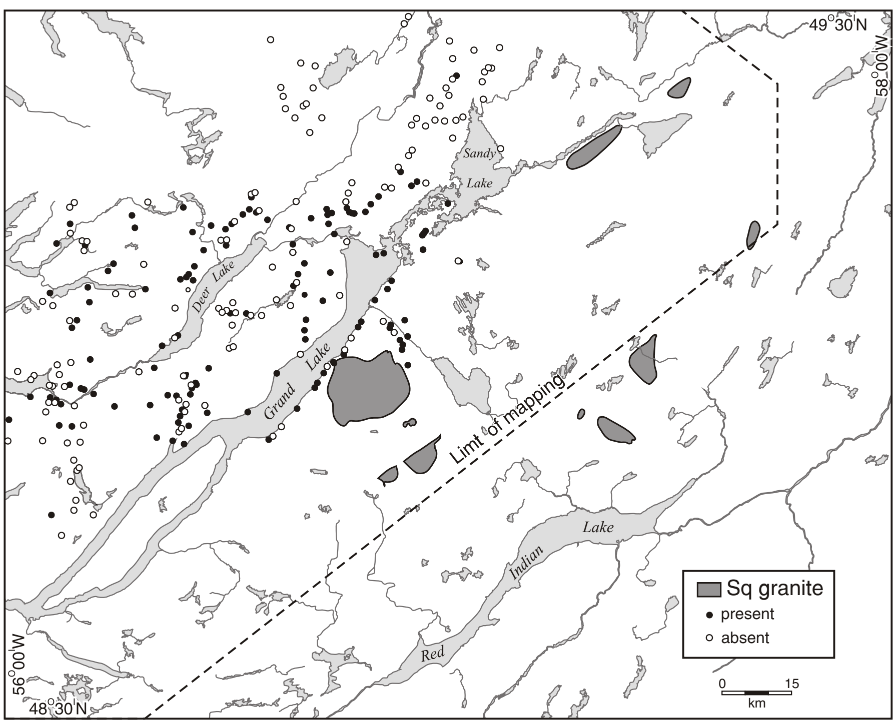


Figure 49. *Dispersal of Sq (Topsails) porphyry clasts across the Humber River basin.*

specific source areas for limestone clasts could not be determined. No other limestone outcrops occur within the study area.

Generally, limestone clasts are found west of their source area (Figure 52), suggesting transport by the regional westward ice flow from The Topsails. The exception is two limestone clasts found within a very compact till in Rocky Brook (*see* Section on Glacial Sediments and Stratigraphy, page 40), about 4.5 km northeast of the nearest source. This till also contains clasts identified as Topsails intrusive suite (units Sm and Sq), Springdale Group (Unit Ssf) and Ordovician granite (Unit Oid), all of which are derived from the east to southeast. Other clasts from within the Rocky Brook till were also derived from The Topsails. Therefore, the limestone may be from the Humber Falls or Rocky Brook formations. The till at Rocky Brook also contains numerous sandstone and siltstone clasts from these formations. These data show the Rocky Brook till was deposited by westward flowing ice from a source in The Topsails.

Gneiss

The largest source of gneiss within the Humber River basin is the Grenville basement that underlies the Long Range Mountains, northwest of the Upper Humber River valley, and along the southern end of Grand Lake. It is a Precambrian, high-grade, medium-grained pink to grey, quartzofeldspathic gneiss (Owen and Erdmer, 1986; Williams and Cawood, 1989). However, there are several other sources of gneiss in the area. The Hungry Mountain Complex, northeast of Hinds Lake is host to a medium grained, white to pink biotite–muscovite gneiss (Whalen and Currie, 1988). Granitoid gneiss and psammitic paragneiss are found in the Caribou Lake complex along the western shore of Grand Lake south of Northern Harbour (Cawood and van Gool, 1992, 1993), and hybrid gneisses are associated with Ordovician granites

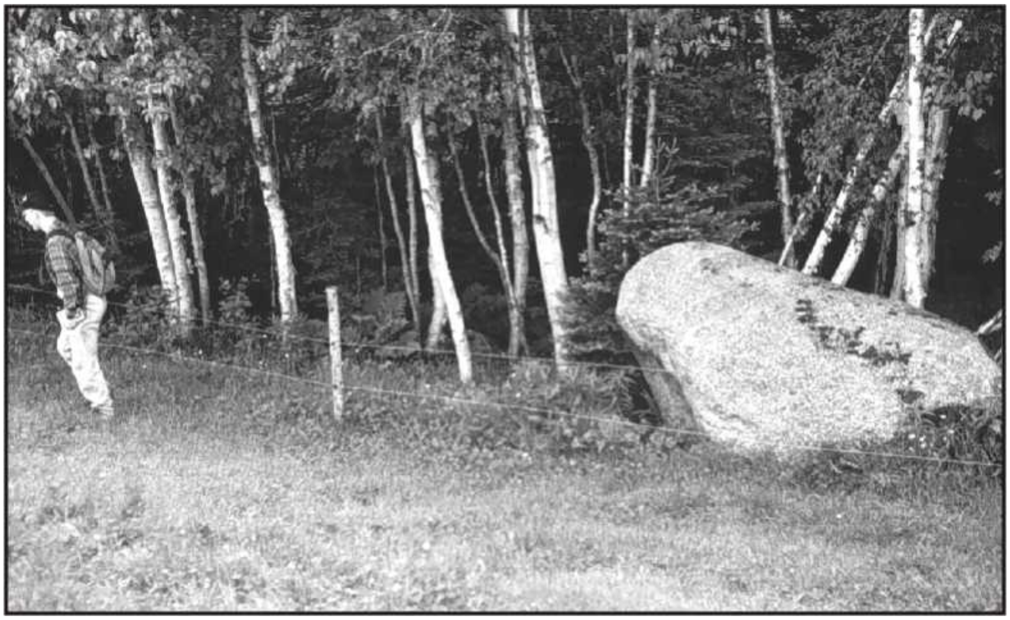


Plate 34. *Topsail granite erratic in cleared field in the Cormack area.*

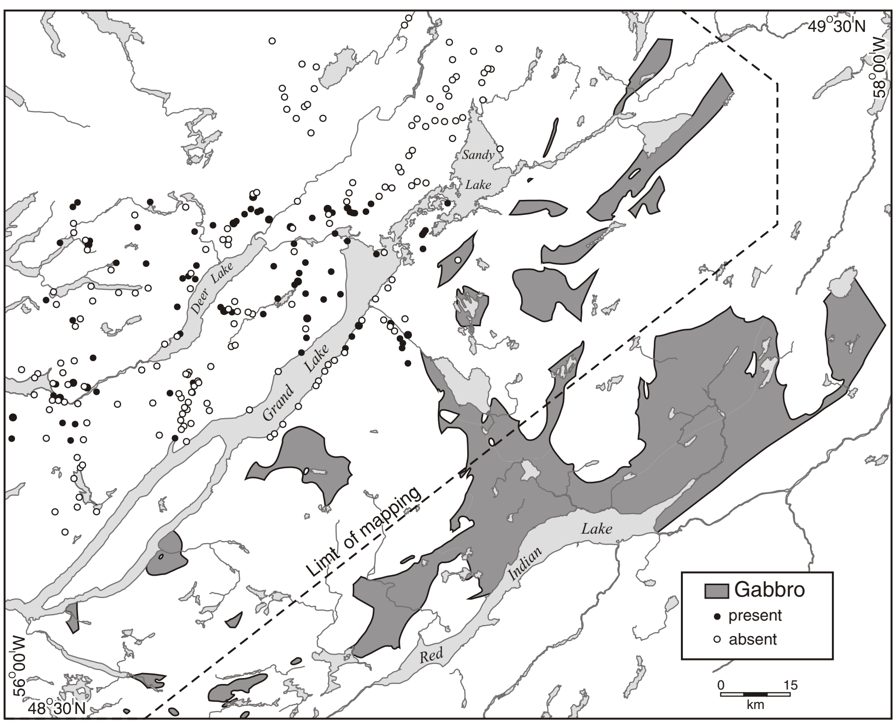


Figure 50. *Dispersal of gabbro clasts across the Humber River basin.*

(Unit Oid) north of Sandy Lake. Similarly, foliated granite clasts may be confused with gneisses. Foliated granites have been reported from the Hughes Lake complex north of Deer Lake (Williams and Cawood, 1989), and the Glover Group (Unit Og) (Whalen and Currie, 1988).

Clasts identified as gneiss are scattered in samples from across the study area (Figure 53). Concentrations of gneiss clasts occur in the Upper Humber River valley, south and west of Adies Pond, on Birchy Ridge, and south of Deer Lake. Other gneiss clasts are found between Deer Lake and Grand Lake, and between Deer Lake and the coast, north of Old Mans Pond. The distribution of clasts in the Upper Humber River valley, their association with Carboniferous sandstone–siltstone clasts, and the lack of clasts from The Topsails suggest southwestward tran-sport down the valley. Also, the large number of gneiss clasts on Birchy Ridge may be explained by southwestward ice flow down the valley, from a source at the north end of the valley. The concentration of sites with gneiss clasts south of Deer Lake, mostly over the Caribou Lake complex, suggests a local source. Other sites across the study area contain gneiss clasts that may have been derived from a number of sources, although those found around Grand Lake are consistent with dispersal from the source in the Hungry Mountain Complex, north of Hinds Lake.

PALEO ICE FLOW IN THE HUMBER RIVER BASIN

The ice-flow history from the Humber River Basin is subdivided into six regions on the basis of physiography and ice-flow patterns found within each. These are: 1) south of Corner Brook along the western edge of the Long Range Mountains; 2) within the Deer Lake valley; 3) between Deer Lake and Grand Lake; 4) the northern part of the Upper Humber River valley; 5) the area over Birchy Ridge; and 6) the Sandy Lake–Birchy Lake valleys. The regions are described separately and later integrated into a discussion of the paleo-ice-flow history of the basin. It should not be assumed that the earliest ice flow event in one area is temporally related to an early flow in another area. Ice-flow trends are derived primarily from striations, unless otherwise noted.

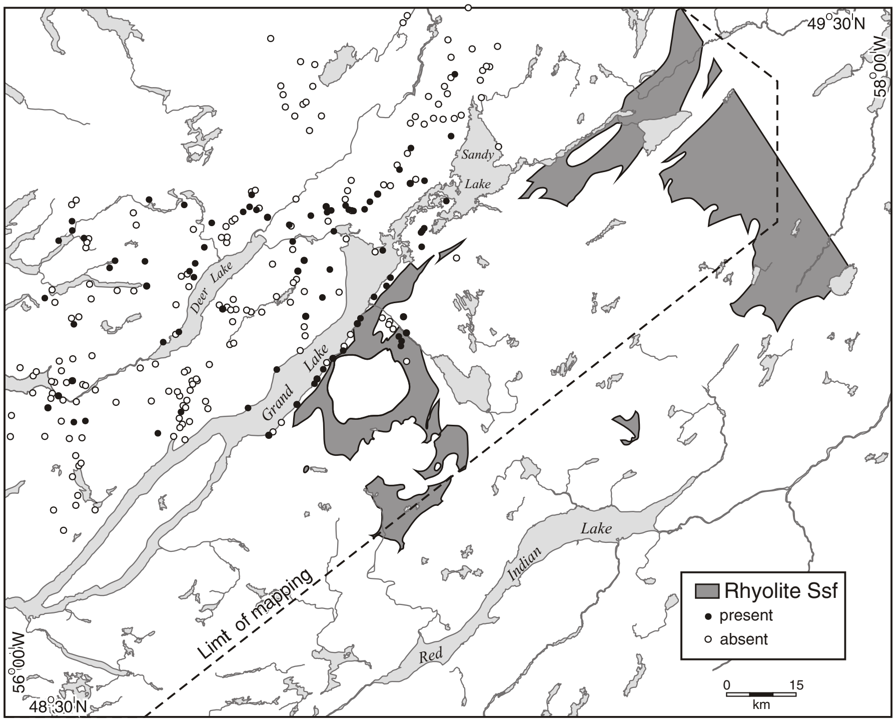


Figure 51. *Dispersal of Springdale Group (Ssf) rhyolite clasts across the Humber River basin.*

The early southward flow was followed by a coastward ice flow from the interior. At the southwest end of Grand Lake, ice flow was westward to southwestward into the Harrys River valley and thence southwestward to St. George's Bay. Farther north, ice moved northwest to westward across the Georges Lake valley, shown by striations that crosscut the earlier southward directed striations. Flow was directed either through the Serpentine Lake valley, or deflected southward along the eastern margins of the Lewis Hills, and out to the coast through the Fox Island River valley. No evidence (e.g., striations, clast provenance) was found to suggest that Topsails-centred ice crossed the Lewis Hills, and there is also no indication that ice flowed east from the Lewis Hills. These highlands are underlain by ultrabasic rocks. The soil supports only sparse vegetation. Lack of vegetation coupled with altitude and a coastal aspect combines to produce an intensely frost weathered environment in which periglacial features (e.g., felsenmeer, gelifluction lobes) are common (Batterson and Liverman, 1995). The presence of erratics and unweathered striated bedrock surfaces found on the Lewis Hills shows the area supported, or was crossed by, glacier ice. Timing of glacier cover is speculative. Surfaces with similar characteristics in Gros Morne National Park that were considered to have remained ice free during the Wisconsinan (Grant, 1987) are now considered to have been ice covered during this period on the basis of cosmogenic isotopic analysis (Gosse and Grant, 1993).

Corner Brook and South

An early south to south–southwest flow is recorded in the area south of Corner Brook, along the western edge of the Long Range Mountains (Figure 54). Generally, the evidence is confined to the Georges Lake valley, although similar trends are recorded as far north as the North Star quarry in Corner Brook (elevation 280 m), and between Cooks Brook and Serpentine Lake (elevation 240 m). South to south–southwest-oriented striations have been recorded as far east as the north end of Corner Brook Lake. The source of this early ice flow is unclear. Strong clast fabrics having a preferred clast orientation trending north–south are found on the north shore of the Humber Arm near Irishtown, and south of Corner Brook. Further support for southward flow is the presence of red micaceous sandstone clasts seen near Pinchgut Lake. Carbonate clasts of local origin are also found (M. Batterson, Department of Mines and Energy, unpublished data). Red Carboniferous clasts indicate that ice flowed across the Deer Lake basin to reach this area. An alternative source for these clasts is the North Brook Formation, exposed at the northern end of Glover Island, however, this is not supported by striation and clast provenance data. Ice flow across this source would also have crossed basalt, diabase and tuff of the Glover Group, possibly gabbro of the Rainy Lake Complex, and granites from the Topsails intrusive suite (Whalen and Currie, 1988). None of these rock types are found associated with the red sandstone clasts at Pinchgut Lake. Similarly, striations do not indicate a flow across Glover Island toward Pinchgut Lake.

Deer Lake Valley

An early southwestward ice flow is recorded along the shores of Deer Lake, and within the Humber River gorge, near Corner Brook (Figure 55). No evidence for this flow is found on the higher ground west or east of the valley, suggesting this was a local, topographically controlled flow.

This early flow was followed by westward to northwestward flow from a source in the interior. Ice crossed Grand Lake and flowed toward Humber Arm at Corner Brook. North of the Steady Brook valley ice flow was westward (270 ± 10°). This reflects the draw-down of ice into the Humber Arm basin. The coastal highlands (North Arm Mountain, Mount Gregory and Table Mountain area) (Figure 1) and the major fjords between Humber Arm and Bonne Bay (Old Mans Pond, Goose Arm, Penguin Arm and North Arm) were a major influence on ice movement. Ice flow west of Deer Lake was either westward along Old Mans Pond and into Goose Arm, or was drawn southwestward into

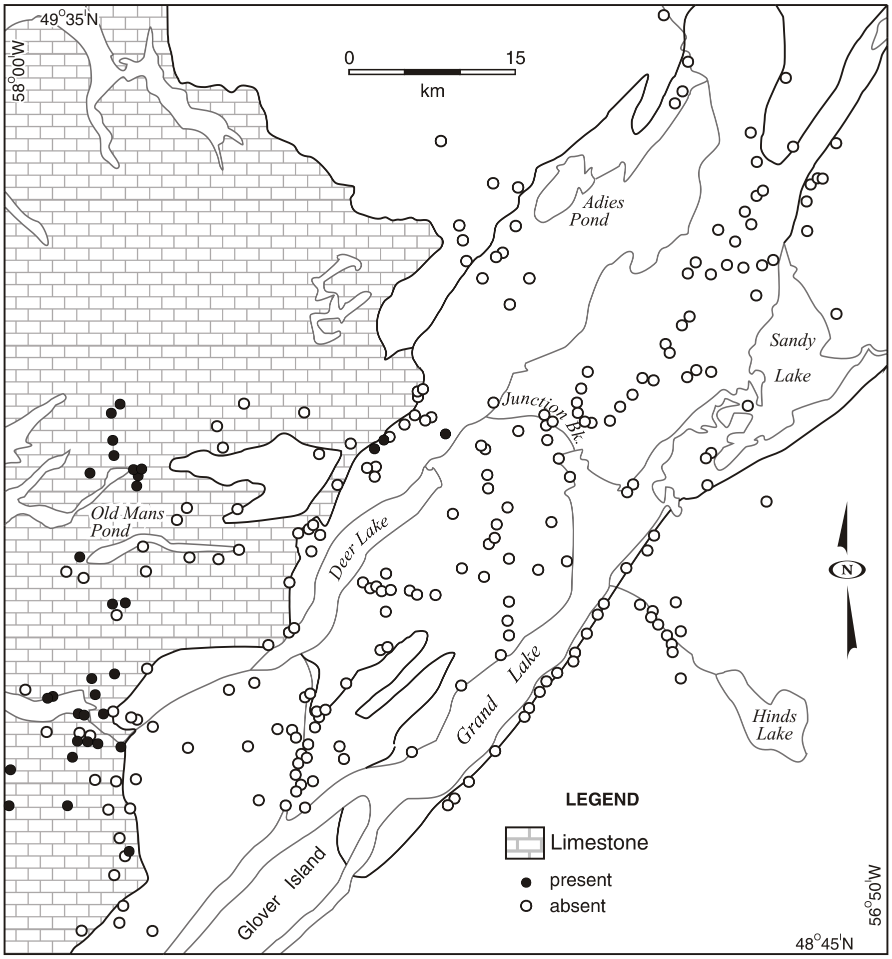


Figure 52. *Dispersal of limestone clasts across the Humber River basin.*

Humber Arm. Northwest of Deer Lake, ice flow was either northwestward toward Bonne Bay or Trout River Pond, or was drawn southwestward into North Arm. There is little published evidence to indicate ice flowing inland off the coastal highlands or them being over-topped by ice flowing from the interior (Taylor *et al.,* 1994).

Regional ice flow from the interior is supported by clast-dispersal patterns, particularly for those rock types found on The Topsails. Preferred clast orientation for tills in the Corner Brook area is toward the Humber Arm, with some indication of flow through the Humber River gorge. Preferred orientations trending westward are parallel to the regional ice-flow patterns that show flow from The Topsails. The exceptions are clast fabrics having a preferred orientation perpendicular to ice-flow events described by the striation record. At the mouth of the Pynn's Brook valley, four clast fabrics taken from glacial diamictons showed preferred orientations perpendicular to striations showing flow down the valley. One of these is from a till overlying sand–gravel that was interpreted to represent local readvance (*see* section Glacial Sediments and Stratigraphy).

Deer Lake–Grand Lake

The highlands between Glide Brook and Grand Lake record striations showing an early east–northeastward ice flow (Figure 56). This flow had its source in The Topsails as shown by the clast provenance of diamictons. Strong fabrics in diamictons in this area also have a preferred clast orientation trending west to northwest.

The early northeast flow from The Topsails was followed by a later south-southwestward flow down the Glide Brook valley toward Deer Lake, and a southward flow east of Glide Brook toward Grand Lake, as interpreted from striations. The source of this ice is uncertain. If it was from the Humber River valley, this should be reflected in the clast provenance of surface clasts. There were no Carboniferous sandstone clasts found in the area. Tills have a preferred clast orientation and clast provenance showing flow from The Topsails, suggesting that the southward ice flow did not rework sediment in the area.

Upper Humber River Valley

The northern part of the Upper Humber River valley has a complex ice-flow history (Figure 57). An early east-ward to east-northeastward ice flow in the Upper Humber River–Taylor Brook area flowing toward White Bay and originating most likely from the Long Range Mountains was identified from striations. These were cross-cut by striations from a southeast to southward ice-flow event that entered the Upper Humber River valley, and subsequently flowed southwestward along the valley axis. Ice flowing into the Upper Humber River basin was recorded east to the Taylor Brook valley. Ice flow was northeastward toward White Bay on the highlands west of White Bay. Generally, ice flow was southward on the uplands northwest of Adies Pond, although striations are oriented eastward along the margins of the Carboniferous basement southwest of Adies Pond. Paleo ice-flow was northwestward or westward towards the coast in the area south of the Middle East Branch valley near Cormack. This is demonstrated by the clast provenance data showing rock types from The Topsails to the south of the valley, but not to the north.

The pattern of striations in the Upper Humber River valley shows that rock types in the area of the Gull Lake intrusive suite were covered by ice flowing eastward towards White Bay, rather than southward down the Hum-

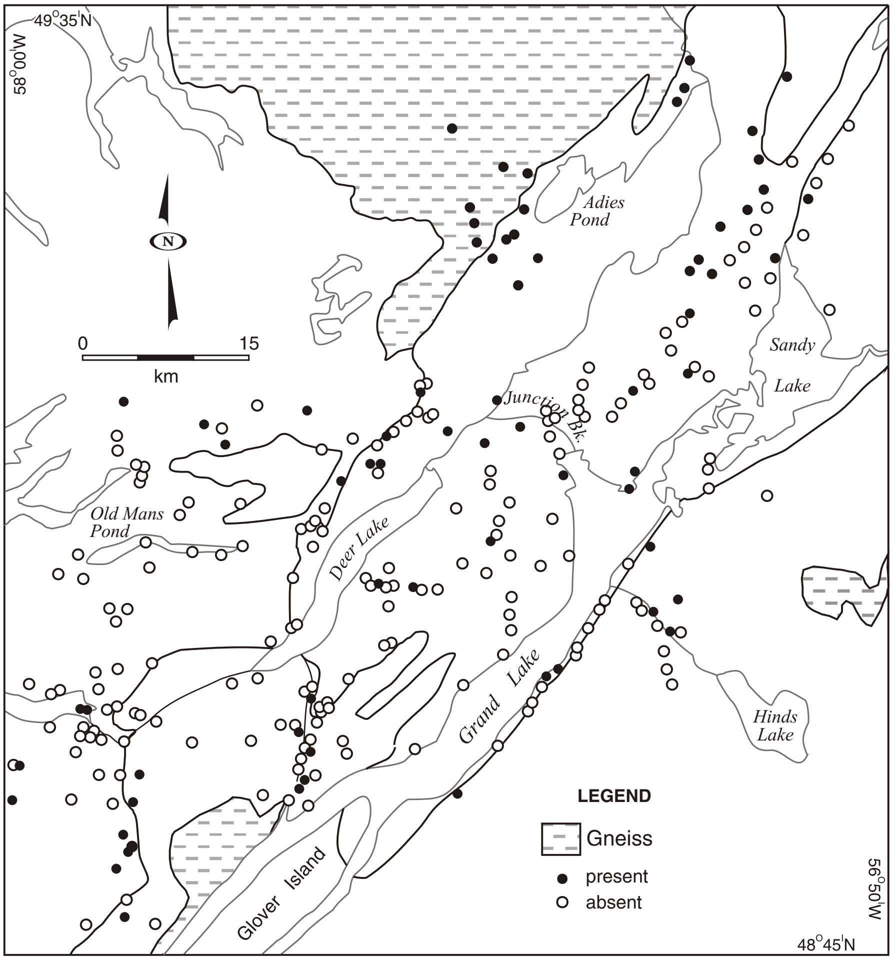


Figure 53. *Dispersal of gneiss clasts across the Humber River basin.*

ber River valley. A source in the Gull Lake intrusive suite had been the preferred source for gabbro clasts found in diamictons near Deer Lake (Vanderveer and Sparkes, 1982), used to demonstrate southwestward ice flow down the Upper Humber River valley. Southward-flowing ice did not cross areas underlain by the Moose Lake granite and the Devil's Room granite that cropout west of White Bay (Saunders and Smyth, 1990). These rocks have similar grain size, colour and mineral composition to some granites on The Topsails. Dispersal of gabbro and granite clasts in the Upper Humber River and areas to the south were therefore derived from a source in The Topsails.

Birchy Ridge

Birchy Ridge is a northeast–southwest trending ridge (200 to 280 m elevation) between the Sandy Lake basin to the east and Upper Humber River valley to the west. Bedrock is mostly grey mudstone and arkosic sandstone of the Anguille Group (Hyde, 1984). This area provides evidence of ice flow with northward, eastward and southward ice-flow indicators (Figure 58). Much of the data were collected from bedrock exposed during construction of the TCH and during mineral exploration for uranium during the late 1970s–early 1980s. These bedrock outcrops are now either partially weathered or overgrown by vegetation.

The northwestern part of the ridge shows an early eastward flow that may be a southern extension of the eastward flow recorded in the Upper Humber River. In the northeast of the ridge, an early southward flow is found, whereas closer to White Bay an early northward flow is recorded.

Striations show a late south–southwestward ice flow along the southern part of the ridge. However, in the north, several sites show northward-directed striations. Other sites in this area record recent south–southwestward and northeastward flow. Evidence of complex ice flow is not restricted to Birchy Ridge. At the north end of Sandy Lake, adjacent to the eastern margin of Birchy Ridge southward oriented striations crosscut earlier north–northeastward striations. Grant (1989b) recorded a similar southward oriented striations in a quarry on the east side of Birchy Ridge. At sites less than 1 km east along the highway, striations indicate ice flow dominantly was northward. The presence of striations indicating southward ice flow suggests that a late flow of ice from a source to the north (Long Range Mountains?). The extent of this flow was undetermined.

Clast provenance of diamictons shows that ice from The Topsails did not overtop Birchy Ridge, although granite (Unit Sp) and porphyry (Unit Sq) clasts from The Topsails are found at the northern and southern ends of the ridge. Drill-core and test-pit data from the Wigwam Brook uranium exploration area on the western flanks of the ridge show dominantly sandstone clasts from the underlying Humber Falls Formation. Non-local clasts included gabbro, granite, and rhyolite, all of which may have been derived from diamictons on Birchy Ridge. Clast fabrics from test pits adjacent to the drill core locations (D. Vanderveer, Department of Mines and Energy, unpublished data, 1987) show weak, girdle fabrics, possibly indicating that sediments were deposited by sediment gravity flow. Clasts identified in drill core as mudstone from the Rocky Brook Formation or granite from the Gales Brook area (Appendix 2) were derived from the north–northeast.

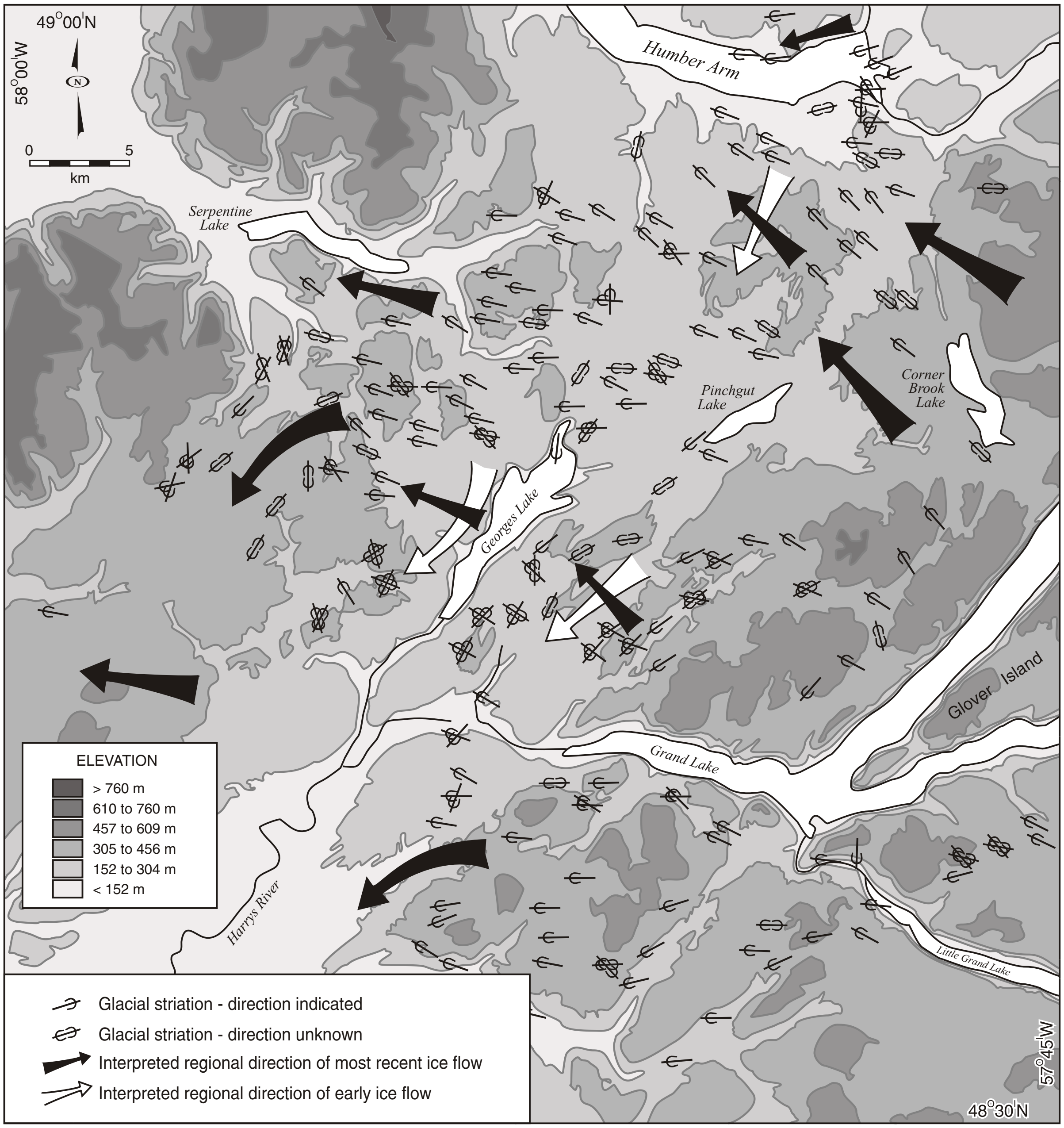


Figure 54. *Ice-flow patterns in the Corner Brook area and south.*

Birchy Lake–Sandy Lake

Along the north shore of Sandy Lake an early northeastward flow is recorded at several sites (Figure 59). A similar ice-flow direction is also recorded at scattered sites near Sheffield Lake and south of Upper Indian Pond. This ice flow was toward Green Bay and Halls Bay.

To the north of The Topsails, ice flow was either northward through the Sandy Lake basin or across the Birchy Lake valley toward White Bay. This ice flow crosscut the earlier northeastward ice-flow event recorded in the valleys. The exception is evidence for southward-flowing ice found at the northern end of Sandy Lake, close to Birchy Ridge. Less than 1 km east, ice-flow indicators record northward flowing ice. This suggests a complex interaction of southward flowing ice from the Long Range Mountains and northward flowing ice from The Topsails. The temporal relationship between these two flow events is not apparent, but clearly could not have occurred contemporaneously. Towards Sheffield Lake, the most recent ice flow is northeastward toward Green Bay and Halls Bay.

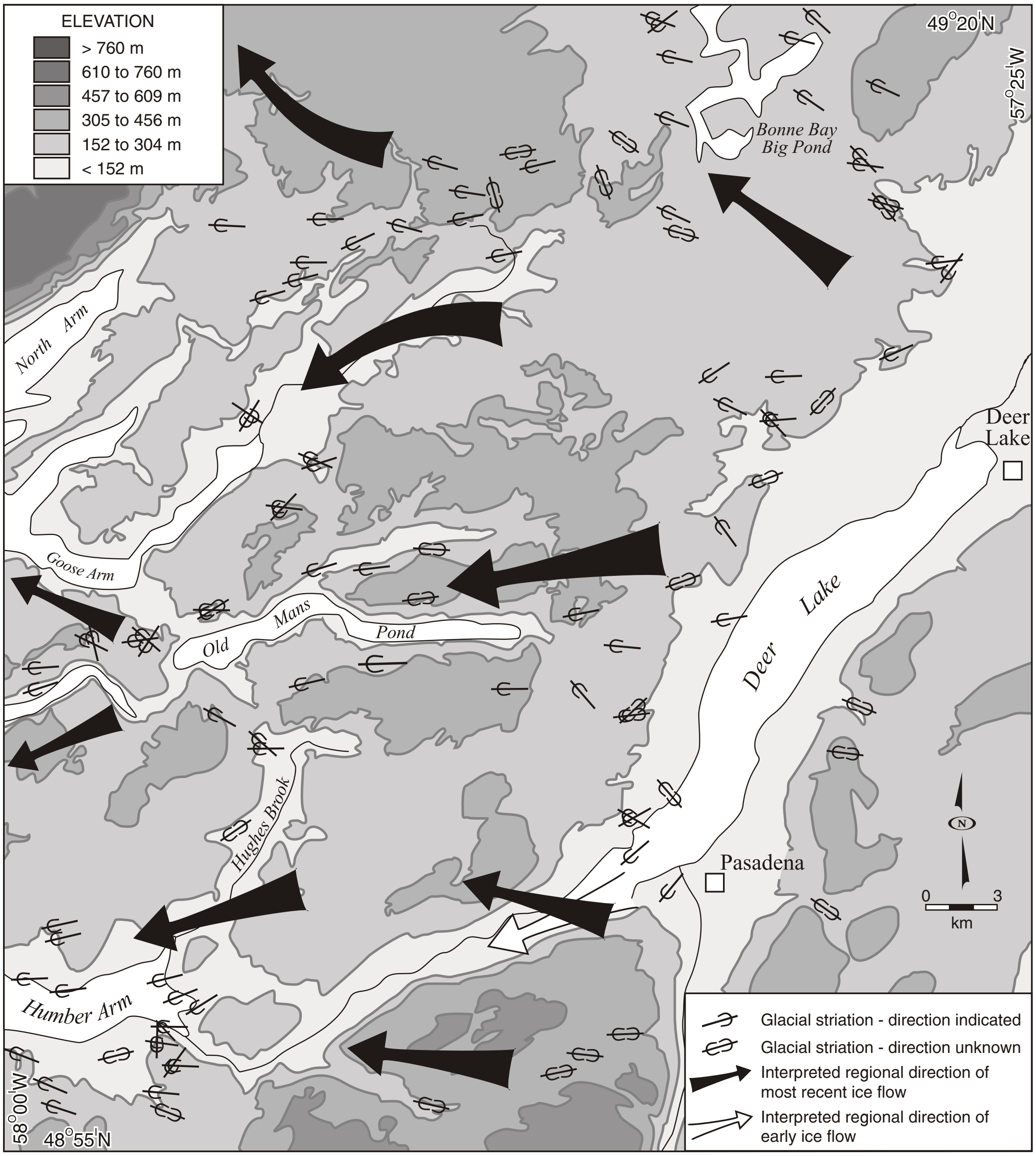


Figure 55. *Ice-flow patterns in the Deer Lake valley.*

SUMMARY OF ICE-FLOW EVENTS

The pattern of glacial striations and clast dispersal suggests a sequence of early flow from a source in the Long Range Mountains that covered the northern and western margins of the basin, followed by a regional radial flow from a dispersal centre on The Topsails that covered most of the basin apart from the Upper Humber River valley and north. Deglaciation produced local, small ice caps on coastal and interior hilltops.

All of the Humber River basin has been glaciated. Coastal highlands such as the Lewis Hills, Blow Me Down Mountain, and North Arm Mountain were thought to have been free of ice during the late Wisconsinan (Grant, 1989a, 1991) or throughout the Quaternary (Coleman, 1926). However, fresh striations and erratics from the Lewis Hills, Blow Me Down Mountain and North Arm Mountain indicate these areas were ice covered. Striations of 293° were reported from near Blow Me Down (elevation 640 m) by Batterson (*in* Taylor *et al.*, 1994). Peridotite clasts derived from a

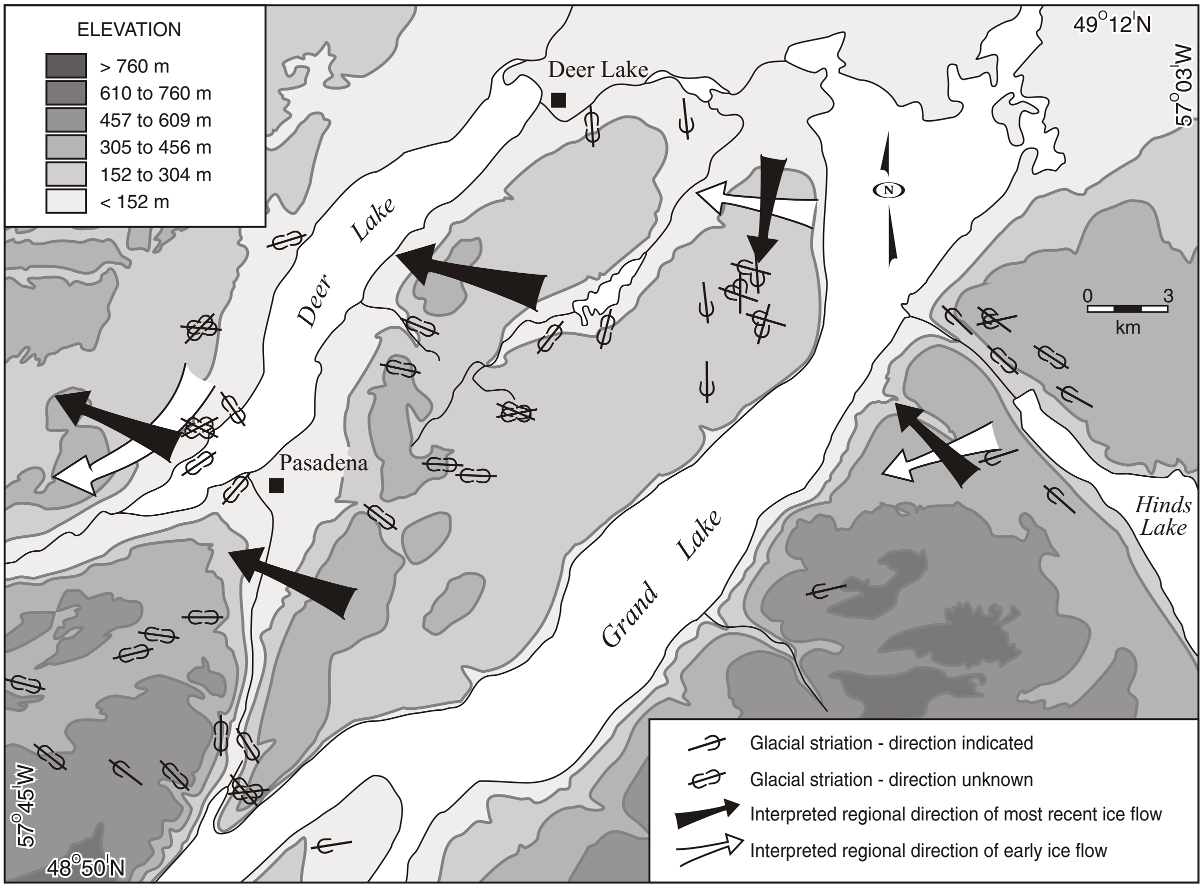


Figure 56. *Ice-flow patterns between Deer Lake and Grand Lake.*

source to the southeast were also found at the site. Taylor *et al*. (1994) also reported striations oriented 290 to 295° in the Frenchman's Cove area (elevation less than 50 m). A clast identified as a granite from the Topsails intrusive suite (probably Unit Sp) was collected on the southern part of the Blow Me Down massif (D. Taylor, Department of Mines and Energy, personal communication, 1995). Several sites on the Lewis Hills show striations, with trends of 240 and 310° from the west and east parts, respectively. The striations had no preserved directional indicators and no clasts were found indicating dispersal from The Topsails. Clasts identified as red Carboniferous sandstone and siltstone clasts were found on North Arm Mountain. This indicates glacial transport across the Deer Lake basin and overtopping of North Arm Mountain.

Glacial Flow from the Long Range Mountains

Ice from the Long Range Mountains covered the western and northern parts of the Humber River basin (Figure 60). Evidence (mostly striations) for an early southward flow is found within the Deer Lake valley, and in the Corner Brook area and south. Clast fabrics from diamicton exposed along the shores of the Humber Arm, and clast provenance data from near Pinchgut Lake provide further evidence for this flow. An early paleo ice-flow, roughly parallel to the west coast, has been described elsewhere. Mihychuk (1986) described southward flow along the coastal plain in the Bellburns area of the Great Northern Peninsula, interpreted as resulting from Laurentide ice. Grant (1994a) noted similarly oriented striations in the same area, and attributed them to piedmont glaciers from the Long Range Mountains. Early southward oriented striations also have been noted on the Port au Port Peninsula (MacClintock and Twenhofel, 1940; Taylor, 1994), formed possibly by ice from the Lewis Hills (Brookes, 1974). Taylor *et al*. (1994) also showed a southward (160°) flow on the coast near Cape Ray in the extreme southwest of Newfoundland, that may have resulted from an ice flow down the Laurentian Channel (Grant, 1987). Evidence for an early southward ice flow along the west coast of Newfoundland is thus fragmentary. All striations are unweathered and, where found, are the youngest.

The ice that eroded the southward striations within the Humber River basin had its source to the north of the basin, probably within the Long Range Mountains (Figure 61). The alternative explanation is that they were produced by Laurentide Ice. This hypothesis would require ice to cross the highlands between North Arm and Bonne Bay, or to flow into Bonne Bay and subsequently flow southward. There are no recorded striations to support either of these ideas. There is also no direct evidence (e.g., erratics) of Laurentide ice

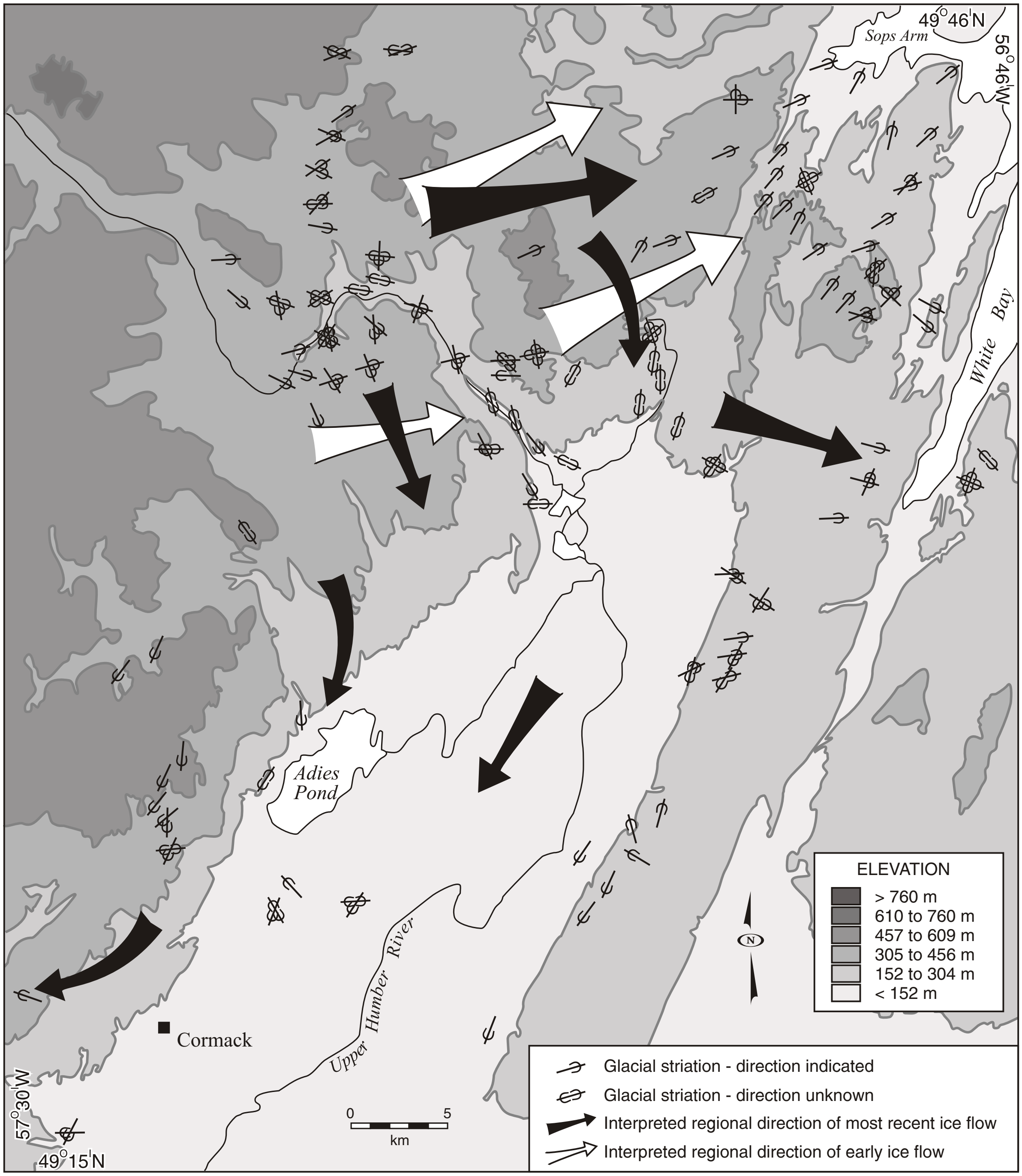


Figure 57. *Ice-flow patterns in the Upper Humber River valley.*

impinging on the coast south of the tip of the Northern Peninsula. The frequency of coast-parallel striations is, however difficult to explain. Although Laurentide ice may not have occupied the Humber River basin, it may have occupied the Esquiman Channel offshore. This ice would have deflected island-based ice southward. This hypothesis remains untested, and requires drilling offshore to identify diamictons with a Labrador clast provenance.

The northern part of the Humber River valley was covered by ice from the Long Range Mountains (Figure 61). Striations record an early ice flow from the mountains toward White Bay, crossed by striations produced by a later flow moving southeast to southwestward into the Humber River valley. In the area roughly defined by the highlands underlain by the Gull Lake intrusive suite, the only ice flow recorded is a northeastward flow into White Bay (Vanderveer and Taylor 1987; Taylor and Vatcher, 1993).

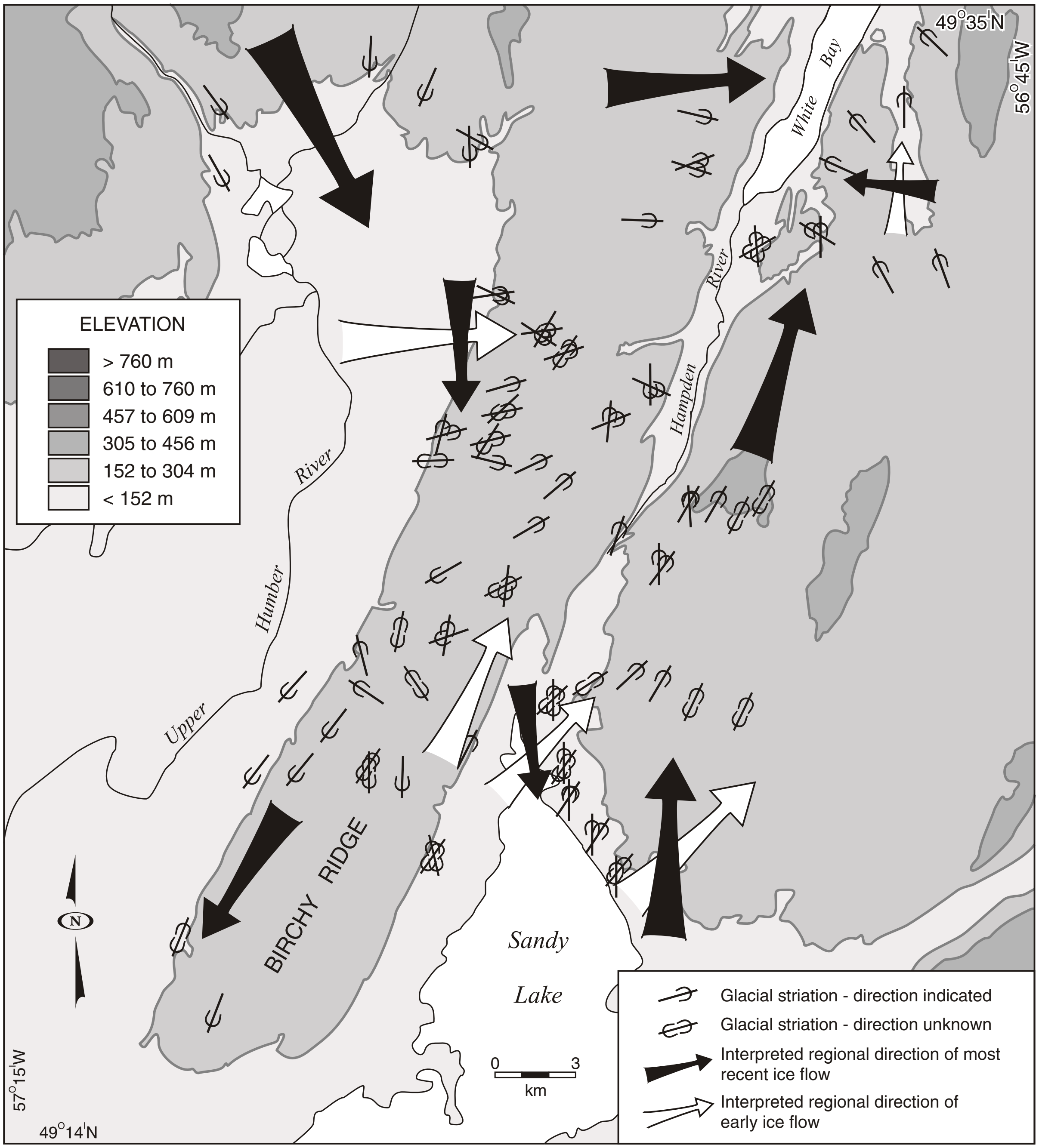


Figure 58. *Ice-flow patterns in the Birchy Ridge area.*

Ice from the Long Range Mountains ice centre entered the Upper Humber River valley and flowed southward toward the Deer Lake valley. This is shown by southwestward oriented striations in the Upper Humber River valley and along the shores of Deer Lake. The textural distribution of diamictons across the Humber River basin also demonstrates this flow. In the Upper Humber River basin diamictons are coarse, although the area is underlain by friable Carboniferous siltstone and sandstone. Similarly, diamicton matrix colour is commonly dark brown (10YR 3/3) rather than reddish brown that is characteristic of Quaternary sediments in many areas underlain by Carboniferous bedrock. Ice moving southwestward down the Humber River from a source in the Long Range Mountains would disperse gneiss clasts over the Carboniferous bedrock, and produce a coarser, brown till. Gneiss clasts dominate the clast assemblage in diamictons found in the Upper Humber River valley.

Ice from the Long Range Mountains remained in the Upper Humber River valley during the late Wisconsinan, as

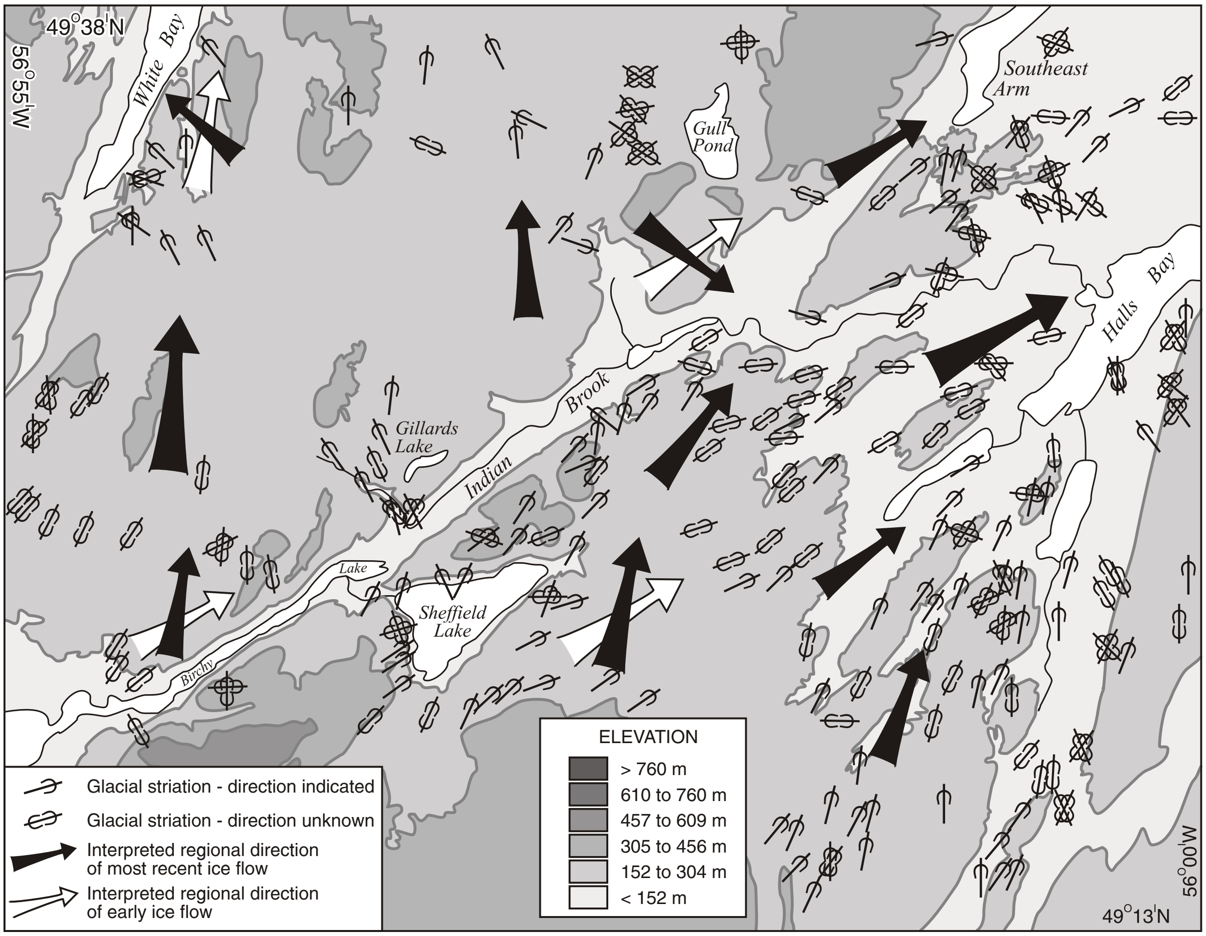


Figure 59. *Ice-flow patterns in the Birchy Lake to Sandy Lake area.*

indicated by the landform, clast fabric, striations and clast provenance data.

Glacial Flow from The Topsails

Most of the Humber River basin was covered by ice from a dispersal centre on The Topsails (Figure 62). Ice flow was generally radial. In the southwest of the basin, ice flow from The Topsails was either southwest toward St. George's Bay, or northwest toward Serpentine Lake or the Humber Arm. Numerous striations indicate ice was deflected southward around the eastern margin of the Lewis Hills.

The west to northwestward flow from The Topsails produced striations that crosscut earlier ones from southward flow down the Georges Lake valley. Similarly, evidence for an early flow down Deer Lake and the Lower Humber River valley is crosscut by a later flow oriented across the valley. Striations and clast dispersal data demonstrate this flow had its source in The Topsails, and extended out to the coast through the major fjords. The coastal highlands deflected ice flow, although striations and non-local clasts show that they were crossed by glaciers.

West of Deer Lake, diamictons are reddish brown (5YR 4/3) although they are underlain by grey limestone. Reddish brown diamictons have been reported as far west as Bonne Bay (Brookes, 1974), although the only local source for red sediments is the Deer Lake basin. Similarly, ice flow from The Topsails explains the relative coarseness of diamictons on the highlands east of Glide Lake, and the common occurrence of granite clasts. The area is underlain by Carboniferous rocks, but lies along a glacial flow path on The Topsails indicated by striations. The position of the Glide Lake highlands (elevation 320 m) as the first uplands west of Grand Lake would likely intercept material carried englacially (cf. Batterson, 1989; Liverman, 1992).

Ice flowed northward out of the Sandy Lake basin, toward White Bay. This is shown by clast dispersal data, particularly those from bedrock types restricted to The Topsails (e.g., units Sp, Sm, Sq and Oib). This corresponds to observations of striations and landforms by MacClintock and Twenhofel (1940) and Grant (1989a). Clast provenance and striation evidence suggests ice from The Topsails flowed northward onto the central and southern parts of Birchy Ridge. There is no evidence to suggest this ice flow

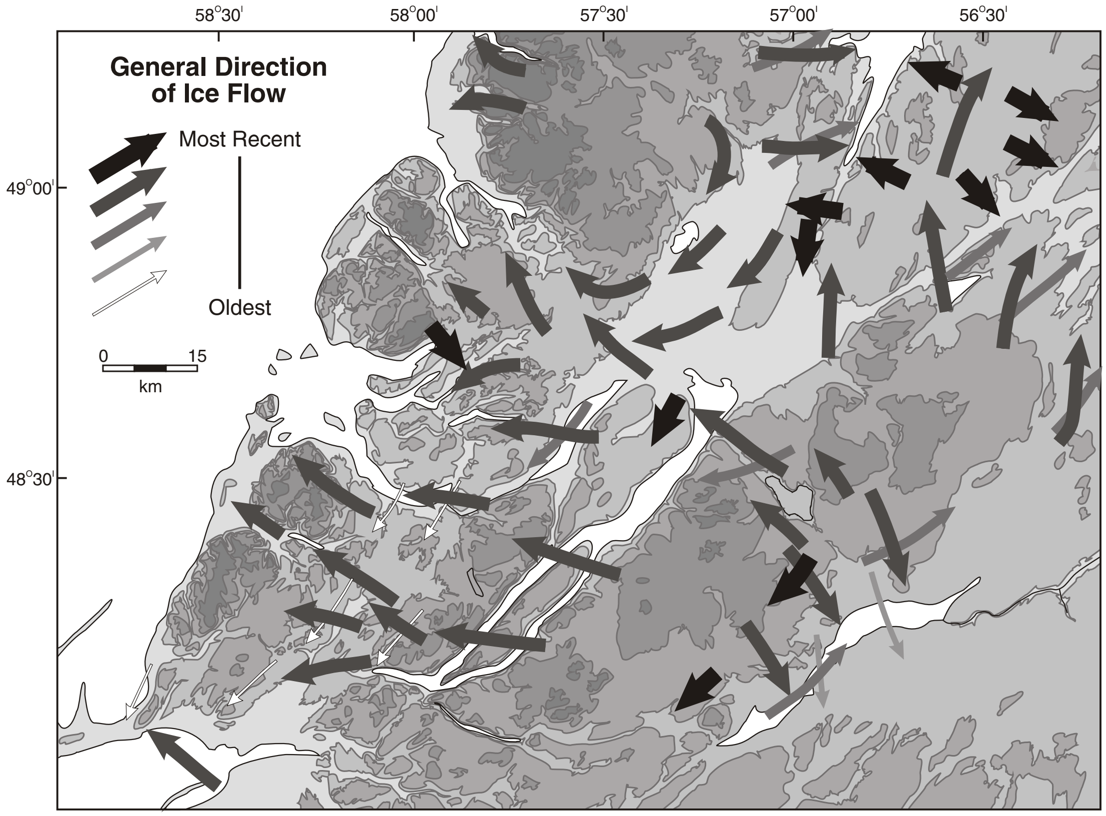


Figure 60. *Summary of ice-flow history across the Humber River basin.*

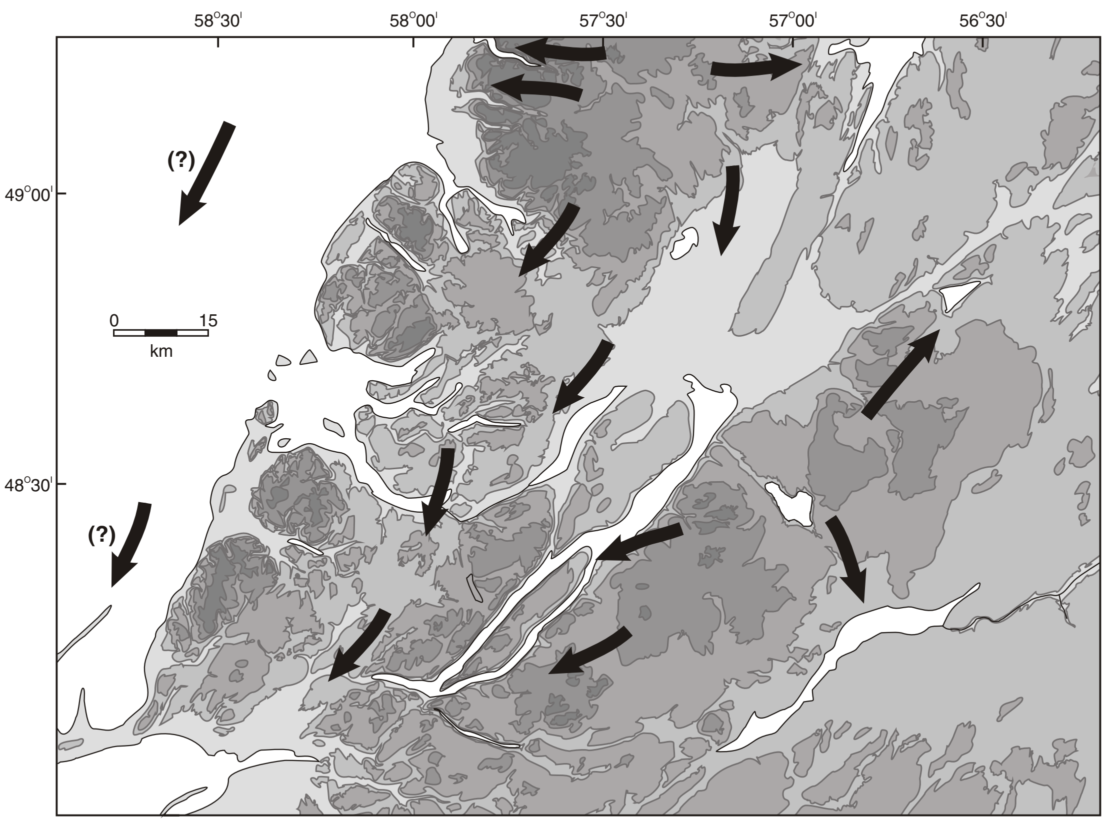


Figure 61. *Early ice flow in the Humber River basin.*

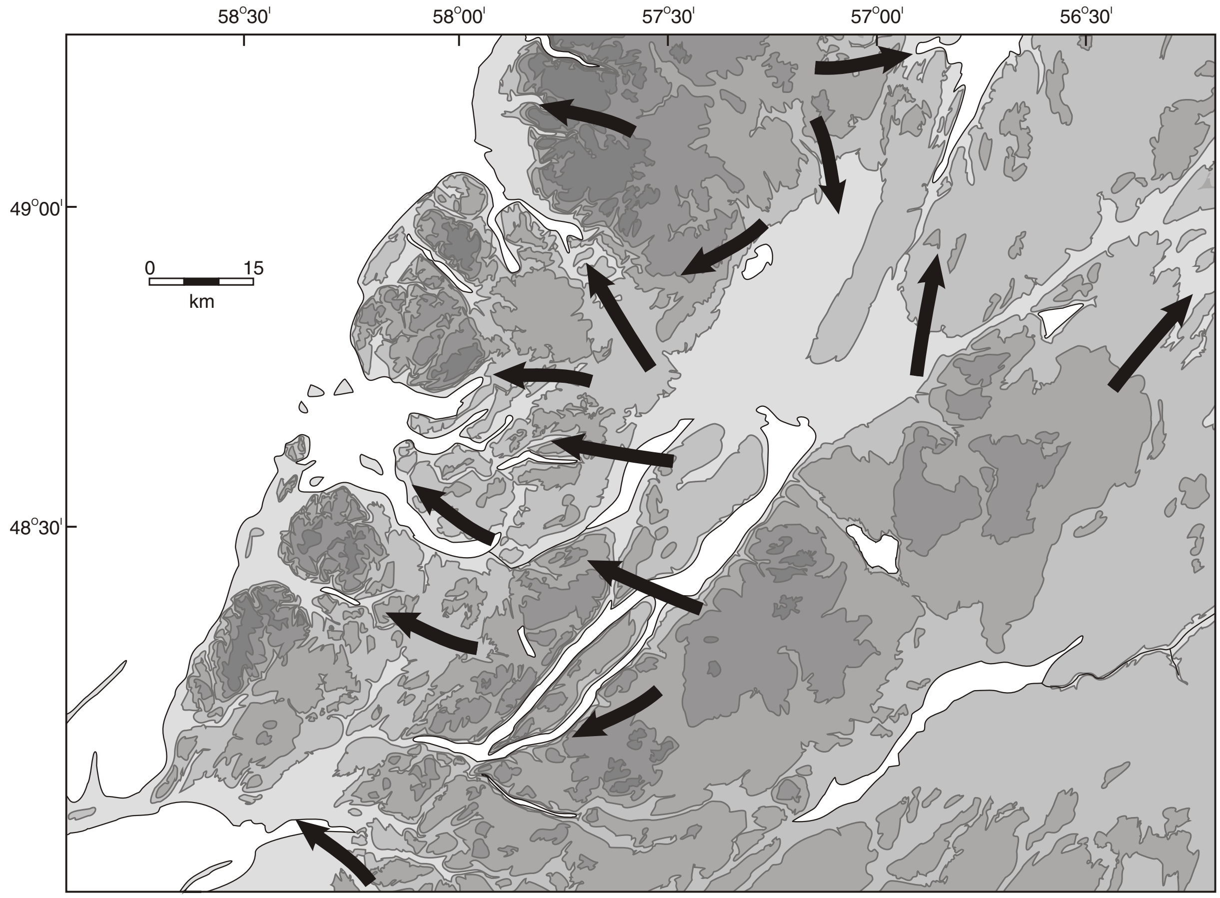


Figure 62. *Ice-flow patterns at the glacial maximum.*

crossed Birchy Ridge into the Upper Humber River basin. Further east, in the Birchy Lake valley, ice flow was northeastward toward the coast.

Insufficient data are available to determine the exact location of The Topsails dispersal centre, although it almost certainly was not static during the late Wisconsinan. Evidence for migrating ice centres has been described from other parts of Newfoundland and Labrador (e.g., Klassen and Thompson, 1993; Catto *et al.*, 1995). On The Topsails, an early westward ice flow is recorded by striations in the Hinds Brook valley, and as far east as the high plateau area northeast of Hinds Lake. Only one ice-flow direction is recorded by striations in other areas of The Topsails. Northwestward ice flow was recorded over the central parts of The Topsails as far east as the Hinds Lake to Buchans Lake area (Klassen, 1994). This ice flow extended down the Hinds Brook valley, and also formed the till ridges in the Goose Pond area, but did not cover the high plateau of The Topsails, which only has evidence for northeastward flowing ice. In the southwest of The Topsails, ice flow generally was westward. Vanderveer and Sparkes (1982) reported southwestward flow in the Star Lake area. There is only one striation site recorded in the central part of The Topsails between Hinds Lake and Rainy Lake. Interpolation from adjacent areas would suggest northwestward to westward ice flow in this area. Based on these limited data, it is likely a major ice dispersal centre was located on The Topsails having an ice divide extending along the southwestern margin of the plateau overlooking the Red Indian Lake valley.

Late Wisconsinan ice flow within the Upper Humber River valley and on adjacent highlands to the west was into and down the basin toward Deer Lake, as shown by striations and clast-provenance data. Clasts from The Topsails were not found in the valley. The southwestward ice flow is dominant as far south as Cormack. Southwest of this point, the increasing proportion of clasts from The Topsails suggests influence of northwestward flowing ice. Striations found west of Adies Pond are oriented increasingly westward the farther south they are traced, eventually becoming confluent with The Topsails-centred ice flowing northwestward toward Bonne Bay (Figure 62).

Late-Stage Ice Caps

Striations and clast fabric data supports the presence of several late-stage ice caps, produced during retreat of ice. These were located on Birchy Ridge, the highlands east of Glide Lake, and highlands adjacent to the modern coast (Figure 63).

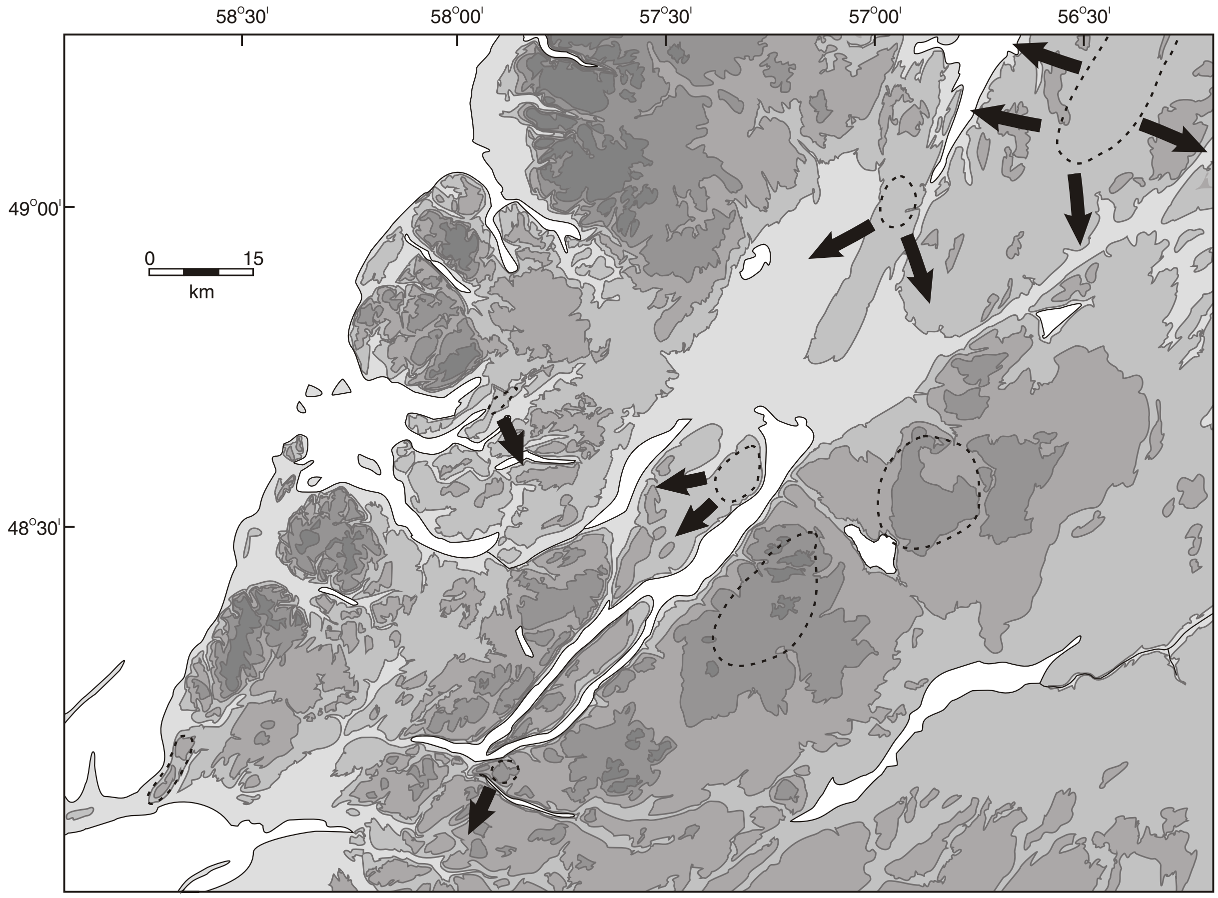


Figure 63. *Late-glacial dispersal patterns in the Humber River basin.*

Striation evidence suggests ice flowed southward over Birchy Ridge, although the presence of clasts from The Topsails shows that at least part of the ridge, especially north of the old TCH, was covered by northward flowing ice from The Topsails ice centre. The ice-flow history over Birchy Ridge is further complicated by southward striations found on the east side of Birchy Ridge at the head of Sandy Lake (Taylor and Vatcher, 1993). These suggest that southward flowing ice from the Long Range ice centre crossed Birchy Ridge and flowed into the northern part of the Sandy Lake basin. This ice flow did not extend as far west as the area underlain by the Gull Lake intrusive suite. Grant (1989b) also recorded a southward flow along the west side of Sandy Lake. Crosscutting striations show that this southward flow followed an earlier regional northward flow from The Topsails, and suggest it was a late-stage event that occurred at a time when ice from The Topsails had retreated from the basin.

Late-stage southward ice flow also is recorded on the highlands east of Glide Lake (Batterson and McGrath, 1993). No evidence was found to link this event with southward-flowing ice in Sandy Lake. Ice from this area flowed down the Pynn's Brook valley, as shown by till overlying sands (*see* section on Glacial Sediments and Stratigraphy, page 40).

Several sites were found that showed striations at variance with the regional ice-flow patterns. Near Goose Arm, striations record a late southeastward flow that postdated a regional southwestward flow into Goose Arm. A late southward ice flow was recorded at the mouth of Little Grand Lake, following the regional westward ice flow. A late south-southwestward flow was recorded at Bonne Bay Little Pond, where the regional flow was northwestward. In the headwaters of the Humber River, striations show a late westward ice flow toward the coastal mountains, where the regional flow was east to southeastward. In each case, the striations recording the late ice flow direction could not be traced over a distance larger than 1 km. These are all interpreted to represent late-stage movements from remnant ice centres that developed on highland areas during the waning stages of the main ice caps. Each of the sites is downslope of a highland area.

SEA-LEVEL HISTORY

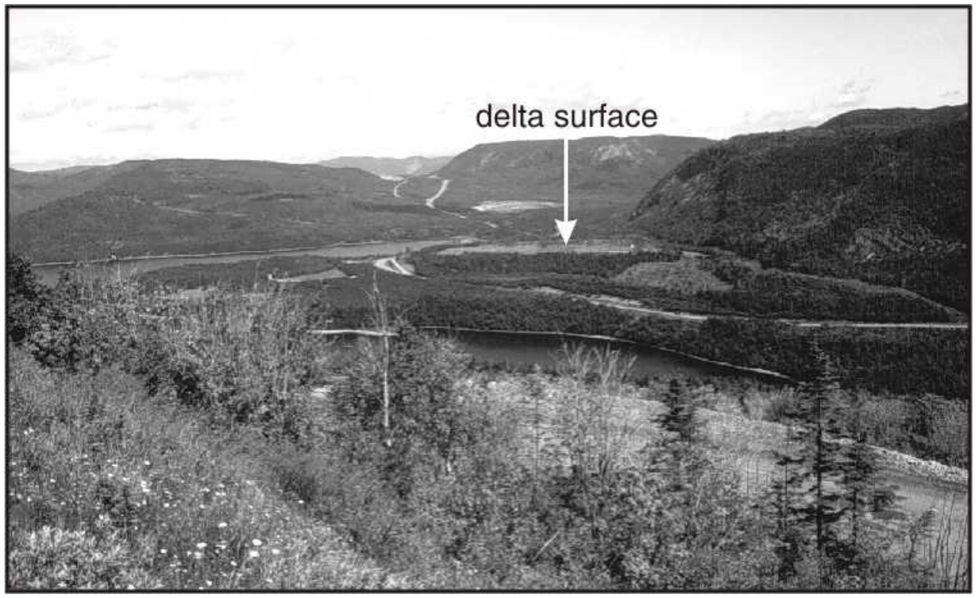


Plate 35. *Raised marine delta at Humbermouth, having a surface elevation of 50 m asl.*

During deglaciation, relative sea level was higher than at present and there was a protracted episode of standing water in the Deer Lake basin at elevations below 50 m asl, resulting in the deposition of deltas along the margins, and rhythmically bedded sediments in the basin. However, the presence of *Balanus hameri* shell fragments in the Humber River gorge, confirms that the Lower Humber River valley was inundated by the sea. The marine limit in the Bay of Islands is defined by raised deltas (Flint, 1940; Brookes, 1974), and dated at circa 12 600 BP. The delta at Humbermouth (49 m) (Plate 35) was an ice-contact feature (Brookes, 1974) that extended on both sides of the valley. Sediment exposed within the delta suggested that, "the snout of a valley glacier in the Humber River valley stood near the fjord head, when a proglacial delta was being built into the sea at 49 m" (Brookes, 1974, p. 18). It was dissected by meltwater, flowing through the Humber River gorge following retreat of ice from the Lower Humber River area. The deltas at Nicholsville, Junction Brook and elsewhere in the Deer Lake area must therefore have formed after those in the Humber Arm. The date of 12 220 ± 90 BP (TO-2885) from *Balanus hameri* fossils in the Humber River gorge suggests that marine invasion of the Deer Lake basin took place shortly after construction of deltas on the coast.

The samples analysed for micropaleontological analysis were taken from prodelta or bottomset beds on the east side of the Deer Lake basin. These prodelta or bottomset environments would be strongly influenced by inflowing meltwater, and likely had low salinities and a high suspended sediment content. This may explain the low faunal contents in sediments from the Deer Lake basin itself.

Seabrook (1962) in a survey of fish species in Deer Lake reported the presence of tom cod (*Microgadus tomcod*). Freshwater occurrences of tom cod are rare, and the Deer Lake fish are considered to be permanently landlocked populations (Scott and Crossman, 1973). Strong current flow through the Humber River gorge would preclude recent migration of cod into Deer Lake. Marine invasion and subsequent isolation of the Deer Lake basin early in the Holocene are a more likely explanation of their presence there.

There is no published relative sea-level curve for the Humber River basin. Regional sea-level studies have been undertaken in the St. George's Bay–Port au Port Bay area (Brookes *et al.,* 1985; Forbes *et al.,* 1993), and around Springdale (MacClintock and Twenhofel, 1940; Jenness, 1960; Tucker, 1974a). These areas form the southwest and northeast margins of the basin.

The pattern of relative sea-level change in areas adjacent to continental ice sheets is controlled by passage of an ice marginal forebulge, produced by crustal displacement (Clark *et al.,* 1978; Quinlan and Beaumont, 1981). Clark *et al*. (1978) suggested Newfoundland was transitional between zones of continual emergence and continual submergence. Quinlan and Beaumont (1981) identified four zones of relative sea-level change in Newfoundland, controlled by location relative to the migrating forebulge. Curves varied between continual emergence (Type A) and continual submergence (Type D). A curve showing initial emergence and subsequent submergence is characteristic of a Type-B curve (Quinlan and Beaumont, 1981, 1982).

One of the best documented and well-constrained sea-level curves in Newfoundland is from the St. George's Bay–Port au Port Bay area. Early work by Brookes (1977a) suggested initial emergence from a marine limit of 27 to 45 m asl, followed by a transition from emergence to submergence at about 11 500 BP, with a -15 m low stand at about 10 000 BP and relative sea level returning to near present at about 5500 BP (Figure 64). Brookes *et al*. (1985) refined this curve, based on radiocarbon dates from marine shells

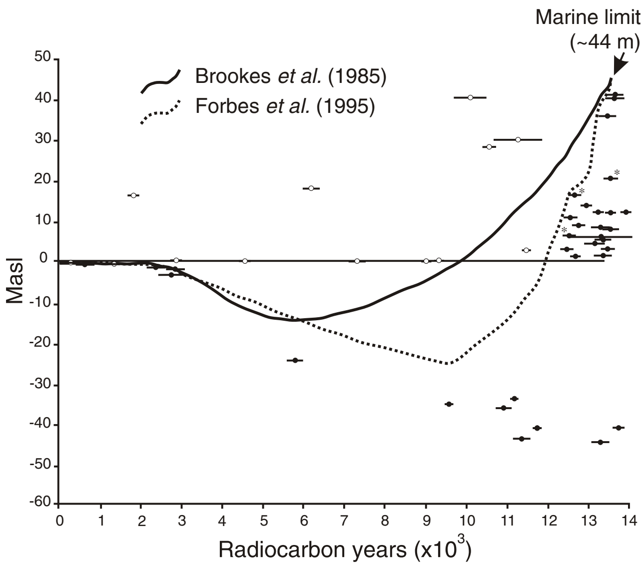


Figure 64. *Relative sea-level curves for St. George's Bay.*

and analysis of foraminifera and pollen, and showed the emergence–submergence transition was between 9800 and 9500 BP and a low stand of 11 to 14 m below present, at about 5800 BP, followed by gradual emergence to the present (Figure 64). Forbes *et al*. (1993) provided a further iteration. Using the elevation of submerged deltas and terraces, and additional radiocarbon dates, they demonstrated that the emergence–submergence transition occurred at about 11 700 BP where a lowstand at 25 m below present occurred at 9500 BP (Figure 64).

The Springdale area has a less well constrained relative sea-level history than St. George's Bay. Marine limit was about 75 m above present, based on the elevation of deltas at Springdale and around Hall's Bay (MacClintock and Twenhofel, 1940; Jenness, 1960) and dated at 12 000 to 11 000 BP (Grant, 1974; Tucker, 1974a); this age for the marine limit has been questioned by Scott *et al.* (1991). The implications are that at least the lower reaches of the Indian Brook valley were ice free at 12 500 BP, and that the marine limit deltas could predate 12 500 BP. There is no data on possible lowstands for this part of the coast, although Liverman (1994) has speculated that the area should have a Type-B curve based on the distribution of radiocarbon dates.

Bonne Bay at the northwestern extension of the study area was deglaciated before 12 400 BP, based on dates from Neddy Harbour (12 400 ± 140 BP, GSC-4553; Table 15), although inner parts of the bay may not have been ice free until some time later, possibly as late as 10 500 BP (Brookes, 1974). Marine limit in Bonne Bay was about 70 m above present (Brookes, 1974).

The preceding data suggest that the margins of the Humber River basin were deglaciated at about the same time, and that the marine limit was higher in the northeast part of the basin compared to the southwest. Isostatic tilt, of originally horizontal surfaces through the Humber River basin should now be inclined upward toward the northeast.

RELATIVE SEA-LEVEL CURVE: HUMBER ARM–BAY OF ISLANDS

A tentative relative sea-level curve was constructed for the Bay of Islands area, based on ^{14}C dates of marine shells (Figure 65; Table 17), and published data on lowstand features. The elevation of marine limit is based on the top of the Hughes Brook delta at 60 m. Shaw and Forbes (1995) identified a sea-level lowstand of 6 m below present at Corner Brook. Although this lowstand is not directly dated, interpolation between those dates derived for St. George's Bay (Forbes *et al.*, 1993; Shaw and Forbes, 1995) and Bonne Bay (Brookes and Stevens, 1985; Grant, 1972) suggests the lowstand was at about 7900 BP. Only one terrestrial date is recorded in the Bay of Islands area, that of 9050 ± 130 BP (GSC-4281) on gyttja at 52 m asl from York Harbour (McNeely and McCuaig, 1991).

Figure 66 shows three relative sea-level curves produced from this data. Dashed line 1 is a smooth, steep curve that incorporates all dates from the area. Extrapolation suggests that a marine limit of 60 m asl should date at about 12 200 BP. However, a marine shell sample from Goose Arm Brook at 50 m asl was ^{14}C dated at 13 070 ± 90 BP (TO-3624).

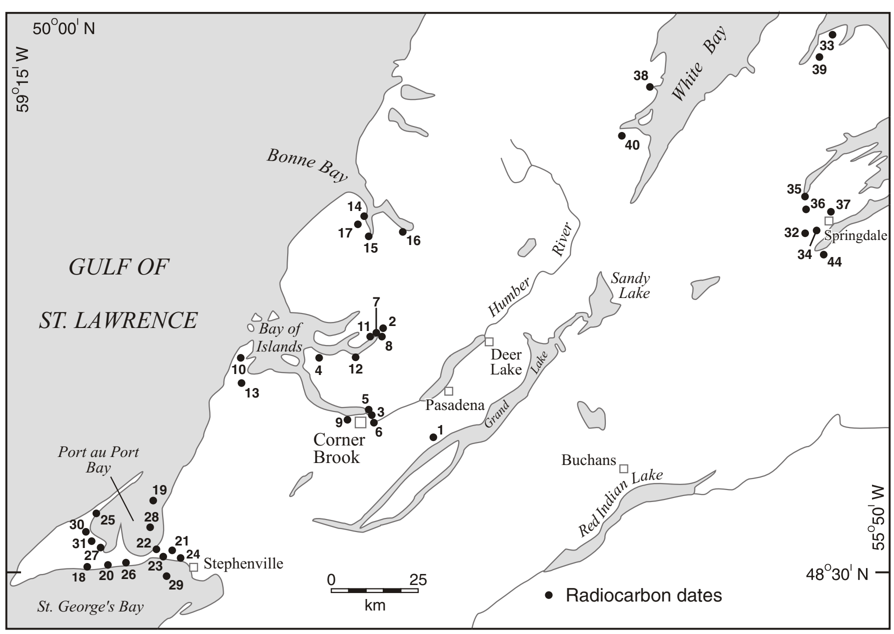


Figure 65. *Location of sites from which radiocarbon dates were derived.*

Table 17. Radiocarbon dates are referred to in the text and have been used to construct relative sea-level curves presented here. The assigned numbers (first column) correspond to sample points in Figure 65. Dates from the University of Toronto (designated TO) are adjusted using a 410 year correction

#	Location	Date (Corr.)	±	Lab. No.	Lat. (°N)	Long. (°W)	Elev. (m)	Source	Material
Humber Arm and Basin									
1	South Brook	13100	220	GSC-5302	48°54.9'	57°37.7'	135	Batterson *et al.*, 1995	Bulk organics
2	Goose Arm	13070	90	TO-3624	49°12.7'	57°49.7'	50	Batterson *et al.*, 1993	Shells (*Mya truncata*)
3	Dancing Point	12700	300	GSC-4272	48°57'	57°53'	15	GSC Paper 87-7	Shells (*Macoma balthica*)
4	Cox's Cove	12600	170	GSC-868	49°07'	58°05'	38	Brookes, 1974	Shells(*Hiatella arctica*)
5	Wild Cove	12450	90	TO-2884	48°58.3'	57°52.7'	18	Batterson *et al.*, 1993	Shells (*Mya truncata*)
6	Humber River gorge	12300	110	GSC-5300	48°56.9'	57°50.8'	13	This report	Shells (*Balanus hameri*)
6	Humber River gorge	12220	90	TO-2885	48°56.9'	57°50.8	13	Batterson *et al.*, 1993	Shells (*Balanus hameri*)
7	Goose Arm	12120	90	TO-3623	49°11.0'	57°51.8'	29	This report	Shells (*Mya truncata*)
8	Goose Arm	12100	130	GSC-5538	49°10.9'	57°51.5'	7	This report	Shells (*Nucalanapernula*)
9	Curling	12090	90	TO-4358	48°57.5'	57°59.3'	10	This report	Shells (*Mya truncata*)
10	Little Port	12000	320	GSC-1462	49°06.7'	58°24.8'	37	Brookes, 1974	Shells(*Mytilus edulis*)
11	Goose Arm	11900	120	GSC-5516	49°10.1'	57°53.1'	27	This report	Shells (*Mya truncata*)
12	Goose Arm	10600	100	GSC-4400	49°07.4'	57°56'	6	GSC Paper 89-7, p. 17	Shells (*Mya truncata*)
1	South Brook	9540	90	TO-5707	48°54.9'	57°37.7'	135	This report	Salix twig
13	York Harbour	9050	130	GSC-4281	49°02.8'	58°22.4	52	GSC Paper 89-7, p. 17	Gyttja
Bonne Bay area									
14	Muddy Brook	12100	160	GSC-4158	49°29.5'	57°55.6'	20	GSC Paper 87-7, p. 5	Shell (*Macoma calcarea*)
15	Glenburnie	11200	150	GSC-4279	49°26.1'	57°53.9'	15	GSC Paper 87-7	Shells (*Cyrtodaria siliqua*)
16	Lomond	11200	130	GSC-4790	49°27.3'	57°45.1'	2	GSC Paper 89-7, p. 17	Shells (*Macoma calcarea*)
16	Lomond	10500	300	GSC-1575	49°27.3'	57°45.3'	4	Brookes 1974	Shells (*Macoma calcarea*)
17	Bonne Bay	1730	140	GSC-1266	49°28.1'	57°57.4'	260	GSC Paper 71-7, p. 260	Peat
St. George's Bay area									
18	Fiods Cove	26600	550	GSC-4563	48°30.9'	58°57.4'	35924	GSC Paper 89-7, p. 18	Shells (*Mya, Astatre, Nuculana*)
19	Port au Port Bay	13710	115	Beta 30002	48°43'	58°43'	–41	Forbes and Shaw 1989	Shells (*Portlandia arctica*)
20	Abraham's Cove	13700	230	GSC-1074	48°31.5'	58°55'	45	GSC Paper 71-7, p. 263	Shells (*Hiatella arctica*)
21	Romaines Brook	13680	100	TO-6137	48°34.1'	58°40.3'	37	This report	Shells(*Mya truncata*)
20	Abraham's Cove	13600	110	GSC-2015	48°32'	58°54'	40	GSC Paper 75-7, p. 6	Shells (*Hiatella arctica*)
20	Abraham's Cove	13600	180	GSC-968	48°31.5'	58°55'	7.5	GSC Paper 71-7, p. 262	Shells (*Hiatella arctica*)
22	Port au Port	13400	290	GSC-1187	48°33.8'	58°42.6'	2	GSC Paper 71-7, p. 261	Shells (*Balanus* sp.)
23	Romaines	13345	230	S-3074	48°33'	58°41'	35953	Grant 1991	Whale bone
24	Stephenville	13300	810	GSC-2063	48°32'	58°37'	35920	Brookes 1977a	Shells
25	Rocky Point	13200	220	GSC-937	48°39.1'	58°57.4'	3.7	GSC Paper 70-2, p. 51	Shells (*Mya arenaria*)
23	Romaines	13100	180	GSC-4095	48°33.2'	58°41'	3	GSC Paper 87-7, p. 6	Shells (*Mya truncata*)
26	Campbells Cove	13300	120	GSC-4346	48°31.5'	58°51.2'	11	GSC Paper 87-7	Shells (*Hiatella arctica*)
27	Piccadilly	13000	110	GSC-4584	48°34.2'	58°54.6'	14	GSC Paper 89-7, p. 18	Shells (*Mya truncata*)
23	Romaines	12800	130	GSC-5030	48°33.2'	58°41.0'	35829	GSC Paper 95-G, p. 13	Shells (*Hiatella arctica*)
23	Romaines	12800	130	GSC-4858	48°33.2'	58°41.0'	35953	GSC Paper 89-7, p. 18	Shells (*Hiatella arctica*)
23	Romaines	12700	110	GSC-4017	48°33.3'	58°41.1'	1	GSC Paper 87-7, p. 6	Plant debris
24	Kippens	12610	100	TO-6138	48°32.7'	58°37.8'	8	This report	Shells(*Hiatella arctica,Macoma Calcarea*)
24	Kippens	12600	120	GSC-5942	48°32.6'	58°38.0'	21	This report	Shells (*Mya truncata*)
24	Stephenville	12600	140	GSC-2295	48°32.6'	58°36.7'	10	Brookes 1977b	Shells
19	Port au Port Bay	11740	100	Beta 30004	48°43'	58°43'	–41	Forbes and Shaw 1989	Shells (*Astarte undata*)
23	Romaines	11500	100	GSC-4291	48°33.3'	58°41.1'	1	GSC Paper 87-7	Peat
28	Port au Port Bay	11300	100	Beta 30005	48°38'	58°43'	–34	Forbes and Shaw 1989	Shells (*Astarte undata*)
28	Port au Port Bay	11165	95	Beta 30003	48°38'	58°43'	–34	Forbes and Shaw 1989	Shells
28	Port au Port Bay	9570	150	GSC-4724	48°38'	58°43'	–34	Forbes and Shaw 1989	Shell (*Spisula polynyma*)
19	Port au Port Bay	5800	210	GSC-1203	48°43'	58°50'	24	GSC Paper 71-7, p. 262	Shells (*Hiatella arctica*)
29	St. George's Bay	3695	95	Beta 30001	48°31'	58°40'	–42	Forbes and Shaw 1989	Polychaete worm tubes
30	Victor's Brook	2840	80	GSC-4243	48°37.8'	58°58.5'	3.2	GSC Paper 87-7	Wood
31	Hynes Brook	2770	300	UQ-646	48°36'	58°57'	2.8	Brookes *et al.,* 1985	Peat
31	Hynes Brook	2365	175	GX-9527	48°36'	58°57'	1.8	Brookes *et al.,* 1985	Peat
Hall's Bay area									
32	Hall's Bay	12470	380	TO-2305	49°30'	56°11'	29	Scott *et al.*, 1991	Shell (*Mya truncata*)
33	Deer Cove Pond	12400	110	GSC-4700	50°0.6'	56°03.3	72	GSC Paper 89-7, p. 13	Shell (*Hiatella arctica*)
34	South Brook	12000	220	GSC-1733	49°25.5'	56°06.5'	20	GSC Paper 83-7, p. 6	Shell (*Balanus* sp.)
35	Southwest Arm	11880	190	GSC-87	49°35'	56°12'	12	GSC Paper 63-21	Shell
36	Kings Point	11800	200	GSC-3957	49°31.2'	56°11.8'	102	GSC Paper 86-7, p. 4	Gyttja
37	Hall's Bay	11340	150	TO-2306	49°30'	56°10'	24	Scott *et al.*, 1991	Organic detritus
38	Jacksons Arm	11200	100	GSC-4247	49°51.7'	56°49.0'	22-28	GSC Paper 87-7, p. 8	Shell (*Mya truncata*)
39	Ming's Bight	10400	160	GSC-3966	49°57.0'	56°05.3'	122	Dyer, 1986	Gyttja
40	Sops Arm	10200	100	GSC-4023	49°43.5'	56°55.7'	27	GSC Paper 87-7, p. 8	Shell (*Mya truncata*)
41	Hall's Bay	7890	80	TO-2304	49°30'	56°11'	27	Scott *et al.*, 1991	Shell (*Mya arenaria*)

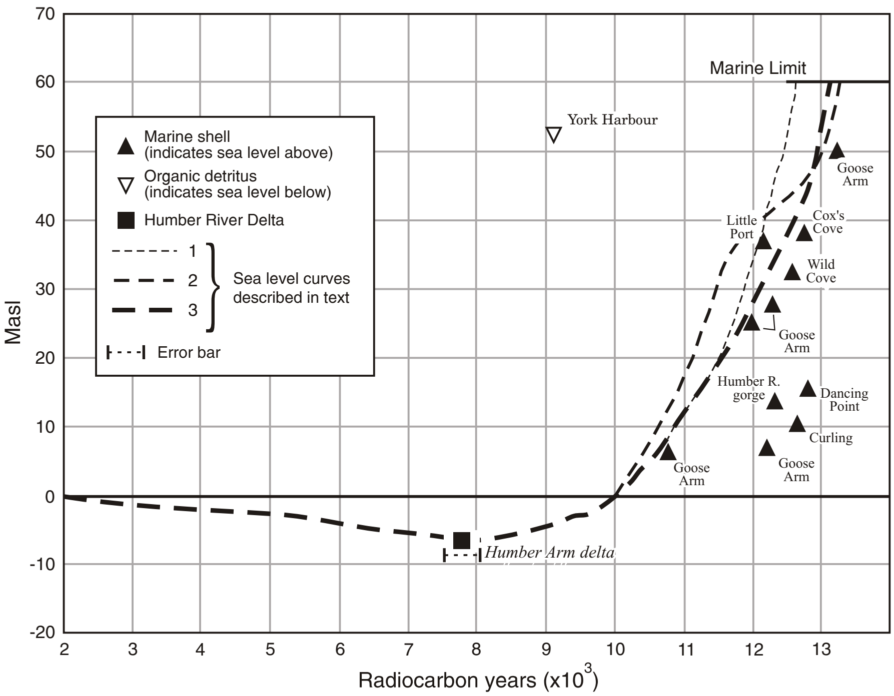


Figure 66. *Relative sea-level curve for the Humber Arm area.*

Dashed line 2 shows a perturbation in the curve around the Little Port date. The older part of the curve is more in accordance with the Goose Arm Brook date, and shows the marine limit at about 13 200 BP. Forbes *et al*. (1993) recognised a similar distortion in their curve for St. George's Bay. The perturbation there occurs at about 12 600 BP at 14 to 16 m asl. Terraces, produced during a stillstand have been recorded around St. George's Bay at this elevation (e.g., Brookes, 1974) although the causes for a stillstand are unknown. The distortion in the Bay of Islands curve occurs at between11 500 and 12 000 BP at about 37 m asl. No associated terraces or deltas have been recorded in the Bay of Islands at this elevation (cf. Brookes, 1974; Flint, 1940). The date from Little Port of 12 000 ± 320 BP (GSC-1462) appears to be anomalous when compared to other dates in the area. The date was interpreted by Brookes (1974) as dating the 47 m asl marine limit for the area. This is similar to the 49 m asl delta at Cox's Cove dated at 12 600 ± 170 BP (GSC-868). Brookes (1974) questions the reliability of the Little Port date (although with the large error bars the dates are statistically similar), and suggests the difference in dates may be attributed to ice lingering in constricted valleys on the outer coast. Such an age difference should also be shown in an elevational difference of delta surfaces, due to the influence of isostatic rebound, but this is not the case. The Little Port date is therefore considered spurious.

Dashed curve 3 incorporates all dates apart from Little Port. The upper part of the curve is based on the Goose Arm Brook date, and places the marine limit at about 13 200 BP. The central part of the curve is based on dates in the Goose Arm area and the delta at Cox's Cove. The lower part is constrained by the date from the central part of Goose Arm. In the absence of conflicting data, a smooth relative sea-level curve is considered a reasonable first approximation of events in the Bay of Islands.

DISCUSSION

The Bay of Islands relative sea-level curve is a Type-B curve using the terminology of Quinlan and Beaumont (1981, 1982), showing initial emergence and subsequent submergence of the coastline. The details of the curve are consistent with the model of relative sea-level change for Newfoundland developed by Liverman (1994), using the ^{14}C date compilation of Batterson *et al.* (1992). Liverman's model uses the distribution of ^{14}C dated marine molluscs to predict the date at which sea level fell below present, which is then tested against existing relative sea-level curves. The model predicts that a sea level fall below that of present should have occurred about 10 500 BP for the Bay of Islands. The relative sea-level curve presented above suggests this transition occurred about 500 years later.

The relative sea-level curve for the Bay of Islands area is in qualitative agreement with that predicted by the maximum ice model of Quinlan and Beaumont (1981, 1982). Their minimum ice model suggested the Bay of Islands should be a Type-C curve where no sea-level features above present are found. Grant (1989a) used this model to support the concept of limited ice over Newfoundland during the late Wisconsinan. The timing of the transition from emergence to submergence, and the elevation and date of the marine limit are quantitatively different from Quinlan and Beaumont's model. Liverman (1994) described similar incompatibilities for Newfoundland, which is characterised mainly by Type-B relative sea-level curves. This is in contrast to the narrow belt of Type-B curves described by Quinlan and Beaumont (1981, 1982), and suggests an underestimation of the area affected, and speed of migration of the collapsing forebulge (Liverman, 1994).

The data shows a disparity between model predictions and field data, that are likely the result of the geological inputs to the initial model. Following the work of Quinlan and Beaumont (1981, 1982), different geophysical inputs have been applied to models considering relative sea-level changes in areas marginal to continental ice sheets (e.g., Lambeck *et al.*, 1990, 1996; Mitrovica and Peltier, 1995; Davis and Mitrovica, 1996). Although beyond the scope of this report, it is perhaps an appropriate time to apply newer ice and Earth inputs in models of relative sea-level change in Newfoundland, to be tested against an increasing body of field data.

QUATERNARY HISTORY

PHASE 1: GLACIATION

Plate 36. *View of North Arm Mountain.*

The entire Humber River basin was glaciated during the late Wisconsinan. Striated bedrock surfaces and exotic clast rock types were found on the highest hills within, and along the margins of the basin (Plate 36). Diamictons are fresh, and striated bedrock surfaces unweathered. The Rocky Brook till, suggested by Vanderveer and Sparkes (1982) to be pre-Wisconsinan, remains undated due to a lack of suitable dateable material. The ^{14}C dating of marine shells stratigraphically above the diamictons shows that glaciation predated about 13 000 BP. Thus, it is most reasonable to assume that tills were deposited during the late Wisconsinan. Exposures with multiple till units are interpreted to have been deposited during a single glacial episode. Definitive Pre-late Wisconsinan deposits were not recognised within the study area.

The Humber River basin was covered by ice advancing from two major dispersal centres, the Long Range Mountains and The Topsails. Initial advance was from the Long Range Mountain ice cap southward into the Upper Humber River valley, flowing into the Deer Lake basin; the southwestern extent of this flow is uncertain. Striation and clast provenance data from the Corner Brook area, near Georges Lake, and near Stephenville indicate an early southward flow, but it is uncertain if these features can be correlated.

No evidence has been found to show this early flow covered the Grand Lake valley, although the lack of striated bedrock surfaces and the lack of distinctive rock types to trace dispersal, does not preclude the possibility. During this early phase of glaciation the extent of Topsails-centred ice remains unresolved.

Striation and clast dispersal patterns are dominated by radial ice flow from an ice centre on The Topsails extending along the southwestern margin of the plateau overlooking the Red Indian Lake valley. In the Hinds Lake area, ice flow was northwestward, whereas flow on the high plateau to the east was northeastward. This constrains the position of the northeast part of the dispersal centre. Topsails-centred ice crossed the Glide Lake highlands and Deer Lake, flowing toward the coast. It covered the lower part of the Upper Humber River valley south of Cormack. In the north of the Humber River basin, ice crossed Sandy Lake and Birchy Lake and flowed toward the coast into White Bay or Hall's Bay, while in the southwest, ice flow was toward St. George's Bay or through Serpentine Lake.

Limited evidence, mostly erratics, shows that ice crossed the coastal highlands from the interior. The elevation of coastal highlands range from 814 m asl for the Lewis Hills, 650 m asl for Blow-me-Down Mountain, to 706 m asl for North Arm Mountain. If the erratics are assumed to have been deposited during the late Wisconsinan, this represents a minimum ice thickness over The Topsails of 260 m (the elevation difference between the two areas). The surface slope of most glaciers is small, between 1:100 and 1:1000 (Drewry, 1986), suggesting an ice thickness of 350 to 1200 m.

The striation pattern shows that ice was deflected by coastal highlands. Ice flow was northwestward near the community of Deer Lake, but was drawn southwestward toward the coastal fjords of North Arm, Goose Arm and Penguin Arm. Similarly, westward ice flow across the Georges Lake valley was deflected southward by the Lewis Hills.

PHASE 2: DEGLACIATION

Deglaciation of the Bay of Islands commenced at about 13 000 BP. The inner Humber Arm was ice free by 12 700 to 12 500 BP, based on dates from Wild Cove and Dancing Point. This is similar timing to deglaciation of the southwest and northeast margins of the Humber River basin (Brookes, 1974; Macpherson and Anderson, 1985; Scott *et al.*, 1991).

The distribution of eskers, meltwater channels and hummocky moraine suggests some directions and areas of ice wastage. The distribution of hummocky moraine in the central and eastern parts of the basin shows ice retreated across the highlands west of the Humber River valley, and stagnated within the Upper Humber River valley, on the Glide Lake highlands, in the lowlands around Howley, and on The Topsails. These areas also contain meltwater channels.

The lack of meltwater channels over the highlands west of Deer Lake, and in the southern part of the Grenville Inlier suggests that ice retreated rapidly from these areas, and did not waste *in situ*. In contrast, the numerous meltwater channels on The Topsails plateau show this to be a site of ice disintegration. Meltwater channels spread radially outward into the Hinds Brook and Kelvin Brook valleys, and into the Sandy Lake–Grand Lake basins from the 340 m asl plateau between Goose Pond and Hinds Lake. The presence of meltwater channels across the slope between this plateau surface and the higher level at about 520 m asl to the east of Goose Pond, suggests ice remained on the lower plateau as the slopes to the upper plateau became ice free.

The radial pattern of meltwater channels also suggests that remnant ice was located on the southern end of Birchy Ridge, on the highlands between the Glide Lake and Deer Lake valleys, overlooking the South Brook valley, and on The Topsails southwest of Hinds Brook. These remnant ice centres remained active, at least for a short period, as indicated by striation patterns. In the Glide Lake–Pynn's Brook area, till found overlying glaciofluvial sand–gravel is suggestive of a local readvance down the Pynn's Brook valley.

PHASE 3: GLACIAL LAKE DEVELOPMENT

Features and sediment related to deposition in standing water are found on the shores of Grand Lake and in adjacent valleys, such as South Brook (Figure 67). They were formed within either a marine or lacustrine depositional environment. The surface elevation of deltas are at least 70 m above known marine features adjacent to the modern coast. Features deposited in a marine environment found at modern elevations of 145 m asl are found in southern Labrador (Grant, 1992), approximately 300 km north of Grand Lake. Sediments around Grand Lake contained no marine macro- or micro-fossils. The sedimentology and elevation of the deposits indicate deposition in a glaciolacustrine environment.

Deposition was either in a single glacial lake, or in a series of smaller ice-marginal lakes. A single lake would require that all shoreline features be at similar elevations, or must show a progressive change related to isostatic deformation; it must be able to include all shoreline features, without unreasonably complex or numerous ice dams; and any postulated spillway must be at the lowest elevation on the lake margin, in an area free of ice dams.

The formation of proglacial lakes during regional stagnation of ice masses in an undulating terrain has been documented in many places, including the Ural Mountains (Arkhipov *et al.*, 1995); the Scottish Highlands (Sissons, 1979; Sissons and Cornish, 1983; and Lowe and Cairns, 1991); the Interior Plateau of British Columbia (Fulton, 1969, 1975); Alberta and northeastern British Columbia (St. Onge, 1972; Mathews, 1980), and in the Mackenzie River valley of the Northwest Territories (Smith, 1992, 1994). In contrast to the strongly developed strandlines of lake basins in some lowland areas, such as the Great Lakes basin (Karrow and Calkin, 1985; Colman *et al.*, 1994; Lewis *et al.*, 1994), and southern Lake Agassiz (Fisher and Smith, 1994; Teller and Clayton, 1983; Teller *et al.*, 1983), lacustrine features in areas of variable topography tend to be discontinuous and scattered. Identification of individual lake stages and recognition of their spatial and temporal relationships is based on the presence and elevations of spillways, strandlines, deltas, and sediments. A lake developed over a broad area is delineated by assemblages of features at similar elevations or a consistent trend of elevations, where affected by isostatic rebound. In contrast, sediment and landforms associated with small, isolated transitory lakes formed at the margins of a downwasting ice mass will be found over a range of elevations, with each lake producing a local series of unrelated features.

The spatial and elevation distribution of features related to higher water levels in the Grand Lake basin (Figure 68) suggest formation in a single glacial lake, occupying most of the basin with a shoreline at approximately 120 m asl in the south to 140 m asl in the north. The name ‘glacial Lake Howley' was proposed for this body of water (Batterson *et al.*, 1993).

Three reconstructions are presented for the extent and duration of glacial Lake Howley: a maximum, minimum, and intermediate view.

Maximum Reconstruction

This is the reconstruction of Batterson *et al.* (1993; Figure 69) that proposed a lake more than 125 km long, and up to 25 km wide having a surface area of about 1850 km^2. Impoundment of a lake having a shoreline at approximately 140 m asl in the Grand Lake basin requires damming of two low points on the basin margin, the Lower Humber River valley (likely in the Humber River gorge) (± 0 m asl) and the Birchy Lake valley (± 110 m asl). The origin of the Humber River gorge ice dam was speculated to be dead ice isolated during retreat of The Topsails ice. The lake was proposed as covering the entire Humber River valley, including the modern day Grand Lake, Sandy Lake, Deer Lake and Birchy Lake valleys having an outlet at the western end of Grand

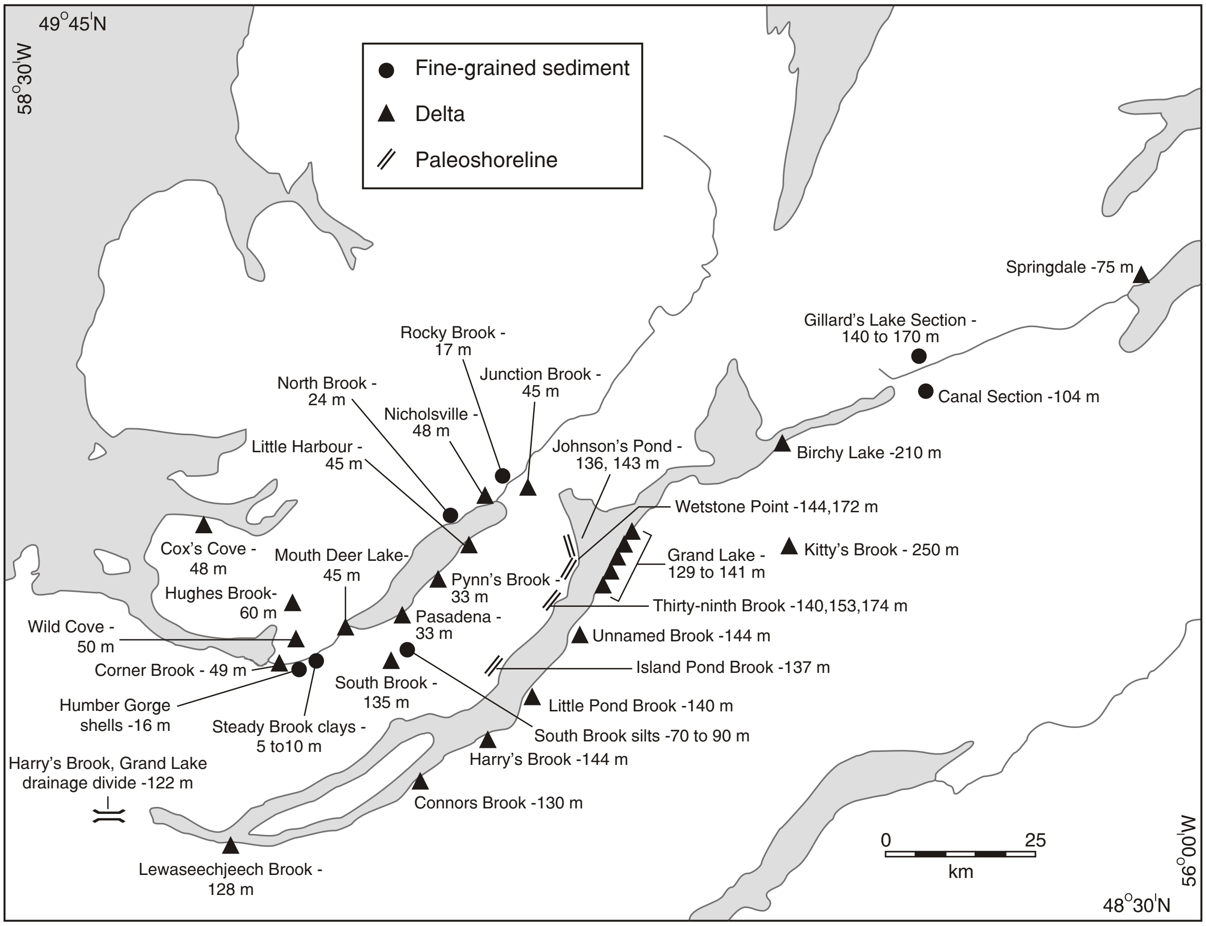


Figure 67. *Location map showing features related to higher water levels in the Humber River basin.*

Lake. This was a hypothesis based on the data known as of 1993, and involved extrapolating the extent of the lake along topographic contours into those areas that had not been extensively mapped at that time.

Lake extent was based mainly on sedimentological and geomorphological descriptions from exposures along the margins of the proposed glacial Lake Howley. Liverman and St. Croix (1989b) described sections in the upper Indian Brook valley near Gillards Lake as interbedded gravels and diamictons, and beds of rhythmically bedded silt and fine sand, containing dropstones. They interpreted the sections as representing deposition within an ice-contact lake.

Lundqvist (1965) described a section on the watershed between Birchy Lake and Indian River at 104 m as composed of lacustrine sediments. Liverman (personal communication, 1995) visited the site in 1989 and described rippled sands and rhythmically bedded silt and clay overlain by gravel and coarse sand. The lower part of the sequence is interpreted to have been deposited in a proglacial lacustrine environment. No diamicton interbeds or dropstones were noted in this exposure. Lundqvist (1965) interpreted this section to represent a small, ice-contact lake in the Indian Brook valley. The location of the section on the drainage divide between Birchy Lake and the Indian Brook valley implies that an ice dam must have been present to the east to preclude eastward drainage. Liverman and St. Croix (1989b) described deltaic sediments at Baie Verte Junction, at about 60 m asl, showing water flow to the west, opposite to the modern drainage direction. This evidence indicates that an ice dam did not exist to the west of the watershed section.

The pattern of differential isostatic rebound across the area is difficult to assess, as the only firm data points are the marine limit estimates on the coast. Estimates exist of 60 m asl for Corner Brook, 40 m asl at Stephenville (Brookes, 1974), and 75 m asl for the Springdale area (Tucker, 1974a), and the times of deglaciation are thought to be similar for both the northeast and west coasts. The effect of isostatic rebound likely increases from southwest to northeast. A lake shoreline formed shortly after deglaciation would probably now show a tilt up to the northeast.

If the Humber River gorge and the Birchy Lake valley were dammed, the lake would drain through the lowest available point on the basin edge. This is the drainage divide at the southwestern end of Grand Lake, which currently lies

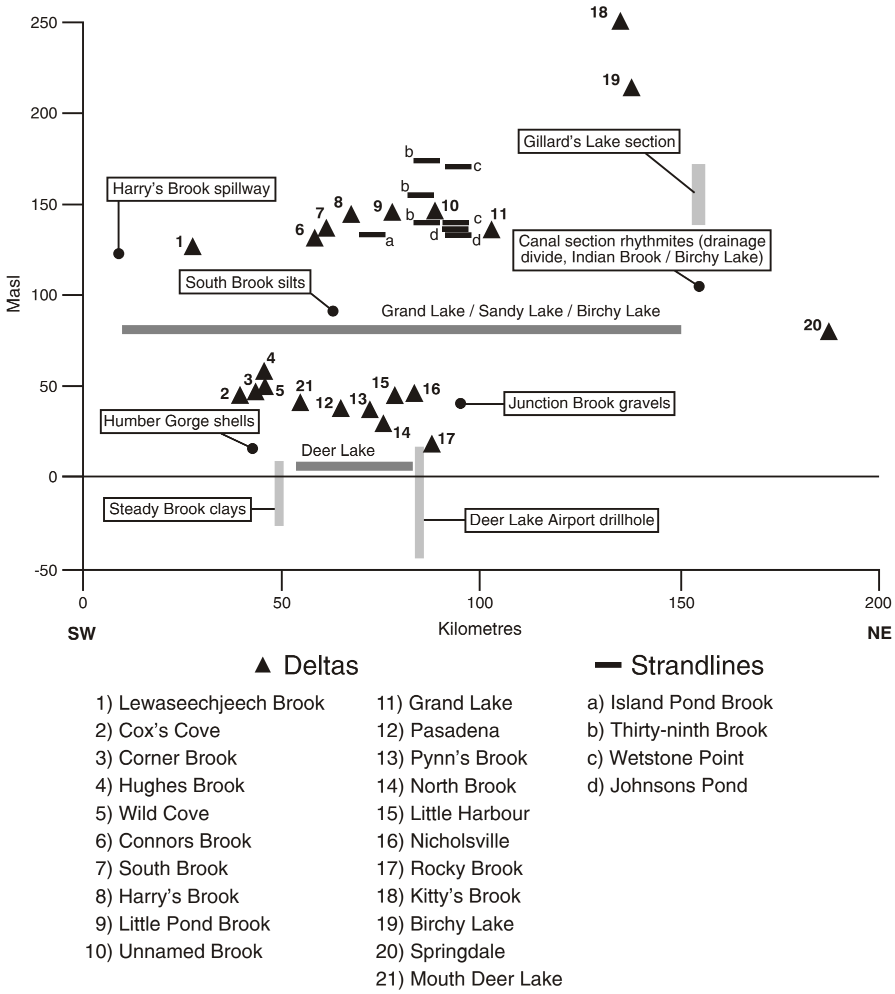


Figure 68. *Spatial and elevational distribution of features related to higher water levels in the Humber River basin.*

at 122 m asl. This divide consists of a low valley between Grand Lake Brook (flowing east into Grand Lake) and Ahwachenjeech Brook, a tributary of the southward flowing Harrys River. Aerial photographs (Plate 37) show a single, well-defined channel, 170 to 400 m wide having a gradient of 1:95, extending about 11.5 km south from Grand Lake toward Harrys River (Plate 38). Grant (1991) recognised the extension of this large proglacial channel down the Harrys River valley, entering St. George's Bay at Stephenville Crossing. This feature is the spillway resulting from overflow of an enlarged Grand Lake. The difference between the elevation of this overflow at 122 m asl and the deltas at 140 m asl at the north end of Grand Lake may be due to a combination of differential isostatic rebound or erosion. Sediment interpreted as lacustrine having a surface elevation of about 120 m asl, were described from the sand pit along the Grand Lake road. These sediment show that glacial Lake Howley extended west of the modern lake margin, current flow was westward away from the lake, and the lake surface was at least 120 m asl. The paleo-channel becomes well established west of the sand pit, indicating the sands were deposited close to the outlet of the lake.

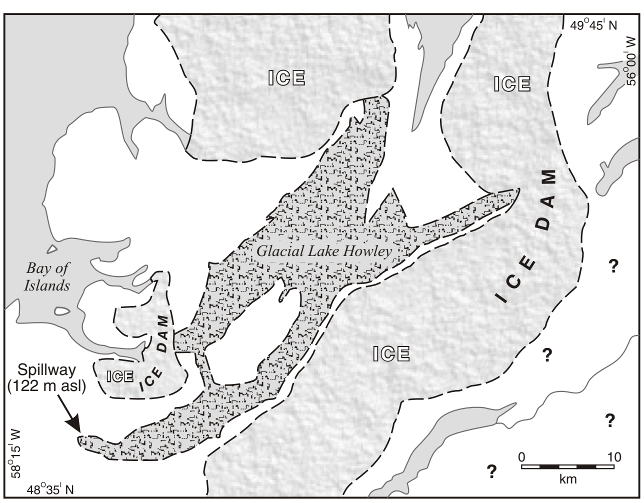


Figure 69. *Maximum reconstruction of glacial Lake Howley (*after *Batterson* et al.*, 1993).*

Sediment interpreted as proglacial outwash was described from the Gallants Pit along the channel margin. Other poorly exposed deposits of sand and gravel are found along the western margins of Grand Lake, up to an elevation of about 135 m. No sections were found exposing glaciofluvial sediments capped by till or other features (e.g., eskers) suggesting subglacial deposition, and these sediments are thus interpreted as sandur deposits. The sandur was produced by melting ice at the western end of Grand Lake, as suggested by Brookes (1974) and Grant (1991). The channel is incised into these sediments and therefore must postdate them. The channel does not cross the regional topographic gradients that may indicate sub-glacial water flow under pressure (Rothlisberger, 1972), and does not have an undulating longitudinal profile typical of tunnel valleys (e.g., Gilbert, 1990; Brennard and Shaw, 1994). Although the channel crosses the low divide between Grand Lake Brook and Ahwachenjeech Brook, it slopes away from glacial Lake Howley. Tunnel valleys are recognised by an anabranching network of channels incised into bedrock or sediment, an undulating linear

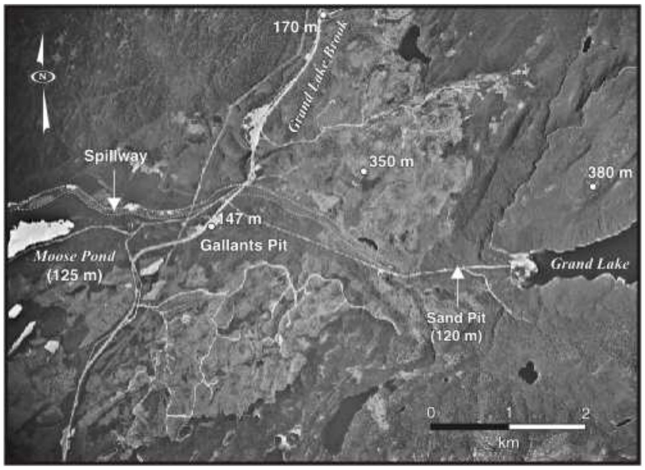


Plate 37. *Vertical aerial photograph of part of the spillway from glacial Lake Howley, extending from the southwestward end of Grand Lake toward Moose Pond.*

Plate 38. *View of Harrys River spillway from near Moose Pond.*

profile, a channel path against the regional gradient, glacial or glaciofluvial deposits, particularly eskers, within the channel, and the truncation of subglacially produced glacial landforms by the channel (Mooers, 1989; Gilbert, 1990; Shaw, 1983; Brennard and Shaw, 1993). The channel at the western end of Grand Lake is incised into sediment and bedrock, but shows none of the other characteristics normally associated with tunnel valleys. Thus, formation as a proglacial channel is more likely than a subglacial one.

Ice retreated rapidly from the western end of Grand Lake following a period of ice-contact sedimentation. This is shown by the lack of sediment along the shores of Grand Lake, except in the raised deltas at the mouths of many of the modern streams.

The reconstruction of Batterson *et al.* (1993) can no longer be supported in total. Field mapping in the Upper Humber River valley (Batterson and Taylor, 1994) showed no evidence of high-level deltas, or the presence of lacustrine sediments. In the Cormack area, surface sediments commonly are diamictons of local provenance. Toward Alder Pond, adjacent to the Humber River, diamictons are overlain by fluvial sands. In the Adies Pond area, a pronounced esker ridge overlies diamicton. The esker, which was also mapped by Grant (1989b), shows that ice retreated up the valley.

An ice-contact delta having a surface elevation of 46 m asl at the mouth of Deer Lake, exposed during road construction in 1994–1995 is also incompatible with the reconstruction presented by Batterson *et al.* (1993), which shows the ice dam seaward of this position.

Ice-flow data also supports ice cover extending southward beyond the Upper Humber River basin. Southward-directed striations are found at the north end of Sandy Lake. Similarly, southward striations are found on Birchy Ridge, and also on the highlands east of Glide Lake, which crosscut striations formed by the regional westward flow from The Topsails. Southward oriented striations in the Sandy Lake basin are only found on the north and west side.

Minimum Reconstruction

A minimum reconstruction is that glacial Lake Howley occupied only the area of the modern Grand Lake basin. The ice-contact delta at the mouth of Deer Lake, and eskers near Adies Pond and adjacent to the Humber River show that ice retreated up the Humber River valley toward the Long Range Mountains. Therefore, the Humber River valley was filled with ice during the glacial Lake Howley phase in Grand Lake, and the surface of glacial Lake Howley was controlled by the elevation of the outflow at the western end of the lake. Some of the high-level beaches recorded on the west side of the lake may have been cut during this early phase of lake development.

During deglaciation, glacial Lake Howley developed rapidly as ice retreated across the Grand Lake basin. The lake was most likely a time-transgressive feature and had a continually changing geometry in response to wasting ice in the Grand Lake–Sandy Lake lowlands suggesting an unstable configuration.

As ice retreated northward, up the valley, from the delta at Deer Lake (Figure 70), it exposed South Brook valley, allowing drainage from glacial Lake Howley (Figure 71). As a consequence, the lake surface dropped to about 145 m asl, the elevation of the col at the southern end of South Brook. This would allow formation of the prominent deltas found south of Hinds Brook along the east side of Grand Lake.

Raised deltas, on the east side of Grand Lake, increase in surface elevation from 128 m asl, at Lewaseechjeech Brook at the western end of Grand Lake, to about 145 m, north of Hinds Brook. This represents an increase of 0.22 m km^{-1} toward the northeast. This value does not necessarily represent isostatic tilt, because individual deltas are likely progressively younger toward the northeast. The rate is similar to a 0.20 m km^{-1} increase between Stephenville and Springdale, calculated from known paleo-sea-level data (Grant 1989a; Scott *et al.*, 1991; Forbes *et al.*, 1993).

The next low point farther up-valley is between the Glide Lake highlands and Birchy Ridge, now occupied by the modern outflow of Grand Lake through Junction Brook; this area must have been dammed by ice. The distribution of meltwater channels on the Glide Lake highlands, and the southern part of Birchy Ridge suggests melting of ice on highlands north and south of Junction Brook. Ice retreat up the valley was rapid, indicated by the similar elevation of the delta at the mouth of Deer Lake and those at the head of the lake. As ice retreated across the Junction Brook lowland it allowed drainage of glacial Lake Howley (Figure 72), through a spillway currently occupied by Junction Brook.

Junction Brook flows through a narrow, steep sided, flat-bottomed channel incised into bedrock, up to 450 m wide and 40 m deep that descends steeply about 30 m over 4.3 km (Plate 39). It has two small tributaries, both of which are less than a kilometre long. Junction Brook likely was formed as ice

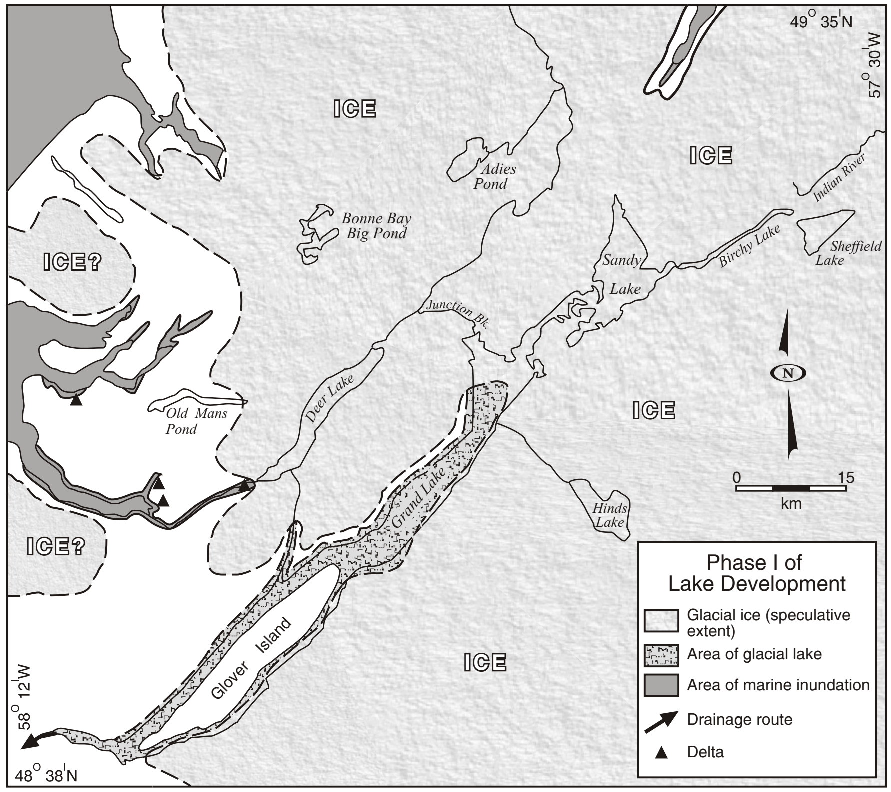


Figure 70. *Minimum reconstruction of glacial Lake Howley (Phase I).*

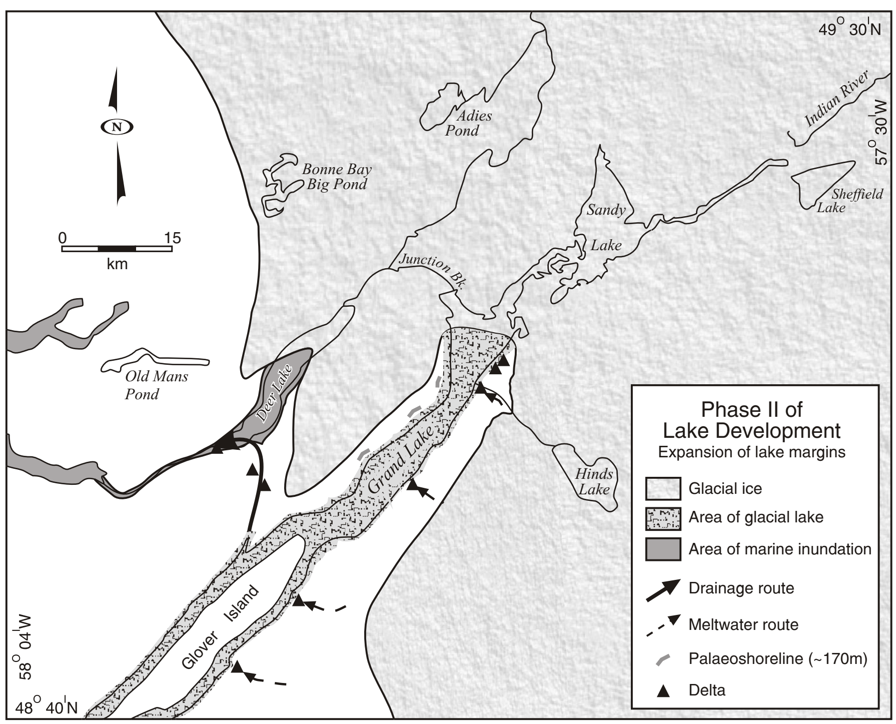


Figure 71. *Phase II of lake development. Expansion of lake margins.*

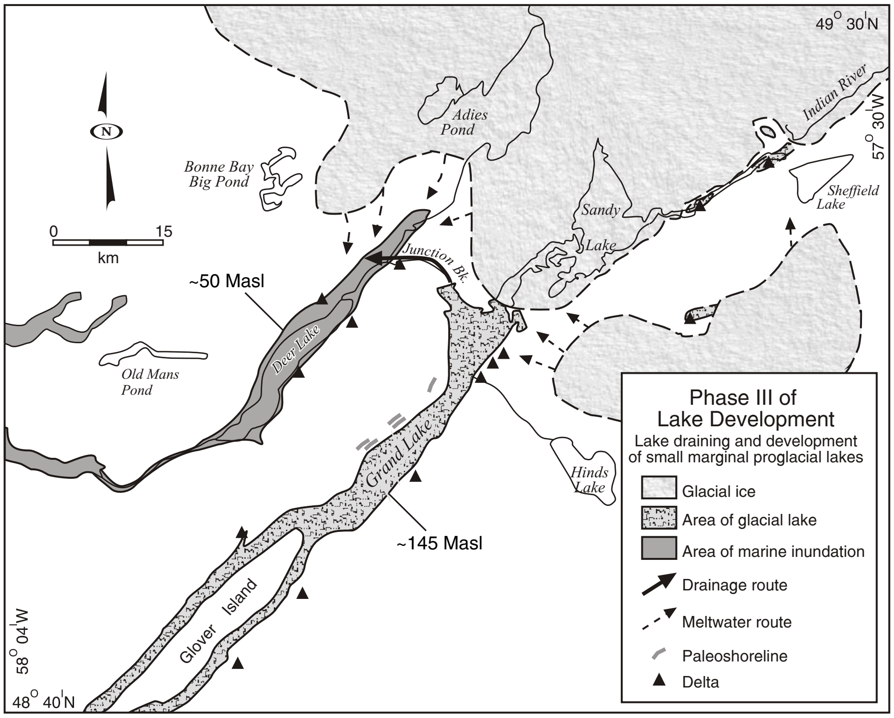


Figure 72. *Phase III of lake development. Lake draining and development of small ice marginal proglacial lakes.*

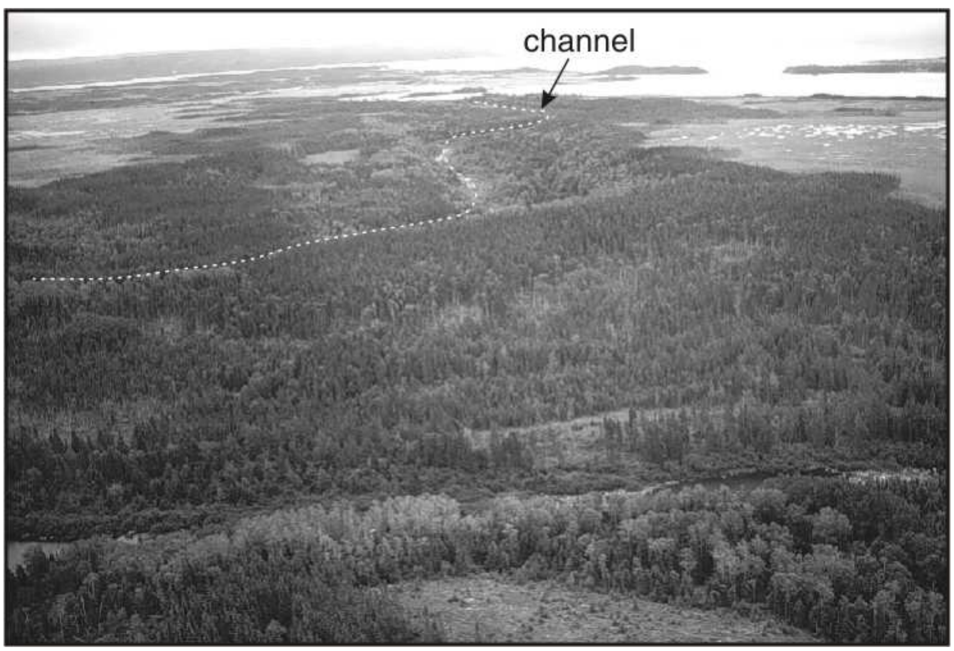


Plate 39. *Oblique aerial view of Junction Brook.*

retreated northward, and its formation accompanied a drop in lake level from 145 m (controlled by the South Brook col), to the pre-modern dam level of about 75 m asl, a drop of 70 m. Some lake drainage may also have been through the tributary south of Junction Brook, which is also incised in its lower reaches.

An estimate of potential discharge through the Junction Brook spillway may be calculated through application of the Manning equation. The Manning equation states:

$$u = \frac{1.009}{n} \cdot R^{2/3} \cdot \sqrt{slope}$$

where u=stream velocity, n=Manning number, R=hydraulic radius of the channel.

The Manning number takes into account the form of the channel (straight or meandering), and the sediment within the channel (clay, sand, gravel). The Junction Brook channel is generally straight and has a gravel base, producing a Manning number of 0.05. The cross sectional area of the channel at the upstream end is about 2500 m^2, giving a value for R of 13.8. The slope is calculated to be 0.0070. Assuming the channel was cut instantaneously and maintained bank-full levels, this gives a velocity of 4.1 m sec^{-1}, and a maximum discharge of 10 200 m^3 sec^{-1}. The surface area of the proglacial lake described under the minimum reconstruction is about 365 km^2. A drop in water level of 71 m produces a volume of 2.59 x 10^{10} m^3, not accounting for the effect of isostasy or basin geometry, or the slope of the basin margins. Given the maximum discharge of 10 200 m^3 sec^{-1}, a lake with the volume calculated would drain in approximately 29 days.

The peak discharge calculated for glacial Lake Howley is comparable to observed discharges on modern glacier dammed lakes (Walder and Costa, 1996). The discharge is small compared to rates calculated for glacial Lake Agassiz discharge that commonly exceeded 100 000 m^3 sec^{-1} (Teller and Thorleifson, 1987; Teller, 1987).

The channel was not cut instantaneously and is unlikely to have maintained bank-full levels during its development, and thus the assumptions used in these calculations are not necessarily valid. However, this exercise shows that the channel was cut over a short period of time and the whole lake could have drained in less than a month.

The minimum view suggests that events in the Birchy Lake valley are not associated with the development of glacial Lake Howley. The presence of rhythmically bedded silts and clays in the Gillards Lake basin and in the canal connecting Indian Brook and Birchy Lake could individually be interpreted as ice-marginal ponding, as suggested by Liverman and St. Croix (1989b) and Lundqvist (1965), respectively. Ice that occupied the northern end of Grand Lake may have separated Birchy Lake from Grand Lake. Ice disintegration is shown by the ice-flow data on the west side of the lake, the dead-ice topography in the Howley area, and an esker mapped west of Howley. The age of these features is unknown. Hummocks in the Howley area contain clasts derived from The Topsails, suggesting this as the source area. There is no evidence to support the view that ice from the Long Range Mountains ice cap extended as far east as Howley.

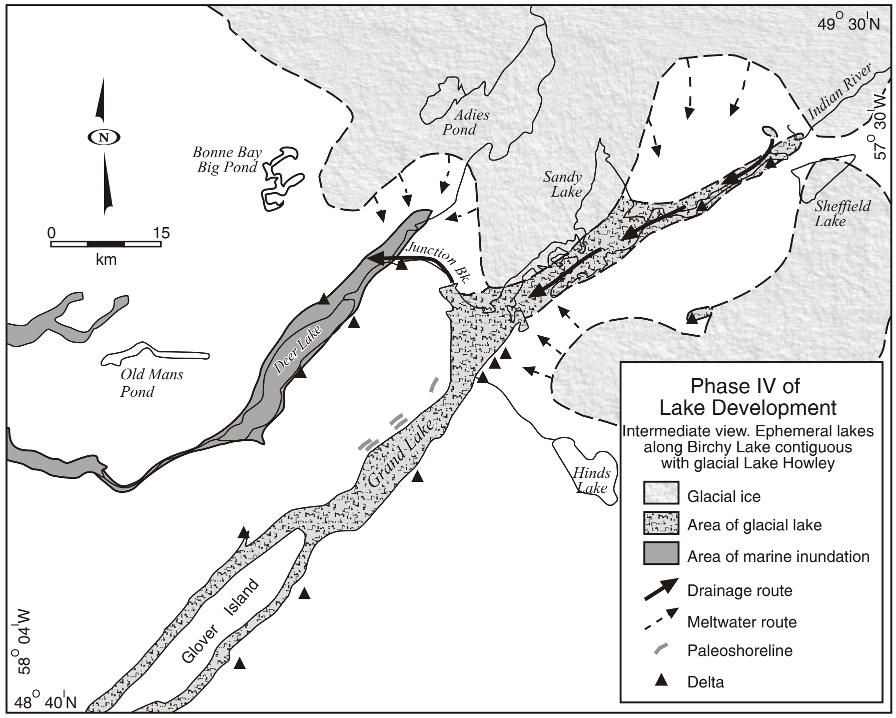


Figure 73. *Phase IV of lake development. Intermediate view.*

Intermediate Reconstruction

This view accepts the reconstruction for the northern part of the Grand Lake basin, but suggests there is a connection with the Birchy Lake valley during the waning stages of glacial Lake Howley (Figure 73). As ice retreated northward from the Junction Brook area, it may have allowed connection of the Grand Lake and Birchy Lake systems. Much of the evidence used to support the minimum view may also be applied to this intermediate reconstruction.

The rhythmically bedded silt and clay in the Gillards Lake area are at an elevation of 160 m. Liverman and St. Croix (1989b) cite the lack of associated deltas or strandline features, and the height of the Gillards Lake section above the valley to argue for deposition in an ice-marginal lake, produced by ice in the Birchy Lake valley. The elevation difference between glaciolacustrine sediments at Gillards Lake and the eastern shore of Grand Lake might be explained by differential isostatic rebound (Figure 68). The presence of rhythmites on the watershed at 104 m asl between Indian Brook and Birchy Lake shows the existence of standing water. This is well above the known marine limit for the adjacent coast. The rhythmites are thin, planar bedded, contain no diamicton lenses or beds, dropstones, and show no evidence of faulting or contortion that suggests deposition was in an ice-proximal environment. The ice dam that formed the lake was east of Birchy Lake. This is shown by westward paleo-current indicators from a gravel pit near Baie Verte junction described by Liverman and St. Croix (1989b). This is opposed to the modern eastward drainage. Southward striations showing flow from the Baie Verte Peninsula into the Indian Brook valley have also been recognised by Grant (1974) and Liverman and St. Croix (1989a, b). Apart from an ice dam to the east, higher water levels at the modern watershed require either an ice dam, or a continuous water body to the west. The absence of other reported rhythmites to the west may suggest the former. Modern colluvial activity along Birchy Lake has likely reworked or covered sediments deposited during deglaciation. Flat-topped deltas are found at the mouths of Sheffield Brook (115 m), at the northern end of Birchy Lake, and Voyins Brook (105 m) at the southern end of Birchy Lake. Neither has exposed sections, and determination of a deltaic origin is on morphological grounds. Limited evidence suggests higher water levels in Birchy Lake during deglaciation.

Isostatic rebound is expected to produce a northwestward tilt on raised marine (and by implication lacustrine) features (*see* earlier discussion). The raised deltas identified along Birchy Lake at 110 to 115 m asl may reasonably be associated with sediments or features at some elevation lower than this in the Grand Lake basin.

Support for the former connection of the modern Grand Lake and Sandy Lake basins comes from the extensive sand and gravel deposits found along the north shore of Grand Lake, up to about 100 m asl, and along The Main Brook area connecting the two lake basins. Generally, paleo-flow indicators support fluvial transport toward the Junction Brook area. Although much of this area has been mapped as ice-contact sediments by Grant (1989b) this is not supported by field evidence. In particular, several of the features mapped as eskers were not verified in the field. Connection of the Sandy Lake and Birchy Lake basins during regional deglaciation is not as readily apparent.

Discussion

The maximum view of glacial Lake Howley proposed by Batterson *et al.* (1993) is considered unlikely on the basis of field work in the Upper Humber River valley, and the identification of an ice-contact delta at the mouth of Deer Lake. Evidence from raised deltas at 100 to 115 m, and the presence of rhythmically bedded silt and clay on the Indian Brook–Birchy Lake drainage divide at 104 m shows that higher lake level than present occurred in the Birchy Lake valley during deglaciation. An ice dam to the east of Baie Verte junction impounded the lake at the eastern end.

The evidence for a separate ice dam in the western part of the Birchy Lake basin is not conclusive. No eskers or other ice contact features have been identified to suggest ice retreated up the valley. Similarly, no channels connecting an independent lake in the Birchy Lake valley with glacial Lake Howley have been found. Finally, the elevation of sediment and features in the two basins is similar. These factors suggest that the connection of the Birchy Lake and Grand Lake basins, at least in the final phases of glacial Lake Howley, is reasonable. As such, the intermediate reconstruction is the one favoured here.

Acceptance of the intermediate reconstruction produces a long, narrow lake, up to 135 km long and 10 km wide having a surface area of 650 km^2. This is considerably smaller than the 1850 km^2 suggested by Batterson *et al.* (1993). A lake of this size, draining exclusively through the Junction Brook spillway would drain in about 52 days.

PHASE 4: MARINE INUNDATION

A period of higher relative sea levels (RSL) on an isostatically depressed coastline followed deglaciation of the Humber River valley. Along the margins of the Humber Arm constructional features such as marine terraces and deltas are found up to an elevation of about 60 m. This marine limit is marked by a delta in the Hughes Brook valley, and wave-cut terraces at about the same elevation near Humber mouth. Marine shells found in fine grained sediments along the modern inner coast have been dated. The highest, and farthest inland are shell fragments found along Goose Arm Brook at 50 m, and dated at 13 070 ± 90 BP (TO-3624) (Table 17). Marine shells are found at the eastern end of the Humber River gorge, indicating that marine inundation occurred through this part of the basin. Fine-grained sediments are common beneath about 30 m inland of the gorge, and deltas are found at the mouth of Deer Lake and at the head of the lake between Nicholsville and Junction Brook. The Junction Brook delta was likely formed during outflow from glacial Lake Howley.

Ice retreated up the Lower Humber River valley, accompanied by marine inundation. It initially stalled at the head of the Humber Arm, producing the delta at Humbermouth, that was subsequently incised as ice retreated through the Humber River gorge. Marine shells in the adjacent Wild Cove valley were radiocarbon dated at 12 500 BP, providing a minimum age for deglaciation for this area. A delta at the head of Wild Cove having a surface elevation of 51 m also shows deglaciation about this time.

Retreat of ice through the Lower Humber River valley stalled at the mouth of modern Deer Lake, where it formed an ice-contact delta. Farther retreat of the ice, and the accompanying marine inundation led to the formation of the deltas at the head of Deer Lake. Ice retreated rapidly from this area, as the deltas show no indication of formation in an ice-proximal environment.

Subsequent isostatic rebound resulted in lowered sea level, and the development of deltas in the Lower Humber River valley at 33 m asl, such as the one on which the community of Pasadena is located. No equivalent delta surfaces were found in the Humber Arm (cf. Flint, 1940).

Marine inundation and subsequent isostatic adjustment accompanied development and final drainage of glacial Lake Howley. They are both time transgressive events, discussed below.

TIMING OF GLACIAL LAKE HOWLEY

The sea-level curve developed for the Bay of Islands area allows an estimation of the age of the deltas at Deer Lake. Projection of these deltas at 45 m onto the proposed relative sea-level curve provides an estimated age of about 12 600 ± 200 BP. The effects of differential isostatic uplift across the basin are uncertain, although the northeastward isostatic tilt suggests delta formation may have been some time (several hundred years?) later. The Deer Lake deltas are about 45 km inland from the modern coast, and on a line oblique to isobases proposed in existing models of relative sea-level history (e.g., Grant, 1989a). Assuming the marine limit in the inner Humber Arm is 60 m asl (elevation of Hughes Brook delta), and the marine limit at Springdale is 75 m asl (Springdale deltas), this would suggest a rise of 0.1 m km^{-1} between Corner Brook and Springdale. Deglaciation at Corner Brook (~13 100 BP) was likely earlier than at Springdale (~12 500 BP). Isostatic tilt was therefore steeper, up to 0.2 m km^{-1}.

The deltas at the head of Deer Lake, 45 km from the modern coast correspond to a position on the RSL curve at 41.5 to 36.5 m asl, giving a date of 12 400 to 12 200 BP. This age estimate corresponds to the date of marine shells in the Humber River gorge, and in the Wild Cove valley, both of which have statistically similar dates of 12 500 to 12 200 BP.

EFFECTS OF CATASTROPHIC DRAINAGE IN THE LOWER HUMBER RIVER VALLEY

Glacial Lake Howley drained through two outlets, apart from the Harrys River exit. The first, through the South Brook valley, lowered water levels, from about 170 m (based on the elevation of strandlines on the west side of the lake), to 145 m, the level of the col at the southern end of South Brook. The areal extent of the early phase of the lake is unclear. No deltas equivalent to the high beaches on the west side have been located on the east side of Grand Lake; the lake may have been of limited extent. It is possible that this early phase extended into the modern South Brook valley. A large flat-topped, steep sided feature, possible ice contact delta on the basis of morphology and sediment type (bedded sand and gravel, plus diamicton beds), is found in the central part of the valley having a surface elevation of 150 m asl. Exposure was from a single backhoe test-pit that showed planar-bedded sand and gravel having beds dipping southeastward down the valley. Two small deltas are also found on the western side of the valley, at surface elevations of about 140 m asl for a delta immediately west of the 150 m delta (Site 91232: Appendix 1), and 145 m for a delta about 2 km north (Site 89006: Appendix 1). Both are poorly exposed, and show inclined beds of sand and gravel dipping into the valley. Further evidence for standing water in the South Brook valley is the presence of greater than 2 m of structureless clayey silt at 90 m asl (Sites 91226 and 91230: Appendix 1). Other small exposures of sand and gravel, overlying diamicton are found on the eastern flanks of the South Brook valley.

The South Brook valley contains fragmentary evidence for standing water, in the form of deltas and fine grained sediment; the location of an ice dam is uncertain. The direction of dipping beds in the 155-m delta indicates deposition from the north, and any proglacial lake may have accompanied retreat of ice up the Lower Humber River valley. The lack of exposures through the valley makes this conclusion speculative.

The effects of discharge through the South Brook valley are unclear. The lower part of the valley contains a well-defined terrace grading southward up the valley. Sediment contain clasts derived from The Topsails. Similarly, a channel is incised through the ice-contact delta at the mouth of Deer Lake and the terrace gravels and the eroded channel are speculatively the result of lake drainage through the South Brook valley.

Final draining of glacial Lake Howley through the Junction Brook lowlands was catastrophic, and introduced an estimated volume of water of 2.6 to 4.5 x 10^{10} m^3 over a period of 30 to 50 days, into an isostatically depressed inner coastal environment where the sea level was about 45 m higher than present; lake drainage formed the Junction Brook spillway. Flow was subsequently into a raised post-glacial sea. The spillway is thus the only subaerially exposed erosional feature. Other erosional features may exist on the floor of modern Deer Lake. A geophysical survey designed to examine lake bottom topography and stratigraphy failed due to mechanical difficulties.

The major effect of lake outflow was depositional. The Junction Brook area has a surface veneer of sandy gravel and large (up to 300 cm diameter) granite boulders, presumably deposited as the outflow lost its competence to transport them. Deposition of these sediments, and the associated delta, trapped most of the coarse-grained sediment from lake discharge. Fine grained sediment was transported as turbid underflow with or without overflow–interflow through the Lower Humber River valley. A short core taken on a reconnaissance seismic mapping and sampling program in the Humber Arm and Bay of Islands (Shaw *et al.*, 1995) revealed a reddish brown buttery clay that may be derived from the Deer Lake basin. The clay contains low foraminifera counts containing early colonizer species (*Cassidulina reniforme* and *Elphidium excavatum*; Vilks *et al.*, 1989; MacLean *et al.*, 1992; Scott *et al.*, 1984). The lower part of this core is undated, but may have been deposited during outflow from glacial Lake Howley. The proximity of the core site to the mouth of the Humber River, and the presence of red sedimentary bedrock in the Carboniferous Deer Lake basin that is drained by the Humber River supports this conclusion. The high percentage of clay through the lower 60 cm of the core suggests deposition by suspension settling, and the consistency of sediment texture suggests rapid deposition. The low proportion of foraminifera tests contained within the clay, and the tolerance ranges of species identified, suggest a high freshwater component to the water. The clay was deposited some time before 5400 BP.

The extent of this clay unit in the Humber Arm and Bay of Islands is uncertain. A near-surface, thin, reddish clay unit having a sparse palynofloral assemblage was mapped in the Bay of Islands (A. Aksu, personal communication, Department of Earth Sciences, Memorial University of Newfoundland, 1998). The clay sediments identified by Aksu and Shaw *et al.* (1995) are similar, and may represent the same discharge event.

Ericson *et al.* (1981) and Barranco *et al.* (1989) described red marine sediments (brick red lutites) in the Laurentian Channel–Gulf of St. Lawrence area that were derived from Permo-Carboniferous red sediments in New Brunswick, Nova Scotia and Prince Edward Island. Bartlett and Molinsky (1972) reported red sediments having low microfossil concentrations, from the Laurentian Channel. All these sediments represent periods of high sedimentation associated with deglaciation.

Cumming (1991) described rapidly deposited (about 200 years based on ^{14}C dating), fine-grained marine sediments having sparse microfauna or microflora in cores from Clode Sound in eastern Newfoundland. The stratigraphy was similar to that described in Humber Arm. Sommerville (1996) described glacial lake sediments from the Terra Nova River valley that drains into Clode Sound. No correlation between these two events has been established.

DISCUSSION: THE IMPLICATIONS OF MARINE INUNDATION AND GLACIAL LAKE HOWLEY FOR REGIONAL DEGLACIATION

During the deglaciation of the west coast, the existence of a proglacial lake, some time before 12 600 to 12 300 BP, in the interior of Newfoundland has several important implications for the established Quaternary history and stratigraphy of the west coast, particularly in the St. George's Bay region.

The presence of a spillway at the western end of Grand Lake implies that meltwater could flow westward without obstruction, and that the Stephenville area was deglaciated before 12 600 to 12 300 BP. This is incompatible with the timing of the Robinsons Head readvance (Brookes, 1977a). Discussion of the evidence and timing of the Robinsons Head readvance is warranted, although it is outside the Humber River basin. These arguments are summarized in Batterson *et al*. (1993, 1995) and Brookes (1995).

The culmination of the Robinsons Head readvance in the Stephenville area is interpreted at about 12 600 BP based on a ^{14}C date of from a section near Kippens (Table 17). The extent of the Robinsons Head moraine has been described by Brookes (1970a, 1974, 1977a, 1987) as extending roughly from Highlands in the south, around St. George's Bay, to just west of Romaines Brook in the north. Robinsons Head Drift is described as "a coarse, loosely-structured till of englacial and supraglacial origin" (Brookes, 1974, p. 22), and ice-contact stratified drift where the outer limits are marked by a kame moraine consisting of "stratified gravel and sand with minor silt and small irregular bodies of till" (Brookes, 1974, p. 22). The Robinsons Head Drift was first described by MacClintock and Twenhofel (1940) and along with the underlying St. George's Bay delta and St. George's River Drift has become established as the accepted Quaternary stratigraphic units in southwestern Newfoundland (e.g., Brookes, 1970a, 1974, 1977a, 1987; Grant 1987, 1989; Proudfoot *et al.,* 1988). Brookes (1977a) suggests the Robinsons Head readvance was climatically induced. Supportive arguments for a regional correlation with the Robinsons Head Drift event (Brookes, 1977a) were from a late Wisconsinan readvance on the Cabot Strait coast of Newfoundland (Brookes, 1975), the Highland Front Morainic system in southeastern Canada (Gadd *et al.,* 1972), the Pineo Ridge moraine (Borns, 1973), and the Inner Port Huron moraine (Evenson *et al.,* 1976). Each of these features was dated at about 12 700 to 12 600 BP, although the cause of readvance at some localities may not have been climate related (Brookes, 1977a).

Of the cases cited by Brookes (1977a), only the Port Huron stade remains the best dated. The readvance of three lobes in the Great Lakes region, the Huron, Michigan and Ontario/Erie lobes, produced a series of moraines (e.g., Port Huron moraine in Michigan, and Wyoming–Banks moraine in Ontario (Karrow, 1989), as well as a series of small moraines on the Niagara Peninsula (Dreimanis and Goldthwait, 1973; Barnett, 1979). Also, it deposited a regional correlated till unit, named the St. Joseph Till (Cooper and Clue, 1974) from the Huron Lobe, the Oak Creek Formation from the Michigan Lobe (Eschman and Mickelson, 1986), and the Halton Till (Karrow, 1959) from the Ontario Lobe. Although the regional correlation of the readvance of three separate lobes suggests a climatic link, the large distances (~400 km) and short time period (200 to 300 years) suggests surging may have been the cause (Hansel *et al.,* 1985). A further argument against a climatic link is that while this part of the Laurentide Ice Sheet was advancing, areas to the east and west of the Great Lakes basin were retreating (Dyke and Prest, 1987).

The Highland Front moraine system of Gadd *et al.* (1972) on the southern margin of the Champlain Sea was formed about 12 500 BP, by southward flowing ice from the retreating Laurentide Ice Sheet. However, Parent and Ochietti (1988) consider the "Highland Front moraine system" to be a grouping of unrelated features, including bedrock controlled ridges, diachronous moraines, eskers, and deltas. Even assuming its existence, there is no stratigraphic evidence for a readvance in this area (LaSalle *et al.,* 1977; Gadd, 1978), and the feature is considered to represent a halt in the general recession of the ice sheet (Gadd *et al.,* 1972; Chauvin *et al.,* 1985). The Pineo Ridge moraine in Maine has been re-dated at about 13 200 to 13 000 BP (Smith, 1985), with the ice being well inland by about 12 600 BP (Borns *et al.,* 1985). Several other moraines have been identified in the Maritimes dating around 12 600 BP. In the Baie des Chaleurs, the Elmtree Interlobate moraine dates around 12 500 BP (Bobrowsky *et al.,* 1987). Rappol (1989) describes moraines in the St. John River valley from the local Appalachian Ice Divide prior to 11 000 BP, and Stea *et al*. (1992) described advance of ice from the Scotian Ice Divide and Antigonish Highlands in the 13 000 to 12 000 BP period that may be correlated with Robinsons Head readvance. However, none of these features has been linked with climatic forcing, and no cooling event that may trigger a readvance has been recorded from paleo-ecological sources.

Perhaps it is significant that all the moraines dating in the 13 000 to 11 500 BP range are found adjacent to the coast, at a time when the main Laurentide Ice Sheet was downwasting. The drawdown of ice into marine basins and the influence of falling relative sea levels may have been the trigger for re-activation of local ice fronts, rather than any climatic trigger. In contrast to the Younger Dryas event, which is recorded in mountaineous (e.g., Clapperton, 1993; Magny and Ruffaldi, 1995; Williamson *et al.*, 1993; Reasoner *et al.*, 1993), as well as coastal areas (Bergstrom, 1995; Cwynar and Levesque, 1995; Anderson and Macpherson, 1994), the advances at about 12 500 BP are only recorded adjacent to coasts and are not recorded in mountaineous areas such as the Rockies.

The extent and timing of the Robinsons Head readvance is problematic. The date of the readvance at 12 600 BP is based on a single radiocarbon date from a section at Kippens interpreted to represent deposition within a kame. In contrast, a section at Romaines, 3.5 km west of Kippens yielded dates of 13 100 ± 180 BP (GSC-4095, *Mya truncata* shells) and 13 345 ± 230 BP (S-3074, whale bone) from silt overlying sediment interpreted as being associated with the Robinsons Head readvance (Grant, 1987, 1991; Proudfoot *et al.*, 1988). This suggests that the area may have been ice free at the proposed time of the existence of the lake.

The section at Kippens contains shell fragments (GSC-2295) found in a sand bed within kame gravels (Brookes, 1977a), which are described as "...unevenly stratified gravels, with many cobbles and boulders" (Brookes, 1970a, p. 136). It seems likely that the assignment of an ice-contact origin was based primarily on local geomorphology. A date of 13 300 ± 810 BP (GSC-2063) from the same horizon as GSC-2295 was rejected by Brookes (1977a) due to the large error bar. Batterson and Janes (1997) interpreted the Kippens section (Plate 40) as initial ice-proximal glaciomarine sedimentation, followed by delta progradation produced by glacial retreat or a shallowing sea. Whole marine shells (*Hiatella arctica*) were found in upper foreset gravels, and subsequently dated at 12 600 ± 120 BP (GSC-5942). Marine shells from an adjacent exposure found within marine muds at 8 m asl, below a marine to freshwater transition, were dated at 12 610 ± 100 BP (TO-6138) (Table 17).

The western extent of the morainic topography around St. George's Bay, associated with the Robinsons Head readvance, is located at Romaines. Batterson and Janes (1997) interpreted the sediment as having been deposited into a gypsum karst depression. Marine silt found in the upper part of the section contain marine macrofauna, including two marine shell samples and a whale bone, previously dated at 13 100 ± 180 BP (GSC-4095), 12 800 ± 130 BP (GSC-4858) and 13 345 ± 230 BP (S-3047) respectively (Grant, 1989b; McNeely and Jorgensen, 1993) (Table 17). Marine shells (*Mya truncata*) found with a diamicton, within the Romaines Brook valley at 37 m asl, were dated at 13 680 ± 100 BP (TO-6137).

The section had been previously interpreted as showing kame deposits of the Robinsons Head readvance overlain by silts (Brookes, 1970a). The radiocarbon dates from Romaines suggest the area was ice free before about 13 000 BP. This is in contrast to Brookes' (1977a) interpretation of the Kippens section. Grant (*in* Blake, 1988, p. 6-7) considered three alternative explanations for this discrepancy: the disappearance of ice from the area; reworking from a lower stratigraphic unit; or a meltwater effect on the ^{14}C dates. In a later discussion of the site, Grant (*in* McNeely and Jorgensen, 1993, p. 14-15) considered the reworking scenario unlikely because the shells are intact, and suggested that an average date on the marine silts overlying the interpreted Robinsons Head readvance sediments is about 13 000 BP.

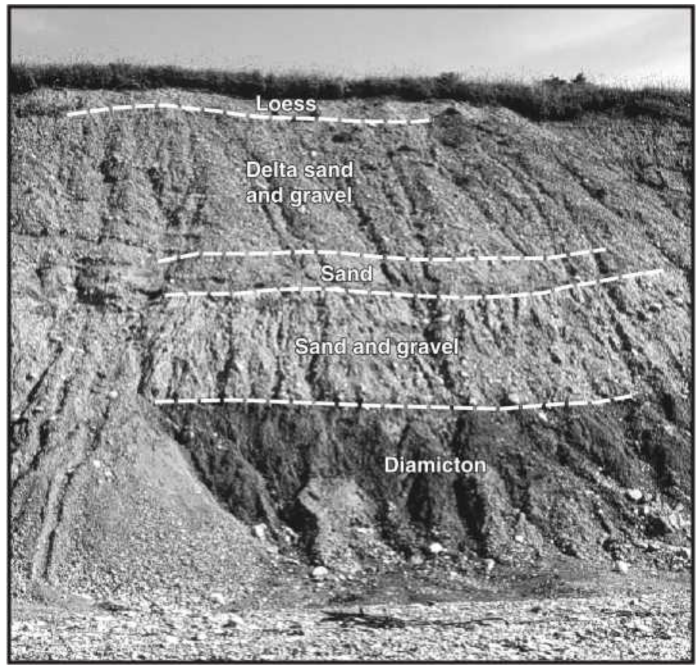


Plate 40. *A 20-m high coastal exposure near Kippens, west of Stephenville.*

The apparent dichotomy between the sedimentologic evidence and the numerous depressions that characterize the surface morphology of the northern coast of St. George's Bay, interpreted by Brookes (1970a) as kettle holes, is resolved by interpretation of the depressions as sinkholes resultant from gypsum karst (Grant, 1989b; Batterson and Janes, 1997). Gypsum is exposed on the surface near Romaines Brook and is also exposed farther east in the Flat Bay area. Bedrock exposures are absent between Kippens and Romaines Brook.

Liverman and Bell (1996) have re-examined some of the sections between Bank Head and Highlands, including the type section for the Robinsons Head readvance at Robinsons Head. In contrast to the tripartite stratigraphy described by MacClintock and Twenhofel (1940) and Brookes (1969, 1974), Liverman and Bell (1996) record a complex stratigraphy having numerous lateral and vertical shifts in sediment type. Liverman and Bell (1996) proposed that sediment in the southern part of St. George's Bay were deposited as grounding line fans from a calving ice mass. No readvance is implied in this interpretation.

Batterson *et al.* (1993) used a date of 13 100 ± 220 BP (GSC-5302) (Table 17) on non-calcareous bulk organics from a small lake at the south end of the South Brook valley (Thane Anderson, Geological Survey of Canada, personal communication, 1994) to further discuss the timing of the inland deglaciation. However, a *Salix* twig from the base of this core was dated at 9540 ± 90 BP (TO-5707). Although this promotes further discussion of the reliability of bulk organic dates using conventional techniques, it provides a little additional information on the timing of glacial Lake Howley.

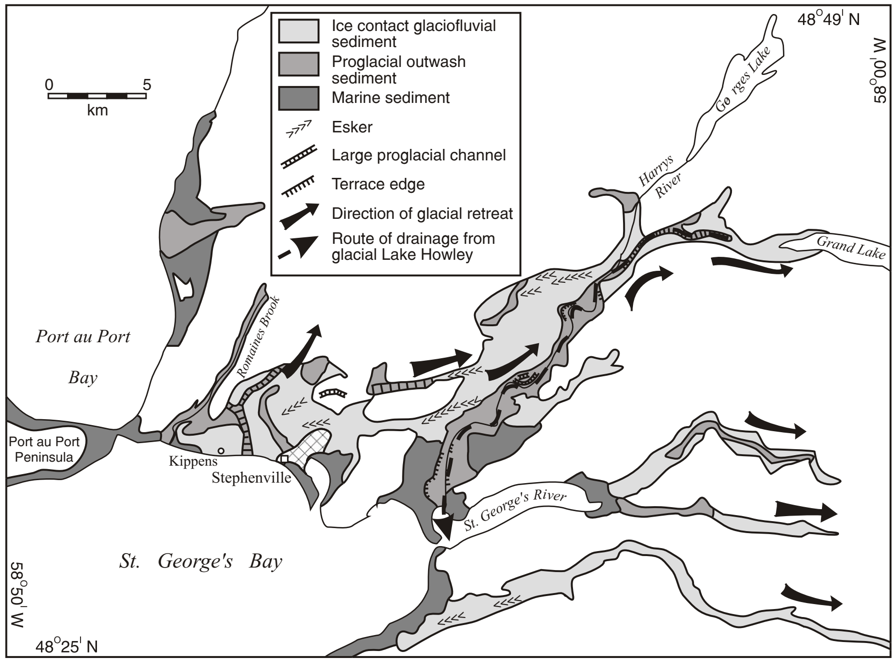


Figure 74. *Drainage route of glacial Lake Howley toward St. George's Bay.*

The area near Stephenville Crossing (Figure 74), where drainage from Grand Lake likely reached the sea, has no well described or dated sections that show that the outlet would have been ice-covered at the time of glacial Lake Howley. The route of drainage from Grand Lake to the coast is marked by a large proglacial channel. Brookes (1974) showed an esker following a similar course to the channel, and makes no reference to the large meandering channel identified by Grant (1991).

The timing of deglaciation at the northwest and northern margins of the Humber River basin must also be considered in discussions of glacial Lake Howley. Pink till derived from Carboniferous sediment in the Deer Lake basin have been described from Bonne Bay (Brookes, 1974, 1995; Liverman and Batterson, 1995). This shows that ice must have flowed across the Deer Lake basin, and ice-flow data from the Humber River basin supports this conclusion. The oldest date from Bonne Bay is 12 100 ± 160 BP (GSC-4158) (Table 17). Brookes (1995) argued that Bonne Bay area was not deglaciated until 12 100 BP, which precludes the existence of glacial Lake Howley prior to 12 200 BP. Radiocarbon dates on marine organisms provide *minimum* dates on deglaciation, although it is commonly assumed that these relate to marine limit features, and by implication date actual deglaciation. To make this conclusion, careful sedimentological examination and interpretation is required, demonstrating that the sediments in which the organisms are found are indeed the lateral equivalent to marine limit deltas. Radiocarbon dates (12 000 to 11 000 BP; Table 17) from the Springdale–Halls Bay area were used to date the numerous ice contact deltas found at the coast (Tucker, 1974a; Grant, 1974). Using similar arguments to Brookes (1995), these dates would preclude the existence of a large deglaciated area inland prior to 12 200 BP. Subsequent discoveries of marine fauna, dated at 12 470 ± 300 BP (TO-2305), up to 10 km inland of the marine features at the modern coast clearly demonstrate that the original interpretation was incorrect, and that the dates relate to a later stage in deglaciation (Scott *et al.*, 1991). At Goose Arm, a date of 10 600 ± 100 BP (GSC-4400) at 6 m asl was interpreted as a minimum date for a glacial stand at this position (Grant *in* McNeely and McCuaig, 1991, p. 17). However, marine shells were later found in Goose Arm Brook at 50 m asl and dated at 13 150 ± 90 BP (TO-3264) (Batterson *et al.*, 1993). This provides a new minimum date for deglaciation of the area. Goose Arm is about 13 km west of the Deer Lake watershed, about half the distance between Bonne Bay and the Deer Lake watershed. The date of 12 220 ± 90 BP (TO-2885) on marine shells in the Humber River gorge shows that the western end of the Humber River basin was ice free before this time.

The dates from the Humber River gorge and from Goose Arm are difficult to reconcile with the assertion by Brookes (1974, 1995) that Bonne Bay was not deglaciated until 12 100 BP. Faced with a similar predicament at Lark Harbour where the date on a delta there was substantially younger than for a delta at a similar elevation at Cox's Cove, Brookes (1974) suggested that the ice lingered in the constricted valley at Lark Harbour; similar reasoning can be suggested for the Bonne Bay area. The radiocarbon dates and the lack of evidence showing ice-proximal conditions suggest that the Deer Lake–Grand Lake basin was ice free before 12 200 BP.

In the absence of similar detailed sedimentological studies, it is suggested that dates in Bonne Bay, and other areas (including the Stephenville and Bay of Islands areas) should be interpreted as minimum dates only.

Glacial Lake Howley was dependent on the presence of ice dams in the Humber River gorge and the Indian Brook valley, and maintenance of the Harrys Brook outlet at 122 m asl. These constraints were unlikely to remain for an extended period, and the life span of the lake was probably short. The thin lacustrine sediment in the Birchy Lake canal section support this conclusion. Also, thicker, fine-grained sediment in the South Brook valley are tentatively identified as lacustrine. The surface sediment in much of the Sandy Lake–Birchy Lake region is sand and gravel (Grant, 1989b), suggesting any lacustrine deposits have been either eroded or buried by subsequent fluvial activity.

The scarcity of lacustrine sediments in the basin indicates a short lived lake that drained rapidly. This conclusion is supported by the absence of successively lower shoreline features, and the lack of terracing of the Grand Lake outlet. If the Indian Brook valley ice dam collapsed prior to the Humber gorge ice dam, then a set of strandline features related to the elevation of the Indian Brook–Birchy Lake drainage divide should have been preserved. The absence of a secondary shoreline indicates that drainage followed the collapse of the Humber gorge dam.

Previous studies have suggested that deglaciation of the Humber River basin followed an orderly retreat from the coast toward remnant ice centres on west coast highlands, beginning about 14 000 to 13 000 BP (Prest *et al.*, 1968; Dyke and Prest, 1987; Grant, 1989a). An alternative style of deglaciation has been presented here that shows a large, low-elevation basin in the interior was deglaciated, possibly at a similar time to ice retreat to coastal positions from the offshore. Due to isolation from marine influence, modern mean summer temperatures are higher, and precipitation and snowfall lower within the Deer Lake–Grand Lake basin than in areas near the coast (Banfield, 1981; Department of Environment and Lands, 1992). Assuming similar patterns existed during deglaciation, isolation from marine influence may have been an important control on melting rates, with ice more persistent in coastal areas. The presence of a large proglacial lake shows that deglaciation was underway before 12 200 BP, possibly as early as 13 000 BP, throughout the Deer Lake–Grand Lake basin and that drainage outlets to the southwest were also ice free at this time. Following the retreat of ice through the Lower Humber River valley, raised postglacial sea levels resulted in flooding of large areas of the basin. Glacier ice was restricted to isolated remnants. Grant (1974) described the final stage of deglaciation as being multiple remnant ice caps derived from the major Newfoundland ice sheet by "a process of shrinkage, separation and migration". The pattern described here is an intermediate stage between this final configuration, and that of the glacial maximum. Early deglaciation of the Deer Lake–Grand Lake basin is required to enable separation of distinct Northern Peninsula, The Topsails, and Baie Verte ice caps (cf. Grant, 1974).

SUMMARY

This report focused on three areas of study; the mapping and detailed description of surficial sediments; the determination of paleo ice-flow using erosional and depositional evidence; and definition of the relative sea-level history. Subsequently, these are integrated to provide a Quaternary history. The major findings of this report are:

1. The entire Humber River basin was glaciated. Striated bedrock surfaces and exotic clast rock types were found on the highest hills within, and along, the margins of the basin. Diamictons were fresh, and striated bedrock surfaces unweathered. The tills were deposited during the late Wisconsinan.

2. Diamictons found across the entire Humber River basin were deposited mostly as lodgement and melt-out tills, or their secondary derivatives. Depositional environments were defined on the basis of sedimentary structures, and clast fabrics. Evidence of deformation was found in sections south of the Pasadena dump. The clastic dykes found within the middle diamicton at the Pasadena dump exposures, and similar features near Deer Lake, are relatively unusual features. They have not been reported elsewhere in Newfoundland.

3. Initial advance of glaciers was from a Long Range Mountain ice cap southward into the Upper Humber River valley, and then flowing into the Deer Lake basin. The southwestern extent of this flow is uncertain. Southward-directed striations are found near Stephenville and farther south, but there is insufficient data to link these features to the advance from the Long Range Mountains.

4. The next phase of ice flow was from an ice centre on The Topsails. In the Hinds Lake area, ice flow was northwestward, whereas flow on the high plateau to the east was northeastward, thus providing data on the northeastern part of the dispersal centre. Topsails-centred ice crossed the Glide Lake highlands and Deer Lake, flowing toward the coast. Ice flow from The Topsails did not cross into the Upper Humber River valley that contained ice from the Long Range Mountains throughout the late Wisconsinan. In the north, ice crossed Sandy Lake and Birchy Lake and flowed toward the coast into White Bay or Hall's Bay (Figure 45), while in the southwest, ice flow was directed toward St. George's Bay or through Serpentine Lake. These flow patterns provide a sequence of ice flows that differs from existing models, but explain the striation, clast fabrics and clast provenance data.

5. The distribution of eskers, meltwater channels and hummocky moraine suggests areas of ice wastage. The presence of hummocks in the central and eastern parts of the basin suggests that ice retreated across the high-

lands west of the Humber River valley, and stagnated within the Upper Humber River valley, on the Glide Lake highlands, in the lowlands around Howley, and on The Topsails. These areas also contain meltwater channels. A radial pattern of meltwater channels also suggests that wasting ice was located on the southern end of Birchy Ridge, on the highlands between the Glide Lake and Deer Lake valleys, overlooking the South Brook valley, and on The Topsails southwest of Hinds Brook. These remnant ice centres remained active, at least for a short period, as indicated by the striation patterns. In the Glide Lake–Pynn's Brook area, till found overlying glaciofluvial sand–gravel suggests a local readvance down the Pynn's Brook valley.

6. The spatial and elevational distribution of features indicating deposition within a body of standing water, well above marine limit, in the Grand Lake basin suggests formation within a single glacial lake. At its maximum extent, this lake occupied most of the basin, including Grand Lake, and likely Sandy Lake and Birchy Lake. Lake level was controlled by discharge through the southwestern end into Harrys River, and configuration of the lake was controlled by retreating ice through Deer Lake and Sandy Lake. The elevation of shoreline features increases northeastward from 128 to 150 m, due to differential isostatic rebound. The name ‘glacial Lake Howley' was proposed for this body of water (Batterson *et al.*, 1993).

7. Glacial Lake Howley drained through a spillway now occupied by Junction Brook, as ice retreated across the lowland between Glide Lake highlands and Birchy Ridge. Drainage through the Junction Brook spillway lowered lake levels by up to 71 m, representing a volume of 2.6–4.1 x 10^{10} m^3 of water. Discharge through this channel is estimated at a maximum of 10 200 m^3 sec^{-1}. This would drain the lake within a 30- to 60-day period, depending on the configuration of the lake.

8. Marine limit on the coast is about 60 m asl, based on the elevation of a delta in the Hughes Brook valley. This is 10 m higher than published records of marine limit for the Humber Arm area.

9. A protracted episode of standing water in the Deer Lake basin at elevations below 50 m asl resulted in the formation of deltas along the basin margins, and deposition of rhythmically bedded sediment within deeper parts of the basin. Although sedimentology, the lack of fauna or flora, and a poorly defined geochemical signature cannot confirm a marine origin, the elevation with respect to marine limit and the presence of a relict population of the marine fish *Microgadus tomcod* in modern Deer Lake, suggest that a marine origin is more likely than lacustrine.

10. Discharge from drainage of glacial Lake Howley would have introduced large quantities of water and sediment into the Deer Lake basin. As a result, most of the coarse-grained material would have been deposited in the Junction Brook delta formed during marine inundation of the Deer Lake basin. Fine-grained sediment would have transported through the Lower Humber River valley and into the Humber Arm. This mechanism may explain the presence of a well sorted, reddish brown clay layer containing a sparse microfaunal assemblage described by Shaw *et al.* (1995)

11. Radiocarbon dating of marine shells in the Humber Arm and tributaries was used to construct a relative sea-level curve for the area. It is a Type-B (Quinlan and Beaumont, 1981) curve, showing rapid relative sea-level fall from a marine limit of 60 m dated at about 13 100 BP, falling below modern sea level at about 10 000 BP, reaching a sea-level lowstand at -6 m at about 7900 BP and recovering since then.

12. The relative sea level curve developed from Humber Arm, was used to date the deltas at the head of Deer Lake at between 12 600 to 12 300 BP. These deltas were formed subsequent to lake drainage. Glacial Lake Howley was formed sometime before these dates.

13. Draining of a lake through the Harrys River valley before 12 600 to 12 300 BP is not compatible with the established chronology for the St. George's Bay area of western Newfoundland. It is argued that the extent and/or timing of the Robinsons Head readvance should be re-evaluated.

14. The style of deglaciation presented here differs from previously accepted models for this part of eastern Canada (Prest *et al.*, 1968; Dyke and Prest, 1987; Grant, 1989a). Rather than a gradual retreat from the coast to remnant ice centres on topographic highs, a style is suggested in which a large, low elevation basin in the interior was deglaciated early, possibly at a similar time to ice retreat to coastal positions from offshore. Early deglaciation of the Deer Lake–Grand Lake basin is required to enable separation of distinct Northern Peninsula, Topsail Hills, and Baie Verte ice caps (cf. Grant, 1974).

FURTHER RESEARCH

The following are several areas where further work is required:

1. The depositional environment of fine-grained sediments in the Lower Humber River valley remains speculative. In this report, they were interpreted as glaciomarine sediments deposited during marine inundation of the Humber River valley. However, the sediments contained only rare foraminifera, no macrofossils, no microflora, and had no distinct geochemical signature. A detailed examination of these sediments is warranted to determine:

a) criteria that may be used to differentiate fine-grained sediments deposited in marine or freshwater environments.

b) whether a marine to freshwater transition can be determined from a detailed examination of sections at, for instance, Wild Cove, or from drill core from the Steady Brook area, where thick sediment is found.

A detailed microfaunal and/or microfloral examination may be considered. Alternatively, the use of strontium isotopes (e.g., Reinhardt, 1996) or carbon isotopes (e.g., Winkler, 1994) may be applicable to the problem in the Humber Arm.

2. The implications of glacial Lake Howley have been discussed in some detail. A major conclusion is that the existing chronology of southwestern Newfoundland needs to be re-examined if the Grand Lake basin was ice free at about 12 500 BP. This would preclude the possibility of Robinsons Head readvance ice covering the Harrys River lowland, the drainage route for the lake. Recent work by Liverman and Bell (1996) and Batterson and Janes (1997) bears on this discussion, but considerably more study, involving detailed section descriptions, is required to resolve this problem.

3. A new relative sea level curve for the Humber Arm area has increased data concerning the sea level history of western Newfoundland. The data does not readily fit any of the existing geophysical models to explain sea-level change in Newfoundland (e.g., Quinlan and Beaumont, 1981, 1982). A new synthesis of this data is required, perhaps adopting the approach of Lambeck *et al*. (1996) to determine the pattern of crustal recovery from the influence of the Laurentide ice sheet and an independent Newfoundland ice cap.

4. The presence of an early southward ice flow requires further investigation. Evidence of southward flow exists in the erosional record from the Northern Peninsula to Cape Ray, with supporting fabric data from the Corner Brook area. The source of this flow is unknown, and the reasons for coast-parallel flow unclear. Alternative hypotheses, including piedmont glaciation onto the coastal platform or southward ice occupying the Esquiman channel require testing.

5. The hypothesis of west coast refugia remains unresolved. Grant (1987) and Brookes (1977b) suggested the presence of nunataks on some west coast uplands. Cosmogenic radionuclide investigations demonstrated that Gros Morne (one of the proposed nunataks) was likely ice covered during the late Wisconsinan (Gosse and Grant, 1993). Complete glacial coverage in the late Wisconsinan must be reconciled with the biological data that suggests refugia, e.g., the distribution of the land-snail *Cepaea hortensis* on the west coast. The location of this refugium (if any) is speculative, but may not be in the west coast highlands.

6. Other areas of interest are: detailed mapping in the South Brook valley. This valley connects the Grand Lake basin with the Lower Humber River valley, through which the early phases of glacial Lake Howley drained; sedimentological descriptions and mapping of the valley connecting the western end of Grand Lake to the Harrys River valley; detailed mapping and sedimentology of deltas at the head of Deer Lake.

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REFERENCES

Agterberg, F.P. and Banerjee, I.
1969: Stochastic model for the deposition of varves in glacial Lake Barlow-Ojibway, Canada. Canadian Journal of Earth Sciences, Volume 6, pages 625-652.

Alam, M. and Piper, D.J.W.
1977: Pre-Wisconsin stratigraphy and paleoclimates off Atlantic Canada and its bearing on glaciation in Québec. Géographie physique et Quaternaire, Volume 31, pages 15-22.

Alam, M., Piper, D.J.W. and Cooke, H.B.S.
1983: Late Quaternary stratigraphy and paleo-oceanography of the Grand Banks, continental margin, eastern Canada. Boreas, Volume 12, pages 253-261.

Allen, J.R.L.
1963: The classification of cross-stratified units, with notes on their origin. Sedimentology, Volume 2, pages 93-114.

1984: Sedimentary structures: their character and physical basis. Unabridged two-volume edition. Developments in Sedimentology 30. Elsevier. Amsterdam, 593 and 663 pages.

1993: Large-scale textural patterns and sedimentary processes on tidal salt marshes in the Severn Estuary, southwest Britain. Sedimentary Geology, Volume 81, pages 299-318.

Alley, D.W.
1975: Drift prospecting and glacial geology in the Sheffield Lake - Indian Pond Area, north-central Newfoundland. Unpublished M.Sc. thesis, Department of Geology, Memorial University of Newfoundland, St. John's, Newfoundland, 215 pages.

Alley, D.W. and Slatt, R.M.
1975: Drift prospecting and glacial geology in the Sheffield Lake, Indian Pond area, north-central Newfoundland. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 75-3, 96 pages.

Åmark, M.
1980: Glacial flutes at Isfallsglaciaren, Tarfala, Swedish Lapland. Geologiska Foreningen i Stockholm Forhandlingar, Volume 102, pages 251-259.

1986: Clastic dikes formed beneath an active glacier. Geologiska Foreningen i Stockholm Forhandlingar, Volume 108, pages 13-20.

Ammann, B. and Lotter, A.F.
1989: Late-glacial radiocarbon- and palynostratigraphy on the Swiss Plateau. Boreas, Volume 18, pages 109-126.

Anderson, T.W. and Lewis, C.F.M.
1992: Climatic influences of deglacial drainage changes in southern Canada at 10 to 8 Ka suggested by pollen evidence. Géographie physique et Quaternaire, Volume 46, pages 255-272.

Anderson, T.W. and MacPherson, J.B.
1994: Lateglacial environmental change: Newfoundland regional review. Journal of Quaternary Science, Volume 9, pages 171-178.

Anderson, T.W. and Stephens, M.A.
1972: Tests for randomness of directions against equatorial and bimodal alternatives. Biometrika, Volume 59, pages 613-621.

Antevs, E.
1922: The recession of the last ice sheet in New England. American Geographical Society, Research Series 11, 120 pages.

Arkhipov, S.A., Ehlers, J., Johnson, R.G. and Wright, H.E. Jr.
1995: Glacial drainage towards the Mediterranean during the Middle and Late Pleistocene. Boreas, Volume 24, pages 196-206.

Ashley, G.M.
1972: Rhythmic sedimentation in glacial Lake Hitchcock, Massachusetts-Connecticut. University of Massachusetts, Geology Publication 10, 148 pages.

1975: Rhythmic sedimentation in Glacial Lake Hitchcock. *In* Glaciofluvial and Glaciolacustrine Sedimentation. *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Palaeontologists and Mineralogists Special Publication 23, pages 304-320.

1988: Classification of glaciolacustrine sediments. *In* Genetic Classification of Glacigenic Deposits. *Edited by* W.P. Goldthwait and C. L. Matsch. A.A. Balkema, Rotterdam, The Netherlands, pages 243-260.

Ashley, G.M., Shaw, J. and Smith, N.D.
1985: Glacial sedimentary environments. Society of Paleontologists and Mineralogists, SEPM Short Course No. 16, 246 pages.

Ashley, G.M., Southard, J.B. and Boothroyd, J.C.
1982: Deposition of climbing ripples; a flume simulation. Sedimentology, Volume 29, pages 67-79.

Babcock, E.A.
1977: A comparison of joints in bedrock and fractures in overlying Pleistocene lacustrine deposits, central Alberta. Canadian Geotechnical Journal, Volume 14, pages 357-366.

Bagnold, R.A.
1954: Experiments on a gravity free dispersion of large solid spheres in a Newtonian fluid under shear. Proceedings of the Royal Society of London, Series A, Volume 225, pages 49-63.

Bajzak, D., Rees, D. and Vanderveer, D.G.
1977: Hinds Lake environmental study. Department of Engineering and Applied Sciences, Memorial University of Newfoundland, St. John's, Newfoundland, 188 pages.

Banerjee, I.
1977: Experimental study on the effect of deceleration on the vertical sequence of sedimentary structures in silty sediments. Journal of Sedimentary Petrology, Volume 47, pages 771-783.

Banfield, C.E.
1981: The climatic environment of Newfoundland. *In* The Natural Environment of Newfoundland Past and Present. *Edited by* A.G. Macpherson and J.B. Macpherson. Memorial University of Newfoundland, St. John's, Newfoundland, pages 83-153.

Barnett, P.J.
1979: Glacial Lake Whittlesey: the probable ice frontal position in the eastern end of the Erie basin. Canadian Journal of Earth Sciences, Volume 16, pages 568-574.

Batterson, M.J.
1989: Quaternary geology and glacial dispersal in the Strange Lake area, Labrador. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 89-3, 51 pages.

1992: Surficial geology and landform classification of the Corner Brook map sheet (NTS 12A/13). Newfoundland Department of Mines and Energy, Open file 12A/0616, Map 92-13.

1994a: Surficial geology and landform classification of the Deer Lake map sheet (NTS 12H/03). Newfoundland Department of Mines and Energy, Open File 12H/1287, Map 94-119.

1994b: Surficial geology and landform classification of the Pasadena map sheet (NTS 12H/04). Newfoundland Department of Mines and Energy, Open file 12H/1288, Map 94-120.

1994c: Surficial geology and landform classification of the Cormack map sheet (NTS 12H/06). Newfoundland Department of Mines and Energy, Open file 12H/1289, Map 94-121.

1994d: Ice flow indicators, Pasadena map sheet, 12H/04. Newfoundland Department of Mines and Energy, Geological Survey Branch, Open file 12H/1284. Scale 1:50 000.

1994e: Ice flow indicators, Deer Lake map sheet, 12H/03. Newfoundland Department of Mines and Energy, Open File 12H/1284, Map 94-82.

1994f: Ice flow indicators, Cormack map sheet, 12H/06. Newfoundland Department of Mines and Energy, Open File 12H/1286, Map 94-83.

1997: A revised configuration for glacial Lake Howley. The Canadian Association of Geographers, Annual Meeting, Programs and Abstracts, St. John's, Newfoundland, August 1997, pages 64-65.

2001: Surficial Geology of the Humber River Basin and Surrounding areas (parts of NTS 12H and 12A), Newfoundland. Newfoundland Department of Mines and

Energy, Geological Survey, Map 201-42, Open File NFLD/2768.

Batterson, M.J. and Janes, J.
1997: Stratigraphy of Late Quaternary sediments exposed in coastal cliffs west of Stephenville. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey, Report 97-1, pages 151-165.

Batterson, M.J. and Liverman, D.G.E.
1995: Landscapes of Newfoundland and Labrador: A collection of aerial photographs. Newfoundland Department of Natural Resources, Geological Survey, Report 95-3, 133 pages.

Batterson, M.J., Liverman, D.G.E. and Kirby, G.E.
1993: Glacial lake development and marine inundation, Deer Lake area, Newfoundland, Canada: topographically controlled deglaciation of an interior basin. Journal of Quaternary Science, Volume 8, pages 327-337.

Batterson, M., Liverman, D. and Taylor, D.
1995: The assessment of geological hazards and disasters in Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 95-1, pages 55-75.

Batterson, M.J. and McGrath, B.
1993: Quaternary geology of the Deer Lake and Pasadena map areas (NTS 12H/3 and 12H/4). *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 93-1, pages 103-112.

Batterson, M.J. and Taylor, D.M.
1990: Glacial history of the Humber River basin. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 90-1, pages 1-6.

Batterson, M.J. and Taylor, R.C.
1994: Quaternary geology of the upper Humber River area, western Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 94-1, pages 1-9.

Batterson, M.J. and Vatcher, S.V.
1992a: Quaternary geology of the Corner Brook-Pasadena area (NTS sheets 12A/13 and 12H/4). *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 92-1, pages 1-12.

1992b: Ice flow indicators, Corner Brook (NTS 12A/13). Newfoundland Department of Mines and Energy, Geological Survey Branch, Open File 12A/0617. Scale 1:50 000.

Batterson, M.J., Wareham, B.J. and St. Croix, L.
1992: Carbon-14 date list for Newfoundland and Labrador. Newfoundland Department of Mines and Energy, Geological Survey Branch, Open File NFLD (2190).

Becker, B., Kromer, B. and Trimhorn, P.
1991: A stable-isotope tree-ring timescale of the Late Glacial/Holocene boundary. Nature, Volume 353, pages 647-649.

Bell, R.A.
1884: Observations on the geology, mineralogy, zoology and botany of the Labrador coast, Hudson's Strait and Bay. Geological Survey of Canada, Annual Reports, 1882-1884, Part DD, pages 5-62.

Bell, T., Liverman, D., Evans, S. and Batterson, M.
1997: Great Northern Peninsula and Southern Labrador field trip guide. Canadian Association of Geographers meeting, St. John's, August 1997, 92 pages.

Belland, R.J.
1981: Ecology and phytogeography of the mosses of the Bonne Bay region, western Newfoundland. Unpublished M.Sc. thesis, Memorial University of Newfoundland, 133 pages.

1987: The disjunct moss element of the Gulf of St. Lawrence region: glacial and postglacial dispersal and migrational histories. Journal of the Hattori Botanical Laboratory, Volume 63, pages 1-76.

Benn, D.I.
1995: Fabric signature of subglacial till deformation, Breidamerkurjökull, Iceland. Sedimentology, Volume 42, pages 735-747.

1996: Subglacial and subaqueous processes near a glacier grounding line: Sedimentological evidence from a former ice-dammed lake, Achnasheen Scotland. Boreas, Volume 25, pages 23-36.

Bergstrom, B.
1995: Stratigraphical evidence of a considerable Younger Dryas glacier advance in southeastern Norway. Norsk Geologisk Tidsskrift, Volume 75, pages 127-136.

Betz, F.
1939: Geology and mineral deposits of the Canada Bay area, northern Newfoundland. Newfoundland Geological Survey, Bulletin 16, 53 pages.

Blackwelder, E.
1930: Striated boulders as evidence of glacial action. Bulletin of the Geological Society of America, Volume 41, 154 pages.

Blake, W. Jr.
1987: Geological Survey of Canada Radiocarbon Dates XXVI. Geological Survey of Canada, Paper 86-7, 60 pages.

1988: Geological Survey of Canada Radiocarbon Dates XXVII. Geological Survey of Canada, Paper 87-7, 100 pages.

Blanchon, P. and Shaw, J.
1993: Catastrophic sea-level rise during deglaciation: Evidence from submerged terraces and glacial landforms. Geological Association of Canada/Mineralogical Association of Canada, Programs and Abstracts, Edmonton, Alberta, page A-10.

Bobrowsky, P.T., Lamothe, M. and Shilts, W.W.
1987: Late Quaternary glaciation and deglaciation in the Baie des Chaleurs area, northern New Brunswick. Geological Association of Canada/Mineralogical Association of Canada, Program with Abstracts, 26 pages.

Böcher, T.W.
1963: Phytogeography of Greenland in the light of recent investigations. *In* North Atlantic Biota and their History. *Edited by* A. Löve and D. Löve. Pergamon Press, London, pages 285-295.

Borns, H.W. Jr.
1973: Late Wisconsin fluctuations of the Laurentide Ice Sheet in southern and eastern New England. Geological Society of America, Memoir 136, pages 37-45.

Borns, H.W. Jr., LaSalle, P. and Thompson, W.B.
1985: Late Pleistocene history of northeastern New England and adjacent Quebec. Geological Society of America, Special Paper 197.

Bostock, H.S.
1970: Physiographic subdivisions of Canada. *In* Geology and Economic Minerals of Canada. *Edited by* R.J.W. Douglas. Economic Geology Report, Number 1, 4th edition, Ottawa, Department of Energy, Mines and Resources, pages 10-30.

Boulton, G.S.
1968: Flow tills and related deposits on some Vestspitsbergen glaciers. Journal of Glaciology, Volume 7, pages 391-412.

1970: On the deposition of subglacial and meltout tills at the margin of certain Svalbard glaciers. Journal of Glaciology, Volume 9, pages 231-245.

1971: Till genesis and fabric in Svalbard, Spitsbergen. *In* Till: A Symposium. *Edited by* R.P. Goldthwait. Ohio State University Press, Columbus, Ohio, pages 41-72.

1972: Modern arctic glaciers as depositional models for former ice sheets. Journal of the Geological Society, Volume 128, pages 361-393.

1976: The development of geotechnical properties in tills. *In* Glacial Till. *Edited by* R.F. Legget. Royal Society of Canada Special Publication 12, pages 292-304.
1978: Boulder shapes and grain size distributions of debris as indicators of transport paths through a glacier and till genesis. Sedimentology, Volume 25, pages 773-799.

1979: Processes of glacier erosion on different substrata. Journal of Glaciology, Volume 23, pages 15-37.

1986: Push-moraines and glacier-contact fans in marine and terrestrial environments. Sedimentology, Volume 33, pages 677-698.

1987: A theory of drumlin formation by subglacial sediment deformation. *In* Drumlin Symposium. *Edited by* J. Menzies and J. Rose. A.A. Balkema, Rotterdam, pages 25-80.

Boulton, G.S. and Hindmarsh, R.C.A.
1987: Sediment deformation beneath glaciers: rheology and geological consequences. Journal of Geophysical Research, Volume 92(B), pages 9059-9082.

Boulton, G.S. and Paul, M.A.
1976: The influence of genetic processes on some geotechnical properties of glacial tills. Quarterly Journal of Engineering Geology, Volume 9, pages 159-194.

Bouma, A.H.
1962: Sedimentology of some flysch deposits: a graphic approach to facies interpretation. Elsevier. Amsterdam, 168 pages.

Bowles, J.E.
1978: Engineering properties of soils and their measurement (2nd edition). McGraw-Hill, New York, 213 pages.

Boyer, S.J. and Pheasant, D.R.
1974: Delimitation of weathering zones in the fiord area of eastern Baffin Island, Canada. Bulletin of the Geological Society of America, Volume 85, pages 805-810.

Brennard, T.A. and Shaw, J.
1994: Tunnel channels and associated landforms, south-central Ontario: their implications for ice-sheet hydrology. Canadian Journal of Earth Sciences, Volume 31, pages 505-522.

Broecker, W.S., Andree, M., Wolfli, W., Oeschger, H., Bonani, G., Kennett, J. and Peteet, D.
1988: The chronology of the last deglaciation: Implications to the cause of the Younger Dryas event. Paleoceanography, Volume 3, pages 1-19.

Brookes, I.A.
1969: Late glacial marine overlap in western Newfoundland. Canadian Journal of Earth Sciences, Volume 6, pages 1397-1404.

1970a: The glaciation of southwestern Newfoundland. Unpublished Ph.D. thesis, McGill University, 208 pages.

1970b: New evidence for an independent Wisconsin-age ice cap over Newfoundland. Canadian Journal of Earth Sciences, Volume 7, pages 1374-1382.

1971: Fossil ice wedges casts in Western Newfoundland. Maritime Sediments, Volume 7, pages 118-122.

1974: Late-Wisconsin glaciation of southwestern Newfoundland (with special reference to the Stephenville map-area). Geological Survey of Canada, Paper 73-40.

1975: Late-Wisconsin readvance of piedmont glaciers, southwest Newfoundland. Maritime Sediments, Volume 11, pages 47-48.

1977a: Radiocarbon age of Robinson's Head moraine, west Newfoundland, and its significance for postglacial sea level change. Canadian Journal of Earth Sciences, Volume 14, pages 2121-2126.

1977b: Geomorphology and Quaternary geology of Codroy Lowlands and adjacent plateaus, southwest Newfoundland. Canadian Journal of Earth Sciences, Volume 14, pages 2101-2126.

1982: Ice marks in Newfoundland: A history of ideas. Géographie physique et Quaternaire, Volume 36, pages 139-163.

1987: Late Quaternary glaciation and sea-level change, southwest Newfoundland, Canada. *In* Geological Society of America. Centennial Field Guide, Northeastern Section, pages 445-450.

1995: Glacial lake development and marine inundation, Deer Lake area, Newfoundland, Canada: topographically controlled deglaciation of an interior basin - discussion. Journal of Quaternary Science, Volume 10, pages 180-182.

Brookes, I.A., Scott, D.B. and McAndrews, J.H.
1985: Postglacial relative sea-level change, Port au Port, west Newfoundland. Canadian Journal of Earth Sciences, Volume 22, pages 1039-1047.

Brookes, I.A. and Stevens, R.K.
1985: Radiocarbon age of rock-boring Hiatella Arctica (Linné) and postglacial sea-level change at Cowhead, Newfoundland. Canadian Journal of Earth Sciences, Volume 22, pages 136-140.

Broster, B.E.
1991: Glacitectonic deformation in sediment and bedrock, Hat Creek, British Columbia. Géographie physique et Quaternaire, Volume 45, pages 5-20.

Broster, B.E., Dreimanis, A. and White, J.C.
1979: A sequence of glacial deformation, erosion and deposition at the ice rock interface during the last glaciation; Cranbrook, B.C. Journal of Glaciology, Volume 23, pages 283-295.

Caldenius, C.C., SON
1932: Les glaciaciones cuaternarias en la Patagonia y Tierra del Fuego. Geografiska Annaler, Volume 14, pages 1-164.

Cant, D.J. and Walker, R.G.
1976: Development of a braided-fluvial facies model for the Devonian Battery Point Sandstone, Quebec. Canadian Journal of Earth Sciences, Volume 13, pages 102-119.

Car, D. and Ayres, L.D.
1991: A thick dacitic debris flow sequence, Lake of the Woods greenstone terrane, central Canada: resedimented products of Archean vulcanian, Plinian and dome-building eruptions. Precambrian Research, Volume 50, pages 239-260.

Catto, N.R.
1984: Glacigenic deposits at the Edmonton Convention Centre, Edmonton, Alberta. Canadian Journal of Earth Sciences, Volume 21, pages 1473-1482.

1987: Lacustrine sedimentation in a proglacial environment, Caribou River valley, Yukon, Canada. Boreas, Volume 16, pages 197-206.

Catto, N.R., Hamlyn, C. and Catto, G.
1995: Field trip guidebook, eastern Avalon Peninsula. CANQUA-CCRG Joint Meeting, Programme, Abstracts, Field Guides, St. John's, Newfoundland, pages EC1-EC9.

Catto, N.R., Patterson, R.J. and Gorman, W.A.
1981: Late Quaternary marine sediments at Chalk River, Ontario. Canadian Journal of Earth Sciences, Volume 18, pages 1261-1267.

Cawood, P.A. and Van Gool, J.A.M.
1992: Stratigraphic, structural, and metamorphic relations along the eastern margin of the Humber Zone, Corner Brook Lake map area, western Newfoundland. *In* Current Research, Part E. Geological Survey of Canada, Paper 92-1E, pages 239-247.

1993: Stratigraphic and structural relations within the western Dunnage Zone, Glover Island region, western Newfoundland. *In* Current Research, Part D. Geological Survey of Canada, Paper 93-1D, pages 29-37.

Chamberlin, T.C.
1888: The rock scourings of the great ice invasions. United States Geological Survey, 7th Annual Report, pages 155-248.

Chase, R.R.P.
1979: Settling behaviour of natural aquatic particulates. Limnology and Oceanography, Volume 24, pages 417-426.

Chauvin, L., Martineau, G. and LaSalle, P.
1985: Deglaciation of the Lower St. Lawrence Region, Québec. *In* Late Pleistocene History of Northeastern New England and Adjacent Quebec. *Edited by* H.W. Borns Jr., P. LaSalle and W.B. Thompson. Geological Society of America, Special Paper 197, pages 111-123.

Cheel, R.J. and Rust, B.R.
1982: Coarse-grained facies of glaciomarine deposits near Ottawa, Canada. *In* Research in Glacial, Glacio-Fluvial and Glacio-Lacustrine systems. *Edited by* R. Davidson-Arnott, W. Nickling and B.D. Fahey. 6th Geulph Symposium on Geomorphology, 1980, pages 279-295.

Church, M. and Ryder, J.M.
1972: Paraglacial sedimentation: a consideration of fluvial processes conditioned by glaciation. Bulletin of the Geological Society of America, Volume 83, pages 3059-3072.

Clapperton, C.M.
1993: Glacier readvances in the Andes at 12500-10000 yr BP: Implications for mechanism of late-glacial climatic change. Journal of Quaternary Science, Volume 8, pages 197-215.

Clark, J.A.
1980: A numerical model of worldwide sea level changes on a viscoelastic earth. *In* Earth Rheology, Isostasy and Eustasy. *Edited by* N.-A. Morner. Wiley and sons, London, pages 525-534.

Clark, J.A., Farrell, W.E. and Peltier, W.R.
1978: Global changes in postglacial sea level: a numerical calculation. Quaternary Research, Volume 9, pages 265-287.

Clark, P.U. and Fitzhugh, W.W.
1992: Postglacial relative sea level history of the Labrador coast and interpretation of the archaeological record. *In* Paleoshorelines and Prehistory: An Investigation of Method. *Edited by* L.L. Johnson. CRC Press, Boca Raton, Florida, pages 189-213.

Clark, P.U. and Hansel, A.K.
1989: Clast ploughing, lodgement and glacier sliding over a soft glacier bed. Boreas, Volume 18, pages 201-207.

Clayton, K.M.
1972: Slopes. Oliver and Boyd, Edinburgh, 288 pages.

Clayton, L., Laird, W. and Kupsch, W.D.
1965: Intersecting minor lineations of the Lake Agassiz Plain. Journal of Geology, Volume 73, pages 652-656.

Clemmensen, L.B. and Houmark-Nielsen, M.
1981: Sedimentary features of a Weichselian glaciolacustrine delta. Boreas, Volume 10, pages 229-245.

Cocksedge, J.E.
1983: Road construction in glaciated terrain. *In* Glacial Geology: An Introduction for Engineers and Earth Scientists. *Edited by* N. Eyles. Pergamon Press, England, pages 302-317.

Colella, A. and Prior, D.B. (editors)
1990: Coarse-grained deltas. International Association of Sedimentologists, Special Publication 10, 357 pages.

Coleman, A.P.
1926: The Pleistocene of Newfoundland. Journal of Geology, Volume 34, pages 193-223.

Collinson, J.D. and Thompson, D.B.
1982: Sedimentary Structures. George Allen and Unwin, London, 194 pages.

Colman, S.M., Clark, J.A., Clayton, L., Hansel, A.K. and Larsen, C.E.
1994: Deglaciation, lake levels, and meltwater discharge in the Lake Michigan basin. Quaternary Science Reviews, Volume 13, pages 879-890.

Colman-Sadd, S.P., Hayes, J.P. and Knight, I.
1990: Geology of the Island of Newfoundland. Newfoundland Department of Mines and Energy, Geological Survey Branch, Map 90-01, scale 1: 1 000 000.

Cooper, A.J. and Clue, J.
1974: Quaternary geology of the Grand Bend area, southern Ontario. Ontario Division of Mines, Preliminary Map P-974.

Cooper, J.R.
1937: Geology and mineral deposits of the Hare Bay area. Newfoundland Geological Survey, Bulletin 9, 36 pages.

Corney, R.E.
1993: A depositional model for chromite-bearing Pleistocene sediments of Port au Port Bay, western Newfoundland. Unpublished M.Sc. thesis, University of Toronto, 69 pages.

Couch, E.I.
1971: Calculation of paleosalinities from boron and clay mineral data. Bulletin of the American Association of Petroleum Geologists, Volume 55, pages 1829-1937.

Cram, G.F.
1900: Newfoundland. Published by G.F. Cram, Chicago, Ill. Map scale 1:1 400 000.

Croot, D.G. and Sims, P.C.
1996: Early stages of till genesis: an example from Fanore, County Clare, Ireland. Boreas, Volume 25, pages 37-46.

Cumming, E.H.
1991: Seismic stratigraphy and Late Quaternary sedimentary history of Bonavista Bay, northeastern Newfoundland. Unpublished M.Sc. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 230 pages.

Cwynar, L.C. and Levesque, A.J.
1995: Chironomid evidence for late-glacial climatic reversals in Maine. Quaternary Research, Volume 43, pages 405-413.

Daly, R.A.
1902: The geology of the northeast coast of Labrador. Memoirs of Museum of Comparative Zoology, Harvard University, Volume 38, pages 205-270.

1921: Post-glacial warping of Newfoundland and Nova Scotia. American Journal of Science, Volume 1, pages 381-391.

1934: The Changing World of the Ice Age. Yale University Press, 271 pages.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjornsdottir, A.E., Jouzel, J. and Bond, G.
1993: Evidence for general instability of past climate from a 250-kyr ice-core record. Nature, Volume 364, pages 218-220.

Davis, A.M.
1984: Ombrotrophic peatlands in Canada: Their origins, development and trans-Atlantic affinities. Chemical Geology, Volume 44, pages 287-309.

1993: The initiation and development of peatlands in Newfoundland and their response to global warming. *In* The Scientific Challenge of Our Changing Environment. *Edited by* J. Hall and M. Wadleigh. Canadian Global Change Program Incidental Report, Series No. IR93-2, Royal Society of Canada, pages 24-25.

Davis, J.L. and Mitrovica, J.X.
1996: Glacial isostatic adjustment and the anomalous tide gauge record of eastern North America. Nature, Volume 379, pages 331-333.

Degeer, G.
1892: On Pleistocene changes of level in eastern North America. Proceedings of the Boston Natural History Society, Volume 25, pages 454-477.

Denton, G. and Hughes, T.J.
1981: The Last Great Ice Sheets. Wiley, New York, 484 pages.

Department of Environment
1995: Water well data for Newfoundland and Labrador 1950-1994. Water Resources Division, Department of Environment, Government of Newfoundland and Labrador, St. John's, Water Resources Report 1-4 Groundwater Series.

Department of Environment and Lands
1992: Water Resources Atlas of Newfoundland. Water Resources Division, Department of Environment and Lands, Government of Newfoundland and Labrador, St. John's 79 pages.

Derbyshire, E. and Jones, P.F.
1980: Systematic fissuring of a matrix-dominated lodgement till at Church Wilne, Derbyshire. Geological Magazine, Volume 117, pages 243-254.

Domack, E.W. and Lawson, D.E.
1985: Pebble fabric in an ice rafted diamicton. Journal of Geology, Volume 93, pages 577-591.

Dowdeswell, J.A., Hambrey, M.J. and Wu, R.
1985: A comparison of clast fabric and shape in Late Precambrian and modern glacigenic sediments. Journal of Sedimentary Petrology, Volume 55, pages 691-704.

Dowdeswell, J.A. and Sharp, M.J.
1986: Characterization of pebble fabrics in modern terrestrial glacigenic sediments. Sedimentology, Volume 33, pages 699-710.

Drake, L.D.
1974: Till fabric control by clast shape. Bulletin of the Geological Society of America, Volume 85, pages 247-250.

Dredge, L.A. and Grant, D.R.
1987: Glacial deformation of bedrock and sediment, Magdalen Islands and Nova Scotia, Canada: Evidence for a regional grounded ice sheet. *In* Tills and Glaciotectonics. *Edited by* J.J.M. van der Meer. A.A. Balkema, Rotterdam, pages 183-195.

Dreimanis, A.
1969: Till wedges as indicators of direction of glacial movement. Geological Society of America, Abstract with Programs 7, pages 52-53.

1982: Two origins of the stratified Catfish Creek Till at Plum Point, Ontario, Canada. Boreas, Volume 11, pages 173-180.

1988: Tills: their genetic terminology and classification. *In* Genetic Classification of Glacigenic Deposits. *Edited by* R.P. Goldthwait and C.L. Matsch. A.A. Balkema, Rotterdam, pages 17-83.

1992: Downward injected till wedges and upward injected till dikes. Sveriges Geologiska Undersökning, Series Ca 81, pages 91-96.

Dreimanis, A. and Goldthwait, R.P.
1973: Wisconsinan glaciation in the Huron, Erie, and Ontario lobes. *In* The Wisconsinan Stage. *Edited by* R.F. Black, R.P. Goldthwait and H.B. Willman. Ohio State University Press, Columbus, Ohio, pages 71-106.

Dreimanis, A. and Lundqvist, J.
1984: What should be called till? Striae, Volume 20, pages 5-10.

Dreimanis, A. and Rappol, M.
1997: Late Wisconsinan sub-glacial clastic intrusive sheets along Lake Erie bluffs, at Bradtville, Ontario, Canada. Sedimentary Geology, Volume 111, pages 225-248.

Drewry, D.
1986: Glacial Geologic Processes. Edward Arnold, London, 276 pages.

Dyke, A.S.
1993: Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. Progress in Physical Geography, Volume 17, pages 223-247.

Dyke, A.S., Dale, J.E. and McNeely, R.N.
1996: Marine molluscs as indicators of environmental changes in glaciated North America and Greenland during the last 18000 years. Géographie physique et Quaternaire, Volume 50, pages 125-184.

Dyke, A.S. and Prest, V.K.
1987: Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. Géographie physique et Quaternaire, Volume 41, pages 237-263.

Dyson, J.L.
1937: Snowslide striations. Journal of Geology, Volume 45, pages 549-557.

Edwards, M.
1986: Glacial environments. *In* Sedimentary Environments and Facies. *Edited by* H. Reading. Blackwell Scientific Publications, pages 445-470.

Edwards, R.L., Beck, J.W., Burr, G.S., Donahue, D.J., Chappell, J.M.A., Bloom, A.L., Druffel, E.R.M. and Taylor, F.W.
1993: A large drop in atmospheric $^{14}C/^{12}C$ and reduced melting in the Younger Dryas, documented with ^{230}Th ages of corals. Science, Volume 260, pages 962-967.

Elson, J.A.
1957: Striated boulder pavements of southern Manitoba, Canada. Bulletin of the Geological Society of America, Volume 67, page 1722.

1961: The geology of tills. *In* Proceedings of the 14th Canadian Soil Science Conference. *Edited by* E. Penner and J. Butler. National Research Council of Canada, Soil and Snow Mechanics Technical memorandum 69, pages 5-36.

1989: Comment on glaciotectonite, deformation till, and comminution till. *In* Genetic Classification of Glacigenic Deposits. *Edited by* R.P. Goldthwait and C.L. Matsch. A.A. Balkema, Rotterdam, pages 85-88.

Embleton, C. and King, C.A.M.
1968: Glacial and Periglacial Geomorphology. Edward Arnold, London, 608 pages.

Engelder, J.T.
1974: Microscopic wear grooves on slickensides: Indicators of paleoseismicity. Journal of Geophysical Research, Volume 79, pages 4387-4392.

England, J.
1987: Glaciation and the evolution of the Canadian high arctic landscape. Geology, Volume 15, pages 419-424.

Environment Canada
1980: Groundwater resource evaluation of the Deer Lake airport. Internal Report of the Newfoundland Water Planning and Management Branch, Inland Waters Directorate, Atlantic Region, Halifax.

Erdmer, P.
1986: Geology of the Long Range Inlier in Sandy Lake map area, western Newfoundland. *In* Current Research, Part B. Geological Survey of Canada, Paper 86-1B, pages 19-27.

Eschman, D.F. and Mickelson, D.M.
1986: Correlation of glacial deposits of the Huron, Lake Michigan and Green Bay lobes in Michigan and Wisconsin. *In* Quaternary Glaciations in the Northern Hemisphere. *Edited by* V. Sibrava, D.Q. Bowen and G.M. Richmond. Pergamon Press, Oxford, pages 53-57.

Evenson, E.B., Farrand, W.R., Eeschman D.F., Mickelson, D.M. and Maher, L.J.
1976: Greatlakean Substage: A replacement for Valderan Substage in the Lake Michigan Basin. Quaternary Research, Volume 6, pages 411-424.

Eyles, N.
1977: Late Wisconsinan glacitectonic structures and evidence of postglacial permafrost in north-central Newfoundland. Canadian Journal of Earth Sciences, Volume 14, pages 2797-2806.

1979: Facies of supraglacial sedimentation on Icelandic and Alpine temperate glaciers. Canadian Journal of Earth Sciences, Volume 16, pages 1341-1362.

1983: Glacial Geology: An introduction for engineers and earth scientists. Pergamon Press, England, 409 pages.

1993: Earth's glacial record and its tectonic setting. Earth Science Reviews, Volume 35, pages 1-248.

Eyles, N., Clark, B. and Clague, J.J.
1987: Coarse grained sediment gravity flow facies in a large supraglacial lake. Sedimentology, Volume 34, pages 193-216.

Eyles, N. and Eyles, C.H.
1992: Glacial depositional systems. *In* Facies Models Response to Sea Level Change. *Edited by* R.G. Walker and N.P. James. Geological Association of Canada, pages 73-100.

Eyles, N., Eyles, C.H. and McCabe, A.M.
1988: Late Pleistocene subaerial debris-flow facies of the Bow Valley, near Banff, Canadian Rocky Mountains. Sedimentology, Volume 35, pages 465-480.

Eyles, N., Eyles, C.H. and Miall, A.D.
1983b. Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences. Sedimentology, Volume 30, pages 393-410.

Eyles, N. and Slatt, R.M.
1977: Ice-marginal sedimentary, glacitectonic, and morphologic features of Pleistocene drift: an example from Newfoundland. Quaternary Research, Volume 8, pages 267-281.

Fairbanks, R.G.
1989: A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. Nature, Volume 342, pages 637-642.

Fairchild, H.L.
1918: Postglacial uplift of northeastern America. Bulletin of the Geological Society of America, Volume 29, pages 187-238.

Falconer, G.
1966: Preservation of vegetation and patterned ground under a thin ice body in northern Baffin Island, N.W.T. Canadian Geographical Branch, Geographical Bulletin, Volume 8, pages 194-200.

Farrand, W.R. and Gajda, R.T.
1962: Isobases on the Wisconsin marine limit. Geographical Bulletin, Volume 17, pages 5-22.

Fernald, M.L.
1911: A botanical expedition to Newfoundland and southern Labrador. Rhodora, Volume 13, pages 108-162.

1925: Persistence of plants in unglaciated areas of boreal America. Memoirs of the American Academy of Arts and Sciences, Volume 15, pages 241-342.

1930: Unglaciated western Newfoundland. Harvard Alumni Bulletin, January 23, 1930, pages 1-6.

Fisher, T.G.
1989: Rogen moraine formation: Examples from three distinct areas within Canada. Unpublished M.Sc. thesis, Department of Geography, Queen's University, 196 pages.

Fisher, T.G. and Smith, D.G.
1994: Glacial Lake Agassiz: Its northwest maximum extent and outlet in Saskatchewan (Emerson Phase). Quaternary Science Reviews, Volume 13, pages 845-858.

Flint, R.F.
1940: Late Quaternary changes of level in western and southern Newfoundland. Geological Society of America Bulletin, Volume 51, pages 1757-1780.

1961: Two tills in southern Connecticut. Bulletin of the Geological Society of America, Volume 72, pages 1687-1691.

1971: Glacial and Quaternary Geology. Wiley, New York, 892 pages.

Flint, R.F., Sanders, J.E. and Rodgers, J.
1960: Diamictite: A substitute term for symmictite. Bulletin of the Geological Society of America, Volume 71, pages 1809-1810.

Folk, R.L. and Ward, W.C.
1957: Brazos River bar: a study in the significance of grain size parameters. Journal of Sedimentary Petrology, Volume 27, pages 3-27.

Forbes, D.L., Shaw, J. and Eddy, B.G.
1993: Late Quaternary Sedimentation and the Postglacial Sea-Level Minimum in Port-Au-Port Bay and Vicinity, West Newfoundland. Atlantic Geology, Volume 29, pages 1-26.

French, H.M.
1976: The Periglacial Environment. Longman, London and New York, 309 pages.

Fulton, R.J.
1969: Glacial lake history, southern interior plateau, British Columbia. Geological Survey of Canada, Paper 69-37.

1975: Quaternary geology and geomorphology, Nicola-Vernon area, British Columbia. Geological Survey of Canada, Memoir 380.

Fulton, R.J., Hodgson, D.A. and Minning, G.V.
1975: Inventory of Quaternary geology, southern Labrador: An example of Quaternary geology-terrain studies in undeveloped areas. Geological Survey of Canada, Paper 74-46.

Gadd, N.R.
1978: Surficial geology of Saint-Sylvestre map-area, Québec. Geological Survey of Canada, Paper 77-16, 9 pages.

Gadd, N.R., McDonald, B.C. and Shilts, W.W.
1972: Deglaciation of southern Québec. Geological Survey of Canada, Paper 71-47, 19 pages.

Gangloff, P.
1983: Les fondements geomorphologiques de la theorie des paleonunataks: le cas des Monts Torngats. Zeitschrift fr Geomorphologie N.F., Volume 47, pages 109-136.

Geikie, A.
1863: On the phenomenon of the glacial drift of Scotland. Transactions of the Geological Society of Glasgow, Volume 1.

Gellatly, A.F., Gordon, J.E., Whalley, W.B. and Hansom, J.D.
1988: Thermal regime and geomorphology of plateau ice caps in northern Norway: Observations and implications. Geology, Volume 16, pages 983-986.

Geodetic Survey of Canada
1978: Vertical Control Data. Quad. No. 48057, 49057, 49058. Surveys and Mapping Branch, Department of Energy, Mines and Resources, Ottawa, Canada.

Gilbert, G.K.
1890: Lake Bonneville. United States Geological Survey, Monograph 1, pages 1-438.

Gilbert, R.
1975: Sedimentation in Lillooet Lake, British Columbia. Canadian Journal of Earth Sciences, Volume 12, pages 1697-1717.

1990: Evidence for subglacial meltwater origin and late Quaternary lacustrine environment of Bateau Channel, eastern Lake Ontario. Canadian Journal of Earth Sciences, Volume 27, pages 939-945.

Gilbert, R. and Shaw, J.
1981: Sedimentation in proglacial Sunwapta Lake, Alberta, Canada. Canadian Journal of Earth Sciences, Volume 18, pages 81-93.

Golder Associates
1983: Hydrogeology of the Humber Valley area. Water Resources Division, Department of Environment, Government of Newfoundland and Labrador, St. John's, Water Resources Report 2-5 Groundwater Series, 55 pages.

Goldthwait, J.W.
1924: Physiography of Nova Scotia. Geological Survey of Canada, Memoir 140, 179 pages.

Goldthwait, R.P. and Matsch, C.L.
1988: Genetic classification of glacigenic deposits. A.A. Balkema, Rotterdam, 294 pages.

Goodarzi, F. and Swaine, D.J.
1994: Paleoenvironmental and environmental implications of the boron content of coal. Geological Survey of Canada, Bulletin 471, 76 pages.

Gosse, J.C. and Grant, D.R.
1993: Weathering zone cosmogenic radionuclide Quaternary surface exposure age chronology, Gros Morne National Park, Western Newfoundland. Progress Report for Gros Morne National Park, 17 pages.

Gosse, J.C., Grant, D.R., Klein, J., Klassen, R.A., Evenson, E.B., Lawn, B. and Middleton, R.
1993: Significance of altitudinal weathering zones in Atlantic Canada, inferred from in situ produced cosmogenic radionuclides. *In* Geological Society of America. Northeastern Section, Program with Abstracts, pages 394.

Grant, D.R.
1969a: Late Pleistocene readvance of piedmont glaciers in western Newfoundland. Maritime Sediments, Volume 5, pages 126-128.

1969b: Surficial deposits, geomorphic features, and Late Pleistocene history of the terminus of the Northern Peninsula of Newfoundland and adjacent Québec-Labrador. Maritime Sediments, Volume 5, pages 123-125.

1972: Postglacial emergence of northern Newfoundland. Geological Survey of Canada, Paper 72-1B, pages 100-102.

1973: Surficial Geology map, Pasadena map sheet (12H/4). Geological Survey of Canada, Open File 180.

1974: Prospecting in Newfoundland and the theory of multiple shrinking ice caps. *In* Report of Activities. Geological Survey of Canada, Paper 74-1B, pages 215-216.

1975: Glacial style and the Quaternary stratigraphic record in the Atlantic Provinces, Canada. Geological Survey of Canada, Paper 75-1B, pages 109-110.

1976: Glacial style and the Quaternary stratigraphic record in the Atlantic Provinces, Canada. *In* Quaternary Stratigraphy of North America. *Edited by* W.C. Mahaney. Dowden, Hutchinson and Ross Inc., Stroudsburg, Pa, USA, pages 33-36.

1977a: Glacial style and ice limits, the Quaternary stratigraphic record, and changes of land and ocean level in the Atlantic Provinces, Canada. Géographie physique et Quaternaire, Volume 31, pages 247-260.

1977b: Altitudinal weathering zones and glacial limits in western Newfoundland, with particular reference to Gros Morne National Park. Geological Survey of Canada, Paper 77-1A, pages 455-463.

1980: Quaternary sea level change in Atlantic Canada as an indication of crustal delevelling. *In* Earth Rheology, Isostasy, and Eustacy. *Edited by* N.A. Morner. Wiley Interscience, Toronto, pages 201-214.

1987: Quaternary geology of Nova Scotia and Newfoundland (including Magdalen Islands). International Union for Quaternary Research, XII INQUA Congress, Ottawa, Excursion Guidebook A-3/C-3, National Research Council of Canada, Publication 27525, 62 pages.

1989a: Quaternary geology of the Atlantic Appalachian region of Canada. *In* Quaternary Geology of Canada and Greenland. *Edited by* R.J. Fulton. Geological Survey of Canada, Geology of Canada No. 1, pages 391-440.

1989b: Surficial geology, Sandy Lake-Bay of Islands, Newfoundland. Geological Survey of Canada, Map 1664A, Scale 1:250 000.

1991: Surficial geology, Stephenville-Port aux Basques, Newfoundland. Geological Survey of Canada, Map 1737A, Scale 1:250 000.

1992: Quaternary geology of St. Anthony-Blanc Sablon area, Newfoundland and Québec. Geological Survey of Canada, Memoir 427, 60 pages.

1994a: Quaternary geology of Port Saunders map area, Newfoundland. Geological Survey of Canada, Paper 91-20, 59 pages.

1994b: Quaternary geology of Gros Morne National Park area, western Newfoundland, Canada: glaciation, sea-level change, and mass wasting. Field Trip Guidebook, International Union of Geological Sciences, Commission on Geological Sciences for Environmental Planning, Corner Brook, Newfoundland, July 1994.

Gravenor, C.P. and Coyle, D.
1985: Origin and magnetic fabric of glacial varves, Nottawasaga River, Ontario, Canada. Canadian Journal of Earth Sciences, Volume 22, pages 291-294.

Gravenor, C.P., Von Brunn, V. and Dreimanis, A.
1984: Nature and classification of waterlain glaciogenic sediments, exemplified by Pleistocene, Late Paleozoic and Late Precambrian deposits. Earth Science Reviews, Volume 20, pages 105-166.

Gustavson, T.C., Ashley, G.M. and Boothroyd, J.C.
1975: Depositional sequences in glaciolacustrine deltas. *In* Glaciofluvial and Glaciolacustrine Sedimentation. *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Paleontologists and Mineralogists Special Publication No. 23, pages 264-280.

Haldorsen, S.
1981: Grain-size distribution of subglacial till and its relation to glacial crushing and abrasion. Boreas, Volume 10, pages 91-105.

Haldorsen, S. and Shaw, J.
1982: The problem of recognising melt-out till. Boreas, Volume 11, pages 261-277.

Hancock, P.L. and Barka, A.A.
1987: Kinematic indicators on active normal faults in western Turkey. Journal of Structural Geology, Volume 9, pages 573-584.

Hansel, A. and Johnson, W.H.
1992: Fluctuations of the Lake Michigan lobe during the late Wisconsin subepisode. *In* Quaternary Stratigraphy, Glacial Morphology and Environmental Changes. *Edited by* A.-M. Robertson, B. Ringberg, U. Miller and L. Brunnberg. Sveriges Geologiska Undersökning, Research Paper Series Ca 81, pages 133-144.

Hansel, A., *et al.*
1985:

Harder, H.
1970: Boron content of sediments as a tool in facies analysis. Sedimentary Geology, Volume 4, pages 153-175.

Harms, J.C., Southard, J.B. and Walker, R.G.
1982: Structures and sequences in clastic rocks. Society of Economic Paleontologists and Mineralogists, Short Course 9, 249 pages.

Harris, C. and Bothamley, K.
1984: Englacial deltaic sediments as evidence for basal freezing and marginal shearing, Leirbreen, southern Norway. Journal of Glaciology, Volume 30, pages 30-34.

Harrison, P.
1957: A clay-till fabric, its character and origin. Journal of Geology, Volume 65, pages 275-308.

Hart, J.K.
1994: Till fabric associated with deformable beds. Earth Surface Processes and Landforms, Volume 19, pages 15-32.

1995: Subglacial erosion, deposition and deformation associated with deformable beds. Progress in Physical Geography, Volume 19, pages 173-191.

Hart, J.K. and Boulton, G.S.
1991: The interrelation of glaciotectonic and glaciodepositional processes within the glacial environment. Quaternary Science Reviews, Volume 10, pages 335-350.

Hart, J.K. and Roberts, D.H.
1994: Criteria to distinguish between subglacial glaciotectonic and glaciomarine sedimentation, I. Deformation styles and sedimentology. Sedimentary Geology, Volume 91, pages 191-213.

Hartshorn, J.H.
1958: Flowtill in southeastern Massachusetts. Bulletin of the Geological Society of America, Volume 69, pages 477-482.

Hein, F.J. and Walker, R.G.
1977: Bar evolution and development of stratification in the gravelly, braided, Kicking Horse River, British Columbia. Canadian Journal of Earth Sciences, Volume 14, pages 562-570.

1982: The Cambro-Ordovician Cap Enragé Formation, Québec, Canada: conglomeratic deposits of a braided submarine channel with terraces. Sedimentology, Volume 29, pages 309-329.

Hicock, S.R.
1991: On subglacial stone pavements in till. Journal of Geology, Volume 99, pages 607-619.

1992: Lobal interactions and rheologic superposition in subglacial till near Bradtville, Ontario, Canada. Boreas, Volume 21, pages 73-88.

Hicock, S.R. and Dreimanis, A.
1985: Glaciotectonic structures as useful indicators in glacial deposits: four Canadian case studies. Canadian Journal of Earth Sciences, Volume 22, pages 339-346.

1989: Sunnybrook drift indicates a grounded early Wisconsin glacier in the Lake Ontario basin. Geology, Volume 17, pages 169-172.

1992: Deformation till in the Great Lakes region - Implications for rapid flow along the south-central margin of the Laurentide Ice Sheet. Canadian Journal of Earth Sciences, Volume 29, pages 1565-1579.

Hicock, S.R., Goff, J.R., Lian, O.B. and Little, E.C.
1996: On the interpretation of subglacial till fabric. Journal of Sedimentary Research, Volume 66, pages 928-934.

Holmes, C.D.
1941: Till fabric. Bulletin of the Geological Society of America, Volume 52, pages 1299-1354.

Hooke, R. Leb.
1970: Morphology of the ice-sheet margin near Thule, Greenland. Journal of Glaciology, Volume 9, pages 303-325.

Humlum, O.
1978: A large till wedge in Denmark: implications for the subglacial thermal regime. Bulletin of the Geological Society of Denmark, Volume 27, pages 63-71.

1981: Observations on debris in the basal transport zone of Myrdalsjokull, Iceland. Annals of Glaciology, Volume 2, pages 71-78.

Hunter, R.E.
1977: Terminology of cross stratified sedimentary layers and climbing ripple structures. Journal of Sedimentary Petrology, Volume 47, pages 697-706.

Hyde, R.S.
1979: Geology of the Carboniferous strata in portions of the Deer Lake basin, western Newfoundland. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 79-6, 43 pages.

1982: Geology of the Carboniferous Deer Lake basin. Mineral Development Division, Department of Mines and Energy, Map 82-7.

1984: Geology and mineralization of the Carboniferous Deer Lake Basin, western Newfoundland. *In* Mineral Deposits of Newfoundland - A 1984 Perspective. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 84-3, pages 19-26.

Iverson, N.R.
1991: Morphology of glacial striae: Implications for abrasion of glacier beds and fault surfaces. Geological Society of America Bulletin, Volume 103, pages 1308-1316.

Ives, J.D.
1960: The deglaciation of Labrador-Ungava - an outline. Cahiers de Geographie, Quebec, Volume 4, pages 324-343.

1978: The maximum extent of the Laurentide ice sheet along the east coast of North America during the Last Glaciation. Arctic, Volume 31, pages 24-53.

Jackson, R.G.
1976: Depositional model of point bars in the lower Wabash River. Journal of Sedimentary Petrology, Volume 46, pages 579-594.

Jenness, S.E.
1960: Late Pleistocene glaciation of eastern Newfoundland. Geological Society of America Bulletin, Volume 71, pages 161-180.

Johansson,
1963: Orientation of pebbles in running water. A laboratory study. Geografiska Annaler 45, pages 85-112.

Jopling, A.V. and Walker, R.G.
1968: Morphology and origin of ripple drift cross lamination with examples from the Pleistocene of Massachusetts. Journal of Sedimentary Petrology, Volume 38, pages 971-984.

Jukes, J.B.
1842: Excursions in and about Newfoundland during the years 1839 and 1840. J. Murray, London, 2 volumes, 322 and 354 pages.

Karrow, P.F.
1959: Pleistocene geology of the Hamilton map-area. Ontario Department of Mines, Geological Circular No. 8, 6 pages.

1989: Quaternary Geology of the Great Lakes subregion. *In* Quaternary Geology of Canada and Greenland. *Edited by* R.J. Fulton. Geological Survey of Canada, Geology of Canada, No. 1, pages 319-389.

Karrow, P.F. and Calkin, P.E. (editors)
1985: Quaternary evolution of the Great Lakes. Geological Association of Canada, Special Paper 30, 258 pages.

Kerr, A.
1994: Magmatic, hydrothermal and surficial processes in the development of multicoloured dimension-stone granites of the Topsail Plateau area (NTS 12H/02). *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 94-1, pages 147-166.

Kerr, J.H.
1870: Observations on ice-marks in Newfoundland. Quarterly Journal of the Geological Society of London, Volume 26, pages 704-705.

Kirby, G.E.
1988: Soils of the Pasadena-Deer Lake area, Newfoundland. Newfoundland Soil Survey, Agriculture Canada, Report 17, 154 pages.

Klassen, R.A.
1994: A preliminary interpretation of glacial history derived from glacial striations, central Newfoundland. *In* Current Research. Geological Survey of Canada, Paper 1994-D, pages 13-22.

Klassen, R.A. and Murton, J.B.
1996: Quaternary geology of the Buchans area, Newfoundland: implications for mineral exploration. Canadian Journal of Earth Sciences, Volume 33, pages 363-377.

Klassen, R.A. and Thompson, F.J.
1993: Glacial history, drift composition, and mineral exploration, central Labrador. Geological Survey of Canada, Bulletin 435, 76 pages.

Kleman, J.
1990: On the use of glacial striae for reconstruction of paleo-ice sheet flow patterns. Geografiska Annaler 72A, pages 217-236.

1992: The palimpsest glacial landscape in northwestern Sweden. Geografiska Annaler 74A, pages 305-325.

1994: Preservation of landforms under ice sheets and ice caps. Geomorphology, Volume 9, pages 19-32.

Kleman, J. and Borgström, I.
1996: Reconstruction of palaeo-ice sheets: The use of geomorphological data. Earth Surface Processes and Landforms, Volume 21, pages 893-909.

Knight, I.
1994: Geology of Cambrian-Ordovician platformal rocks of the Pasadena map area (NTS 12H/4). *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 94-1, pages 175-186.

1995: Preliminary 1:50 000 mapping of Lower Paleozoic parautochthonous sedimentary rocks of the Corner Brook area. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 95-1, pages 257-265.

Knudsen, K.L.
1971: Late Quaternary foraminifera from the Løkken area. Bulletin of the Geological Society of Denmark, Volume 21, pages 130-158.

Kruger, F.C.
1938: A clastic dike of glacial origin. American Journal of Science, Volume 35, pages 305-307.

Kruger, J.
1979: Structures and textures in till indicating subglacial deposition. Boreas, Volume 8, pages 323-340.

1984: Clasts with stoss-lee form in lodgement tills: A discussion. Journal of Glaciology, Volume 30, pages 241-243.

Krumbein, W.C.
1939: Preferred orientation of pebbles in sedimentary deposits. Journal of Geology, Volume 47, pages 673-706.

Kuenen, Ph.H.
1950: Turbidity currents of high density. Report of 18th International Geological Congress, London, Pt. 8, pages 44-52.

1953: Significant features of graded bedding. Bulletin of the American Association of Petroleum Geologists 37, pages 1044-1066.

1966a: Experimental turbidite lamination in an circular flume. Journal of Geology, Volume 74, pages 523-545.

1966b: Turbidity currents. *In* The Encyclopedia of Oceanography. *Edited by* R.W. Fairbridge. Encyclopedia of Earth Sciences Series, Volume 1, Reinhold Publishing Corporation, New York, pages 943-948.

Kuenen, Ph.H. and Migliorini, C.I.
1950: Turbidity currents as a cause of graded bedding. Journal of Geology, Volume 58, pages 91-125.

Lambeck, K.
1990: Late Pleistocene, Holocene and present sea-levels: constraints on future change. Paleogeography, Paleoclimatology, Paleoecology, Volume 89, pages 205-217.

Lambeck, K., Johnston, P. and Nakada, M.
1990: Holocene glacial rebound and sea-level change in NW Europe. Geophysics Journal International, Volume 103, pages 451-468.

Lambeck, K., Johnston, P., Smither, C. and Nakada, M.
1996: Glacial rebound of the British Isles .2. Constraints on mantle viscosity. Geophysical Journal International, Volume 125, pages 340-354.

Larsen, E. and Mangerud, J.
1992: Subglacially formed clastic dikes. Sveriges Geologiska Undersökning, Series Ca 81, pages 163-170.

LaSalle, P., Martineau, G. and Chauvin, L.
1977: Morphology, stratigraphy and deglaciation in Beauce-Notre Dame Mountains-Laurentide Park area. Ministère des Richesses naturelles du Québec, DPV-516, 74 pages.

Lawson, D.E.
1979: A comparison of the pebble orientations in ice and deposits of the Matanuska Glacier, Alaska. Journal of Geology, Volume 87, pages 629-645.

1981: Distinguishing characteristics of diamictons at the margin of the Matanuska Glacier, Alaska. Annals of Glaciology, Volume 2, pages 78-84.

1982: Mobilisation, movement, and deposition of active sub-aerial sediment flows, Matanuska Glacier, Alaska. Journal of Geology, Volume 90, pages 279-300.

1988: Glacigenic resedimentation: Classification concepts and application to mass-movement processes and deposits. *In* Genetic Classification of Glacigenic Deposits. *Edited by* R.P. Goldthwait and C.L. Matsch. A.A. Balkema, Rotterdam, pages 147-169.

Lawson, T.J.
1996: Glacial striae and former ice movement: the evidence from Assynt, Sutherland. Scottish Journal of Geology, Volume 32, pages 59-65.

Leeder, M.R.
1982: Sedimentology: Process and Product. George Allen and Unwin, London, 344 pages.

Lerman, A.
1966: Boron in clays and estimation of paleosalinities. Sedimentology, Volume 6, pages 247-266.

Levesque, A.J., Mayle, F.E., Walker, I.R. and Cwynar, L.C.
1993: A previously unrecognized late-glacial cold event in eastern North-America. Nature, Volume 361, pages 623-626.

Levinson, A.A. and Ludwick, J.C.
1966: Speculation on the incorporation of boron into argillaceous sediments. Geochimica Cosmochimica Acta, Volume 30, pages 855-861.

Lewis, C.F.M., Moore, T.C., Rea, D.K., Dettman, D.L., Smith, A.M. and Mayer, L.A.
1994: Lakes of the Huron basin: Their record of runoff from the Laurentide Ice Sheet. Quaternary Science Reviews, Volume 13, pages 891-922.

Lindroth, C.H.
1963: The faunal history of Newfoundland illustrated by Carabid beetles. Opuscula Entomologica, Supplement 23, Lund, Entomologiska Sallskapet, 112 pages.

Link, P.K. and Gostin, V.A.
1981: Facies and paleogeography of Sturtian glacial strata (Late Precambrian), South Australia. American Journal of Science, Volume 281, pages 353-374.

Liverman, D.G.E.
1991: Sedimentology and history of a Late Wisconsinan glacial lake, Grand Prairie, Alberta, Canada. Boreas, Volume 20, pages 241-257.

1992: Application of regional Quaternary mapping to mineral exploration, northeastern Newfoundland, Canada. Transactions of the Institution of Mining and Metallurgy Section B - Applied Earth Science, Volume 101, pages B89-B98.

1994: Relative sea-level history and isostatic rebound in Newfoundland, Canada. Boreas, Volume 23, pages 217-230.

Liverman, D.G.E. and Batterson, M.J.
1995: West Coast of Newfoundland field trip guide. CANQUA-CCRG Joint Meeting, Programme, Abstracts, Field Guides, St. John's, Newfoundland, pages WC1-WC78.

Liverman, D.G.E. and Bell, T.
1996: Late Quaternary glacial and glaciomarine sediments in southern St. George's Bay. Current Research, Report 96-1, pages 29-40.

Liverman, D.G.E., Catto, N., Batterson, M.J., MacKenzie, C., Scott, S. and Sommerville, A.
In prep.: Evidence of late glacial permafrost in Newfoundland. Géographie physique et Quaternaire.

Liverman, D.G.E., Scott, S. and Vatcher, H.
1991: Surficial Geology and landform classification, Springdale map sheet, NTS 12H/8. Newfoundland Department of Mines and Energy, Open File 012H/1216, Map 91-12.

Liverman, D.G.E. and St. Croix, L.
1989a: Ice flow indicators on the Baie Verte Peninsula. Newfoundland Department of Mines and Energy, Open File map 89-36.

1989b: Quaternary Geology of the Baie Verte Peninsula. *In* Current Research. Newfoundland Department of Mines, Geological Survey of Newfoundland, Report 89-1, pages 237-247.

Liverman, D.G.E., St. Croix, L., Vatcher, H. and Scott, S.
1990: Ice flow indicators, Springdale map sheet (12H/8). Newfoundland Department of Mines and Energy, Geological Survey of Newfoundland, Map 90-123.

Løken, O.H.
1962: On the vertical extent of glaciation in northeastern Labrador-Ungava. Canadian Geographer, Volume 6, pages 106-119.

Lønne, I.
1995: Sedimentary facies and depositional architecture of ice-contact glaciomarine systems. Sedimentary Geology, Volume 98, pages 13-43.

Lotter, A.F.
1991: Absolute dating of the late-glacial period in Switzerland using annually laminated sediments. Quaternary Research, Volume 35, pages 321-330.

Lotter, A.F., Eicher, U., Siegenthaler, U. and Birks, H.J.B.
1992: Late-glacial climatic oscillations as recorded in Swiss lake sediments. Journal of Quaternary Science, Volume 7, pages 187-204.

Lowe, D.R.
1976: Grain flow and grain flow deposits. Journal of Sedimentary Petrology, Volume 46, pages 188-199.

1982: Sediment gravity flows II, Depositional models with special reference to the deposits of high density turbidity currents. Journal of Sedimentary Petrology, Volume 52, pages 279-297.

Lowe, J.J. and Cairns, P.
1991: New pollen-stratigraphic evidence for the deglaciation and lake drainage chronology of the Glen Roy-Glen Spean area. Scottish Journal of Geology, Volume 27, pages 41-56.

Lowe, J.J. and Walker, M.J.C.
1984: Reconstructing Quaternary environments. Longman, New York. 389 pages.

Lubinsky, I.
1980: Marine bivalve molluscs of the Canadian central and eastern Arctic: Faunal composition and zoogeography. Canadian Bulletin of Fisheries and Aquatic Sciences, Bulletin 207, 111 pages.

Lundqvist, J.
1965: Glacial geology in northeastern Newfoundland. Geologiska Foreningen i Stockholm Forhandlingar, Volume 87, pages 285-306.

1977: Till in Sweden. Boreas, Volume 6, pages 73-85.

1990: Glacial morphology as an indicator of the direction of glacial transport. *In* Glacial Indicator Tracing. *Edited by* R. Kujansuu and M. Saarnisto. A.A. Balkema, Rotterdam, pages 61-70.

MacClintock, P. and Twenhofel, W.H.
1940: Wisconsin glaciation of Newfoundland. Bulletin of the Geological Society of America, Volume 51, pages 1729-1756.

MacEachran, D.B.
1990: Stereo™, the sterographic projection software program for the Apple Macintosh® computer. Distributed by Rockware Inc, Wheat Ridge, Colorado, USA.

MacKay, J.R.
1990: Some observations on the growth and deformation of epigenetic, syngenetic and anti-syngenetic ice wedges. Permafrost and Periglacial Processes, Volume 1, pages 15-29.

1992: The frequency of ice-wedge cracking (1967-1987) at Garry Island, western Arctic coast, Canada. Canadian Journal of Earth Sciences, Volume 29, pages 236-248.

MacKiewicz, N.E., Powell, R.D., Carlson, P.R. and Molnia, B.F.
1984: Interlaminated ice-proximal glacimarine sediments in Muir Inlet, Alaska. Marine Geology, Volume 57, pages 113-147.

MacLean, B., Vilks, G. and Deonarine, B.
1992: Depositional environments and history of Late Quaternary sediments in Hudson Strait and Ungava Bay: further evidence from seismic and biostratigraphic data. Géographie physique et Quaternaire, Volume 46, pages 311-329.

MacPherson, J.B.
1981: The development of the vegetation of Newfoundland and climatic change during the Holocene. *In* The Natural Environment of Newfoundland Past and Present. *Edited by* A.G. Macpherson and J.B. Macpherson. Department of Geography, Memorial University of Newfoundland, pages 189-217.

1995: A 6ka reconstruction for the island of Newfoundland from a synthesis of Holocene lake-sediment pollen records. Géographie physique et Quaternaire, Volume 49, pages 163-182.

1996: Delayed deglaciation by downwasting of the northeast Avalon Peninsula, Newfoundland: An application of the early postglacial pollen record. Géographie physique et Quaternaire, Volume 50, pages 201-220.

MacPherson, J.B. and Anderson, T.W.
1985: Further evidence of late glacial climatic fluctuations from Newfoundland: pollen stratigraphy from a north coast site. *In* Current Research, Part B. Geological Survey of Canada, Paper 85-1, pages 383-390.

Magny, M. and Ruffaldi, P.
1995: Younger Dryas and early Holocene lake-level fluctuations in the Jura mountains, France. Boreas, Volume 24, pages 155-172.

Mark, D.M.
1973: Analysis of axial orientation data, including till fabrics. Bulletin of the Geological Society of America, Volume 84, pages 1369-1374.

1974: On the interpretation of till fabrics. Geology, Volume 2, pages 101-104.

Mathews, W.H.
1980: Retreat of the last ice sheets in northeastern British Columbia and adjacent Alberta. Geological Survey of Canada, Bulletin 331.

Matsch, C.L. and Schneider, A.F.
1986: Stratigraphy and correlation of the glacial deposits of the glacial lobe complex in Minnesota and northwestern Wisconsin. *In* Quaternary Glaciations in the Northern Hemisphere. *Edited by* V. Sibrava, D.Q. Bowen and G.M. Richmond. Pergamon Press, Oxford, pages 59-64.

McCarroll, D. and Nesje, A.
1993: The vertical extent of ice sheets in Nordfjord, western Norway - measuring degree of rock surface weathering. Boreas, Volume 22, pages 255-265.

McKee, E.D.
1965: Experiments on ripple lamination. *In* Primary Sedimentary Structures and their Hydrodynamic Interpretation. *Edited by* G.V. Middleton. Society of Economic Paleontologists and Mineralogists, Special Publication 12, pages 66-83.

McNeely, R. and Jorgensen, P.K.
1993: Geological Survey of Canada Radiocarbon Dates XXXI. Geological Survey of Canada, Paper 91-7.

McNeely, R. and McCuaig, S.
1991: Geological Survey of Canada Radiocarbon Dates XXIX. Geological Survey of Canada, Paper 89-7.

Menzies, J.
1989: Subglacial hydraulic conditions and their possible impact upon subglacial bed formation. Sedimentary Geology, Volume 62, pages 125-150.

1990a: Brecciated diamictons from Mohawk Bay, S. Ontario, Canada. Sedimentology, Volume 37, pages 481-493.

1990b: Sand intraclasts within a diamicton mélange, southern Niagara Peninsula, Ontario, Canada. Journal of Quaternary Science 5, pages 189-206.

Miall, A.D.
1977: A review of the braided river depositional environment. Earth Science Reviews, Volume 13, pages 1-62.

1978: Lithofacies types and vertical profile models in braided river deposits: a summary. *In* Fluvial Sedimentology. *Edited by* A.D. Miall. Canadian Society of Petroleum Geologists, pages 597-604.

1983: Principles of sedimentary basin analysis. Springer Verlag, New York.

1984: Deltas. *In* Facies Models, second edition. *Edited by* R.G. Walker. Geoscience Canada, Reprint Series 1, pages 105-118.

1992: Alluvial deposits. *In* Facies Models Response to Sea Level Change. *Edited by* R.G. Walker and N.P. James. Geological Association of Canada, pages 119-142.

Mickelson, D.M.
1971: Glacial geology of the Burroughs Glacier area, southeastern Alaska. Ohio State University, Institute of Polar Studies, Report 40, 149 pages.

Middleton, G.V.
1967: Experiments on density and turbidity currents. III Deposition of sediment. Canadian Journal of Earth Sciences, Volume 4, pages 475-505.

1993: Sediment deposition from turbidity currents. Annual Review of Earth and Planetary Science, Volume 21, pages 89-114.

Middleton, G.V. and Hampton, M.A.
1973: Part 1. Sediment gravity flows: mechanics of flow and deposition. *In* Turbidites and Deep Water Sedimentation. *Edited by* G.V. Middleton and A.H. Bouma. SEPM Pacific Section, Short Course Notes, Anaheim, California. 38 pages.

1976: Subaqueous sediment transport and deposition by sediment gravity flows. *In* Marine Sediment Transport and Environmental Management. *Edited by* D.J. Stanley and D.J.P. Swift. Wiley, New York, pages 197-218.

Miller, G.H. and Dyke, A.S.
1974: Proposed extent of late Wisconsin Laurentide ice on eastern Baffin Island. Geology, Volume 2, pages 125-130.

Mills, H.H.
1977: Differentiation of glacier environments by sediment characteristics: Athabasca glacier, Alberta, Canada. Journal of Sedimentary Petrology, Volume 47, pages 728-737.

1984: Clast orientation in Mount St. Helens debris-flow deposits, North Fork Toutle River, Washington. Journal of Sedimentary Petrology, Volume 54, pages 626-634.

Milne, J.H.
1874: Notes on the physical features and mineralogy of Newfoundland. Quarterly Journal of the Geological Society of London, Volume 30, pages 722-745.

1876: Ice and ice-work in Newfoundland. Geological Magazine, Volume 3, pages 303-308, 345-350, 403-410.

Mitrovica, J.X. and Peltier, W.R.
1995: Constraints on mantle viscosity based upon the inversion of post-glacial uplift data from the Hudson Bay region. Geophysics Journal International, Volume 122, pages 353-377.

Molnia, B.F.
1983: Subarctic glacial-marine sedimentation: a model. *In* Glacial Marine Sedimentation. *Edited by* B.F. Molnia, Plenum Press, New York, pages 95-144.

Mooers, H.D.
1989: On the formation of the tunnel valleys of the Superior Lobe, central Minnesota. Quaternary Research, Volume 32, pages 24-35.

Moore, R.C.
1964: Treatise on invertebrate paleontology. Geological Society of America and University of Kansas, 511 pages.

Mörner, N.-A.
1973: A new find of till wedges in Nova Scotia, Canada. Geologiska Foreningen i Stockholm Forhandlingar, Volume 95, pages 272-273.

Moss, A.J.
1963: The physical nature of common sandy and pebbly deposits. Part II. American Journal of Science, Volume 261, pages 297-343.

Mosser, C.
1983: The use of B, Li and Sn in determining the origin of some sedimentary clays. Chemical Geology, Volume 38, pages 129-139.

Muller, E.H.
1983: Dewatering during lodgement of till. *In* Tills and Related Deposits. *Edited by* E.B. Evenson, Ch. Schlüchter andJ. Rabassa. A.A. Balkema, Rotterdam, pages 13-18.

Murray, A.
1866: Report on the geology of Newfoundland for 1865, with a prefatory report and an appendix by Sir W.E. Logan. Montreal, J. Lovell.

1882: Glaciation in Newfoundland. Proceedings and Transactions of the Royal Society of Canada, Volume 1, pages 55-76.

Murray, A. and Howley, J.P.
1881: Geological Survey of Newfoundland. Edward Stanford, London, 534 pages.

1918: Reports of Geological Survey of Newfoundland From 1881 to 1909. Robinson and Company Limited Press, St. John's, Nfld, 704 pages.

Murray, J.W.
1991: Ecology and palaeoecology of benthic foraminifera. Longman Scientific and Technical, Harlow, Essex, 387 pages.

Neale, E.R.W. and Nash, W.A.
1963: Sandy Lake (east half), Newfoundland. Geological Survey of Canada, Paper 62028, 40 pages.

Nemec, W. and Steel, R.J. (editors)
1988: Fan Deltas: Sedimentology and tectonic settings. Blackie, London, 464 pages.

Nilsson-Cantell, K.-A.
1978: Cirripedia thoracica and acrothoracica. Universitetsforlagat Oslo, Norway, 134 pages.

Oberbeck, V.R., Marshall, J.R. and Aggarwal, H.
1993: Impacts, tillites, and the breakup of Gondwanaland. Journal of Geology, Volume 101, pages 1-19.

Ovenshine, A.T.
1970: Observations of ice rafting in Glacier Bay, Alaska, and identification of ice rafted deposits. Bulletin of the Geological Society of America, Volume 81, pages 891-894.

Owen, J.V.
1986: Geology of the Silver Mountain area, western Newfoundland. *In* Current Research, Part B. Geological Survey of Canada, Paper 86-1B, pages 515-522.

Owen, J.V. and Erdmer, P.
1986: Precambrian and Paleozoic metamorphism in the Long Range Inlier, western Newfoundland. *In* Current Research. Geological Survey of Canada, Paper 86-1B, pages 29-38.

Parent, M. and Occhietti, S.
1988: Late Wisconsinan deglaciation and Champlain Sea invasion in the St. Lawrence valley, Québec. Géographie physique et Quaternaire, Volume 42, pages 215-246.

Peltier, W.R.
1974: The impulse response of a Maxwell Earth. Reviews of Geophysics and Space Physics, Volume 12, pages 649-669.

1976: Global sea level rise and glacial isostatic adjustment: An analysis of data from the east coast of North America. Geophysical Research Letters, Volume 23, pages 717-720.

Peltier, W.R. and Andrews, J.T.
1976: Glacial-isostatic adjustment-I. The forward problem. Geophysical Journal of the Royal Astronomical Society, Volume 46, pages 605-646.

Peteet, D.
1995: Global Younger Dryas? Quaternary International, Volume 28, pages 93-104.

Péwé, T.L.
1966: Ice wedges in Alaska - classification, distribution and climatic significance. *In* Proceedings, 1st International Permafrost Conference. National Academy of Science, National Research Council of Canada, Publication 1287, pages 76-81.

Picard, M.D. and High, J.R.
1973: Sedimentary structures of ephemeral streams. Elsevier, Amsterdam, 223 pages.

Pilsbury, H.A.
1916: The sessile barnacles (cirripedia) contained in the collections of US National Museum: Including a monograph of the American species. Smithsonian Institution, United States National Museum, Bulletin 93, 366 pages.

Piper, D.J.W., Mudie, P.J., Fader, G.B., Josenhans, H.W., MacLean, B. and Vilks, G.L.
1990: Quaternary geology, Chapter 10. *In* Geology of the Continental Margin of Eastern Canada. *Edited by* M.J. Keen and G.L. Williams. Geological Survey of Canada, Geology of Canada, No. 2, pages 475-607.

Postma, G.
1995: Sea-level-related architectural trends in coarse-grained delta complexes. Sedimentary Geology, Volume 98, pages 3-12.

Potter, P.E. and Pettijohn, F.J.
1963: Palaeocurrents and basin analysis. Springer, Berlin, 425 pages.

Powell, R.D.
1981: A model for sedimentation by tidewater glaciers. Annals of Glaciology, Volume 2, pages 129-134.

1983: Glacial-marine sedimentation processes and lithofacies of temperate tidewater glaciers. *In* Glacial Marine Sedimentation. *Edited by* B.F. Molnia, Plenum Press, New York, pages 185-231.

1984: Glacimarine processes and inductive lithofacies modeling of ice shelf and tidewater glacier sediments based on Quaternary examples. Marine Geology, Volume 57, pages 1-52.

1990: Processes at glacial grounding-line fans and their growth to ice-contact deltas. *In* Glacimarine Environments: Processes and Sediments. *Edited by* J.A. Dowdeswell and J.D. Scourse. Geological Society of London, Special Publication, Number 53, pages 53-74.

1991: Grounding line systems as second-order controls on fluctuations of tidewater glaciers. *In* Glacial Marine Sedimentation; Paleoclimatic Significance. *Edited by* J.B. Anderson and G.M. Ashley, Geological Society of America, Special Paper 261, pages 75-94.

Powell, R.D. and Molnia, B.F.
1989: Glacimarine sedimentary processes, facies and morphology of the south-south-east Alaska shelf and fjords. *In* Modern Glacimarine Environments: Glacial and Marine Controls of Modern Lithofacies and Biofacies. *Edited by* R.D. Powell and A. Elverhøi. Marine Geology, Volume 85, pages 359-390.

Power, W.L. and Tullis, T.E.
1989: The relationship between slickensides surfaces in fine-grained quartz and the seismic cycle. Journal of Structural Geology, Volume 11, pages 879-894.

Prest, V.K.
1970: Quaternary geology of Canada. *In* Geology and Economic Minerals of Canada. *Edited by* R.J.W. Douglas. Economic Geology Report, Number 1, 4th edition, Ottawa, Department of Energy, Mines and Resources, 838 pages.

Prest, V.K., Grant, D.R. and Rampton, V.N.
1968: Glacial map of Canada. Geological Survey of Canada, Map 1253A, Scale 1:5 000 000.

Price, R.J.
1973: Glacial and fluvioglacial landforms. Oliver and Boyd, Edinburgh, 242 pages.

Prior, D.B. and Bornhold, B.D.
1990: The underwater development of Holocene fan-deltas. *In* Coarse-grained Deltas. *Edited by* A. Colella and D.B. Prior. International Association of Sedimentologists, Special Publication 10, pages 75-90.

Proudfoot, D.N., Grant, D.R. and Batterson, M.J.
1988: Quaternary geology of western Newfoundland. Field Trip Guidebook, Geological Association of Canada, St. John's, Newfoundland, 53 pages.

Quinlan, G. and Beaumont C.
1981: A comparison of observed and theoretical post-glacial relative sea level in Atlantic Canada. Canadian Journal of Earth Sciences, Volume 18, pages 1146-1163.

1982: The deglaciation of Atlantic Canada as reconstructed from the postglacial relative sea level record. Canadian Journal of Earth Sciences, Volume 19, pages 2232-2246.

Rappol, M.
1985: Clast-fabric strength in tills and debris flows compared for different environments. Geologie en Mijnbouw, Volume 64, pages 327-332.

1989: Glacial history and stratigraphy of northwestern New Brunswick. Géographie physique et Quaternaire, Volume 43, pages 191-206.

Rawlence, D.J.
1988: The post-glacial diatom history of Splan Pond. Journal of Paleolimnology, Volume 1, pages 51-60.

1992: Paleophycology of Long Lake, Saint John County, New Brunswick, Canada, based on diatom distribution in sediments. Canadian Journal of Botany, Volume 70, pages 229-239.

Reasoner, M.A., Osborn, G. and Rutter, N.W.
1993: The age of the Crowfoot Moraine system in the Canadian Rocky Mountains; A possible Younger Dryas time equivalent glacial event. CANQUA Program with abstracts and field guides, Victoria, British Columbia, A37 pages.

Reinhardt, E.G.
1996: A combined micropaleontological and strontium isotope methodology as a paleoceanographic tool. Unpublished Ph. D. thesis, Department of Earth Science, Carleton University, Ontario, 195 pages.

Ricketts, M.J.
1987: Wild Cove clay, western Newfoundland: A review and recent work. *In* Current Research. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 87-1, pages 57-62.

1993: Aggregate resources graph program (Graph v 3). Newfoundland Department of Mines and Energy, Geological Survey Branch. Open File Nfld (2270).

Roberts, B.A.
1993: Ecology of landscapes with serpentinized rocks. *In* The Scientific Challenge of Our Changing Environment. *Edited by* J. Hall and M. Wadleigh. Canadian Global Change Program Incidental Report Series No. IR93-2, Royal Society of Canada, pages 78-79.

Rodrigues, C.G.
1992: Successions of Invertebrate Microfossils and the Late Quaternary Deglaciation of the Central St-Lawrence Lowland, Canada and United-States. Quaternary Science Reviews, Volume 11, pages 503-534.

Rogerson, R.J.
1979: Drift prospecting in the Deer Lake lowlands.

Westfield Minerals Limited, Reidville, Newfoundland. Newfoundland Department of Mines and Energy. Unpublished Report. Open File 12H(559), 12 pages.

1981: The tectonic evolution and surface morphology of Newfoundland. *In* The Natural Environment of Newfoundland Past and Present. *Edited by* A.G. Macpherson and J.B. Macpherson. Memorial University of Newfoundland, pages 24-55.

1982: The glaciation of Newfoundland and Labrador. *In* Prospecting in Areas of Glaciated Terrain - 1982. *Edited by* P.H. Davenport. Canadian Institute of Mining and Metallurgy, pages 37-56.

Rogerson, R.J. and Tucker, C.M.
1972: Observations on the glacial history of the Avalon Peninsula, Newfoundland. Maritime Sediments, Volume 8, pages 25-31.

Röthlisberger, H.
1972: Water pressure in intra- and subglacial channels. Journal of Glaciology, Volume 11, pages 177-203.

Rowe, J.S.
1972: Forest regions of Canada. Department of the Environment, Canadian Forestry Service, Publication 1300.

Ruddiman, W.F.
1987: Synthesis: The ocean / ice sheet record. *In* North America and Adjacent Oceans during the Last Deglaciation. *Edited by* W.F. Ruddiman and H.E. Wright, Jr. Geological Society of America, The Geology of North America, volume K-3, pp. 463-478.

Rust, B.R.
1975: Fabric and structure in glaciofluvial gravels. *In* Glaciofluvial and Glaciolacustrine Sedimentation. *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Paleontologists and Mineralogists, Special Publication 23, pages 238-248.

1977: Mass flow deposits in a Quaternary succession near Ottawa, Canada: Diagnostic criteria for subaqueous outwash. Canadian Journal of Earth Sciences, Volume 14, pages 175-184.

1978: Depositional models for braided alluvium. *In* Fluvial Sedimentology. *Edited by* A.D. Miall. Canadian Society of Petroleum Geologists, pages 605-625.

Rust, B.R. and Koster, E.H.
1984: Coarse alluvial deposits. *In* Facies Models, second edition. *Edited by* R.G. Walker. Geoscience Canada, Reprint Series 1, pages 53-69.

Rust, B.R. and Romanelli, R.
1975: Late Quaternary subaqueous outwash deposits near Ottawa, Canada. *In* Glaciofluvial and Glaciolacustrine Sedimentation. *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Paleontologists and Mineralogists 23, pages 177-192.

Ryder, J.M.
1971a: The stratigraphy and morphology of paraglacial alluvial fans in south-central British Columbia. Canadian Journal of Earth Sciences, Volume 8, pages 279-298.

1971b: Some aspects of the morphometry of paraglacial fans in south-central British Columbia. Canadian Journal of Earth Sciences, Volume 8, pages 1252-1264.

Saunders, C.M. and Smyth, W.R.
1990: Geochemical characteristics of the Gull Lake Intrusive Suite and Devils Room granite, western White Bay, Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 90-1, pages 183-199.

Sayles, R.W.
1919: Seasonal deposition in aqueo-glacial sediments. Memoirs of Museum of Comparative Zoology, Harvard University, Volume 47, pages 1-67.

Schermerhorn, L.J.G.
1974: Late Precambrian mixtites: glacial and/or nonglacial? American Journal of Science, Volume 274, pages 673-824.

Scott, D.B., Boyd, R. and Medioli, F.S.
1987: Relative sea-level changes in Atlantic Canada: Observed level and sedimentological changes vs. theoretical models. *In* Sea Level Fluctuations and Coastal Evolution. *Edited by* D. Nummendal, O.H. Pilkey andJ.D. Howard. Society of Economic Paleontologists and Mineralogists, Special Publication No. 41, pages 87-96.

Scott, D.B., Mudie, P.J., Vilks, G. and Younger, D.C.
1984: Latest Pleistocene-Holocene palaeoceanographic trends on the continental margin of eastern Canada: Foraminiferal, Dinoflagellate and pollen evidence. Marine micropaleontology, Volume 9, pages 181-218.

Scott, D.B., Schafer, C.T. and Medioli, F.S.
1980: Eastern Canadian estuarine foraminifera: a framework for comparison. Journal of Foraminiferal Research, Volume 10, pages 205-234.

Scott, K.M.
1988: Origins, behaviour, and sedimentology of lahars and lahar-runout flows in the Toutlt-Cowlitz River system. United States Geological Survey, Professional Paper 1447-A, pages A1-A74.

Scott, S., Catto, N.R. and Liverman, D.G.E.
1991: Sedimentology of Quaternary marine sediments

in the Hall's Bay area, Newfoundland. Atlantic Geology, Volume 27, pages 181-191.

Scott, W.B. and Crossman, E.J.
1973: Freshwater fishes of Canada. Department of Fisheries and Oceans, Fisheries Research Board, Bulletin 184, 966 pages.

Seabrook, W.D.
1962: A survey of nine lakes on the Island of Newfoundland. Environment Canada, Fisheries and Marine Services, 70 pages.

Sharpe, R.P.
1938: Pleistocene glaciation in the Ruby-East Humboldt Range, northeastern Nevada. Journal of Geomorphology, Volume 1, pages 296-323.

Shaw, J.
1975: Sedimentary successions in Pleistocene ice-marginal lakes. *In* Glaciofluvial and Glaciolacustrine Sedimentation. *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Paleontologists and Mineralogists Special Publication 23, pages 281-303.

1977: Till body morphology and structure related to glacier flow. Boreas, Volume 6, pages 189-201.

1979: Genesis of the Sveg tills and Rogen moraines of central Sweden: a model of basal meltout. Boreas, Volume 8, pages 409-426.

1982: Meltout till in the Edmonton area, Alberta, Canada. Canadian Journal of Earth Sciences, Volume 19, pages 1548-1569.

1983: Forms associated with boulders in melt-out till. *In* Tills and Related Deposits. *Edited by* E.B. Evenson, Ch. Schlüchter and J. Rabassa. A.A. Balkema, Rotterdam, pages 3-13.

1987: Glacial sedimentary processes and environmental reconstruction based on lithofacies. Sedimentology, Volume 34, pages 103-116.

1988: Sublimation till. *In* Genetic Classification of Glacigenic Deposits. *Edited by* R.P. Goldthwait and C.L. Matsch. A.A. Balkema, Rotterdam, pages 141-142.

1989: Drumlins, subglacial meltwater floods, and ocean responses. Geology, Volume 17, pages 853-856.

Shaw, J. and Archer, J.
1978: Winter turbidity current deposits in Late Pleistocene glaciolacustrine varves, Okanagan Valley, British Columbia, Canada. Boreas, Volume 7, pages 123-130.

Shaw, J., Gilbert, R. and Archer, J.
1978: Proglacial lacustrine sedimentation during winter. Arctic and Alpine Research, Volume 10, pages 689-699.

Shaw, J. and Forbes, D.L.
1995: The postglacial relative sea-level lowstand in Newfoundland. Canadian Journal of Earth Sciences, Volume 32, pages 1308-1330.

Shaw, J., Forbes, D.L., Ceman, J.A., Asprey, D.E., Beaver, B., Wile, B., Frobel, D. and Jodrey, F.
1995: Marine geological surveys in Chedabucto and St. George's Bays, Nova Scotia, and Bay of Islands, Newfoundland. Geological Survey of Canada, Open File 3230.

Shimp, N.F., Witters, J., Potter, P.E. and Schleicher, J.A.
1969: Distinguishing marine and fresh-water muds. Journal of Geology, Volume 77, pages 566-580.

Shultz, A.W.
1984: Subaerial debris-flow deposition in the upper Paleozoic Cutler Formation, western Colorado. Journal of Sedimentary Petrology, Volume 54, pages 759-772.

Sissons, J.B.
1979: Catastrophic lake drainage in Glen Spean and the Great Glen, Scotland. Journal of the Geological Society of London, Volume 136, pages 215-224.

Sissons, J.B. and Cornish, R.
1983: Fluvial landforms associated with ice-dammed lake drainage in upper Glen Roy, Scotland. Proceedings of the Geological Association, Volume 94, pages 45-52.

Sladen, J.A. and Wrigley, W.
1983: Geotechnical properties of lodgement till. *In* Glacial Geology: An Introduction for Engineers and Earth Scientists. *Edited by* N. Eyles. Pergamon Press, England, pages 302-317.

Smith, D.G.
1992: Glacial Lake Mackenzie, Mackenzie Valley, Northwest Territories, Canada. Canadian Journal of Earth Sciences, Volume 29, pages 1756-1766.

1994: Glacial Lake McConnell: Paleogeography, age, duration, and associated river deltas, Mackenzie River basin, western Canada. Quaternary Science Reviews, Volume 13, pages 829-843.

Smith, G.A. and Lowe, D.R.
1991: Lahars: Volcano-hydrologic events and deposition in the debris flow-hyperconcentrated flow continuum. *In* Sedimentation in Volcanic Settings. *Edited by* R.V. Fisher and G.A. Smith. Society of Economic Palaeontologists and Mineralogists Special Publication 45, pages 59-70.

Smith, G.W.
1985: Chronology of Late Wisconsinan deglaciation of coastal Maine. *In* Late Pleistocene History of Northeastern New England and Adjacent Quebec. *Edited by* H.W. Borns Jr, P. LaSalle and W.B. Thompson. Geological Society of America, Special Paper 197, pages 29-44.

Smith, N.D.
1974: Sedimentology and bar formation in the upper Kicking Horse River, a braided outwash stream. Journal of Geology, Volume 81, pages 205-223.

1978: Sedimentation processes and patterns in a glacier fed lake with low sediment input. Canadian Journal of Earth Sciences, Volume 15, pages 741-756.

Smith, N.D. and Syvitski, J.
1982: Sedimentation in a glacier fed lake: the role of pelletization in deposition of fine grained suspensates. Journal of Sedimentary Petrology, Volume 52, pages 503-513.

Smith, N.D., Vendl, M.A. and Kennedy, S.K.
1982: Comparison of sedimentation regimes in four glacier-fed lakes of western Alberta. *In* 6th Guelph Symposium on Geomorphological Research in Glacial, Glaciofluvial and Glaciolacustrine Systems. *Edited by* R. Davidson-Arnott, W. Nickling and B.D. Facey. Geobooks, Norwich, pages 203-238.

Smyth, W.R. and Schillereff, H.S.
1982: The pre-Carboniferous geology of southwest White Bay. *In* Current Research. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 82-1, pages 78-98.

Sommerville, A.
1997: The late Quaternary history of Terra Nova National Park and vicinity, northeast Newfoundland. Unpublished M.Sc. thesis. Department of Geography, Memorial University of Newfoundland, St. John's, Newfoundland.

Sorby, H.C.
1859: On the structures produced by the currents present during the deposition of stratified rocks. Geologist, Volume 2, pages 137-147.

Spalletti, L.A., Cingolani, C.A., Varela, R. and Cuerda, A.J.
1989: Sediment gravity flow deposits of an Ordovician deep-sea fan system (western Precordillera, Argentina). Sedimentary Geology, Volume 61, pages 287-301.

Sparkes, B.G.
1985: Quaternary mapping, Central Volcanic Belt. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 85-1, pages 94-98.

1987: Glacial geology and till geochemistry of the Buchans (12A/15) map area, Newfoundland. Mineral Development Division, Department of Mines and Energy, Open File 12A/15 (396), 11 pages.

Splettstoesser, J.F.
1985: Note on rock striations caused by penguin feet, Falkland Islands. Arctic and Alpine Research, Volume 17, pages 107-111.

St. Croix, L. and Taylor, D.M.
1991: Regional striation survey and deglacial history of the Notre Dame Bay area, Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 91-1, pages 61-68.

St. Onge, D.A.
1972: Sequence of glacial lakes in north-central Alberta. Geological Survey of Canada, Bulletin 213, 16 pages.

Stea, R.R., Finck, P.W. and Wightman, D.M.
1985: Quaternary geology and till geochemistry of the western part of Cumberland County, Nova Scotia (sheet 9). Geological Survey of Canada, Paper 85-17.

Stea, R.R., Forbes, D.L. and Mott, R.J.
1992: Quaternary geology and coastal evolution of Nova Scotia. Geological Association of Canada, Field Excursion A-6 Guidebook, 125 pages.

Stea, R.R. and Fowler, J.H.
1979: Minor and trace-element variations in Wisconsinan tills, Eastern Shore Region, Nova Scotia. Nova Scotia Department of Mines and Energy, Paper 79-4, 30 pages.

Sugden, D.E. and John, B.S.
1976: Glaciers and Landscape: A geomorphological approach. Edward Arnold, London, 376 pages.

Sugden, D.E. and Watts, S.H.
1977: Tors, felsenmeer and glaciation in northern Cumberland Peninsula, Baffin Island. Canadian Journal of Earth Sciences, Volume 14, pages 2817-2823.

Syvitski, J.P.M., Andrews, J.T. and Dowdeswell, J.A.
1996: Sediment deposition in an iceberg-dominated glacimarine environment, East Greenland: basin fill implications. Global and Planetary Change 12, pages 251-270.

Tanner, V.
1940: The glaciation of Long Range of western Newfoundland: a brief contribution. Geologiska Foreningen i Stockholm Forhandlingar, Volume 62, pages 361-368.

Taylor, D.M.
1994: Late Wisconsinan ice-flow patterns in southwestern Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 94-1, pages 47-51.

Taylor, D.M., St. Croix, L. and Vatcher, S.V.
1994: Newfoundland Striation Database. Newfoundland Department of Mines and Energy, Geological Survey Branch, Open File Nfld 2195.

Taylor, D.M. and Vatcher, S.V.
1993: Late Wisconsinan deglacial ice-flow patterns in west-central Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 93-1, pages 175-179.

Teller, J.T.
1987: Proglacial lakes and the southern margin of the Laurentide Ice Sheet. *In* North America and Adjacent Oceans during the Last Deglaciation. *Edited by* W.F. Ruddiman and H.E. Wright Jr. Geological Society of America, The Geology of North America, volume K-3, pp. 39-69.

1995: History and drainage of large ice-dammed lakes along the Laurentide Ice Sheet. Quaternary International, Volume 28, pages 83-92.

Teller, J.T. and Clayton, L.
1983: Glacial Lake Agassiz. Geological Association of Canada, Special Paper 26.

Teller, J.T. and Kehew, A.E.
1994: Introduction to the late glacial history of large proglacial lakes and meltwater runoff along the Laurentide Ice Sheet. Quaternary Science Reviews, Volume 13, pages 795-799.

Teller, J.T. and Thorleifson, L.H.
1987: Catastrophic flooding into the Great Lakes from Lake Agassiz. *In* Catastrophic Flooding. *Edited by* L. Mayer and D. Hash, Allen and Unwin, Boston, pp. 121-138.

Teller, J.T., Thorleifson, L.H., Dredge, L.A., Hobbs, H.C. and Schreiner, B.T.
1983: Maximum extent and major features of Lake Agassiz. *In* Glacial Lake Agassiz. *Edited by* J.T. Teller and L. Clayton. Geological Association of Canada, Special Paper 26, pages 43-45.

Thomas, G.S.P. and Connell, R.J.
1985: Iceberg drop, dump, and grounding structures from Pleistocene glacio-lacustrine sediments, Scotland. Journal of Sedimentary Petrology, Volume 55, pages 243-249.

Tucker, C.M.
1974a. A series of raised Pleistocene deltas in Halls Bay, Newfoundland. Maritime Sediments, Volume 10, pages 1-7.

1974b. Unusual till ridges in the Kitty's Brook-Chain Lakes area of west-central Newfoundland. Maritime Sediments, Volume 10, pages 8-13.

1976: Quaternary studies in Newfoundland: a short review. Maritime Sediments, Volume 12, pages 61-73.

Twenhofel, W.H.
1912: The physiography of Newfoundland. American Journal of Science, Volume 33, pages 1-25.

Twenhofel, W.H. and MacClintock, P.
1940: Surface of Newfoundland. Bulletin of the Geological Society of America, Volume 51, pages 1665-1728.

Van Loon, A.J.
1983: The stress systems in mudflows during deposition, as revealed by the fabric of some Carboniferous pebbly mudstones in Spain. Geologie en Mijnbouw, Volume 62, pages 493-498.

Vanderveer, D.G.
1975: The surficial geology of the St. John's area, Newfoundland, with special emphasis on the gravel resources. Newfoundland Department of Mines and Energy, Mineral Development Division, Open File 1N(232).

1977: Clay deposits of Newfoundland and Labrador. Mineral Development Division, Newfoundland and Labrador. Department of Mines and Energy, Report 77-9, 41 pages.

1981: Quaternary mapping - upper Humber River area. *In* Current Research. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 81-1, pages 192-194.

1987: Surficial and glacial geology for the Deer Lake-Hampden area, Newfoundland (NTS 12H/3, 6, 10, 11). Newfoundland Department of Mines and Energy, Geological Survey Branch, Open File 12H(945).

Vanderveer, D.G. and Sparkes, B.G.
1982: Regional Quaternary mapping: an aid to mineral exploration in west-central Newfoundland. *In* Prospecting in Areas of Glaciated Terrain-1982. *Edited by* P.H. Davenport. Canadian Institute of Mining and Metallurgy, Geology Division, pages 284-299.

Vanderveer, D.G. and Taylor, D.M.
1987: Quaternary mapping - glacial dispersal studies, Sops Arm area, Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 87-1, pages 31-38.

Vilks, G., MacLean, B., Deonarine, B., Currie, C.G. and Morgan, K.
1989: Late Quaternary paleoceanography and sedimentary environments in Hudson Strait. Géographie physique et Quaternaire, Volume 43, pages 161-178.

Virkkala, K.
1960: On the striations and glacier movements in the Tampere region, southern Finland. Society Géologie de Finlande, Comptes Rendus, Volume 32, pages 159-176.

Wadell, H.
1936: Volume, shape and shape position of rock fragments in openwork gravel. Geografiska Annaler, Volume 18, pages 74-92.

Walcott, R.I.
1970: Isostatic response to loading of the crust in Canada. Canadian Journal of Earth Sciences, Volume 7, pages 716-727.

1972: Past sea levels, eustasy, and deformation of the earth. Quaternary Research, Volume 2, pages 1-14.

Walder, J.S. and Costa, J.E.
1996: Outburst floods from glacier-dammed lakes: The effect of mode of lake drainage on flood magnitude. Earth Surface Processes and Landforms, Volume 21, pages 701-723.

Walker, R.G.
1978: Deep water sandstone facies and ancient submarine fans: models for exploration for stratigraphic traps. Bulletin of the American Association of Petroleum Geologists, Volume 62, pages 932-966.

1984: Facies Models. Geoscience Canada, Reprint Series 1, 317 pages.

Washburn, A.L.
1947: Reconnaissance geology of portions of Victoria Island and adjacent regions, Arctic Canada. Geological Society of America, Memoir 22, 142 pages.

1980: Geocryology. A survey of periglacial processes and environments. John Wiley and sons, New York, 406 pages.

Wells, E.D. and Pollett, F.C.
1983: Peatlands. *In* Biogeography and Ecology of the Island of Newfoundland. *Edited by* G.R. South. W. Junk publishers, The Hague, pages 207-265.

Whalen, J.B. and Currie, K.L.
1988: Geology, Topsails igneous terrane, Newfoundland. Geological Survey of Canada, Map 1680A, scale 1:200 000.

Williams, H. and Cawood, P.A.
1989: Geology, Humber Arm Allochthon, Newfoundland. Geological Survey of Canada, Map 1678A. Scale 1:250 000.

Williamson, D., Taieb, M., Damnati, B., Icole, M. and Thouveny, N.
1993: Equatorial extension of the younger Dryas event: rock magnetic evidence from Lake Magadi (Kenya). Global and Planetary Change, Volume 7, pages 235-242.

Wilson, S.E., Walker, I.R., Mott, R.J. and Smol, J.P.
1993: Climatic and limnological changes associated with the Younger Dryas in Atlantic Canada. Climate Dynamics, Volume 8, pages 177-187.

Winkler, M.G.
1994: Development of the Gull Pond Chain of Lakes and the Herring River Basin, Cape Cod National Seashore. Final Report. Part II. North Atlantic Region. National Park Service. 239 pp.

Wolfe, A.P. and Butler, D.L.
1994: Late-glacial and early Holocene environments at Pine Hill Pond, eastern Newfoundland. Evidence from pollen and diatoms. Boreas, Volume 23, pages 53-65.

Woodcock, N.H.
1977: Specification of fabric shapes using an eigenvalue method. Bulletin of the Geological Society of America, Volume 88, pages 1231-1236.

Woodcock, N.H. and Naylor, M.A.
1983: Randomness testing in three-dimensional orientation data. Journal of Structural Geology, Volume 5, pages 539-548.

Wright, H.E., Jr.
1989: The amphi-Atlantic distribution of the Younger Dryas paleoclimatic oscillation. Quaternary Science Reviews, Volume 8, pages 295-306.

Wynne-Edwards, V.C.
1937: Isolated arctic-alpine floras in eastern North America: a discussion of their glacial and recent history. Transactions of the Royal Society of Canada, series 3, Volume 31, section V, pages 1-26.

1939: Some factors in the isolation of rare alpine plants. Transactions of the Royal Society of Canada, series 3, Volume 33, section V, pages 35-42.

APPENDIX 1: SITE AND SAMPLE DESCRIPTION

The following listing provides data on individual sites examined across the Humber River basin. Data is presented in the following format:

Yr -	The year in which data was collected (1991, 1992, 1993, 1994).
Site -	The site number assigned during field work.
NTS -	The National Topographic System map number.
East -	The easting on the Universal Transverse Mercator Grid, Zone 21.
North -	The northing on the Universal Transverse Mercator Grid, Zone 21.
Elev -	The elevation of the site above sea level. This was either estimated from a topographic map (accuracy ± 5 m), or from an altimeter (accuracy ± 2 m).
Sed -	The sediment type found at a site, designated by the following letters: D (diamicton), S (sand), SC (silt-clay), SG (sand and gravel), SS (sand-silt).
Munsell Colour -	The moist colour of sediment using the Munsell Soil Colour Chart is shown first. The dry colour of sediment using the Munsell Soil Colour is shown beneath.
Sand -	The percentage of sand found within sediment matrix.
Silt -	The percentage of silt found within sediment matrix.
Clay -	The percentage of clay found within sediment matrix.
Mean -	The graphic mean of particle sizes, derived from Folk and Ward (1957), expressed in ø sizes.
S.D. -	The inclusive graphic standard deviation of sediments, as a measure of sorting, derived from Folk and Ward (1957) and expressed in ø sizes.
Sample -	The sample number assigned to a sediment or pebble sample. Pebble data is compiled in Appendix 3.
S_1 -	The principal eigenvalue from clast fabric analysis. It describes the strength of the strongest cluster.
S_3 -	The eigenvalue of the weakest cluster (i.e., perpendicular to the principal eigenvector).
Comments -	Site description, including brief section description, where applicable

Site	NTS	East North	Elev m	Sed	Munsell colour	Sand	Silt	Clay	Mean S.D.	Sample	S_1 S_3	Comments
89-6	12A13	454350 5424620	145	SG								Interbedded sand, gravelly sand, diamicton. Beds dip at 8° towards 120°. Some normal faulting. Ice wedge cast.
91-1	12A13	433870 5422980	79									Striae 258 ± 4.
91-2	12A13	432930 5421720	238									Striae 284 ± 6.
91-3	12A13	433970 5421830	235									Striae 270 ± 3. Striae 214 ± 5.
91-4	12A13	433920 5422100	229	D	2.5Y 7/2 2.5Y 8/2	48.8	31.9	19.3	2.33 4.62	914000	0.57 0.12	Light grey diamicton. Sandy to silty matrix. Massive.
91-5	12A13	434150 5422880	58	SG						914001		Sandy gravel. Matrix coarse sand.
91-6	12A13	434870 5422070	82	D	5Y 6/3 10YR 8/2	67.3	28.4	4.3	1.16 3.66	914081	0.85 0.04	Silty to fine sandy diamicton overlain by sandy gravel. Striae at west end pit 295 ± 5.
91-7	12A13	435840 5421820	61	GS						914002		Interbedded gravelly sand and gravel.
91-8	12A13	435860 5421950	30	S								Mostly f-c sand. Faulted. Planar interbedded. Beds dip at 24° toward 118°.
91-9	12A13	435130 5422520	21	GS								Dawe's Pit. See Chapter 4.
91-10	12A13	434500 5423900	37	GS								3 m pebbly to gravelly sand. Edge of terrace.
91-11	12A13	435050 5424400	1									Striae 243 ± 2.
91-12	12A13	435040 5424600	2	SC								6 m exposure. Silt, clay and fine sand.
91-13	12A13	434860 5424620	7	SC						914005		7 m silt-clay.
91-14	12A13	434510 5424660	7	D	10YR 4/2 10YR 6/2	44.4	37.6	18.0	3.82 3.91	914006	0.64 0.16	7 m diamicton. Matrix fine sand to silt.
91-15	12A13	434250 5424760	2									Striae 244 ± 3.
91-16	12A13	434020 5424790	8	D	2.5Y 6/4 2.5Y 8/2	54.2	39.4	6.4	2.02 4.4	914007		9 m+ diamicton over bedrock. Matrix fine sand and silt.
91-17	12A13	433840 5424930	2									Striae and grooves. 255 ± 3.
91-18	12A13	433560 5425400	8	SC	5YR 5/3					914008		6 m pebbly sand; 5-7 mm calcite cement; Interbedded silt, clay and fine sand. Bedrock striated 246 ± 6.
91-19	12A13	433570 5425680	15	SC								8 m of silt-clay overlying 12 m sandy gravel, separated by calcite cement.
91-20	12A13	440270 5423700	12	GS	2.5YR 4/2					914009		15 cm gravelly sand. Overlies 1.2 m+ clay and silt.
91-21	12A13	439500 5424120	76	PS	2.5Y 4/4 2.5Y 7/2	73.3	23.9	2.9	1.9 3.34	914010		Pebbly sand.
91-22	12A13	437700 5425070	37	SG								Sand and gravel.
91-23	12A13	437300 5424960	37	D	10YR 5/4 2.5Y 7/4	65.3	27.2	7.4	1.4 3.92	914011		1 m. Sandy diamicton.
91-24	12A13	437020 5424780	18	SG								Sandy gravel overlain by 1 m of gravelly sand.
91-25	12A13	435770 5424650	18	SG	5YR 5/2 10YR 7/3	67.0	25.7	7.3	0.84 5.29	914012		Wild Cove section. See Chapter 4.
91-25	12A13	435770 5424650	18	SG	7.5YR 6/4 10YR 7/3	71.6	24.1	4.4	2.8 3.82	914013		
91-26	12A13	435660 5424590	30	G								Exposure 12 m high. Silty gravel.
91-27	12A13	435590 5426480	156	D	10YR 6/4 10YR 8/3	37.3	46.9	15.8	1.14 4.91	914015	0.83 0.06	Diamicton.
91-28	12A13	435680 5426410	128									Striae 225 ± 4; 280 ± 3.
91-29	12H04	435920 5427950	49	SG								2 to 5 m planar bedded sand and gravel.
91-30	12A13	435030 5427260	43	SG								Interbedded coarse to medium sand and sandy pebble gravel.
91-31	12H04	435250 5428020	49	D	10YR 4/3 7.5YR 7/4	49.0	37.3	13.7	2.93 4.28	914138		Diamicton, grades laterally to sand-gravel.
91-32	12A13	431600 5426330	79	D	10YR 4/3 2.5Y 7/4	69.2	26.1	4.7	1.06 3.63	914016		Diamicton. Overlies interbedded sand and gravel.
91-33	12A13	431100 5425680	2	D	10YR 5/1 10YR 7/1	60.7	33.2	6.1	2.74 3.25	914017	0.58 0.08	30 m terrace. Diamicton. Silt and fine sand. Overlain by sand and gravel.
91-34	12A13	429530 5425030	55									Striae 244 ± 4.
91-35	12A13	429350 5427230	204									Striae 260 ± 2.
91-36	12A13	429020 5426970	152	D	2.5Y 4/2 10YR 7/2	76.2	12.1	11.7	-0.19 3.89	914018	0.79 0.02	1.8 m thick diamicton. Overlain by 1.2 m brown diamicton.
91-36	12A13	429020 5426970	152	D	7.5YR 4/2 10YR 7/2	65.8	23.2	11.1	0.4 4.12	914019	0.79 0.02	
91-37	12A13	429050 5427710	146									Striae 254 ± 10.
91-38	12A13	433050 5420850	155									Striae 274 ± 3.
91-39	12A13	433450 5420600	149	GS	2.5Y 6/2 2.5Y 7/2	85.3	13.7	1.0	0.31 2.79	914020		6 m gravelly sand.

91-40	12A13	433450 5420600	137	D	2.5Y 6/4 10YR 7/1	63.6	28.9	7.5	0.97 4.14	914021	0.52 0.06	Upper 2 of 7 m visible. Diamicton.
91-41	12A13	433400 5420400	177									Striae 292 ± 4. 6 m exposure gravelly sand overlies diamicton.
91-42	12A13	434150 5417900	207									Striae 318 ± 6.
91-43	12A13	427350 5421050	198									Striae 310 ± 4.
91-44	12A13	427650 5419375	219	D	5Y 6/1 5Y 7/1	51.3	20.5	28.2	2.93 4.49	914022	0.92 0.03	2 m diamicton.
91-45	12A13	428400 5419775	223									Striae 296 ± 2.
91-46	12A13	433250 5419200	226	GS	2.5Y 6/4 2.5Y 8/2	73.6	21.7	4.7	-0.86 3.52	914023		Gravelly sand.
91-47	12A13	433050 5416175	284									Striae 320 ± 6.
91-48	12A13	432950 5416050	287	D	5Y 6/2 5Y 7/1	48.4	36.4	15.3	2.04 4.47	914024	0.64 0.08	5 m diamicton.
91-49	12A13	439070 5405600	387									Striae 316 ± 3.
91-50	12A13	440940 5404370	351	D	5Y 6/2 5Y 7/2	63.5	29.9	6.6	0.65 3.86	914025	0.71 0.08	3 m diamicton.
91-51	12A13	438190 5406670	363	D	5Y 6/3 2.5Y 7/2	59.9	24.7	15.4	2.73 3.99	914026		1.5 m diamicton.
91-52	12A13	437350 5409650	341	D	5Y 6/3 5Y 7/2	51.2	36.0	12.8	2.6 4.29	914027		2 m diamicton.
91-53	12A13	438370 5411380	299	D	2.5Y 4/2 2.5Y 7/2	77.6	19.0	3.4	-0.61 3.41	914028	0.73 0.06	6 m diamicton.
91-54	12A13	437960 5421560	45	D	5YR 4/3 5YR 5/3	24.9	18.6	56.4	5.9 4.4	914029	0.77 0.09	1 m+ of gravelly sand;1 m diamicton, interbedded with silt-clay with rare pebbles; and 1.5 m sand.
91-55	12A13	437860 5421570	31	S								3 m well sorted, rippled sand.
91-56	12A13	438030 5421840	37									Striae 208-028.
91-57	12A13	432730 5416170	308									Striae 324 ± 2.
91-58	12A13	432450 5417150	287									Striae 320 ± 4.
91-59	12A13	432670 5418190	311									Striae 322 ± 3.
91-60	12A13	438860 5415830	421	D	2.5Y 5/4 2.5Y 7/2	80.1	19.0	0.9	0.65 3.09	914030	0.85 0.05	7 m diamicton.
91-61	12A13	438380 5414180	445	D								Silt - fine sand diamicton.
91-62	12A13	438710 5411800	360	D	2.5Y 6/4 2.5Y 7/2	83.3	14.2	2.5	0.75 2.99	914031	0.51 0.12	Hummock, 50 m diameter. 7m diamicton.
91-63	12A13	437860 5413000	366	D	2.5Y 5/4 2.5Y 7/2	84.4	14.5	1.1	0.43 2.98	914032		4 m diamicton.
91-64	12A13	436850 5415710	323	D	5Y 6/3 5Y 7/3	57.6	36.0	6.4	3.15 3.26	914033		1.5-2 m diamicton.
91-65	12A13	435190 5418390	274									Striae 297 ± 3.
91-66	12A13	435570 5418470	232	D	2.5Y 5/4 2.5Y 7/4	83.9	14.2	1.9	0.17 3.02	914034	0.57 0.12	4 m diamicton.
91-67	12A13	432800 5415700	280									Striae 310 ± 3.
91-68	12A13	428570 5411620	302									Striae 300 ± 2.
91-69	12H04	455170 5428550	29	S	5YR 4/3 5YR 5/3	85.9	10.2	3.9	2.33 1.85	914035		0-2 m bedrock; 6 m m-c sand; 1 m silt-clay and interbedded fine sand; 1.5 m sandy gravel.
91-70	12H04	455320 5428430	29	SC	5YR 4/3					914037		2 m rhythmically bedded silt-clay. Reddish brown.
91-71	12H04	454370 5429070	31	SG								Grooves 204 ± 4. 4 m sandy pebbly gravel.
91-72	12H04	453340 5428960	24	SG								Interbedded sandy gravel and granules.
91-73	12H04	450530 5427550	61	D	10YR 4/3 10YR 6/4	91.8	7.0	1.2	-1.37 2.7	914038		2 m diamicton.
91-74	12H04	449150 5427520	21	SG								30 cm sandy gravel; 50-80 cm interbedded c, m, f sand, 5-50 cm pebbly diamicton, 5 cm well sorted c sand and >15 cm gravelly sand.
91-75	12A13	433610 5416140	287									Striae 310 ± 2.
91-76	12A13	436020 5412940	314									Grooves 310 ± 3. Grooves 245 ± 3.
91-77	12A13	435300 5410880	375									Striae 308 ± 3.
91-78	12A13	434340 5404350	256	D	5Y 5/3 5Y 6/4	70.6	21.4	8.1	0.62 3.71	914039		1 m diamicton.
91-79	12A13	434510 5413170	317									Grooves 310 ± 2.
91-80	12A13	458500 5421620	311	D	5Y 5/3 2.5Y 7/2	77.1	15.3	7.6	0.87 3.09	914040	0.73 0.12	Diamicton.
91-81	12A13	458870 5420460	256	D	2.5Y 6/4 10YR 3/2	62.5	28.1	9.4	1.57 4	914041		3 m diamicton.
91-82	12A13	455950 5418410	293	D	10YR 3/3 2.5Y 7/4	78.3	18.3	3.4	0.62 2.94	914042	0.79 0.03	2 m very dark greyish brown diamicton. Sharp contact over >1 m dark greyish brown diamicton.

91-82	12A13	455950 5418410	293	D	5Y 5/3 2.5Y 7/2	49.6	32.2	18.2	3.42 4.02	914043		
91-83	12H04	460730 5427920	274									Grooves 308 ± 3.
91-84	12H04	459100 5427480	259	D	10YR 4/3 2.5Y 7/2	78.0	13.1	9.0	-1.69 3.55	914044		75 cm diamicton.
91-85	12A13	457180 5424900	170	D	5YR 4/6 7.5YR 6/6	87.6	11.3	1.1	-0.86 2.9	914045		Diamicton.
91-86	12A13	456220 5423370	154	D	2.5Y 5/2 2.5Y 7/2	51.5	31.1	17.4	1.27 5.46	914046		75 cm diamicton.
91-87	12A13	453430 5416050	126	D	5Y 5/3 5Y 7/2	82.0	13.1	4.9	-0.12 3.08	914047		12 m exp., 4-5 m diamicton.; 6-8 m interbedded gravelly sand, sandy gravel, and diamicton.
91-87	12A13	453430 5416050	126	SG	5Y 4/2 5Y 7/3	98.6	1.4	0.0		914048		
91-88	12A13	453850 5417260	180	SG	5Y 4/2 5Y 6/3	95.5	4.5	0.0		914049		Sandy gravel.
91-89	12A13	454820 5417980	200	D	5Y 4/2 5Y 5/3	64.8	24.7	10.4	-0.09 4.18	914050	0.89 0.02	3 m diamicton.
91-90	12A13	454020 5418140	150	D	5Y 5/3 2.5Y 6/4	77.3	19.2	3.5	0.7 3.23	914051		8 m diamicton.
91-91	12A13	454320 5418930	120	D	5Y 4/2 5Y 5/3	57.5	28.1	14.5	1.21 4.85	914052		2 m exposure. >0.5 m olive grey diamicton. Overlain by 0.8-1.0 m diamicton. Reddish brown.
91-91	12A13	454320 5418930	120	D	5YR 4/3 7.5YR 6/4	35.1	33.0	32.0	4.54 4.26	914053		
91-92	12A13	454350 5419200	150	SG								3 m sand-gravel. Grades laterally into sandy diamicton.
91-93	12A13	454460 5419520	155	SG								5 m sandy gravel to gravelly sand.
91-94	12A13	455640 5426570	65	D	10YR 4/2 7.5YR 7/4	66.4	24.6	9.0	1.47 4.24	914054	0.79 0.05	Dark brownish grey diamicton; olive grey diamicton; dark reddish brown diamicton. Striae and grooves 205 ± 5
91-94	12A13	455640 5426570	65	D	5Y 4/2 2.5Y 6/2	57.0	31.3	11.7	1.98 4.08	914055	0.73 0.12	
91-94	12A13	455640 5426570	65	D	5YR 4/2 7.5YR 6/4	58.6	30.9	10.5	2.27 4.05	914056	0.74 0.1	
91-94	12A13	455640 5426570	65	D						914054a	0.92 0.01	
91-94	12A13	455640 5426570	65	D						914055a	0.72 0.13	
91-94	12A13	455640 5426570	65	D						914056a	0.75 0.09	
91-94	12A13	455640 5426570	65	D						914056b	0.69 0.04	
91-95	12H04	460680 5436930	91	D	10YR 5/4 10YR 3/3	65.3	29.7	5.0	-0.31 4.45	914057		Diamicton.
91-96	12H04	460650 5436920	91	SS	10YR 4/2 7.5YR 6/4	16.6	75.3	8.1	4.31 3.53	914058		Fine sand and silt with some pebbles and cobbles.
91-97	12A13	454600 5419930	145	D						914059		2-3 m diamicton.
91-98	12A13	454930 5420950	135	D	5Y 4/2 2.5Y 6/4	70.5	25.4	4.0	0.41 3.85	914060	0.82 0.04	2 m diamicton.
91-99	12A13	454620 5422830	140	GS								4 m pebbly sand. See site 91235.
91-100	12A13	454390 5422480	220	D	10YR 4/2 10YR 6/4	69.7	23.8	6.5	0.41 3.85	914061		1 m diamicton.
91-101	12A13	455060 5424650	85	S								2.5 m sand, with layers of manganese stain and cemented sand and silty clay.
91-102	12A13	455200 5425330	86	S	7.5YR 5/4 7.5YR 6/4	65.8	32.7	1.4	3.98 1.15	914062		2 m fine sand.
91-103	12H04	460410 5437270	33	GS								7 m interbedded gravelly sand and sandy gravel.
91-104	12A13	455480 5426320	84	D	10YR 5/3 10YR 7/3	57.5	24.6	18.0	3.78 5.01	914063	0.67 0.14	6 m exposure. Diamicton. Overlain by interbedded v. f - m sands and diamicton.
91-104	12A13	455480 5426320	84	D						914063a	0.9 0.04	
91-105	12A13	455420 5426180	88	D	2.5Y 5/2 10YR 7/3	65.3	28.1	6.6	0.52 4.21	914064	0.76 0.07	7 m exposure. Greyish brown diamicton; 45 cm sand, manganese stained and gravelly sand; 1.2 m diamicton, and 0-3m sand- gravel.
91-105	12A13	455420 5426180	88	D	5YR 4/3 7.5YR 6/4	57.0	32.5	10.4	3.12 3.82	914065	0.78 0.07	
91-106	12A13	451690 5416330	570									Striae 315 ± 2.
91-107	12A13	449540 5416440	529									Striae 310 ± 2.
91-108	12A13	448940 5416490	558									Striae 315 ± 2.
91-109	12A13	446970 5417150	492									Striae 318 ± 2.
91-110	12A13	446270 5417320	480									Striae 310 ± 3.
91-111	12A13	450920 5416550	505	D						914066		Diamicton. Thin.
91-112	12A13	454770 5415630	420									Striae 268 ± 2, 290 ± 2, 310 ± 2.
91-113	12A13	455230 5415530	335	D						914067		Erratics collected from hilltop.
91-114	12A13	454930 5415580	354									Striae 250 ± 5.

91-115	12A13	440870 5421330	205	SG								Interbedded c sand/granule, f-m sand, and sandy gravel.
91-116	12A13	444320 5420420	380									Striae 282 ± 3.
91-117	12A13	444270 5421470	380	SG	10YR 4/3 2.5Y 7/4	86.4	11.9	1.8	-1.71 2.66	914068		6 m high terrace. Interbedded granule/pebble gravel and sandy gravel.
91-118	12A13	449680 5421660	380									Striae 262 ± 3.
91-119	12A13	450730 5422480	397									Striae 272 ± 4.
91-120	12A13	449880 5421670	378	D	10YR 4/3 10YR 7/6	86.3	12.0	1.8	-1.84 2.87	914069		Thin diamicton.
91-121	12A13	452650 5423230	350	D	10YR 4/3 10YR 6/3	60.4	24.9	14.7	1.12 5	914070		Thin diamicton. Striae 280 ± 3.
91-122	12A13	454030 5423150	247	D	2.5Y 7/2 10YR 4/2	55.2	34.3	10.6	1.43 4.61	914071		1 m diamicton.
91-123	12A13	454740 5423120	135	GS								1 m interbedded granule gravel, gravelly sand, sand, and diamicton
91-124	12A13	439750 5418560	530									Striae 272 ± 3 to 282 ± 3.
91-125	12A13	439400 5418510	460	D	10YR 4/4 10YR 6/4	69.0	25.1	5.9	-1.3 4.3	914072		75 cm diamicton.
91-126	12A13	437570 5418320	435	D						914073		Pebbles collected from road.
91-127	12H04	454260 5432650	31	D	5YR 4/4 5YR 5/4	56.5	38.1	5.5	3.78 2.71	914074		4 m interbedded sandy gravel, diamicton and sand.
91-128	12H04	454220 5432590	26									Striae 322 ± 6.
91-129	12H04	452870 5431890	107									Grooves 244 ± 4.
91-130	12H04	452660 5431760	110									Striae 244 ± 10. Striae 282 ± 5.
91-131	12H04	452330 5431470	91									Striae 224 ± 3. Striae 335 ± 7.
91-132	12H04	451900 5431100	46									Grooves 270 ± 4.
91-133	12H04	451420 5430680	43	GS	5YR 4/3 5YR 6/6	95.7	3.3	1.0	-0.21 2.21	914075		5 m interbedded sand and pebbly sand.
91-134	12H04	451770 5431030	43	SG	5YR 3/4 7.5YR 5/6	90.7	8.6	0.7	1.22 2.51	914076		Partially cemented sandy gravel exposed in stream bed.
91-135	12H04	453750 5432270	43	D	5YR 3/4 5YR 5/6	76.0	18.9	5.2	0.62 3.98	914077		0.5 m diamicton overlying rotted bedrock.
91-136	12B09	417500 5391350	120	S								12 m well sorted medium sand.
91-137	12B09	413760 5492180	146	S								12 m gravelly sand and boulder gravel.
91-138	12A13	437950 5421850	13	SG								Humber gorge section. See Chapter 4.
91-139	12A13	436550 5421540	15	S								2-3 m very fine sand; 50-75 cm bed open work gravel; 1 m planar sandy gravel, and pebble gravel.
91-140	12A13	436750 5402300	488									Grooves 308 ± 3.
91-141	12A13	435930 5401130	466	D	10YR 5/8 10YR 6/4	76.1	20.5	3.4	0.84 3.2	914082	0.77 0.03	6 m diamicton.
91-142	12A13	437880 5405100	357	D	5Y5/3 5Y 7/2	38.7	55.9	5.4	-0.41 4.38	914083	0.67 0.05	4 m diamicton.
91-143	12H04	461850 5446850	53	SG	5YR 3/3 5YR 6/4	91.7	5.8	2.5	-1.7 2.49	914084		Sandy gravel.
91-144	12H04	461800 5445200	47	S	5YR 4/3 5YR 6/4	99.4	0.6	0.0		914085		5 m exposure. 4 m interbedded medium to coarse sand/granules. Overlain by 1 m sandy gravel.
91-145	12H04	460650 5443500	9	SC	5YR 4/3 5YR 7/4	11.8	27.1	61.1	7.93 2.53	914086		12 m exposure. 1 m pebble gravel; 2 m planar f and v fine sand; 3 m interbedded silt, clay and sand; and 1.5 m pebbly sand.
91-145	12H04	460650 5443500	9	S	5YR 5/4 5YR 6/4	99.9	0.1	0.0		914087		
91-146	12H04	456850 5443550	274	D						914088		Pebble sample from road surface.
91-147	12H04	455600 5441850	183	D	10YR 4/4 10YR 6/4	74.4	17.5	8.1	-0.08 3.67	914089		2 m diamicton.
91-148	12H04	458350 5445900	137									Striae 250 ± 3.
91-149	12H04	458300 5446100	128	D	5YR 4/4 7.5YR 6/4	56.3	37.0	6.8	1.77 4.43	914090		75 cm diamicton.
91-150	12H04	454650 5441500	183	D	5YR 3/4 10YR 6/4	81.5	14.1	4.4	1.15 3.05	914091		2 m exposure. 50 cm+ sand. Overlain by 1.2 m diamicton.
91-150	12H04	454650 5441500	183	S	7.5YR 4/4 7.5YR 7/4	78.6	19.6	1.7	3.06 1.68	914092		
91-151	12H04	454800 5441550	180									Striae 258 ± 6.
91-152	12H04	455950 5442300	177	GS	10YR 4/4 10YR 7/3	78.8	19.4	1.8	0.8 3.29	914093		6 m exposure, basal 3.5 m obscured. >50 cm gravelly sand. Overlain by 2 m sandy gravel.
91-152	12H04	455950 5442300	177	SG	7.5YR 4/4 7.5YR 6/6	77.7	19.5	2.9	-1.17 3.45	914094		
91-153	12H04	456650 5441350	123	D	5YR 4/4 7.5YR 6/4	86.3	9.1	4.6	-0.21 3	914095		2m sandy diamicton.
91-154	12H04	456600 5440100	27									Grooves 252 ± 2.

91-155	12H04	455850 5439850	76	D	5YR 4/4 7.5YR 6/4	59.5	31.6	8.9	2.01 4.05	914096		75 cm diamicton.
91-156	12H04	457250 5448540	165									Striae 270 ± 2.
91-157	12H04	456900 5448790	189									Striae 292 ± 3.
91-158	12H04	456560 5448950	223	D						914097		Pebble sample.
91-159	12H04	463510 5451050	168									Striae 242 ± 4.
91-160	12H04	460610 5448970	195									Grooves 224 ± 2.
91-161	12H04	459870 5448970	213									Striae 240 ± 3.
91-162	12H04	463170 5450610	143	D	5YR 4/6 5YR 6/6	66.3	27.7	6.0	2.49 3.36	914098		2 m diamicton.
91-163	12H04	462720 5450250	119	D	2.5YR 4/4 2.5YR 6/6	48.3	33.8	17.9	3.67 3.91	914099	0.74 0.07	Same as site 89059. 7 m exposure. 2 m + reddish grey diamicton; 3 m reddish brown diamicton.
91-163	12H04	462720 5450250	119	D	5YR 5/3 7.5YR 6/4	49.3	35.7	15.0	3.82 3.58	914100	0.69 0.07	
91-163	12H04	462720 5450250	119	D						914099a	0.67 0.09	
91-163	12H04	462720 5450250	119	D						914100a	0.61 0.1	
91-164	12H04	455110 5452840	174	D	7.5YR 5/4 7.5YR 7/4	59.2	28.6	12.2	2.26 4.21	914101		1.5 m diamicton.
91-165	12H04	459480 5449980	213	D	7.5YR 4/4 10YR 7/4	65.7	28.1	6.3	2.71 3.27	914102		2 m diamicton.
91-166	12H04	461750 5449410	152	D	7.5YR 5/4 7.5YR 8/2	68.5	27.3	4.2	2.34 3.45	914103		2 m diamicton.
91-167	12H04	440450 5428870	213	D	10YR 5/3 10YR 6/4	48.1	42.3	9.6	3 3.93	914104		2 m diamicton.
91-168	12H04	438670 5428620	152	D								1.5 m diamicton.
91-169	12A13	437400 5428380	98	D	5Y 5/3 2.5Y 7/2	69.6	22.2	8.2	0.85 4.15	914105		3-6 m diamicton.
91-170	12H04	436640 5428980	46	SG								12 m exposure, mostly slumped. Interbedded sand, gravelly sand and sandy gravel.
91-171	12H04	437180 5430190	37	SG								4 m terrace. Interbedded sands and sandy pebble gravel.
91-172	12H04	438380 5432870	43	S								4 m exposure interbedded sand and pebble sand.
91-173	12H04	437650 5430880	46	S								Hughes Brook section. See Chapter 4.
91-174	12H04	436910 5430790	46	S								12 m exp. Slumped. Interbeds of cross bedded f sand, sandy gravel and sandy granule gravel.
91-175	12H04	437450 5431520	64									Striae 236 ± 3.
91-176	12H04	437570 5433870	152	D	7.5YR 6/4 5YR 4/3	51.1	40.9	8.0	3.24 3.52	914106	0.82 0.04	2 m diamicton.
91-177	12H04	434560 5437350	183	D	10YR 6/3 10YR 4/2	61.1	25.2	13.7	2.47 4.26	914107	0.59 0.12	6 m exposure. 2 m diamicton. Overlain by 4 m diamicton.
91-177	12H04	434560 5437350	183	D	10YR 7/3 10YR 4/3	75.8	15.1	9.1	0.23 3.8	914108	0.64 0.08	
91-178	12H04	434270 5437680	122	SG								4 m interbedded sand, and sandy pebble gravel. Beds dip 20° to 280°.
91-179	12H04	434110 5439350	137	D	5Y 7/3 5Y 5/3	61.5	24.7	13.9	0.68 4.73	914109		4 m diamicton.
91-180	12H04	437200 5434820	183	D	5Y 7/3 5Y 5/3	37.8	41.3	20.9	3.91 4.25	914110		2 m diamicton.
91-181	12H04	436350 5421650	15	SC	5YR 6/3 2.5YR 3/4	5.3	37.1	57.7	8.85 3.39	914111		15 m exposures. Planar bedded sand and sandy gravels; 1 m planar f sand; and sandy gravel.
91-182	12A13	437970 5421850	15	S								1 m gravelly sand overlain by 1 m rippled sand.
91-183	12A13	438280 5422040	15	SC	5YR 4/3					914112		Reddish brown silty clay. Overlain by interbedded f and v fine sands, and sand and gravel.
91-184	12A13	440980 5423430	40	D	2.5Y 7/2 5Y 4/2	78.5	18.8	2.8	-0.64 3.55	914113		1.5 m diamicton.
91-185	12A13	443400 5425270	20	SC	5YR 6/3 5YR 4/3	1.4	36.0	62.6	8.18 1.76	914114		2 m reddish brown silty clay interbedded with f sand.
91-186	12A13	445900 5425770	37	SG								2 m fine sand overlain by 50 cm sandy gravel.
91-187	12A13	448120 5426920	45	D	2.5Y 7/2 5Y 5/3	61.7	34.2	4.2	0.95 3.96	914115	0.6 0.06	3 m interbedded f and m sands; thin sandy gravel. In places 2 m diamicton, overlain by 2-3 m inter-bedded sand and sandy pebble gravel.
91-188	12A13	446650 5426640	15	SC	5YR 6/3 5YR 4/3	32.1	50.4	17.5	5.64 2.2	914116		Silt-clay, reddish brown. Overlain by 15 m gravelly sand.
91-189	12A13	447440 5426400	30	SG								3 m interbedded f and m sands, and sandy gravel.
91-190	12H04	448380 5427500	46.2	SG								Sandy gravel.
91-191	12H04	462350 5430650	330	D	10YR 6/4 10YR 4/3	73.0	18.6	8.4	-1.25 4.06	914117		1 m diamicton.
91-192	12H04	457760 5432170	13	SC	5YR 6/3 5YR 4/3	1.6	34.5	64.0	8.27 1.89	914118		3 m exposure. 1 m reddish brown silty clay. Overlain by 2 m sandy gravel.

91-192	12H04	457760 5432170	13	SL	5YR 6/3 5YR 4/3	7.9	87.9	4.2	5.46 1.28	924051		
91-193	12H04	462900 5437820	320	D	10YR 6/3 7.5YR 4/4	74.3	23.2	2.5	0.71 3.79	914119	0.79 0.02	6 m diamicton.
91-194	12H04	461390 5436320	152	SS	10YR 6/3 7.5YR 4/4	61.8	20.9	17.3	-1.91 3.54	914120		6 m sandy gravel. Same as Site 91240
91-195	12H04	461950 5436630	146	SG								2 m interbedded fine, medium sands, granule gravel and sandy gravel.
91-196	12H04	463330 5436230	195	D	7.5YR 6/4 5YR 4/3	60.0	32.4	7.6	2.44 3.95	914121		2 m diamicton.
91-197	12H04	462970 5430870	287	D	10YR 6/3 10YR 4/3	79.0	18.4	2.6	0.05 3.47	914122		3-4 m exposure. 2 m diamicton. Overlain by 2-3 m sandy gravel.
91-197	12H04	462970 5430870	287	SG	10YR 6/4 10YR 3/3	91.1	8.4	0.5	-0.59 3.15	914123		
91-198	12H04	462800 5434280	229	D	7.5YR 6/4 5YR 4/3	62.6	23.2	14.2	2.51 4.29	914124		2 m diamicton.
91-199	12H04	457180 5430750	52	GS								1 m pebbly sand overlying 7 m of soft sandstone bedrock.
91-200	12H04	461890 5439980	15	GS								1 m gravel.
91-201	12H04	461740 5439560	31	GS								3 m terrace. Granule gravel to pebbles.
91-202	12A13	441870 5425150	9	S								8 m interbedded m-c sand cross beds.
91-203	12A13	442840 5425530	9	SC	5YR 6/3 5YR 4/3	1.8	73.6	24.7	6.7 1.62	914125		10 m exposure. 7 m slumped. Shows silty clay; 50 cm pebbly sand, and 50 cm f sand.
91-204	12H04	440320 5437900	300	D	5Y 6/3 5Y 4/2	67.7	21.8	10.5	0.56 4.37	914126		2 m diamicton.
91-205	12H04	440070 5440240	105	D	2.5Y 6/4 10YR 4/3	66.2	28.1	5.7	1.52 3.65	914127		2 m diamicton.
91-206	12H04	440910 5439490	98	SG								4 m sandy gravel.
91-207	12H04	444450 5439230	100	D	2.5Y 6/4 2.5Y4/2	66.8	23.3	10.0	0.83 4.16	914128		2 m diamicton.
91-208	12H04	447100 5439170	100	D	2.5Y 6/2 2.5Y4/2	62.8	22.8	14.4	0.63 4.81	914129		Diamicton.
91-209	12H04	447600 5439150	93	SG								2 m sandy gravel.
91-210	12H04	449140 5439950	160	D	10YR 6/3 10YR 3/3	68.4	14.7	16.9	0.4 4.36	914130	0.56 0.09	Two diamictons.
91-210	12H04	449140 5439950	160	D	7.5YR 6/2 5YR 4/3	47.6	41.6	10.8	2.09 5.02	914131	0.67 0.05	
91-211	12H04	453800 5436970	143	D	7.5YR 6/4 5YR 4/3	64.5	33.0	2.5	1.15 4.32	914132		1.5 m diamicton.
91-212	12H04	450220 5437170	210	D	10YR 7/4 7.5YR 4/4	81.4	17.2	1.4	1.41 3.01	914133		2 m diamicton.
91-213	12A13	432250 5422870	110									Striae 274 ± 4.
91-214	12A13	431450 5422540	76									Grooves 256 ± 3.
91-215	12A13	431030 5422970	15	GS	10YR 6/4 10YR 5/3	78.5	14.9	6.6	2.25 3.13	914134		15 m exposure. 3 m+ pebbly sand, 20 cm gravelly sand; 300 cm diamicton; 30 cm interbedded sand, silt and diamicton; 30 cm sandy gravel; 150 cm sand; 350 cm sandy gravel.
91-215	12A13	431030 5422970	15	D	2.5YR 6/0 2.5YR 4/0	52.1	37.2	10.7	2.39 4.08	914135	0.68 0.1	
91-215	12A13	431030 5422970	15	SG	10YR 7/3 10YR 4/3	95.4	3.1	1.5	-0.6 2.98	914136		
91-216	12A13	433430 5423340	46									Striae 285 ± 15.
91-217	12A13	429400 5415750	339									Striae 310 ± 2.
91-218	12A13	427480 5416020	305	D	5Y 6/3 5Y 5/3	57.9	27.1	15.0	1.86 4.4	914137		2 m diamicton.
91-219	12A13	436400 5424600	50.2	D	10YR 7/3 7.5YR 5/4	66.8	27.9	5.4	0.89 4.77	914139		0.2 m+ diamicton. Overlain by 3.4 m of interbedded sand and gravel.
91-220	12A13	437350 5424900	40.7	D	10YR 6/4 10YR 4/4	65.0	29.9	5.1	0.79 4.16	914140	0.76 0.05	Wild Cove section. See Chapter 4.
91-220	12A13	437350 5424900	40.7	S	10YR 6/3 10YR 4/3	83.5	11.2	5.3	2.8 1.89	914141		
91-220	12A13	437350 5424900	40.7	SC	10YR 6/3 7.5YR 5/4	13.1	74.6	12.3	5.45 2.77	914142		
91-221	12A13	437800 5425060	53.2	SG	10YR 5/4 7.5YR 3/2	91.1	7.9	1.0	-1.8 1.87	914143		1.5 m+ interbedded gravel and sandy gravel. Overlain by 1.5 m sandy gravel.
91-222	12A13	438980 5424300	62.2	D	10YR 6/3 10YR 4/3	75.2	20.5	4.2	0.85 3.5	914144	0.65 0.07	3.5 m+ diamicton.
91-223	12A13	439430 5424130	71.2	D	10YR 6/4 10YR 4/3	73.9	21.1	5.1	2.13 3.43	914145	0.61 0.05	3.2 m+ diamicton.
91-224	12A13	455450 5426300	103	D	7.5YR 6/6 7.5YR 4/4	67.7	21.7	10.6	2.01 4.04	914146	0.77 0.07	See also 91104. Diamicton.
91-225	12A13	455140 5426280	125	D	7.5YR 6/4 5YR 4/3	60.1	21.9	18.0	2.57 4.39	914147	0.73 0.05	2 m+ diamicton.
91-226	12A13	455470 5424950	79.7	SC	7.5YR 6/4 5YR 4/4	18.8	48.1	33.1	6.26 3.13	914148		Interbedded sands, silts and pebbly sands. Some silt-clay beds.
91-226	12A13	455470 5424950	79.7	SG	10YR 6/4 10YR 4/4	91.6	6.6	1.8	1.77 1.55	914149		

91-227	12A13	456340 5424850	112	D	7.5YR 6/4 5YR 4/4	54.5	32.9	12.6	3.61 3.6	914150	0.89 0.04	50 cm+ sand. Overlain by 2.5 m diamicton.
91-227	12A13	456340 5424850	112	S	7.5YR 5/4 5YR 4/4	84.8	11.3	3.8	1.38 3.09	914151		
91-228	12A13	457030 5425170	152	GS	10YR 6/3 7.5YR 4/4	88.9	9.2	1.9	0.43 2.76	914152		1 m gravelly sand. Overlain by 3 cm silt and 2 m sandy gravel.
91-228	12A13	457030 5425170	152	SG	7.5YR 6/4 5YR 4/4	76.9	17.2	5.8	-0.61 3.89	914153		
91-229	12A13	456450 5424350	136	SG	7.5YR 5/4 7.5YR 4/4	95.0	3.8	1.2	-2.38 3.01	914154		3 m interbedded sands and sandy gravels. Beds dip 30° towards 310°.
91-230	12A13	455400 5424100	88.7	SC	7.5YR 6/4 7.5YR 4/4	5.8	57.1	37.1	7.19 1.78	914155		Generally over 2 m+ silt-clay, brown. Overlain by 80 cm gravelly sand.
91-231	12A13	456050 5423370	129	GS	7.5YR 6/4 5YR 5/4	87.8	10.0	2.2	0.98 2.51	914156		1.5 m+ pebbly sand. Overlain by 50 cm silty diamicton, and 1 m diamicton.
91-231	12A13	456050 5423370	129	D	7.5YR 5/4 5YR 4/4	63.0	26.7	10.3	0.98 4.28	914157	0.84 0.05	
91-231	12A13	456050 5423370	129	D	5YR 6/4 5YR 4/4	36.0	47.6	16.4	5.43 2.53	914158		
91-232	12A13	455700 5422650	150	GS	7.5YR 5/4 7.5YR 4/6	97.6	1.9	0.5	-0.99 2.26	914159		Interbedded sand and gravel and diamicton. Gravel beds dip 25° to 130°.
91-233	12A13	455530 5421790	133	D	2.5Y 7/2 10YR 5/3	50.9	31.7	17.4	2.43 4.49	914160	0.8 0.04	1.5 m diamicton. Brown. Overlain by 1.5 m sandy diamicton.
91-233	12A13	455530 5421790	133	D	2.5Y 6/2 10YR 4/2	79.9	17.3	2.8	-0.23 3.28	914161		
91-234	12A13	455330 5420630	135	SG	10YR 4/2 10YR 3/3	93.1	5.7	1.3	-0.81 2.48	914162	0.62 0.06	3 m sandy gravel.
91-235	12A13	454620 5422830	140	SG	10YR 5/4 10YR 3/3	98.1	1.7	0.2	-1.36 1.94	914163		See site 91099. 3 m interbedded gravelly sand, sandy gravel and gravel.
91-236	12A13	454420 5417450	154	SG	5Y 7/2 5Y 5/2	87.0	10.6	2.4	-1.57 2.49	914164		50 cm+ sandy gravel; 1.2 m olive diamicton; 1.2 m brown diamicton.
91-236	12A13	454420 5417450	154	D	5Y 7/2 5Y 5/3	59.3	26.3	14.4	0.89 4.42	914165	0.67 0.05	
91-236	12A13	454420 5417450	154	D	2.5Y 6/4 5Y 4/2	48.7	32.5	18.8	1.93 5.34	914166	0.69 0.04	
91-237	12A13	454350 5417330	152	GS	2.5Y 7/2 5Y 4/2	80.0	16.5	3.5	0.32 3.01	914167		30 cm+ gravelly sand. Overlain by 1 m sandy gravel, and 2 m sandy gravel.
91-237	12A13	454350 5417330	152	SG	2.5Y 6/2 5Y 4/2	89.1	8.4	2.5	0.26 2.96	914168		
91-237	12A13	454350 5417330	152	SG	2.5Y 6/4 10YR 5/4	87.4	11.2	1.5	-1.74 2.84	914169		
91-238	12H04	461660 5436920	140	SS	10YR 6/3 10YR 3/3	67.8	27.9	4.3	2.96 3.11	914170		2.5 m gravelly sand. Overlain by <1 cm manganese layer, and 30 cm diamicton.
91-239	12H04	461920 5436600	161	SG	10YR 6/3 10YR 6/4	69.8	28.2	1.9	3.87 1.43	914171		1 m+ sand and sandy gravel. Overlain by 2 m diamicton.
91-239	12H04	461920 5436600	161	D	7.5YR 6/4 5YR 4/3	27.7	37.3	35.1	4.48 4.58	914172	0.67 0.13	
91-240	12H04	461390 5436320	163	SG	10YR 5/2 7.5YR 3/2	86.0	10.3	3.7	-1.32 2.47	914173		Same as site 91194. 6 m sandy gravel.
91-241	12H04	461930 5447720	66.5	D	10YR 6/3 10YR 4/3	48.6	36.2	15.2	1.11 4.63	914174	0.47 0.17	2.5 m diamicton, reddish brown to greenish grey.
91-242	12H04	461030 5447670	86	D	10YR 6/3 7.5YR 3/2	66.1	27.1	6.8	-0.05 4.96	914175	0.72 0.08	4 m sandy diamicton.
91-243	12H04	460170 5447470	94	S	7.5YR 6/4 7.5YR 5/4	87.8	10.3	1.8	2.44 1.72	914176		2 m+ interbedded f and f-m sand. Overlain by 1 m sandy gravel.
92-1	12H03	475230 5451130	50	D	5YR 6/3 5YR 4/4	74.8	25.0	0.3	1.71 3.02	924000		Diamicton at base 5 m test pit. Surface disturbed
92-2	12H03	489370 5432060	310									Striae 252 ± 3.
92-3	12H03	487150 5430520	337									Striae 251 ± 3.
92-4	12H03	465710 5454940	128	SG								3-4 m exposure, slumped. Sandy gravel.
92-5	12H03	465870 5454600	138	SG								6 m exposure. m-c sand and pebble gravel.
92-6	12H03	465830 5454440	162	D						924001		Striae 254 ± 6. Clasts taken from surface.
92-7	12H03	467030 5452330	80	D	10YR 6/4 2.5YR 4/4	63.4	31.3	5.4	2.43 3.39	924002	0.7 0.09	1 m+ diamicton.
92-8	12H03	468360 5450890	13	D	2.5YR 5/4 2.5YR 3/4	52.0	43.0	5.1	3.66 2.45	924003	0.72 0.04	1 m+ diamicton. Overlain by 1 m boulder gravel.
92-9	12H03	468350 5450860	13	D	2.5YR 5/4 2.5YR 3/4	73.6	25.9	0.5	2.81 2.14	924004	0.7 0.08	1 m+ diamicton.
92-10	12H03	470560 5452030	8	SS	5YR 6/3 5YR 4/2	12.9	86.5	0.6	5.35 1.11	924005		2-5 m exposure at base sandy gravel terrace. Interbedded sand and silt.
92-10	12H03	470560 5452030	8	SL	5YR 6/3 5YR 4/2	6.2	85.9	8.0	6.42 1.27	924006		
92-11	12H03	490150 5445620	88	SG								M-c sand and granule gravel.
92-12	12H03	489700 5445800	89	S								2.5 m well sorted m-c cross-bedded sand.
92-13	12H03	481130 5449290	90	S								Veneer gravelly sand.
92-14	12H03	492850 5446180	102	GS						924007	0.66 0.05	Hummock 25 m diameter. Gravelly sand.
92-15	12H03	494880 5446920	100	SG						924008		Hummock 75 m diameter. 4 m exposure, slumped. 1 m+ sandy gravel overlain by 1 m gravelly sand.

92-16	12H03	496750 5447530	101	SG	10YR 6/3 10YR 4/3	97.6	2.5	0.0	-1.58 2.42	924009		Hummock 60 m diameter. 1.5 m sandy gravel.
92-17	12H03	498770 5448430	100	SG						924010		Hummock 75 m diameter. 3.5 m sandy gravel.
92-18	12H03	478470 5451960	55	D	5YR 6/3 5YR 4/3	79.7	20.2	0.1	1.15 2.9	924011	0.48 0.12	2 m+ sandy diamicton; 60 cm reddish brown diamicton; and 80 cm sand.
92-18	12H03	478470 5451960	55	D	2.5YR 5/4 5YR 3/4	65.2	33.3	1.5	1.61 3.39	924012	0.66 0.05	
92-18	12H03	478470 5451960	55	S	2.5YR 5/4 2.5YR 3/4	99.1	0.9	0.0	0.52 1.41	924013		
92-19	12H03	463870 5451070	155	D	7.5YR 7/2 10YR 4/3	60.3	38.8	0.9	2.18 3.25	924014		2 m diamicton.
92-20	12H03	464480 5451730	162	D	10YR 6/3 5YR 4/3	71.0	28.2	0.8	1.59 3	924015	0.81 0.06	2 m diamicton.
92-21	12H03	465500 5452530	119	D	10YR 6/3 7.5YR 4/4	71.3	27.5	1.2	1.37 3.3	924016	0.72 0.03	2 m diamicton.
92-22	12H03	465760 5452830	118	D								1 m reddish brown diamicton.
92-23	12H03	466180 5452560	100	SG	10YR 6/3 10YR 4/3	86.8	13.2	0.0	-0.45 3.02	924017		75 cm sandy gravel - gravelly sand.
92-24	12H03	466590 5452050	90	D	7.5YR 7/2 5YR 4/3	70.8	27.9	1.3	1.82 3.13	924018	0.72 0.05	2 - 2.5 m sandy diamicton. More gravelly towards surface.
92-25	12H03	468190 5450170	10	D								Diamicton; sand; 1.5 m sandy gravel.
92-26	12H03	467090 5448820	8	SS	5YR 6/3 5YR 4/3	29.0	65.1	6.0	5.02 1.83	924019		10 m exposure. Slumped. Interbedded silts, clays, fine and medium sands.
92-27	12H03	466720 5448580	42	SG								4 m+ interbedded m, c sands, and granule gravel; 2.5 m interbedded pebble-cobble gravels, and sandy gravels; 1.5 m interbedded f, very f and m sands, and sandy pebble gravels.
92-28	12H03	466260 5448000	27	SG								16.5 m terrace slumped. Interbedded m-c sands and gravelly sands.
92-29	12H03	470000 5447840	40	SG	5YR 6/3 5YR 4/3					924020		Interbedded m-c sands, and sandy gravels.
92-30	12H03	470270 5448580	30	SG								3-4 m exposure interbedded f, rippled sand and sandy gravel.
92-31	12H03	471140 5451080	17	S								3 m interbedded f and m sands. Overlain by planar bedded, sandy granule gravel.
92-32	12H03	474450 5451880	30	SG								3 m rippled f sand. Overlain by 2m+ sandy gravel.
92-33	12H03	472350 5447020	84	D	7.5YR 7/2 5YR 4/3	71.3	28.6	0.0	1.82 2.65	924021	0.66 0.04	3 m diamicton.
92-34	12H03	473200 5447550	85	D	10YR 7/3 10YR 4/3	78.4	21.2	0.4	0.88 2.86	924022		Gravelly sand over 1 m+ diamicton. Overlain by 1 m gravelly sand.
92-34	12H03	473200 5447550	85	GS	10YR 6/3 10YR 4/3	94.2	5.8	0.0	-0.47 2.49	924023		
92-35	12H03	477200 5447150	92	D	10YR 7/3 7.5YR 4/4	82.9	17.0	0.0	1.16 2.67	924024		1 m+ sandy diamicton.
92-36	12H03	478890 5448650	90	D	10YR 7/3 10YR 4/3	84.4	15.4	0.2	0.86 2.58	924025	0.64 0.07	Junction Brook pit. North side 1.5 m exp. Sandy gravel over 1 m+ diamicton.
92-37	12H03	491360 5446180	98	S	10YR 7/3 10YR 4/3	92.5	7.1	0.4	1.22 2.26	924026		1 m+ interbedded f sands, very f sand, and m-c sand beds; 80 cm sandy boulder gravel.
92-38	12H03	495070 5450060	94	GS	10YR 7/3 7.5YR 4/4	99.5	0.5	0.0	-0.32 1.63	924027		1 m pebbly sand.
92-39	12H03	495040 5453840	90	D	7.5YR 7/2 5YR 4/3	58.4	41.3	0.4	1.9 3.87	924028		3 m diamicton.
92-40	12H03	496760 5453450	95	S	7.5YR 6/4 5YR 4/3	93.0	6.7	0.3	1.52 1.6	924029		Sand.
92-41	12H03	470120 5445020	152									Striae 345 ± 3.
92-42	12H03	473220 5445300	142									Striae 354 ± 3.
92-43	12H03	475700 5439410	300									Striae 282 ± 3. Striae 020-200 ± 4. No age relationship.
92-44	12H03	476700 5436520	292									Striae 197 ± 3. Striae 287 ± 4. 197 crosses 287.
92-45	12H03	476820 5436750	260									Grooves. 294 ± 5. Striae. 240-060 ± 3.
92-46	12H03	476480 5438630	298									Striae 177 ± 3. Striae 286 ± 2. 177 crosscuts 286.
92-47	12H03	475470 5439100	290									Striae 262 ± 3.
92-48	12H03	475800 5438340	315									Striae, grooves 184 ± 4.
92-49	12H03	475830 5437850	312									Striae 179 ± 3.
92-50	12H03	475750 5437530	295									Striae 177 ± 3. Striae & grooves 289 ± 5. 177 crosses 289.
92-51	12H03	474250 5437080	272									Striae 170 ± 7.
92-52	12H03	474280 5433630	222									Striae 176 ± 4.
92-53	12H03	473860 5437700	254									Striae 204 ± 2.
92-54	12H03	474820 5438030	308									Striae 185 ± 3. Striae 289 ± 2.

92-55	12H03	470770 5436040	153									Grooves 200 ± 3.
92-56	12H03	470020 5444760	168	D	7.5YR 7/2 5YR 4/3	75.6	23.3	1.2	1.22 3.12	924229	0.79 0.05	1 m diamicton.
92-57	12H03	478250 5442570	300	D	10YR 6/3 10YR 4/3	82.3	17.1	0.7	0.52 3	924030		Poorly exposed. Sandy diamicton.
92-58	12H03	479730 5438970	250	D	7.5YR 7/2 7.5YR 4/2	79.0	20.0	1.0	-0.37 3.52	924031	0.71 0.06	1 m diamicton.
92-59	12H03	477070 5438130	265	D	10YR 6/2 10YR 4/2	77.6	20.0	2.5	-0.5 3.5	924032	0.82 0.08	1.5 m diamicton.
92-60	12H03	474420 5435130	260	D	10YR 6/2 10YR 3/2	62.6	35.4	1.9	-0.31 3.93	924033	0.68 0.06	1 m+ grey diamicton. Overlain by brown diamicton.
92-60	12H03	474420 5435130	260	D	10YR 6/3 7.5YR 4/2	66.0	32.1	1.9	0.93 3.68	924034	0.66 0.09	
92-61	12H03	474180 5433470	218	D	10YR 6/3 10YR 3/3	75.3	24.6	0.1	0.45 3.2	924035	0.78 0.03	2 m diamicton.
92-62	12H03	474320 5432040	263	D	10YR 6/2 10YR 3/3	55.0	44.0	1.1	1.16 3.61	924036	0.88 0.04	2 m diamicton.
92-63	12H03	474200 5439070	220	D	10YR 6/2 10YR 3/2	50.4	42.4	7.2	1.28 3.84	924037	0.75 0.02	1 m+ dark brown diamicton. Overlain by 25 cm light brown diamicton and 1 m gravelly sand.
92-63	12H03	474200 5439070	220	D	10YR 6/3 10YR 3/3	63.0	32.2	4.8	1.23 3.87	924038		
92-63	12H03	474200 5439070	220	GS	10YR 5/3 10YR 3/2					924039		
92-64	12H03	463580 5430040	318									Striae 266 ± 3.
92-65	12H03	463650 5429860	303									Striae 267 ± 3.
92-66	12H03	463870 5429750	290									Striae 266 ± 3.
92-67	12H03	464820 5429600	282									Striae and grooves 270 ± 2.
92-68	12H03	486030 5436450	225									Grooves 322 ± 6.
92-69	12H03	487580 5434750	220									Grooves 314 ± 2.
92-70	12H03	491030 5427500	373									Striae 297 ± 5.
92-71	12H03	489380 5432010	310									Striae 320 ± 5.
92-72	12H03	489440 5432060	310									Striae and grooves. 310 ± 4. Striae 258 ± 3. 310 crosses 258.
92-73	12H03	488820 5442480	90	GS	7.5YR 6/4 7.5YR 5/4	99.9	0.1	0.0		924040		15-20 m exp. Slumped. 5 m+ interbedded m, f and c sands and gravelly sands; 5 m+ gravelly sand.
92-74	12H03	488480 5441950	90	GS								10 m exp. Slumped. 1 m+ gravelly sand.; 1.5 m interbedded f, m-c, c sands and sandy gravel.
92-75	12H03	488050 5441480	90	GS								6 m exp. Slumped. 2 m+ interbedded gravelly sand, f, m, and c sands.
92-76	12H03	487900 5441280	90	GS		99.6	0.4	0.0		934041		2 m+ interbedded f sand, m-c sand, c sand and gravelly sands.
92-77	12H03	487800 5441230	90	GS								2 m+ interbedded f sand, m-c sand, c sand and gravelly sands.
92-78	12H03	485820 5439030	90	S	7.5YR 5/4 7.5YR 3/4	99.0	1.0	0.0		934042		3 m+ sand. Silt-clay interbeds.
92-78	12H03	485820 5439030	90	SL						934043		
92-78	12H03	485820 5439030	90	SL						934044		
92-79	12H03	485280 5438220	90	GS		99.8	0.2	0.0		934045		18 m exp. Slumped. Interbedded f, c, m sand and gravelly sand; 3 m+ interbedded sandy gravel, m-f sand, and pebbly sand.
92-80	12H03	466650 5432320	265									Striae 286 ± 3. Striae 263 ± 4. 286 crosses 263.
92-81	12H03	466950 5431980	250									Striae 292 ± 5.
92-82	12H03	491950 5443630	98	SG								4 m interbedded sandy gravel and c-m sands.
92-83	12H03	489200 5441850	130	SG								Ridge 6 m high and 25 m wide. Trends 250-070. Interbedded f, m-c sands and gravelly sand.
92-84	12H03	487250 5439970	125	GS								15 m exposure, mostly slumped. 2 m+ interbedded f and m sands; discontinuous gravelly sand.
92-85	12H03	485480 5437570	140	SS	10YR 7/2 10YR 4/2	12.0	85.8	2.2	5.35 1.24	924041		8 m exposure. Slumped. F sand and silt and cobbles, overlain by fine sands.
92-86	12H03	486500 5436860	300									Striae 320 ± 3. Striae 252 ± 7. 252 crosses 320.
92-87	12H03	487250 5436350	335									Striae 260 ± 3. Striae 300 ± 3. 260 crosses 300.
92-88	12H03	489680 5434750	340									Striae 310 ± 3.
92-89	12H03	489960 5435060	360	D	10YR 6/1 10YR 4/2	70.3	29.3	0.5	1.83 2.91	924042	0.79 0.03	5 m exp. Bottom 2 m slumped. Sandy diamicton plus lenses.; sandy diamicton, and diamicton
92-89	12H03	489960 5435060	360	D	7.5YR 6/2 7.5YR 4/2	74.0	25.6	0.5	1.04 3.36	924043	0.6 0.07	
92-89	12H03	489960 5435060	360	D	10YR 6/2 10YR 4/2	73.3	26.4	0.4	0.93 3.09	924109	0.58 0.06	
92-90	12H03	490720 5433220	325									Grooves 300 ± 3.

92-91	12H03	487100 5435350	220	SG								4 m exposure. Hummock. Sandy pebble gravel with sand lenses.
92-92	12H03	487200 5434960	230	SG								4 m exposure, lower 2 m slumped. Sandy gravel.
92-93	12H03	487720 5434370	260	D	10YR 6/2 10YR 3/2	78.8	20.7	0.4	1.23 2.74	924044	0.78 0.06	1 m diamicton.
92-94	12H03	487780 5434320	265	GS	7.5YR 6/2 7.5YR 4/2	85.2	14.8	0.0	0.35 2.86	924045		Gravelly sand.
92-95	12H03	488850 5433040	287	GS								Hummock. Gravelly sand.
92-96	12H03	489210 5431930	320	D	10YR 6/2 10YR 4/2	79.3	20.7	0.0	0.99 2.8	924046	0.78 0.05	3 m diamicton.
92-97	12H03	490450 5427970	365	D	10YR 6/2 10YR 3/2	78.2	21.1	0.7	0.55 3.01	924047	0.84 0.03	2 m diamicton.
92-98	12H03	490220 5428150	370									Striae 312 ± 3.
92-99	12H03	488820 5490030	370	GS	10YR 6/2 10YR 4/2	83.3	16.7	0.0	-0.22 3.05	924048		2 m gravelly sand.
92-100	12H03	488780 5429170	360	SG								Hummock. 5 m sandy boulder gravel.
92-101	12H03	468270 5435550	212									Striae and grooves 215 ± 5.
92-102	12A13	435120 5422580	40	SC	5YR 5/3 5YR 3/3	24.7	61.6	13.7	5.25 2.31	924049		Dawe's Pit. See Chapter 4.
92-103	12H04	437650 5430880	58	GS								See site 91173. Interbedded sands and gravelly sands.
92-104	12H04	449120 5427550	25	SL	5YR 6/3 5YR 4/3	3.2	89.1	7.7	6.16 1.11	924050		Interbedded c sands, granule gravel and sandy gravels. F sand and silt at base.
92-105	12A13	455700 5427520	45	SG								3 m interbedded sands and sandy gravels.
92-106	12A13	455720 5426350	50	SG								6 m interbedded sandy gravel, f-m, f, and gravelly sand beds.
92-107	12H04	462740 5440180	30	S	5YR 6/3 5YR 4/3	63.4	36.6	0.0	3.72 1.17	924052		1 m interbedded f sand, very f sand and silt. Overlain by sandy gravel.
92-108	12H03	465300 5442670	40	SG								1 m+ sandy gravel.
92-109	12H03	465050 5442570	28	SG								4 m interbedded m-f, c-m, c sand-granule gravels, and sandy gravels.
92-110	12H03	499940 5443530	240	GS	10YR 5/3 7.5YR 3/4	95.8	4.2	0.0	-1.22 2.26	924053		5 m exposure Slumped. Gravelly sand.
92-111	12H03	499470 5443720	235	D	10YR 7/2 10YR 5/3	84.9	14.9	0.3	1.41 2.65	924054		1 m+ sandy diamicton. Overlain by 1 m sandy gravel.
92-112	12H03	498380 5444400	205	D	10YR 7/2 10YR 5/2	81.4	18.2	0.4	0.72 2.79	924055	0.58 0.06	4 m sandy diamicton.
92-113	12H03	495460 5445120	145	SG								50 cm+ m-c sand. Overlain by 80 cm sandy gravel.
92-114	12H03	494820 5444920	140	SG								Ridge. 1 m+ gravelly sand. Overlain by 1 m sandy gravel.
92-115	12H04	437940 5453600	205	D						924056		Striae 253 ± 8.
92-116	12H04	435480 5446640	26.5	SS	5YR 5/4	35.4	54.9	9.7	4.67 2.83	924057		Silty clay to clayey silt.
92-116	12H04	435480 5446640	26.5	SS	5YR 4/3	38.1	53.2	8.7	4.76 2.54	924083		
92-117	12H04	456170 5450100	210									Striae 238 ± 3.
92-118	12H04	455550 5449230	218									Striae 255 ± 5.
92-119	12H04	447020 5451580	220	D	10YR 6/3 10YR 4/3	71.9	27.9	0.2	2.24 2.64	924058		50 cm+ diamicton.
92-120	12H04	447480 5449630	260	D	7.5YR 6/4 5YR 4/3	75.3	24.4	0.3	1.09 3.22	924059		1.5 m sandy diamicton.
92-121	12H04	451550 5455080	190	D	7.5YR 6/4 7.5YR 3/4	82.9	16.9	0.2	0.7 2.82	924060		1 m diamicton.
92-122	12H04	449520 5453680	155	D	10YR 5/3 10YR 3/3	75.7	23.0	1.3	-0.25 3.78	924061		2 m diamicton.
92-123	12H04	447320 5455040	265									Striae 255 ± 3.
92-124	12H04	446380 5454180	70	S	7.5YR 6/4 5YR 4/3	86.8	13.2	0.0	2.06 1.88	924062		4 m exp. Sand; 10 cm diamicton.; gravelly sand.
92-124	12H04	446380 5454180	70	D	5YR 6/4 5YR 4/3	62.9	36.7	0.3	1.99 3.19	924063		
92-124	12H04	446380 5454180	70	GS	7.5YR 6/4 7.5YR 3/4	88.3	11.7	0.1	-0.08 2.83	924064		
92-125	12H04	444420 5452670	45							924065		Brachiopods and some bivalves sample.
92-126	12H04	444240 5452780	65	D	10YR 6/3 7.5YR 4/4	80.7	19.1	0.2	1.01 2.86	924066		1 m diamicton.
92-127	12H04	445120 5450780	38							924067		Brachiopods and bivalves sample.
92-128	12H04	438360 5435050	85	D	10YR 6/3 10YR 4/3	71.1	28.6	0.3	0.33 3.42	924068		1 m diamicton.
92-129	12H04	438070 5434600	78	SG								1.5 m f sand and granule gravel.
92-130	12H04	436980 5435130	262									Striae and grooves 304 ± 3.

92-131	12H04	431380 5439520	310									Striae and grooves 305 ± 6.
92-132	12H04	431180 5439620	308									Striae 334 ± 3.
92-133	12H04	431080 5440020	295									Striae 295 ± 3.
92-134	12H04	430740 5440170	248									Striae and grooves 290 ± 3. Striae 315 ± 3. 315 crosses 290.
92-135	12H04	430450 5440180	255									Striae 308 ± 4.
92-136	12H04	430770 5439500	322									Striae and grooves 290 ± 4.
92-137	12H04	430340 5439250	310									Striae 296 ± 3. Striae 327 ± 4. 327 crosses 296.
92-138	12H04	429960 5438970	310									Striae and grooves 280 ± 3.
92-139	12H04	428210 5438130	268									Striae and grooves 264 ± 4.
92-140	12H04	427880 5437890	270									Striae 266 ± 3.
92-141	12H04	427000 5437250	230									Striae 254 ± 3.
92-142	12H04	431800 5439170	255									Striae 292 ± 5. Striae 260 ± 3. Striae 224 ± 4. Oldest 292, youngest 224.
92-143	12H04	438120 5444650	280									Striae 274 ± 4.
92-144	12H04	438200 5444850	280									Striae 270 ± 8. Striae 215 ± 15. 270 crosses 215.
92-145	12H04	439480 5446030	185	D	2.5Y 7/2 2.5Y 5/2	58.8	40.4	0.9	0.96 3.59	924069	0.53 0.07	4 m diamicton.
92-146	12H04	439430 5446500	150									Striae 255 ± 3. Striae 289 ± 3. 289 crosses 253.
92-147	12H04	439650 5446870	135	D	10YR 7/2 2.5Y 5/2	60.8	39.0	0.2	0.3 3.45	924070	0.68 0.09	1 m+ diamicton. Overlain by 3 m sandy diamicton.
92-147	12H04	439650 5446870	135	D	10YR 6/2 2.5Y 4/2	74.5	23.9	1.6	-0.07 3.42	924071	0.56 0.07	
92-148	12H04	439860 5447200	95	SG								2 m+ gravelly sand; 5-20 cm f sand and silt; 2.5 m sandy gravel.
92-149	12H04	439930 5447470	90	GS	10YR 6/3 10YR 4/3	87.9	12.2	0.0	1.16 2.54	924072		4 m exposure gravelly sand.
92-150	12H04	439950 5447690	90	D	10YR 6/3 10YR 3/3	63.3	34.4	2.4	2.05 3.3	924073		1 m diamicton.
92-151	12H04	444120 5443970	165	D	10YR 7/2 10YR 4/2	69.9	29.1	1.1	0.96 3.36	924074		2 m diamicton.
92-152	12H04	448870 5443870	240	D	2.5Y 4/2	60.5	38.7	0.8	0.2 3.85	924075	0.77 0.05	5 m exposure, slumped. Diamicton.
92-153	12H04	446200 5443670	160	D	2.5Y 4/2	74.5	24.7	0.9	0.31 3.34	924076		20 cm+ grey diamicton; 1.5 m brown diamicton.
92-154	12H04	443250 5442830	152	D	2.5Y 4/2	69.5	30.3	0.1	1.03 3.46	924077		2 m diamicton.
92-155	12H04	440800 5442250	132	GS								Gravelly sand - sandy gravel.
92-156	12H04	440170 5442150	130	SG								1 m+ sandy gravel; 30 cm granules and c sand.; 1 m sand.
92-157	12H04	439670 5442200	172									Striae 250 ± 4.
92-158	12H04	438680 5447550	40	SS	5YR 4/3	23.1	68.3	8.6	5.03 1.83	924078		6 m sandy gravel, with 2 cm interbed of f sand, very f sand and silt.
92-159	12H04	439180 5447570	60	D	10YR 6/3 10YR 4/3	78.5	21.0	0.5	0.27 3.07	924079		1 m diamicton.
92-160	12H04	438340 5447700	40	SG								1 m+ sandy gravel. Overlain by 1 m gravelly sand.
92-161	12H04	437730 5447800	20	SC	5YR 4/3	3.3	88.6	8.2	5.91 1.15	924080		10-15 m terrace. 70 cm+ silt-clay; 80 cm sandy gravel.
92-162	12H04	437500 5447950	10	SC	5YR 4/2	3.6	91.5	4.9	5.91 1.05	924082		3 m silt-clay. Shell sample.
92-163	12H04	435150 5447150	42	D	10YR 7/2 10YR 5/3	42.5	53.6	4.0	1.73 3.96	924084	0.89 0.04	4 m diamicton.
92-164	12H04	435880 5447550	30	SC	5YR 4/3	9.8	87.7	2.5	5.21 1.17	924085		1 m+ silty clay.
92-165	12H04	436660 5447980	50	D	5YR 6/4 5YR 4/3	55.8	40.3	3.9	2.03 3.5	924086		1 m diamicton. Overlain by 60 cm sandy gravel.
92-166	12H04	436920 5448300	72	D	7.5YR 6/4 7.5YR 4/4	70.9	28.6	0.5	2.54 2.62	924087		1.5 m diamicton.
92-167	12H03	469770 5438270	145	D	5YR 6/3 5YR 3/3	62.3	37.5	0.3	2.09 2.94	924088		50 cm diamicton.
92-168	12H03	469140 5443430	240	D	5YR 6/3 5YR 4/3	76.4	23.5	0.2	1.55 2.76	924089		50 cm diamicton.
92-169	12H04	460350 5435670	120	D	2.5YR 5/4 2.5YR 3/4	56.1	43.6	0.3	1.72 3.14	924090		1.5 m diamicton.
92-170	12H04	460720 5435920	160	D	10YR 6/3 10YR 3/3	67.7	30.4	1.9	0.47 3.85	924091		1 m sandy diamicton.
92-171	12H03	463700 5436270	220	D	5YR 6/3 5YR 3/3	62.2	37.2	0.6	1.97 3.17	924092	0.88 0.04	2 m diamicton.
92-172	12H03	464250 5436180	198	S								75 cm well sorted f sand. Contains diamicton interbeds.

92-173	12H03	465030 5436120	187	SG								4 m exposure Slumped. 1 m + pebble gravel overlain by 1 m sand overlain by sandy gravel.
92-174	12H03	465980 5435980	158	SG								50 cm+ gravelly sand overlain by 1 m sandy gravel.
92-175	12H03	467430 5435800	150	D	5YR 6/3 5YR 3/3	61.8	36.9	1.3	1.41 3.53	924093		1.5 m diamicton.
92-176	12H03	469150 5436080	175	D	7.5YR 5/4 5YR 3/3	70.9	28.6	0.5	0.64 3.27	924094	0.57 0.04	4 m diamicton.
92-177	12H03	465730 5435830	168	SG								10 cm + sandy gravel. Overlain by 25 cm fine sand. Overlain by 50 cm sandy gravel.
92-178	12H03	483120 5434820	100	D	7.5YR 7/2 7.5YR 4/2	70.8	27.0	2.3	1 3.54	924095		20 m exp. Slumped. 50 cm+ diamicton.; 1 m f sand grading down into 30 cm silt and clay; 6 m interbedded sandy gravel.
92-178	12H03	483120 5434820	100	SC	7.5YR 7/2 7.5YR 4/2	32.5	62.0	5.6	4.95 1.98	924096		
92-178	12H03	483120 5434820	100	D	5YR 6/3 5YR 3/3	73.9	24.8	1.3	1.25 3.38	924097		
92-178	12H03	483120 5434820	100	SG	7.5YR 6/2 7.5YR 3/2	92.8				934038		
92-178	12H03	483120 5434820	100	D	5YR 5/3 5YR 3/3	74.7	19.8	5.5	1.55 3.47	934039	0.59 0.17	
92-179	12H03	482630 5433970	100	SG		99.9	0.1	0.0		934037		30 m exposure Slumped. 10 m+ sandy gravel.
92-180	12H03	482120 5433030	90									Grooves 208 ± 4.
92-181	12H03	481780 5432720	90									Striae 214 ± 3.
92-182	12H03	481750 5432630	100	SG								13 m exposure. Sandy gravel. Bedrock at base striated 208 ± 4.
92-183	12H03	481530 5432250	100	SG		99.9	0.1	0.0		934036		Conglomerate overlain by sandy gravel and gravelly sands.
92-184	12H03	480520 5430380	100	GS		99.3	0.7	0.0		934035		10 m exposure Slumped. Gravelly sand.
92-185	12H03	477750 5427770	95	SG		99.9	0.1	0.0		934034		6 m exposure Slumped. Interbedded gravelly sand and sandy gravel.
92-186	12A13	443630 5425350	20	SC	5YR 4/3	0.6	77.0	22.5	6.63 1.54	924098		1 m+ silt-clay.
92-187	12H04	437150 5448070	29	SC	5YR 4/3	25.5	50.6	24.0	5.59 2.75	924099		1 m+ silty clay.
92-188	12H04	436920 5448320	65									Striae 206 ± 6. Striae 126 ± 4. 126 crosses 206.
92-189	12H04	437320 5448770	100	D	5YR 6/3 5YR 4/3	54.2	43.4	2.5	1.53 3.73	924100	0.6 0.1	1.5 m+ diamicton. Overlain by 2.5 m sand and pebble sand.
92-190	12H04	439630 5451200	50	SC	5YR 4/3	4.6	48.5	46.9	7.39 2.08	924101		1 m+ silt-clay. Reddish brown. Overlain by 4 m sandy gravel.
92-191	12H04	437280 5450250	150	D	5YR 6/3 5YR 4/3	55.0	39.6	5.4	2.81 3.11	924102	0.85 0.04	4 m diamicton. Reddish brown.
92-192	12H04	436920 5451240	180	D	7.5YR 6/4 7.5YR 4/4	80.2	19.6	0.3	1.43 2.75	924103		2 m diamicton. Brown.
92-193	12H04	437220 5452870	180	D	10YR 8/2 10YR 5/2	56.5	42.1	1.4	1.59 3.5	924104	0.74 0.06	3 m diamicton. Greyish brown.
92-194	12H03	489270 5444220	90	SS	5YR 3/3	20.8	79.1	0.2	4.67 1.11	924105		6 m exp. Slumped. F-m rippled sand; interbedded m-c sands, and sandy pebble gravels; rippled very f sand and silt.; rippled m-f sand.
92-195	12H03	489660 5431200	330	D	10YR 6/2 10YR 3/2	73.7	26.2	0.1	1.17 3.03	924106	0.53 0.05	1.5 m+ sandy diamicton. Overlain by 1.5 m diamicton.
92-195	12H03	489660 5431200	330	D	10YR 6/3 10YR 3/2	82.3	17.7	0.0	-0.06 3.07	924107	0.88 0.03	
92-196	12H03	489540 5430420	355	D	10YR 6/2 10YR 3/2	83.5	16.5	0.0	-0.28 3.04	924108		1 m diamicton.
92-197	12H04	462950 5430880	290	D	7.5YR 6/2 7.5YR 3/2	80.2	19.2	0.6	-0.25 3.51	924110	0.83 0.05	2 m diamicton.
92-198	12H03	471780 5437120	145	D	5YR 5/3 5YR 3/2	55.9	43.2	1.0	0.69 3.55	924111	0.57 0.11	1 m diamicton.
92-199	12H03	474180 5438680	197	SG								4 m sandy gravel.
92-200	12H03	473800 5438830	200	D	10YR 6/2 10YR 3/2	71.0	24.5	4.5	-0.16 3.42	924112		3 m diamicton.
92-201	12H03	485850 5454560	132	D	7.5YR 6/2 7.5YR 4/2	63.2	36.6	0.3	2.34 3.08	924113	0.61 0.05	Pit 3 m deep. 1.5 m+ diamicton. Overlain by 1.5 m diamicton.
92-201	12H03	485850 5454560	132	D	10YR 6/3 7.5YR 4/2	54.2	44.7	1.2	1.85 3.79	924114	0.53 0.14	
92-202	12H03	484620 5453340	132	D	5YR 6/3 5YR 4/3	66.5	31.9	1.6	2.43 2.77	924115	0.46 0.1	2.5 m diamicton.
92-203	12H03	483780 5452140	122	D	5YR 6/3 5YR 3/3	67.6	30.9	1.5	1.61 3.48	924116	0.73 0.11	2 m diamicton. Gravelly sand lens in middle of unit.
92-204	12H03	482020 5451880	90	GS	5YR 5/3 5YR 3/3	90.7	9.3	0.0	1.66 2.08	924117	0.5 0.07	3 m gravelly sand.
92-205	12H03	472350 5445750	140	D	5YR 6/3 5YR 4/3	41.4	57.3	1.3	2.29 3.85	924118	0.78 0.06	2.5 m diamicton.
92-206	12H03	474170 5444020	128	D	5YR 5/3 5YR 3/3	66.2	33.4	0.5	1.67 3.26	924119	0.72 0.06	50 cm+ diamicton. Overlain by 2 m diamicton.
92-206	12H03	474170 5444020	128	D	5YR 6/3 5YR 3/3	53.0	46.6	0.4	3.54 1.93	924120		
92-207	12H03	473170 5442320	160	D	7.5YR 7/2 7.5YR 4/2	45.5	53.4	1.1	3.8 2.28	924121	0.64 0.09	1 m+ brown diamicton. Overlain by 2 m reddish brown diamicton.
92-207	12H03	473170 5442320	160	D	5YR 6/3 5YR 4/3	66.5	32.6	0.8	2.05 3.03	924122	0.61 0.06	

92-208	12H03	472950 5441000	150	D	7.5YR 6/2 5YR 4/2	51.4	48.4	0.3	1.17 4.01	924123		Dark reddish grey diamicton. Overlain by 1 m dark reddish brown diamicton.
92-208	12H03	472950 5441000	150	D	5YR 5/3 5YR 3/3	51.9	47.1	1.1	2.29 3.28	924124	0.86 0.05	
92-209	12H03	472450 5440480	162	D	5YR 6/3 5YR 3/3	68.5	30.3	1.2	2.47 2.87	924125	0.57 0.09	2.8 m diamicton.
92-210	12H04	461470 5436350	150	D	10YR 6/3 10YR 3/3	69.8	29.7	0.5	-0.32 3.69	924126	0.51 0.12	Ridge. 2.5 m+ diamicton.
92-211	12H04	461220 5436300	150	D	7.5YR 6/2 7.5YR 3/2	81.2	18.3	0.5	1.07 2.8	924127	0.51 0.07	1.3 m+ brown. diamicton.; 1.2 m grey diamicton.
92-211	12H04	461220 5436300	150	D	5YR 6/1 5YR 4/1	69.2	27.4	3.4	0.95 3.96	924128		
92-212	12H04	462500 5436150	160	D	7.5YR 6/4 5YR 3/3	48.9	50.7	0.5	2 3.53	924129	0.84 0.04	2.5 m+ diamicton.
92-213	12H03	477820 5451680	75	D	7.5YR 6/4 7.5YR 4/4	80.1	18.7	1.2	1.23 2.83	924130	0.49 0.22	1.5 m+ sandy diamicton. Brown. Overlain by 70 cm sandy gravel.
92-214	12H03	477970 5450250	100	D	7.5YR 7/2 7.5YR 4/4	77.9	21.2	1.0	1.84 2.61	924131	0.55 0.17	Ridge. 2 m sandy diamicton.
92-215	12H03	480860 5452730	90	D	7.5YR 7/2 5YR 4/3	76.3	23.2	0.5	1.64 2.83	924132	0.54 0.1	Ridge. 2.5 m sandy diamicton.
92-216	12H03	481260 5452080	90	D	7.5YR 7/2 5YR 4/3	78.8	21.1	0.2	2.12 2.16	924133	0.52 0.08	2 m diamicton.
92-217	12H03	493250 5449130	120	D	7.5YR 7/2 5YR 4/3	70.0	29.1	0.9	2.21 3.02	924134	0.67 0.1	2 m diamicton.
92-218	12H03	492950 5448720	122	D	7.5YR 6/2 7.5YR 4/2	66.4	32.7	1.0	2.48 2.78	924135	0.69 0.05	2.5 m diamicton.
92-219	12H03	492750 5448450	115	D	7.5YR 6/2 7.5YR 4/2	72.0	27.4	0.6	1.62 3.21	924136	0.49 0.1	2.5 m diamicton.
92-220	12H03	471980 5437370	140	D	5YR 6/2 5YR 3/2	43.1	54.5	2.4	0.82 5.65	924137	0.71 0.05	2 m diamicton.
92-221	12H04	446180 5443320	175	D	2.5Y 6/2 2.5Y 4/2	56.3	36.5	7.2	2.4 3.6	924138	0.73 0.02	1 m+ dark greyish brown diamicton. Overlain by 1.5 m sandy diamicton. Dark greyish brown.
92-221	12H04	446180 5443320	175	D	2.5Y 6/2 2.5Y 4/2	70.6	26.4	3.1	0.77 3.43	924139	0.74 0.04	
92-222	12H04	432950 5438030	142	D	2.5Y 6/2 10YR 4/2	55.7	41.2	3.2	2.91 3.07	924140	0.65 0.07	2 m diamicton.
92-223	12H04	458420 5448030	150									Striae and grooves 285 ± 3.
92-224	12H04	455720 5454860	200	D	5YR 6/3 5YR 4/3	60.7	36.8	2.6	2.46 3.3	924141		2 m diamicton.
92-225	12H04	456950 5454270	178	SG								1 m+ sandy gravel.
92-226	12A13	442840 5425530	9	SC		2.1	65.5	32.5	6.84 1.99	924142		
93-1	12H06	471890 5467500	255									Striae 227 ± 3. Striae 090 ± 3. 090 x by 227.
93-2	12H06	471090 5468920	327									Striae 267 ± 4.
93-3	12H06	471180 5468400	320	D	5YR 5/6 5YR 3/4	84.5	15.3	0.2	-1.31 3.21	934000		Striae 275 ± 5. 1 m+ diamicton.
93-4	12H06	471550 5468140	297									Striae 256 ± 3.
93-5	12H06	472680 5466950	182	D	5YR 7/4 5YR 3/4	67.3	30.6	2.2	2.24 3.16	934001	0.69 0.08	3 m diamicton.
93-6	12H06	481350 5467300	71									Grooves 215 ± 3. Grooves 158 ± 6. 158 x by 215.
93-7	12H06	476850 5466730	90									Striae 208 ± 8. Striae 336 ± 10. 336 x by 208.
93-8	12H03	480370 5429680	90	D	7.5YR 7/2 7.5YR 3/2	86.1	10.4	3.5	-0.84 3.29	934002		10m exposure. Gravelly sand, reddish brown silt-clay on clasts; sandy gravel, and 30 cm diamicton.
93-8	12H03	480370 5429680	90	D	7.5YR 7/2 5YR 3/3	72.9	15.2	11.9	-3.22 2.38	934003		
93-8	12H03	480370 5429680	90	SG						934004		
93-9	12H03	478750 5428450	90	SG						934005		Grindstone Point section. See chapter 4.
93-9	12H03	478750 5428450	90	SS	5YR 7/1 7.5YR 3/4	59.4	36.9	3.6	3.62 2.17	934006		
93-9	12H03	478750 5428450	90	S		99.1	0.9	0.0		934007		
93-10	12A14	477210 5426720	90	SG						934008		30 m exposure Lower 15 m slumped. 1 m+ sandy gravel.; 60 cm interbedded m-f sands and gravelly sand; 60 cm cemented c sand and gravelly sand.
93-11	12A14	476220 5425180	90	D	7.5YR 6/2 7.5YR 3/2	86.5	10.6	3.0	-0.11 2.93	934009		12 m exposure Slumped. 1.2 m+ sandy diamicton; 1.2 m sandy gravel; 1.1 m diamicton.
93-11	12A14	476220 5425180	90	SG		98.7	1.3	0.0		934010		
93-11	12A14	476220 5425180	90	D	5YR 7/2 5YR 3/3	76.4	20.8	2.7	-0.2 3.13	934011	0.56 0.01	
93-11	12A14	476220 5425180	90	D	7.5YR 7/2 7.5YR 4/2	65.3	27.3	7.5	1.14 4.04	934012	0.51 0.21	
93-12	12A14	475820 5424420	90	SG	7.5 6/4 5YR 3/3	93.0	3.7	3.2	-0.33 2.15	934013		12m exposure Slumped. 2 m+ sandy gravel; 40-80 cm interbedded f, m, c, sand; 5 m sandy gravel.
93-13	12A14	473020 5421120	90	SL	7.5YR 7/2 7.5YR 3/4	8.1	90.4	1.5	5.57 1.53	934014		Little Pond Brook section. See Chapter 4.

93-13	12A14	473020 5421120	90	SS	5YR 6/2 5YR 4/3	24.4	75.5	0.1	4.86 1.24	934015		
93-14	12H03	473540 5429690	150	SG	2.5Y 7/4 2.5Y 4/2	83.1	16.1	0.8	-2.78 3.74	934029		Beach ridges on hillside. Beaches at 54.5m, 67m, and 88.5m above lake level. Uppermost beach is sandy gravel.
93-15	12A14	469820 5426680	90	SC	5YR 5/2 5YR 3/3	35.3	39.8	25.0	5.44 3.26	934016		8 m exposure 3 m+ silty sand. Overlain by 2 m gravelly sand, and 1m+ diamicton.
93-15	12A14	469820 5426680	90	D	5YR 5/3 7.5YR 3/4	63.9	26.5	9.7	1.06 4.64	934017		
93-16	12A14	465250 5420570	90	SG		99.9	0.1	0.0		934018		12m exposure in terrace. Interbedded sandy gravel and gravelly sand. Possible beaches 141m asl
93-17	12A14	468550 5416230	90	SG						934019		20m exposure Slumped. >1m pebbly sand. Overlain by 1.2m sandy gravel.
93-18	12H03	487250 5439970	125	S		99.9	0.1	0.0	2.3 0.8	934020		Alder Brook section. See Chapter 4.
93-19	12H03	490430 5432500	210	D	5YR 7/1 5YR 4/1	68.6	31.2	0.2	1.28 3.11	934021	0.48 0.16	1 m+ dark grey diamicton. Overlain by 2 m dark reddish brown diamicton.
93-19	12H03	490430 5432500	210	D	5YR 6/1 5YR 3/2	80.7	19.1	0.2	0.93 2.68	934022	0.57 0.13	
93-20	12H03	486580 5434880	380	D	5YR 6/1 5YR 4/2	85.9	13.8	0.4	0.47 2.63	934023	0.59 0.16	5m exposure. 3m diamicton. Overlain by 10 cm sand and 1 m+ diamicton.
93-21	12H03	487620 5433840	350	D	7.5YR 6/4 7.5YR 3/4	80.2	19.8	0.0	0.8 2.82	934024		4.5 m diamicton.
93-22	12H03	486660 5445610	90	GS	7.5YR 6/4 7.5YR 3/4	96.9	3.1	0.0		934025		10 m exposure Slumped. 5m+ gravelly sand. Overlain by 2.5 m is interbedded c, m, f sand.
93-22	12H03	486660 5445610	90	SL						934026		
93-23	12H03	485450 5445370	90	SG		99.6	0.4	0.0		934027		35 m exposure. Slumped. 16 m interbedded c, m, f sand, overlain by 16 m sand and gravel.
93-24	12H03	484250 5446150	90	GS								5 m exposure 2 m+ sand overlain by 1 m pebbly sand.
93-25	12H03	480850 5447430	90	SG		99.8	0.2	0.0		934028		15 m exposure. Slumped. Upper 3 m shows sand over sandy gravel over interbedded c sand + granules, c and m sand.
93-26	12H03	480030 5435430	150									Beaches at elevation 144 m and 172 m asl.
93-27	12H03	480310 5438260	150									Beach at 56 m above lake.
93-28	12A14	469220 5416770	90	D	10YR 7/3 10YR 3/4	69.7	23.3	7.0	0.65 4.03	934030	0.61 0.12	10 m exposure. Diamicton. Overlain by 2 m sandy diamicton.
93-28	12A14	469220 5416770	90	D	5YR 6/4 5YR 3/3	62.9	30.4	6.7	1.65 4.01	934031		
93-29	12A14	470550 5418270	90	GS		97.8	2.2			934032		12 m exposure 1 m+ sandy diamicton. Overlain by 4 m gravelly sand.
93-29	12A14	470550 5418270	90	D	5YR 6/2 5YR 3/4	75.1	21.4	3.5	0.37 3.39	934033		
93-30	12A14	475250 5423400	140									Delta. Surface 53 m above lake. Possible higher surface 83 m above lake.
93-31	12H03	479950 5428950	140									Delta surface 43 m above lake.
93-32	12H03	484700 5436200	130									Delta surfaces, some unclear and dissected, 31m, 43m, 74m above lake.
93-33	12H03	480270 5436120	140									Beach terraces at 62m.
93-34	12H03	468420 5450900	30	SG								Rocky Brook section. See chapter 4.
93-35	12A13	437230 5424880	50	D	10YR 7/3 10YR 3/6	69.3	25.6	5.1	1.07 3.72	934046	0.82 0.08	2 m diamicton.
93-36	12A13	436420 5424580	55	GS								4 m interbedded sands and gravelly sand.
93-37	12A13	435120 5422500	30	SG						934047		Same as site 92037
93-37	12A13	435120 5422500	30	SG						934048		
93-37	12A13	435120 5422500	30	SC	10R 6/3 2.5YR 2/4	3.9	28.9	67.2	8.23 1.78	934049		
93-38	12A13	427620 5423000	10							934050		5.5 m+ interbedded clays, fine sands and gravelly sand. Shell sample.
93-38	12A13	427620 5423000	10	SC	2.5YR 6/2 2.5YR 4/2	7.1	55.3	37.6	7.24 2.11	934051		
93-39	12H06	471100 5469070	325									Striae 272 ± 3.
93-40	12H06	470950 5469170	335									Striae 257±3. Striae 210 ± 4. 210 crossed by 257.
93-41	12H06	470730 5470050	380									Striae 270 ± 3.
93-42	12H06	470740 5470700	430									Striae 205 ± 4. Striae 235 ± 4. 205 crossed by 235.
93-43	12H06	470380 5471650	445									Striae 222 ± 3.
93-44	12H06	470460 5471750	435	D	10Y 7/1 10YR 4/2	69.2	27.8	3.0	1.17 3.87	934052		1 m+ diamicton.
93-45	12H06	470820 5470350	305	D	10YR 6/2 10YR 4/2	89.6	9.9	0.5	0.3 2.48	934053	0.61 0.08	2.5 m sandy diamicton.
93-46	12H06	473870 5468850	130	SG								Esker. Sandy gravel.
93-47	12H06	469000 5477140	510									Striae and grooves. 230 ± 10.

93-48	12H06	469050 5478390	435									Grooves 185 ± 4.
93-49	12H06	468850 5479280	420	D						934054		Striae and grooves. 205 ± 5. Pebble sample taken from surface.
93-50	12H06	470650 5480540	470									Striae 205 ± 3.
93-51	12H06	472370 5482600	347	SG								3 m exposure, slumped. Sandy gravel.
93-52	12H06	471600 5481770	355									Grooves 200-020.
93-53	12H06	473600 5475600	395	D						934055		Pebble sample.
93-54	12H06	475860 5475140	220	D						934056		Pebble sample.
93-55	12H06	491880 5466870	125	D	7.5YR 5/4 7.5YR 3/4	73.0	25.0	2.0	-0.03 3.61	934057		1 m+ diamicton.
93-56	12H06	492340 5467830	118									Grooves 185.
93-57	12H06	492570 5468320	108									Grooves 192-012.
93-58	12H06	497800 5471080	270									Striae 210 ± 5.
93-59	12H06	497920 5471160	270									Striae 200 ± 3.
93-60	12H06	498370 5470890	270									Striae 210 ± 4.
93-61	12H06	496930 5470500	240	D						934058		Pebble sample.
93-62	12H06	495520 5468840	240	D						934059		Pebble sample.
93-63	12H06	492750 5469670	90									Striae 215 ± 4.
93-64	12H06	489320 5458450	220	D						934060		Pebble sample.
93-65	12H06	489320 5457780	240									Striae 200 ± 6.
93-66	12H06	488630 5460450	139									Striae 200 ± 5.
93-67	12H06	490580 5461020	210	D	7.5YR 6/0 7.5YR 3/2	49.7	36.9	13.4	2.02 4.68	934061	0.65 0.1	2 m exposure. Upper 80-100 cm road fill. Beneath is 1 m+ diamicton.
93-68	12H06	491140 5461870	228	D						934062		1 m+ diamicton.
93-69	12H06	496370 5466480	254									Grooves 272 ± 3.
93-70	12H06	481060 5467070	72									Striae 214 ± 5. Striae 270 ± 3. 214 crosses 270.
93-71	12H06	480550 5466770	70									Grooves and striae 203 ± 3.
93-72	12H06	495260 5470570	160									Grooves 345 ± 5.
93-73	12H06	496620 5471650	172									Striae 195 ± 5.
93-74	12H06	498120 5473600	183									Grooves and striae 230 ± 3.
93-75	12B09	419800 5391130	90	SG								15 m exposure, slumped. 1 m+ sandy gravel.
93-76	12B09	417500 5391350	107	S								5 m+ sand. Cross-bedded and faulted units included.
93-77	12H06	497520 5476300	150	D	7.5YR 6/4 7.5YR 4/4	71.7	25.7	2.6	0.79 3.6	934064	0.74 0.07	1.5 m diamicton.
93-78	12H06	497020 5479070	80	D						934065		Pebble sample.
93-79	12H03	474320 5450780	50	GS								1.5 m+ sand overlain by 1.2 m gravelly sand
93-80	12H06	491900 5456900	153	D	5YR 6/3 5YR 3/4	60.6	37.1	2.3	1.91 3.54	934066	0.7 0.11	1.5 m diamicton.
93-81	12H06	497520 5463770	185	D						934067		Pebble sample.
93-82	12H06	499130 5467090	155	D	2.5YR 6/0 2.5YR 3/0	45.0	32.6	22.4	1.9 4.79	934068	0.52 0.19	2 m grey diamicton. Overlain by 50 cm brown diamicton.
93-82	12H06	499130 5467090	155	D	2.5YR 6/2 5YR 2.5/1	50.3	36.7	13.0	3.06 4.14	934069		
93-83	12H07	505130 5480030	242									Striae 184 ± 3. Striae 112 ± 8. 184 crosses 112.
93-84	12H11	499550 5489730	192									Striae 210 ± 3. Striae 118 ± 3. 210 crosses 118.
93-85	12H11	497800 5491750	250									Striae 205 ± 3.
93-86	12H10	500320 5484030	142									Striae 085 ± 4.
93-87	12H03	465800 5454930	120	D		41.2	45.2	13.6	3.88 3.72	934070	0.7 0.11	3 m + diamicton. Overlain by 1 m f-m, mod -well sorted sand.
93-88	12H11	489220 5486620	170	SS		36.4	56.0	7.6	4.94 2.26	934071		15 cm structureless sand/silt. Overlain by 60 cm diamicton.
93-88	12H11	489220 5486620	170	D	2.5Y 6/2 2.5Y 3/2	75.0	24.7	0.3	0.43 3.58	934072		

93-89	12H11	490230 5488670	125									Striae 152 ± 3.
93-90	12H11	491080 5485690	90	D	7.5YR 6/0 5YR 4/1	69.0	28.5	2.5	1.49 3.67	934073		30 cm+ diamicton. Overlain by 1 m sand and gravel.
93-91	12H11	491480 5484200	80	GS	10R 5/4 10R 3/3	90.4				934074		80 cm gravelly sand.
93-92	12H06	485820 5474970	80	D	10R 6/3 10R 3/4	53.5	45.2	1.3	2.79 2.87	934075		Ridge. 1 m+ diamicton. Overlain by 1 m gravelly sand.
93-93	12H10	500250 5484170	140	D	10R 6/2 10R 3/2	81.7				934076		4 m exp. Slumped. 2 m+ sandy diamicton.
93-94	12H10	500520 5487070	145	D	5YR 6/2 5YR 3/2	88.7				934077	0.48 0.24	2 m diamicton.
93-95	12H06	486530 5475930	80	D	10R 6/3 10R 3/6	88.5	10.6	1.0	0.04 2.64	934078		50 cm+ diamicton. Overlain by 5-10 cm reddish brown sandy silt and 50 cm gravelly sand.
93-95	12H06	486530 5475930	80	SS	10R 4/8 10R 4/4	27.4	63.7	8.9	3.73 3.52	934079		
93-95	12H06	486530 5475930	80	GS	10R5/6 10R 3/6	99.1	0.9	0.0		934080		
93-96	12H06	488550 5479740	87	D	10R 6/3 10R 3/3	69.9	28.9	1.3	1.58 3.21	934081		50 cm grey diamicton. Overlain by 50 cm gravelly sand.
93-96	12H06	488550 5479740	87	GS	5YR 5/4 5YR 3/3	87.3	12.7	0.0	0.93 2.27	934082		
93-97	12H06	489880 5481800	92	D	5YR 6/3 5YR 3/3	78.1	19.9	2.0	0.18 3.4	934083		Hummock. 30 cm+ sandy diamicton; 20 cm red sandy gravel; 40 cm red grey to grey diamicton.
93-97	12H06	489880 5481800	92	D	10R 6/3 10R 3/4	80.9	18.4	0.8	-0.28 3.51	934084		
93-97	12H06	489880 5481800	92	D	5YR 6/2 5YR 3/1	65.7	34.3	0.0	2.39 2.68	934085		
93-98	12H11	490550 5482930	90	D	2.5YR 5/4 2.5YR 3/4	79.8	18.7	1.5	1.07 2.77	934086		50 cm+ reddish diamicton. Overlain by 1m sandy gravel.
93-98	12H11	490550 5482930	90	SG	7.5YR 5/6 5YR 3/3	99.0	1.0	0.0		934087		
93-99	12H11	491700 5483350	80	SG		99.6	0.4	0.0		934088		3m interbedded sandy gravel and sand.
93-100	12H07	500270 5480080	292									Grooves and striae 072 ± 4.
93-101	12H07	500870 5477630	230	D	5YR 6/4 5YR 3/4	66.1	31.4	2.6	1.57 3.36	934089	0.8 0.04	1.5m diamicton.
93-102	12H07	504900 5478030	185	D	2.5Y 7/2 2.5Y 4/4	56.1	36.5	7.4	-0.01 4.71	934090	0.7 0.08	2m diamicton.
93-103	12H07	503130 5474850	200	D								Drumlin. 1 m+ diamicton; 40 cm sandy diamicton; 2 m+ sandy diamicton.
93-104	12H07	502540 5474200	200	D	2.5Y 5/2 2.5Y 3/2	70.4	26.9	2.8	1.24 3.34	934091		5 m diamicton.
93-105	12H07	502170 5472650	160	D	10YR 6/2 10YR 3/2	73.0	26.4	0.6	0.94 3.17	934092	0.63 0.03	3.5 m exposure, slumped. 1.5 m+ sandy diamicton.
93-106	12H07	511870 5457720	125	SG								1 m+ interbedded sand and sandy gravel.
93-107	12H07	511220 5455300	280	SG								1 m+ sandy gravel.
93-108	12H07	518370 5459580	440	D								veneer sandy diamicton.
93-109	12H07	520550 5458400	380	D								50 cm+ diamicton
93-110	12H07	517320 5458130	270	SG								5 m sandy gravel.
93-111	12H07	515460 5458480	210									Possible delta.
93-112	12H07	517030 5461330	90	SS								30 cm+ fine sand to silt. Overlain by 80 cm diamicton. Overlain by 50 cm sandy gravel.
93-113	12H07	521500 5465250	160	D								6 m diamicton.
93-114	12H07	521660 5465370	150	GS								6 m gravelly sand.
93-115	12H06	466400 5455170	110	D	7.5YR 6/2 7.5YR 3/2	80.2	18.1	1.7	1.21 2.89	934093		3 m diamicton. Overlain by 60 cm gravelly sand.
93-116	12H06	466620 5455290	110	GS								4 m exposure, slumped. Gravelly sand.
93-117	12H06	471040 5458950	108	D	10YR 6/2 10YR 3/4	78.8	19.5	1.7	-0.35 2.86	934094		75 cm+ diamicton.
93-118	12H06	471520 5460000	122	D	10YR 6/2 10YR 3/4	63.0	30.3	6.7	0.72 3.82	934095		1.5 m diamicton.
93-119	12H06	471550 5461050	120	GS								3 m interbedded sands and gravelly sand.
93-120	12H06	472160 5462950	130	D	10YR 6/2 10YR 3/4	80.2	15.1	4.8	0.56 3.09	934096		1 m+ diamicton.
93-121	12H06	473470 5461830	92	D	10YR 6/1 10YR 3/2	76.7	17.9	5.5	1.66 3.23	934097		1 m+ diamicton.
93-122	12H03	488300 5433120	320	D		91.8				934103	0.7 0.06	4m sandy diamicton.
93-123	12H06	474900 5468100	110	GS		99.1	0.9	0.0		934104		Esker. 5 m+ interbedded c., m. and f., sand; 9 cm v. fine sand to silt, and 150 cm gravelly sand.
93-123	12H06	474900 5468100	110	SS	5YR 6/3 5YR 3/3	12.9	79.6	7.6	5.67 1.58	934105		
93-123	12H06	474900 5468100	110	S		99.7	0.3	0.0		934106		

93-124	12H06	474450 5468620	102	D	2.5YR 4/8 2.5YR 3/4	80.9	16.6	2.5	0.79 3.55	934107		1 m+ diamicton.
93-125	12H06	486810 5456030	142	D	5YR 6/2 5YR 4/2	59.4	32.2	8.5	2.01 4.27	934108	0.51 0.12	Ridge. 3 m+ diamicton.
93-126	12H06	487880 5455890	133	D	10YR 7/2 10YR 4/3	65.3	32.4	2.3	1.81 3.47	934109	0.59 0.09	2m diamicton.
93-127	12H06	489370 5458030	230	D	10YR 7/2 10YR 4/3	56.3	37.0	6.7	1.56 4.06	934110	0.62 0.15	180 cm diamicton.
93-128	12H06	488950 5459170	179	D	5YR 3/4 5YR 3/2	67.9	30.4	1.7	0.45 3.61	934111	0.73 0.07	Ridge. 50 cm+ brown red diamicton; 50 cm brown diamicton, and 50 cm reddish brown diamicton.
93-128	12H06	488950 5459170	179	D	7.5YR 7/2 7.5YR 4/2	72.8	25.7	1.5	1.37 3.34	934112		
93-128	12H06	488950 5459170	179	D	2.5YR 5/4 2.5YR 3/4	60.7	34.5	4.9	2.27 3.43	934113		
93-129	12H06	491060 5465820	120	D	5YR 6/3 5YR 4/3	62.1	37.1	0.8	1.89 3.49	934114	0.58 0.1	2m diamicton.
93-130	12H06	493250 5465670	228	D	10YR 6/2 10YR 4/4	50.1	36.5	13.4	2.28 4.38	934115	0.57 0.14	1.5 m diamicton.
93-131	12H06	494730 5466600	270	D	2.5Y 5/2 2.5Y 4/2	57.9	30.2	11.9	2.46 4.01	934116	0.49 0.1	270 cm diamicton.
93-131	12H06	494730 5466600	270	D	2.5Y 6/2 2.5Y 3/2	52.5	39.2	8.3	1.5 4.17	934117		
93-132	12H06	496250 5466450	247	D	10YR 5/3 10YR 3/2	55.1	31.9	12.9	2.18 4.35	934118	0.46 0.18	150 cm. Diamicton.
93-133	12H06	498030 5466570	187	D	2.5Y 6/0 2.5Y 2/0	48.1	34.6	17.4	2.5 4.33	934119	0.58 0.08	70 cm+ grey diamicton. Overlain by 150 cm reddish grey diamicton.
93-133	12H06	498030 5466570	187	D	10YR 6/2 10YR 4/2	50.2	35.1	14.7	2.31 4.4	934120	0.62 0.08	
93-133	12H06	498030 5466570	187	D	7.5YR 6/0 7.5YR 4/0	27.8	54.2	18.1	3.54 4.32	934121		
93-134	12H06	491840 5466870	130	D	5YR 4/6 5YR 3/3	74.6	22.2	3.3	1.3 3.39	934122	0.46 0.1	180 cm diamicton.
93-135	12H06	493930 5470070	130	D	2.5YR 4/2 2.5YR 2/2	84.5	14.2	1.3	0.81 2.65	934123		160 cm diamicton.
93-136	12H06	495670 5470420	160	D	2.5YR 6/2 2.5YR 4/2	29.4	61.3	9.3	5.29 1.93	934124		190 cm. 25 cm grey-green sandstone; 50 cm reddish brown diamicton.; 50 cm sandy diamicton.
93-136	12H06	470420 5495670	160	D	5YR 6/2 5YR 4/2	87.7				934125	0.58 0.15	
93-137	12H06	496470 5471670	178	D	10YR 6/2 10YR 4/2	68.9	26.6	4.6	1.78 3.41	934126	0.62 0.06	130 cm diamicton.
93-138	12H06	497500 5473070	178	D	10YR 6/3 10YR 4/3	64.0	32.9	3.1	1.66 3.52	934127	0.49 0.18	260 cm diamicton.
93-139	12H07	503660 5474790	210	D	10YR 5/2 10YR 3/2	86.5				934128	0.59 0.08	Ridge. 1 m diamicton.
93-140	12H07	506080 5475820	182	D	2.5Y 4/4 2.5Y 3/2	88.4				934129		Ridge. 160 cm diamicton.
93-141	12H07	503140 5474870	200	D	10YR 6/2 10YR 4/2	22.6	51.6	25.9	4.29 4.4	934130	0.62 0.04	Drumlin. 320 cm+ diamicton. Overlain by 140 cm sandy diamicton, and sandy diamicton.
93-141	12H07	503140 5474870	200	D	10YR 6/2 10YR 3/2	70.9	27.3	1.9	1.76 3.17	934133	0.69 0.14	
93-141	12H07	503140 5474870	200	D	10YR 6/2 10YR 3/3	71.1	27.5	1.5	1.65 3	934134		
93-142	12H07	502340 5469890	140	D	10YR 7/1 10YR 4/1	95.8	4.2	0.0		934131	0.51 0.11	350 cm gravelly sand.
93-143	12H07	505060 5462100	110	D	10YR 7/2 10YR 4/2	64.4	32.6	3.0	2.65 2.91	934132	0.72 0.1	320 cm diamicton.
93-144	12H06	475020 5464650	101	D	10YR 6/2 10YR 3/2	75.3	21.7	3.0	1.34 2.91	934135	0.53 0.11	290 cm diamicton.
93-145	12H06	476920 5466970	100	D	10YR 7/1 10YR 3/3	71.5	24.2	4.3	1.61 3.34	934136	0.51 0.16	210 cm. Diamicton.
93-146	12H06	473980 5468820	130	D		82.2	16.5	1.4	1.78 2.32	934137	0.57 0.12	Hummock. 160 cm+ sandy diamicton. Overlain by 130 cm gravelly sand.
93-147	12H06	474470 5469170	115	D	2.5YR 4/8 2.5YR 3/6	84.2	14.1	1.7	0.42 2.86	934138	0.62 0.17	170 cm diamicton.
93-148	12H06	475540 5471680	100	D	2.5YR 5/6 2.5YR 3/4	91.3	8.0	0.7	0.24 2.14	934139	0.76 0.08	200 cm gravelly sand.
93-149	12H03	480700 5453720	80	D	10R 6/2 10R 4/2	73.7	22.3	4.0	2.15 3.2	934140	0.48 0.13	340 cm diamicton.
93-150	12H06	481070 5455100	70	D	7.5YR 8/2 7.5YR 5/3	69.7	26.3	4.0	2.01 3.36	934141	0.52 0.12	210 cm diamicton.
93-151	12H06	481650 5456730	63	D	10R 6/4 10R 4/6	84.3	13.6	2.1	1.48 2.32	934142	0.54 0.14	Ridge. 250 cm diamicton.
93-152	12H06	482530 5456370	88	D	7.5YR 7/2 7.5YR 4/2	70.1	28.6	1.2	1.32 3.39	934143		100 cm diamicton.
93-156	12A13	446900 5404350	580									Striae / grooves 290 ± 4.
93-157	12H04	460900 5444050	20	GS	7.5YR 5/2 7.5YR 3/4	98.5	1.5	0.0		934098		North Brook section. See chapter 4.
93-157	12H04	460900 5444050	20	GS		99.2	0.8	0.0		934099		
93-157	12H04	460900 5444050	20	S		99.8	0.2	0.0		934100		
93-157	12H04	460900 5444050	20	SC						934101		
93-157	12H04	460900 5444050	20	SS						934102		

S2-174	12D09	416300 5980260	244	D	7.5YR 3/0 7.5YR 2/0	75.0	25.3	1.7	1.01 3.17	934062	0.75 0.08	2 m diamicton.
S4-5	12A12	434520 5422000	M	D						94-1007	0.78 0.06	5 m+ diamicton, overlain by 7 m sand and 4 m gravelly sand.

APPENDIX 2: OVERBURDEN THICKNESS

The following listing provides data on overburden thickness across the Humber River basin, and is presented in the following format:

Place - Location where data was collected.

NTS - The National Topographic System map number.

East - The easting on the Universal Transverse Mercator Grid, Zone 21.

North - The northing on the Universal Transverse Mercator Grid, Zone 21.

Depth - Depth of overburden in metres.

Comments - Description of sediment encountered, or other pertinent data include by driller or drill core logger. The following abbreviations are used: H.F. refers to Humber Falls Formation; R.B. refers to Rocky Brook Formation; sdst - sandstone; conglom - conglomerate; slst - siltstone

Source - Source from which data was derived. References are found in the reference list. Water well data refers to the 1995 Department of Environment Report titled ‘Water Well Data for Newfoundland and Labrador 1950-1994'.

Sources identified by a map number followed by a number in parentheses e.g., 12H(709), are from assessment reports filed with the Geological Survey, Newfoundland Department of Mines and Energy, St. John's. All files are non-confidential, and may be attained by writing to the Geoscience Publications and Information Section, or telephone (709) 729-6193.

Sources designated WST refer to drillcore data provided by the Provincial Department of Works, Services and Transportation.

Place	NTS	East	North	Depth	Comments	Source
Cormack	12H06	466431	5453338	3	gravel	Water well data
Cormack	12H06	471750	5461200	7	sand	Water well data
Cormack	12H06	470741	5459207	2	clay	Water well data
Cormack	12H06	468450	5456500	9	clay	Water well data
Cormack	12H06	474075	5463700	6	clay	Water well data
Cormack	12H06	468850	5458000	8	gravel	Water well data
Cormack	12H06	466150	5452000	4	clay	Water well data
Cormack	12H06	471669	5461477	4	gravel	Water well data
Cormack	12H06	471126	5461337	15	clay (13) over gravel	Water well data
Cormack	12H06	468612	5458214	3	gravel	Water well data
Cormack	12H06	464250	5455300	50	sand and gravel (13) over sand	Water well data
Cormack	12H06	466441	5455245	5	clay, sand, gravel	Water well data
Cormack	12H06	466418	5455227	5	clay, sand, gravel	Water well data
Cormack	12H06	464250	5455175	11	clay	Water well data
Cormack	12H06	468500	5458015	2	gravel	Water well data
Cormack	12H06	468450	5457325	10	gravel	Water well data
Cormack	12H06	468491	5458470	3	gravel	Water well data
Cormack	12H06	468474	5458437	6	gravel	Water well data
Cormack	12H06	468907	5458376	3	clay	Water well data
Cormack	12H06	468922	5458361	0		Water well data
Cormack	12H06	471850	5461155	7	sand	Water well data
Cormack	12H06	471850	5461150	18	gravel	Water well data
Cormack	12H06	471427	5461234	20	gravel	Water well data
Cormack	12H06	466415	5453463	2	clay	Water well data
Cormack	12H06	468175	5456500	10	clay	Water well data
Cormack	12H06	469100	5458450	10	clay	Water well data
Cormack	12H06	473250	5462900	12	gravel	Water well data
Cormack	12H06	473000	5462800	9	clay	Water well data
Cormack	12H06	473250	5463350	5	gravel	Water well data
Cormack	12H06	468450	5458200	8	gravel	Water well data
Cormack	12H06	468650	5457700	9	clay	Water well data
Cormack	12H06	468372	5459295	40	sand	Water well data
Cormack	12H06	470875	5461075	5	clay	Water well data

Cormack	12H06	467550	5456725	3	gravel	Water well data
Cormack	12H06	472475	5462075	4	clay	Water well data
Cormack	12H06	468915	5458045	6	clay	Water well data
Cormack	12H06	468925	5458030	9	topsail	Water well data
Cormack	12H06	468916	5458040	10	topsoil	Water well data
Cormack	12H06	468286	5457673	6	gravel	Water well data
Cormack	12H06	468282	5457682	10	gravel	Water well data
Cormack	12H06	471236	5461401	6	clay	Water well data
Cormack	12H06	471212	5461398	5	clay	Water well data
Cormack	12H06	471161	5461375	21	clay	Water well data
Cormack	12H06	471245	5461467	6	clay	Water well data
Cormack	12H06	468550	5456550	9	gravel	Water well data
Corner Brook	12A13	399658	5434689	13	gravel	Water well data
Corner Brook	12A13	434093	5417765	8	gravel	Water well data
Corner Brook	12A13	412950	5399800	25	sand (7) over gravel	Water well data
Corner Brook	12A13	433969	5417661	90	sand	Water well data
Corner Brook	12A13	433756	5417704	19	clay	Water well data
Deer Lake	12H03	466540	5448491	40	sand	Water well data
Deer Lake	12H03	470000	5447500	10	gravel (8) over clay	Water well data
Deer Lake	12H03	465150	5446200	5	sand/clay	Water well data
Howley	12H03	491664	5446438	15	gravel (13) over sand	Water well data
Humber Village	12A13	444602	5426912	5	clay	Water well data
Humber Village	12A13	444151	5426075	27	sand	Water well data
Pasadena	12H04	457400	5430890	85	sand	Water well data
Pasadena	12H04	457547	5431353	38	clay (4) over sand	Water well data
Pynn's Brook	12H04	460293	5437039	38	sand. Salty at 74 m	Water well data
Pynn's Brook	12H04	460933	5438280	5	clay	Water well data
Pynn's Brook	12H04	460505	5436946	62	sand	Water well data
Pynn's Brook	12H04	461005	5438649	5	gravel	Water well data
Pynn's Brook	12H04	460401	5437253	61	gravel (6), sand (12), clay (14) over sand	Water well data
Reidville	12H03	468750	5450260	25	sand	Water well data
South Brook	12H04	448009	5426998	58	sand	Water well data
St. Judes	12H03	466300	5443800	76	sand	Water well data

Steady Brook	12A13	450132	5428091	23	sand (15) over gravel	Water well data
Cormack	12H06	471050	5464750	6.7		12H(573)
Cormack	12H06	475100	5467800	6.8		12H(576)
Cormack	12H06	469300	5462100	6.3		12H(582)
Alder Pond	12H06	495600	5479350	8.4		12H(583)
Adies Pond	12H06	470850	5481800	9.8		12H(584)
Sandy Lake	12H06	499900	5464800	17.5		12H(590)
Sandy Lake	12H06	497600	5461000	9		12H(592)
Sandy Lake	12H06	498800	5461000	19		12H(592)
Sandy Lake	12H06	497000	5458000	15.3		12H(593)
Cormack	12H06	475000	5464700	5.2		12H(597)
Sandy Lake	12H06	495800	5458800	12.4		12H(599)
Wigwam Brook	12H06	495450	5474400	9.5	5YR 3/2. Limestone chips	12H(709)
Wigwam Brook	12H06	494640	5473950	3.4	conglomerate and red mudstone chips	12H(709)
Wigwam Brook	12H06	494980	5474800	5.4	H.F. sandstone and minor rhyolite	12H(709)
Wigwam Brook	12H06	495280	5474700	4		12H(709)
Wigwam Brook	12H06	495390	5474670	8.9		12H(709)
Wigwam Brook	12H06	495560	5474790	7.8		12H(709)
Wigwam Brook	12H06	495530	5474630	4.6		12H(709)
Wigwam Brook	12H06	495590	5474600	3		12H(709)
Wigwam Brook	12H06	496180	5475100	6	granite and mudstone chips	12H(709)
Wigwam Brook	12H06	496610	5476740	4.8	H.F. sandstone chips	12H(709)
Wigwam Brook	12H06	496390	5476650	5.8		12H(709)
Wigwam Brook	12H06	496230	5476460	4.7	granite, rhyolite and H.F. sdst chips	12H(709)
Wigwam Brook	12H06	496460	5476360	5.1	granite and H.F. sdst chips	12H(709)
Wigwam Brook	12H06	496630	5476540	4.7	granite and H.F. sdst chips	12H(709)
Wigwam Brook	12H06	496910	5477000	3.7	gabbro and H.F. sdst chips	12H(709)
Wigwam Brook	12H06	496830	5477320	5.1	H.F. sdst, limestone and gabbro chips	12H(709)
Wigwam Brook	12H06	496300	5475320	5.8	H.F. sdst, rhyolite, intrusive chips	12H(709)
Wigwam Brook	12H06	495880	5474380	4.9	granite chips	12H(709)

Wigwam Brook	12H06	496500	5476850	6.5		12H(709)
Wigwam Brook	12H06	496520	5476810	6	H.F. sdst	12H(709)
Wigwam Brook	12H06	496550	5476770	6.1	H.F. sdst	12H(709)
Wigwam Brook	12H06	496570	5476720	5.5	H.F. sdst	12H(709)
Wigwam Brook	12H06	496470	5476890	8.9	R.B. H.F., gabbro	12H(709)
Wigwam Brook	12H06	496450	5476940	14.2	R.B., H.F., gabbro, granite, gneiss	12H(709)
Wigwam Brook	12H06	496420	5476980	15	felsic volcanic, sdst, conglomerate	12H(709)
Wigwam Brook	12H06	496570	5477060	11	H.F. sdst, conglom, felsic vol with qtz eyes	12H(709)
Wigwam Brook	12H06	496530	5476960	13.4	H.F. sdst and conglomerate	12H(709)
Wigwam Brook	12H06	496420	5476890	11.1	H.F. sdst and conglom, rhyolite mudstone	12H(709)
Wigwam Brook	12H06	496380	5476860	14.2	H.F. sdst, granite	12H(709)
Wigwam Brook	12H06	496460	5476920	9.3	H.F. sdst, conglomerate	12H(709)
Wigwam Brook	12H06	496440	5476960	12.3	H.F. sdst	12H(709)
Wigwam Brook	12H06	496450	5477030	17.9	H.F. sdst, mudstone	12H(709)
Wigwam Brook	12H06	496800	5477020	10.6	H.F. sdst, conglomerate	12H(709)
Wigwam Brook	12H06	496840	5476990	7.7	H.F. conglom, gabbro, limestone, granite	12H(709)
Wigwam Brook	12H06	496550	5477420	11.3	H.F. sdst, mudstone	12H(709)
Wigwam Brook	12H06	496510	5477450	11.7	Gabbro, H.F. sdst and conglom	12H(709)
Wigwam Brook	12H06	496980	5476910	22.4	gabbro, granite, lmst, R.B., H.F. sdst	12H(709)
Wigwam Brook	12H06	496940	5476940	18.9	granite, gabbro, H.F. sdst, limestone	12H(709)
Wigwam Brook	12H06	496680	5476560	7.2	granite, gabbro, sdst, mudstone	12H(709)
Wigwam Brook	12H06	493480	5473080	6.2		12H(709)
Wigwam Brook	12H06	493770	5472350	6.5	H.F. sdst, granite, limestone	12H(709)
Wigwam Brook	12H06	493500	5472200	10.4	H.F. sdst, granite, volcanics	12H(709)
Wigwam Brook	12H06	492960	5471010	9	granite, gabbro, H.F. sdst, rhyolite	12H(709)
Wigwam Brook	12H06	496440	5476490	3	H.F. sdst	12H(709)
Wigwam Brook	12H06	496040	5476400	7.6	H.F. sdst, conglom; gabbro	12H(709)
Wigwam Brook	12H06	496070	5476360	10.5	H.F. sdst	12H(709)
Wigwam Brook	12H06	496110	5476320	7.9		12H(709)

Wigwam Brook	12H06	496140	5476290	7.9	H.F. sdst	12H(709)
Wigwam Brook	12H06	496170	5476210	9.6	sdst	12H(709)
Wigwam Brook	12H06	496200	5476160	11.1	H.F. sdst, gabbro	12H(709)
Wigwam Brook	12H06	496330	5476030	8.3	granite, H.F. sdst	12H(709)
Wigwam Brook	12H06	495760	5475770	7.4	H.F. sdst, rhyolite, quartz (conglom?)	12H(709)
Wigwam Brook	12H06	496060	5475550	3.9	sdst, gabbro, granite	12H(709)
Wigwam Brook	12H06	495900	5475130	3.1	rhyolite	12H(709)
Wigwam Brook	12H06	495570	5475310	6	conglom, R.B. slst, H.F. sdst	12H(709)
Wigwam Brook	12H06	496590	5477340	11.7	H.F. sdst, conglom	12H(709)
Wigwam Brook	12H06	496620	5477290	9	H.F. sdst, granite, gabbro	12H(709)
Wigwam Brook	12H06	496650	5477250	10.6	volcanics, H.F. sdst	12H(709)
Wigwam Brook	12H06	496690	5477210	7.7	granite, sdst	12H(709)
Wigwam Brook	12H06	496720	5477160	9	H.F. sdst	12H(709)
Wigwam Brook	12H06	477200	5477370	14.4	gabbro, H.F. sdst, granite, rhyolite. Red brown to grey red till	12H(709)
Wigwam Brook	12H06	497190	5477380	14.8	gabbro, granite, H.F. sdst. Grey red sandy to red sandy clay till	12H(709)
Wigwam Brook	12H06	495630	5474730	5.8	felsic volcanic	12H(709)
Wigwam Brook	12H06	495670	5474530	3.3	granite, rhyolite, sdst	12H(709)
Wigwam Brook	12H06	494560	5475250	4.7	gabbro, sdst	12H(709)
Wigwam Brook	12H06	494860	5475260	6.4	H.F. sdst, biotite granite	12H(709)
Wigwam Brook	12H06	495710	5474590	5.7	HF sdst, NB lmst, R.B. shale	12H(709)
Wigwam Brook	12H06	495690	5474510	3		12H(709)
Wigwam Brook	12H06	493480	5473080	8.1	granite, gabbro, HF sdst	12H(709)
Wigwam Brook	12H06	493770	5472350	12	granite, rhyolite, HF sdst, mafic volcanic	12H(709)
Wigwam Brook	12H06	493500	5472200	8	HF sdst	12H(709)
Wigwam Brook	12H06	492960	5471010	6.3	HF sdst, gabbro, RB mudstone. Sandy pale red 5Y 6/2 till	12H(709)
Wigwam Brook	12H06	496200	5475050	5	granite, NB conglom	12H(709)
Wigwam Brook	12H06	496360	5475240	5		12H(709)
Wigwam Brook	12H06	496330	5475210	4		12H(709)
Wigwam Brook	12H06	495780	5474470	5.5	gabbro, granite	12H(709)
Wigwam Brook	12H06	495840	5474430	7.4	HF sdst, gabbro, slate, granite	12H(709)

Wigwam Brook	12H06	495880	5474420	6.9	HF sdst, Gales Brook granite, NB conglom	12H(709)
Wigwam Brook	12H06	495750	5474500	3.5		12H(709)
Wigwam Brook	12H06	495760	5474490	5.4	red brown clay till	12H(709)
Cormack	12H06	465500	5455500	12.1		12H(706)
Junction Brook	12H03	478200	5448350	13.7		12H(640)
Hinds Brook	12H03	484660	5435550	2.1	granite	12H(226)
Hinds Brook	12H03	484550	5435500	7.9	clay, sand and andesite boulder	12H(226)
Hinds Brook	12H03	484550	5435270	8.5		12H(226)
Hinds Brook	12H03	484000	5434740	7.3		12H(226)
Hinds Brook	12H03	484200	5434720	3	andesite bo and sand	12H(226)
Hinds Brook	12H03	484020	5434680	1.5		12H(226)
Wigwam Brook	12H06	497100	5476800	9.5	average of 11 holes	12H(1114)
Wigwam Brook	12H06	496500	5476200	4.6	average of 11 holes	12H(1114)
Wigwam Brook	12H06	493200	5472800	2.7	average of 21 holes	12H(549)
Glover Island	12A12	442500	5394600	6.7	average of 5 holes	12A(638)
Glover Island	12A12	442100	5394600	4.6	average of 2 holes	12A(638)
Glover Island	12A12	441800	5394300	5.5	average of 8 holes	12A(638)
Glover Island	12A12	444300	5396400	3.0	average of 2 holes	12A(638)
Glover Island	12A12	445300	5397500	4.4	average of 3 holes	12A(638)
Corner Brook	12A13	430800	5421680	4	average of 10 holes. Mostly silt and sand	NGL 2019
Deer Lake	12H03	469250	5448200	2.5	average 5 pits. No bedrock.	PWC, 1989
Corner Brook	12A13	429480	5422850	11.3	average 5 holes. Sand and gravel	Golder, 1983
Corner Brook	12A13	429420	5422740	8	average 5 holes. Sand and gravel	Golder, 1983
Corner Brook	12A13	431000	5423250	12.5	average 5 holes. Sand and gravel	Geotechnical Associates, 1979
Deer Lake	12H03	469200	5448200	5	average 6 holes	Golder, 1980
Corner Brook	12A13	430000	5422360	8.3	average 8 holes. Sand, gravel, clay	Golder, 1991
Corner Brook	12A13	431330	5420900	4	average 5 pits. Till	Golder, 1986
Corner Brook	12A13	427800	5419050	3.6	average 3 holes. Till	Geotech, 1980
Corner Brook	12A13	431550	5420200	2.4	average 24 holes. Till	Newfoundland Design, 1970
Corner Brook	12A13	430720	5421070	6.4	grey till. Borehole 1	Newfoundland Design, 1968
Corner Brook	12A13	430530	5420960	6.1	grey till. borehole 5	Newfoundland Design, 1968
Corner Brook	12A13	430660	5420770	6.4	grey till. Borehole 10	Newfoundland Design, 1968

Corner Brook	12A13	430470	5420780	4	grey till. No rock. Borehole 12	Newfoundland Design, 1968
Corner Brook	12A13	430580	5420550	6.7	grey till. Borehole 19	Newfoundland Design, 1968
Corner Brook	12A13	430670	5421270	3.4	grey till. Borehole 22	Newfoundland Design, 1968
Corner Brook	12A13	430370	5421220	4	grey till. No rock. Borehole 24	Newfoundland Design, 1968
Corner Brook	12A13	432950	5421750	4.7	till. Average 3 boreholes	Geotechnical Associates, 1973
Corner Brook	12A13	433000	5421770	1.9	till. Average 3 boreholes	Geotechnical Associates, 1973
Corner Brook	12A13	427150	5419550	4.6	Till. No bedrock. Borehole 1	Geotechnical Associates, 1980
Corner Brook	12A13	427250	5419560	3.6	Till. No bedrock. Borehole 2	Geotechnical Associates, 1980
Corner Brook	12A13	427400	5419540	3.6	Till. No bedrock. Borehole 3	Geotechnical Associates, 1980
Corner Brook	12A13	427420	5419520	3.6	Till. No bedrock. Borehole 4	Geotechnical Associates, 1980
Corner Brook	12A13	427680	5419550	4.6	Till. No bedrock. Borehole 5	Geotechnical Associates, 1980
Corner Brook	12A13	427700	5419430	4.6	Till. No bedrock. Borehole 6	Geotechnical Associates, 1980
Corner Brook	12A13	427820	5419580	4.6	Till. No bedrock. Borehole 7	Geotechnical Associates, 1980
Corner Brook	12A13	428000	5419570	4.6	Till. No bedrock. Borehole 8	Geotechnical Associates, 1980
Corner Brook	12A13	428160	5419630	3.6	Till. No bedrock. Borehole 9	Geotechnical Associates, 1980
Corner Brook	12A13	428280	5419620	4.6	Till. No bedrock. Borehole 10	Geotechnical Associates, 1980
Corner Brook	12A13	427240	5419780	4.6	Till. No bedrock. Borehole 11	Geotechnical Associates, 1980
Corner Brook	12A13	427350	5419860	4.6	Till. No bedrock. Borehole 12	Geotechnical Associates, 1980
Corner Brook	12A13	427420	5419800	4.6	Till. No bedrock. Borehole 13	Geotechnical Associates, 1980
Corner Brook	12A13	427480	5419750	4.6	Till. No bedrock. Borehole 14	Geotechnical Associates, 1980
Corner Brook	12A13	427420	5419920	4.6	Till. No bedrock. Borehole 15	Geotechnical Associates, 1980
Corner Brook	12A13	427150	5419270	4.6	Till. No bedrock. Borehole 16	Geotechnical Associates, 1980
Corner Brook	12A13	427450	5419150	4.6	Till. No bedrock. Borehole 17	Geotechnical Associates, 1980
Corner Brook	12A13	427660	5419170	4.6	Till. No bedrock. Borehole 18	Geotechnical Associates, 1980
Corner Brook	12A13	427920	5421000	3	Till. No bedrock. Average 6 pits	NGL, 1989
Corner Brook	12A13	435840	5417880	4.3	Till?. Borehole 1	Geotechnical Associates, 1976
Corner Brook	12A13	435740	5418050	3.4	Till?. Borehole 2	Geotechnical Associates, 1976
Corner Brook	12A13	435660	5418180	6.1	Till?. Borehole 3	Geotechnical Associates, 1976
Corner Brook	12A13	435620	5417300	1.2	Till?. Borehole 4	Geotechnical Associates, 1976
Corner Brook	12A13	435550	5417420	4.3	Till?. Borehole 5	Geotechnical Associates, 1976
Corner Brook	12A13	435320	5417700	2.1	Till?. Borehole 7	Geotechnical Associates, 1976
Corner Brook	12A13	435240	5417700	2.1	Till?. Borehole 8	Geotechnical Associates, 1976
Corner Brook	12A13	435170	5417780	2.1	Till?. Borehole 9	Geotechnical Associates, 1976
Corner Brook	12A13	435000	5417920	1.8	Till?. Borehole 10	Geotechnical Associates, 1976

Corner Brook	12A13	434900	5419040	1.8	Till?. Borehole 11	Geotechnical Associates, 1976
Corner Brook	12A13	434720	5419130	1.5	Till?. Borehole 12	Geotechnical Associates, 1976
Corner Brook	12A13	434520	5419260	1.2	Till?. Borehole 13	Geotechnical Associates, 1976
Corner Brook	12A13	432750	5422850	15.8	Borehole 2. No bedrock. Till?	WST, 1991
Steady Brook	12A13	437600	5422150	25.6	10 m sandy gravel over silt-clay. 1963 hold showed 38 m.	WST, 1990
Steady Brook	12A13	437750	5422230	18.3	No bedrock. Sandy gravel	WST, 1990
Corner Brook	12A13	435870	5421940	25.3	5.9 m silty clay over sand and gravel over limestone	WST, 1991
Corner Brook	12A13	435850	5422030	27.3	10.9 m silty clay over sand and sand-gravel over limestone	WST, 1991
Little Rapids	12A13	444170	5425570	20.4	No bedrock. Clay.	WST, 1994
Little Rapids	12A13	447300	5426700	30.4	Est. location. 3 m gravelly sand over clay	WST, 1994
Little Rapids	12A13	444220	5425480	12.9	Borehole 1. Mostly sand. No bedrock.	WST, 1995
Steady Brook	12A13	443980	5425450	29.5	Borehole 3. 15.2 m sand and gravel over clay. No bedrock.	WST, 1995
Deer Lake	12H03	471170	5451300	54.9	23 m silty sand over 27 m silt-clay over 5 m sand-gravel.	Environment Canada, 1980
Deer Lake	12H03	471200	5451200	60.9	No bedrock. silty clay to sandy clay	Environment Canada, 1980
Steady Brook	12A13	438700	5421950	120	mostly silt-clay	Golder, 1983
Howley	12H03	491600	5443900	15.2	mostly sand, some clay near base. Location est.	Murray and Howley, 1881
Howley	12H03	493500	5445500	13.1	No bedrock. Sand/gravel. Location est.	Murray and Howley, 1881
Howley	12H03	492400	5443850	35.1	6 m clay overlain by sand-gravel. Location est.	Murray and Howley, 1918

APPENDIX 3: CLAST ROCK TYPE DETERMINATIONS FROM DIAMICTON SAMPLES

Data presented here is of clast rock type identification from 351 samples across the Humber River basin. Most samples are diamicton. A field description of sites from which samples were collected is found in Appendix 1.

Clasts were identified in the field, and subdivided into rock type or, where positively identified, into source areas. Specific identification was dominantly to rock types outcropping on The Topsails, and terminology follows that of Whalen and Currie (1988). Where no source could be identified, only the rock type is recorded. The reader is referred to the section Introduction to the Geology of the Humber River Basin for a complete discussion of the bedrock geology of the Humber River basin.

A total of 19 723 clasts were identified, and subdivided into 41 categories. In the following tables rock types or groups have been abbreviated. These are described below:

Sample	The sample number assigned. See Appendix 1 for details.
Site	The site number assigned. See Appendix 1 for details.

Clasts identified to a specific area

Sq, Sm, Sp, Ssy	Topsails Intrusive Suite clasts using the nomenclature taken from Whalen and Currie (1988).
Tps	Clasts from The Topsails but not assigned to a specific area.
Ss, Ssm, Ssf	Springdale Group clasts with nomenclature taken from Whalen and Currie (1988).
Oi, Oib, Oic, Oid	Ordovician granite and granodiorite with nomenclature taken from Whalen and Currie (1988).
Ohm	Rocks of the Hungry Mountain Complex with nomenclature taken from Whalen and Currie (1988).
MM	Mount Musgrave Group. Nomenclature from Williams and Cawood (1989).
HL	Hughes Lake Complex. Nomenclature from Williams and Cawood (1989).

Rock types not assigned to a specific area:

Shl	Shale
Slst	Siltstone
Sdst	Sandstone
Arks	Arkose
Cng	Conglomerate
Lst	Limestone
Dlmt	Dolomite
Scst	Schist
Gnss	Gneiss
Grnt	Granite
Gbr	Gabbro
Ultr	Ultramafics
Tuf	Tuff
Rhy	Rhyolite
Bslt	Basalt
Qzte	Quartzite
Mtsd	Metasediment
Mar	Marble
Am	Amphibolite
Vl br	Volcanic breccia
Ac vl	Acid volcanic
Gngr	Granite/granodiorite
Prph	Porphyry
Qtz peb	Quartz pebble
Unk	Unknown
Tot	Total

Sample and site data is found within Appendix 1, except for the following. These are samples originally collected by Doug Vanderveer, Department of Mines and Energy, but pebble contents are identified as part of this research. All data is unpublished.

Sample	Site #	NTS	Easting	Northing
860725	80012	12H03	471800	5449680
860726	80012	12H03	471800	5449680
860727	80012	12H03	471800	5449680
860728	80020	12H03	477670	5452570
860729	80020	12H03	477670	5452570
860730	80023	12H03	480760	5452730
860731	80023	12H03	480760	5452730
860732	80024	12H03	481730	5451920
860734	80027	12H03	485860	5454540
860741	80128	12H06	491030	5456250
860742	80130	12H06	493300	5456680
860743	80130	12H06	493300	5456680
860744	86002	12H03	465930	5454970
860747	86001	12H03	472100	5449480
860749	80007	12H03	472870	5453770
864501	80128	12H06	491030	5456250
864502	80128	12H06	491030	5456250
864503	80128	12H06	491030	5456250
864504	80012	12H03	471800	5449680
864505	80012	12H03	471800	5449680
870001	80020	12H03	477670	5452570
870002	80020	12H03	477670	5452570
870003	80020	12H03	477670	5452570
870004	80020	12H03	477670	5452570
870005	80128	12H06	491030	5456250
870007	80128	12H06	491030	5456250
870008	80128	12H06	491030	5456250
870009	80012	12H03	471800	5449680
870010	80012	12H03	471800	5449680
870020	80130	12H06	493300	5456680
870021	80130	12H06	493300	5456680