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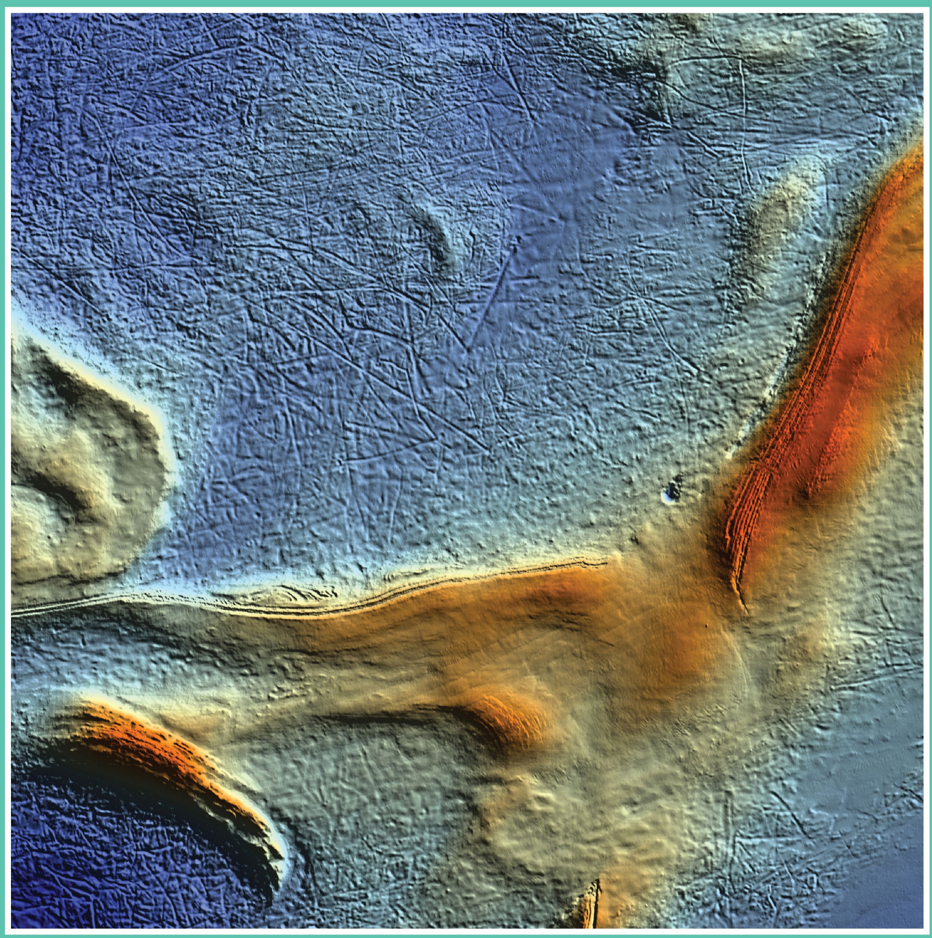

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J. Shaw and D.P. Potter

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Cover illustration
Multibeam-sonar image showing complex bedrock structure on the seafloor, outer St. George's Bay, Newfoundland

Critical review
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Abstract

This bulletin is a synthesis of several decades of Geological Survey of Canada research into the surficial geology of the inner continental shelf around the Island of Newfoundland, and the geomorphology of adjacent coastlines. Based on analyses of data from multibeam-sonar surveys, other marine surveys, and from coastal mapping, the island is divided into fourteen regions, each with distinctive surficial geology and coastal geomorphology. The evidence of glaciation is dominant in all regions, and includes: 1) streamlined landforms formed in the onset areas of ice streams; 2) arcuate submarine moraines formed by glacial standstills at the mouths of west and south coast fiords; 3) transverse moraines formed by standstills within fiords; and 4) swarms of De Geer moraines formed during ice retreat. The seafloor of the inner shelf is imprinted by relict iceberg furrows and pits in all fourteen regions, but a modern population is recognized only off the east and northeast coasts. Evidence of lowered postglacial sea levels is found in submerged deltas and wave-cut platforms at various depths off the southwest and south coasts. The zone of modern sediment mobility is relatively wide on the shallow inner shelves off southwest and northeast Island of Newfoundland, and relatively narrow elsewhere. The multibeam-sonar imagery reveals a wide range of submarine landforms and processes, including active submarine fans in west coast fiords, an inner-shelf submarine canyon off the southwest coast, and large-scale erosion of the seafloor in Placentia Bay and off Bay of Islands. Anthropogenic modification of the seafloor is most intense near Corner Brook (Bay of Islands) and in parts of Placentia Bay, and includes submarine landslides triggered by port construction.

Résumé

Le présent bulletin est une synthèse de plusieurs décennies de recherches menées par la Commission géologique du Canada sur la géologie des matériaux superficiels de la plate-forme continentale interne ceinturant l'île de Terre-Neuve, et sur la géomorphologie du littoral adjacent. D'après des analyses de données de levés réalisés par sonar multifaisceaux, d'autres types de levés marins et de la cartographie du littoral, l'île est divisée en quatorze régions, chacune dotée d'une géologie des matériaux superficiels et d'une géomorphologie côtière caractéristiques. Des signes évidents des glaciations sont présents dans toutes les régions, notamment : 1) des formes de relief fuselées dans les secteurs d'origine des courants glaciaires; 2) des moraines sous-marines de forme arquée formées lors de haltes glaciaires aux embouchures des fjords des côtes ouest et sud; 3) des moraines transversales formées lors de haltes glaciaires à l'intérieur des fjords; et 4) des essaims de moraines de De Geer formées durant le recul du front glaciaire. Des sillons et des cavités reliques creusés par des icebergs marquent le fond marin de la plate-forme continentale interne dans l'ensemble des quatorze régions, mais une population récente n'est reconnue qu'au large des côtes est et nord-est. Des signes de niveaux de la mer plus bas durant la période postglaciaire sont livrés par des deltas submergés et des plates-formes érodées par les vagues à diverses profondeurs au large des côtes sud-ouest et sud. La zone de mobilité récente des sédiments est relativement large dans les milieux peu profonds de la plate-forme continentale interne au large des côtes sud-ouest et nord-est de l'île de Terre-Neuve, alors qu'elle est relativement étroite ailleurs. Les images captées par sonar multifaisceaux révèlent une grande diversité des formes de relief et des processus sous-marins, dont des cônes sous-marins actifs dans les fjords de la côte ouest, un canyon sous-marin dans la plate-forme continentale interne au large de la côte sud-ouest, et une érosion à grande échelle du fond marin dans la baie Placentia et au large de la Bay of Islands. Les modifications d'origine anthropique du fond marin sont plus intenses près de Corner Brook (Bay of Islands) et dans des parties de la baie Placentia, et elles comprennent des glissements sous-marins déclenchés par la construction du port.

SUMMARY

Multibeam-sonar data collected since 1996, together with geophysical and sampling data, are used to describe the geomorphology and surficial geology of the inner shelf adjacent to the coasts of the Island of Newfoundland. Fourteen regions are discussed defined on the basis of geomorphology and surficial materials.

St. Georges Bay

In the inner bay copious deposits of acoustically stratified glaciomarine sediments that infill glacially overdeepened valleys are overlain by gas-charged postglacial mud. A submarine moraine extending across the bay was subaerially exposed during the early Holocene relative sea-level lowstand. Eroding Quaternary deposits at the coast nourish strand plains and their associated subaqueous platforms. In the middle and outer bay, the Quaternary cover is largely restricted to northeast-trending valleys, and Carboniferous bedrock ridges with highly complex structure are draped by an iceberg-turbated glaciomarine veneer.

Cape Anguille to Cape Ray

On a gently shelving inner shelf extensive bedrock exposures are separated by thin bodies of mobile sand. Some inner shelf sand bodies extend to the coast and link with sandy beaches and dunes. A large, active submarine canyon just west of Cape Ray connects with the Laurentian Channel.

Cape Ray to Bay de Loup

Crystalline basement rocks form a shallow, irregular, wave-dominated platform with fluted morainal deposits on its outer edge. Fiords contain submerged deltas at their heads and thick glaciomarine and postglacial mud deposits in deep water.

Bay de Loup to Hermitage Bay

Fiords up to 790 m deep have arcuate moraines at their mouths dating to ca. 14 ka. Submerged deltas at the fiord heads formed during the early Holocene relative sea-level lowstand. Offshore, Carboniferous bedrock is overlain by patches of glacial diamict, draped glaciomarine mud, and a thick blanket of postglacial mud.

SOMMAIRE

Les données acquises par sonar multifaisceaux depuis 1996, jumelées à des données géophysiques et des données d'échantillonnage, sont utilisées pour décrire la géomorphologie et la géologie des matériaux superficiels de la plate-forme continentale interne adjacente au littoral de l'île de Terre-Neuve. Nous traitons de quatorze régions définies en fonction de leur géomorphologie et de leurs matériaux superficiels.

Baie St. Georges

Dans la partie interne de la baie, d'amples dépôts de sédiments glaciomarins stratifiés en vitesse acoustique remplissent des vallées surcreusées par le passage des glaciers et sont recouverts de boue postglaciaire chargée de gaz. Une moraine sous-marine, s'étendant en travers de la baie, a été exposée dans des conditions subaériennes lors de l'épisode de bas niveau relatif de la mer de l'Holocène précoce. L'érosion des dépôts du Quaternaire sur la côte nourrit les plaines d'estran et leurs plates-formes subaquatiques associées. Dans les parties médiane et externe de la baie, les sédiments de couverture du Quaternaire sont en grande partie circonscrits à des vallées de direction nord-est, et des crêtes du substratum rocheux du Carbonifère, présentant des structures très complexes, sont moulées par un placage de sédiments glaciomarins modifiés par l'action des icebergs.

Du cap Anguille au cap Ray

Sur une plate-forme continentale interne à pente douce, de vastes affleurements du substratum rocheux sont séparés par de minces accumulations de sable mobile. Certaines de ces accumulations sableuses de la plate-forme continentale interne s'étendent jusqu'à la côte et présentent un passage continu aux plages sablonneuses et aux dunes. Un large canyon sous-marin actif, juste à l'ouest du cap Ray, se connecte au chenal Laurentien.

Du cap Ray à la baie de Loup

Des roches du socle cristallin forment une plate-forme peu profonde, irrégulière et soumise à l'action dominante des vagues, qui présente à sa bordure externe des dépôts morainiques cannelés. Des fjords renferment des deltas submergés à leur extrémité amont et, en eau profonde, d'épais dépôts de boue glaciomarine et postglaciaire.

De la baie de Loup à la baie Hermitage

Des fjords dont la profondeur peut atteindre 790 m présentent à leur embouchure des moraines arquées qui remontent à environ 14 ka. Des deltas submergés à l'extrémité amont des fjords se sont formés pendant l'épisode de bas niveau relatif de la mer de l'Holocène précoce. Au large, le substratum rocheux du Carbonifère est recouvert de plaques de diamicton glaciaire, de boue glaciomarine moulant son substrat et d'une épaisse nappe de boue postglaciaire.

Fortune Bay

This region encompasses the highly indented rocky coast of the area informally known as 'Hermitage peninsula', the more linear steep rocky coasts extending to the head of Fortune Bay, and the low-relief coasts of the Burin Peninsula, with till outcrops and several substantial sand and gravel barrier-beach complexes. Offshore, a deep channel extending to the southwest contains thick Quaternary deposits.

Baie Fortune

Cette région renferme la côte rocheuse très échancrée d'un secteur auquel on réfère informellement par le nom de «péninsule Hermitage», les côtes rocheuses raides et plus linéaires s'étendant jusqu'au fond de la baie Fortune, ainsi que les côtes de plus faible relief de la péninsule Burin, avec des affleurements de till et plusieurs importants complexes de cordons littoraux de sable et de gravier. Au large, un chenal profond s'étend vers le sud-ouest et contient d'épais dépôts du Quaternaire.

Placentia Bay

In the upper bay, deep channels separated by islands host thick Quaternary sediments, with bedrock predominant on channel flanks. The wide and shallow outer bay contains glaciomarine mud overlain by pockmarked postglacial mud. Notable aspects of the bay are: a megaflute zone off the coast of the Avalon Peninsula; streamlined glacial landforms in the southwest, with superimposed moraines; a wide, submerged littoral platform on the east side of the bay, formed during and after the postglacial relative sea-level lowstand; coastal inlets containing thick postglacial mud deposits, with strong anthropogenic impacts; and the gravel strand plain at Placentia that formed over the past several millennia in a regime of rising relative sea level.

Baie Placentia

Dans la partie interne de la baie, des chenaux profonds séparant des îles renferment d'épais sédiments du Quaternaire et leurs flancs sont surtout formés du substratum rocheux. La partie externe de la baie, large et peu profonde, contient de la boue glaciomarine recouverte de boue postglaciaire criblée de dépressions d'échappement de gaz. Les aspects notables de la baie sont les suivants : une zone à mégacannelures au large de la presqu'île Avalon; des reliefs glaciaires fuselés au sud-ouest auxquels se superposent des moraines; une plate-forme littorale large et submergée du côté est de la baie, formée pendant et après la période de bas niveau relatif de la mer durant la période postglaciaire; des bras de mer côtiers contenant d'épais dépôts de boue postglaciaire, qui portent la trace d'importants impacts anthropiques; et la plaine d'estran graveleuse à Placentia, qui s'est formée sur plusieurs millénaires dans un régime de hausse du niveau relatif de la mer.

Avalon Peninsula

Coasts are steep and rocky, with sandy substrates immediately offshore. Modern iceberg impact has pitted the seafloor farther offshore in places, forming furrows and pits. St. John's Harbour contains postglacial mud, together with an early Holocene submerged delta the depth of which was determined by a sill at the harbour entrance. By contrast, the early Holocene submerged delta at Holyrood (Conception Bay) and an adjacent submerged erosional platform register the postglacial relative sea-level lowstand depth.

Presqu'île Avalon

Les côtes sont raides et rocheuses, et on trouve des substrats sablonneux immédiatement au large. Les impacts d'icebergs récents ont marqué le fond marin à plus grande distance des côtes à certains endroits, en formant des sillons et des cavités. Le port de St. John's contient de la boue postglaciaire, ainsi qu'un delta submergé de l'Holocène précoce, dont la profondeur a été déterminée par celle d'un seuil à l'entrée du port. En revanche, le delta submergé de l'Holocène précoce à Holyrood (baie Conception) et une plate-forme d'érosion submergée adjacente rendent compte de la profondeur du bas niveau relatif de la mer dans la période postglaciaire.

Trinity Bay to Bonavista Bay

The coastline is indented by a series of fiords hosting thick deposits of glaciomarine and postglacial mud. Inner fiords have an unusual pitted topography comparable to terrestrial dead-ice topography. Outside the fiords, streamlined glacial landforms indicate the movement of glacier ice into Trinity Bay, a former ice-stream conduit.

De la baie Trinity à la baie Bonavista

La côte est échancrée par une série de fjords qui renferment d'épais dépôts de boue glaciomarine et postglaciaire. La partie interne des fjords présente le relief inhabituel d'un fond marin criblé de cavités comparable à celui produit par la glace morte en milieu continental. À l'extérieur des fjords, les reliefs glaciaires fuselés rendent compte du mouvement de la glace de glacier dans la baie Trinity, un ancien conduit de courant glaciaire.

Cape Freels to Dog Bay

The coast is sandy, with dune-ridge forelands, beaches, and coastal dunes with radiocarbon ages extending back several millennia. The relatively wide, shallow inner shelf is characterized by bedrock outcrops and expansive sheets of mobile sand and gravel. In deeper water, thin glaciomarine and postglacial sediments are imprinted by iceberg furrows and pits; modern iceberg impacts in shallower water are effaced by waves.

Du cap Freels à la baie Dog

La côte est sablonneuse, avec des arrière-plages où percent des crêtes de dune, des plages et des dunes côtières dont les âges radiocarbone remontent à plusieurs millénaires. La plate-forme continentale interne peu profonde et relativement large est caractérisée par des affleurements du substratum rocheux et de vastes nappes de sable et de gravier mobiles. Dans les eaux plus profondes, de minces sédiments glaciomarins et postglaciaires portent les traces de sillons et de cavités produits par des icebergs; les impacts d'icebergs récents dans les eaux moins profondes sont effacés par les vagues.

Notre Dame Bay

Numerous fiords extending northeast into Notre Dame Bay host thick deposits of Quaternary sediments, including transverse moraines located at fiord constrictions. The littoral fringe is heavily impacted by icebergs, resulting in a furrowed and pitted seafloor, with postglacial mud in depressions and gravelly substrates in surrounding areas. Outside the fiords, at the boundary between crystalline basement rocks and younger strata, streamlined glacial landforms form a convergent pattern consistent with the onset area of a former ice stream in Notre Dame Channel.

Baie Notre Dame

De nombreux fjords s'étendant vers le nord-est dans la baie Notre Dame renferment d'épais dépôts de sédiments du Quaternaire, dont des moraines transversales dans les zones de resserrement des fjords. La frange littorale est intensément frappée par les icebergs, ce qui produit un fond marin criblé de sillons et de cavités où de la boue postglaciaire occupe les cavités, alors que les zones environnantes sont constituées d'un substrat graveleux. À l'extérieur des fjords, à la limite entre les roches du socle cristallin et les strates plus récentes, des reliefs glaciaires fuselés forment une configuration convergente compatible avec l'existence du secteur d'origine d'un ancien courant glaciaire dans le chenal Notre Dame.

White Bay to the Strait of Belle Isle

Sand and gravel deposits are located between bedrock outcrops in shallow water, and thick Quaternary deposits are found in deeper water. A series of relatively small, fiord-like inlets contain deposits of glaciomarine and postglacial mud. Streamlined glacial landforms in several areas are consistent with the flow of glacier ice out of White Bay toward Notre Dame Bay. The seafloor is heavily imprinted by iceberg furrows and pits, both relict and modern.

De la baie White au détroit de Belle Isle

Des dépôts de sable et de gravier sont présents entre des affleurements du substratum rocheux dans les eaux peu profondes et l'on trouve d'épais dépôts du Quaternaire dans les eaux plus profondes. Une série de bras de mer relativement petits, ressemblant à des fjords, contiennent des dépôts de boue glaciomarine et postglaciaire. Des reliefs glaciaires fuselés en plusieurs endroits sont compatibles avec l'écoulement d'un glacier hors de la baie White en direction de la baie Notre Dame. Le fond marin est fortement marqué par des sillons et des cavités produits par des icebergs, anciens et récents.

Strait of Belle Isle to Bonne Bay

This region is characterized by extremely low relief in the coastal regions and gentle slopes in the offshore. Relative sea level has been falling since deglaciation. No new multibeam data were available for this publication. Pre-existing maps show a current-swept seafloor with sandy bedforms, bedrock outcrops, lag gravels, and fields of minor moraines.

Du détroit de Belle Isle à la baie Bonne

Cette région est caractérisée par un relief extrêmement faible dans les régions côtières et des pentes douces au large. Le niveau relatif de la mer diminue depuis la déglaciation. Aucune nouvelle donnée de levés par sonar multifaisceaux n'était disponible pour cette publication. Les cartes existantes montrent un fond marin balayé par les courants avec des figures sédimentaires de fond sableuses, des affleurements du substratum rocheux, des graviers de déflation et des champs de moraines mineures.

Bonne Bay to Port au Port Bay

Several major fiords contain thick packages of glaciomarine sediments overlain by postglacial mud, with large arcuate submarine moraines at fiord mouths. Where the moraines reach the coast at the entrances to Bonne Bay and Bay of Islands, submarine fans actively transport sand into deep water. Whereas Bonne Bay is relatively pristine, Bay of Islands shows many effects of human activity. The inner shelf outside the fiords is wave dominated, with sand and gravel, whereas farther offshore the seafloor is strongly imprinted by relict iceberg furrows.

Port au Port Bay

Port au Port Bay contains relatively thick deposits of glaciomarine mud, gas-charged postglacial sediments, and a large submerged early Holocene delta. Outside the bay the inner shelf is distinguished by linear bedrock ridges in the north and complex bedrock topography in the south.

The dominant imprint in all regions is that of glaciation. In Placentia Bay, Bonavista Bay, and Notre Dame Bay streamlined glacial landforms indicate converging and accelerating flow of glacier ice into troughs on the continental shelf. On the south and west coasts, submarine moraines at fiord mouths show that ice margins became established at modern coasts at ca. 14 ka (conventional radiocarbon years). Whereas fiords on the southeast, east, and north coasts do not have moraines at their mouths, they do contain transverse moraines located at constrictions.

In general the data confirm the large spatial variations in postglacial relative sea levels described in previous work, with minor changes in some areas, notable southern Placentia Bay. In all regions relict furrows and grounding pits are observed, but in regions off the east and northeast coasts modern impact is also evident at the seafloor. Changes in postglacial sedimentation patterns indicate changes in ocean circulation in the Gulf of St. Lawrence and Placentia Bay. Anthropogenic modification of the seafloor is most intense in Bay of Islands and parts of Placentia Bay.

De la baie Bonne à la baie Port au Port

Plusieurs fjords d'importance contiennent d'épais ensembles de sédiments glaciomarins recouverts de boue postglaciaire et l'on trouve de vastes moraines sous-marines arquées à l'embouchure des fjords. Aux endroits où les moraines atteignent la côte aux entrées de la baie Bonne et de la Bay of Islands, des cônes sous-marins assurent le transport actif de sable dans les eaux profondes. Bien que la baie Bonne ait relativement été conservée dans son état naturel, la Bay of Islands montre de nombreux effets de l'activité humaine. La plate-forme continentale interne, à l'extérieur des fjords, est soumise à l'action dominante des vagues, avec un couvert de sable et de gravier, tandis que plus loin au large des côtes le fond marin est fortement marqué par des sillons d'icebergs reliques.

Baie Port au Port

La baie Port au Port contient des dépôts relativement épais de boue glaciomarine et de sédiments postglaciaires chargés de gaz ainsi qu'un vaste delta submergé de l'Holocène précoce. À l'extérieur de la baie, la plate-forme continentale interne se distingue par des crêtes linéaires du substratum rocheux au nord, et une topographie complexe du substratum rocheux au sud.

L'empreinte dominante dans toutes les régions est celle des glaciations. Dans les baies Placentia, Bonavista et Notre Dame, des reliefs glaciaires fuselés sont des indicateurs de directions convergentes de l'écoulement des glaces et de l'accélération de celles-ci dans des cuvettes de la plate-forme continentale. Sur les côtes sud et ouest, des moraines sous-marines à l'embouchure des fjords montrent que les marges glaciaires se sont installées le long des côtes actuelles à environ 14 ka (selon la datation au radiocarbone classique). Bien que les fjords sur les côtes sud-est, est et nord ne présentent pas de moraines à leurs embouchures, ils contiennent néanmoins des moraines transversales dans les zones de rétrécissement.

En règle générale, les données confirment la grande variation spatiale des niveaux relatifs de la mer dans la période postglaciaire décrits dans les études antérieures, avec quelques changements mineurs en certains endroits, notamment dans la partie sud de la baie Placentia. Dans toutes les régions, on observe des sillons et des cavités d'échouage d'icebergs reliques, mais dans les régions au large des côtes est et nord-est, des impacts récents sont également évidents sur le fond marin. Les changements dans les configurations de dépôt postglaciaires indiquent également des changements dans la circulation océanique dans le golfe du SaintLaurent et la baie Placentia. La modification anthropique du fond marin est plus intense dans la Bay of Islands et dans certaines parties de la baie Placentia.

INTRODUCTION

Background and objectives

Some of the earliest information on the marine geology of the Island of Newfoundland coastal waters is found in publications by Quaternary scientists who, while principally studying events on land, were aware of the information that might lie offshore. Thus Flint (1940, p. 1773) examined hydrographic charts of south coast fiords and detected "… sizeable deltas with the breaks of the crests of their distal slopes occurring at depths ranging between 30 feet and 12 feet below sea level." He thought that the deltas had undergone submergence since their formation. Similarly, Grant (1987) identified an arcuate submarine moraine off Bay of Islands on the basis of bathymetry, and claimed it marked the limit of Late Wisconsinan ice.

The growth of marine geological research at the Bedford Institute of Oceanography, Dartmouth, Nova Scotia, from the 1960s onward led to the publication of a series of surficial geology maps based on sparse networks of bathymetric and geophysical profiles. The most relevant to the present study was the map by Fader et al. (1982) dealing with the Laurentian Channel and the western half of The Grand Banks of Newfoundland; however, a quick perusal of this map shows little coverage in coastal areas, and fiords and other inlets were not examined.

Knowledge of the innermost shelf began to grow from the mid-1980s onward when the Geological Survey of Canada (GSC) commissioned a series of surveys in coastal regions, including surveys that penetrated the many inlets and fiords of the Island of Newfoundland. These surveys were conducted by a range of Canadian Coast Guard research vessels, notably by the 20 m long CSS *Navicula*, and by larger vessels such as CCGS *Hudson,* CSS *Dawson*, and CCGS *Parizeau.* These vessels towed a range of high- and low-resolution seismic-reflection systems, and had the capability to collect piston and gravity cores, grab samples, and bottom photographs. The research was directed toward various ends, and the results were published in GSC Open File reports and scientific papers. The work conducted off northeast Island of Newfoundland and funded under the Canada-Newfoundland Mineral Development Agreement resulted in a 1:250 000 scale surficial geology map (Shaw et al., 1999b) and an accompanying report that included a series of detailed maps. Mapping with such 'standard' techniques has continued to this day, a notable example being the map by Cameron and King (2010), which includes St. Mary's Bay and part of Placentia Bay, but otherwise deals with areas far offshore.

Pioneering surveys with multibeam sonar commenced in 1995, when inner St. George's Bay, southwest Island of Newfoundland, was surveyed. The results were reported in several papers (Shaw and Courtney, 1997; Shaw et al., 1997) and quickly confirmed that even in an area intensely surveyed with more conventional techniques, multibeam-sonar mapping provided a strikingly different view of the seafloor, and revealed complex landforms and processes the presence of which was commonly not evident from the previous work. Subsequently, many inner-shelf coastal areas and fiords have been surveyed, sometimes by GSC scientists on Canadian Coast Guard vessels, with equipment supplied by the Canadian Hydrographic Service (CHS), and sometimes in partnership with CHS, notably in Placentia Bay where mapping was funded under Canada's Oceans Action Plan. Canadian Hydrographic Service has mapped many areas for the purpose of updating charts, and has made data available, usually in the form of raw multibeam data, but in some instances as images of shaded relief. The shaded-relief data cannot be queried, nor given variable illumination angles and differing colour bars, and consequently have been less useful. In several instances mapping was conducted by the Marine Institute, Memorial University, St. John's, Island of Newfoundland.

Consequently a large amount of multibeam-sonar data and imagery from these various sources was available for analysis, and in almost all instances ground-truthing data were also available: high- and low-frequency subbottom profile data, sidescan sonograms, grab samples, cores, and bottom photographs. These combined data sets shed light on the nature of the seafloor in coastal waters of the island, and in particular facilitated study of the offshore extensions of littoral systems such as spits, barriers, and strand plains, previously known from studies on land. In this context the submarine extensions of littoral systems in St. George's Bay are the prime example (Shaw and Courtney, 1997; Shaw et al., 1997).

Whereas only a small portion of the island's inner shelf has been surveyed using multibeam-sonar technology, it is nevertheless timely to attempt to synthesize the data in order to present a coherent picture of the surficial geology of the inner shelf and adjacent coastal areas. The goal of this bulletin is therefore to describe the surficial geology of coastal waters of the Island of Newfoundland, focusing on areas that have been mapped with multibeam-sonar systems, and to include commentary on the coastal geomorphology, particularly where the offshore data further the understanding of onshore systems.

REGIONAL SETTING

Geology

The Island of Newfoundland (Fig. 1, 2) is divided into three geological zones that reflect a history of continental drift and ocean spreading (Colman-Sadd et al., 1990). The western part of the island — the Humber Zone — was originally part of the North American Plate. The central zone is the mobile belt comprised of remnant volcanic arc and the ancient Iapetus Ocean floor. The eastern zone (Avalon

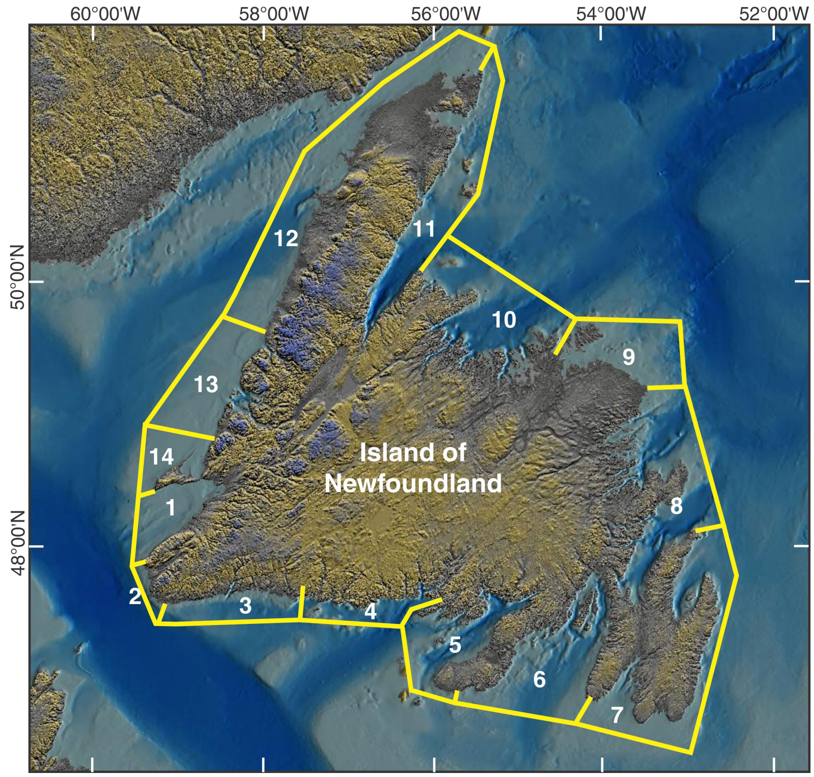


Figure 1. Location of the 14 coastal and inner shelf regions discussed in the text: 1 = St. George's Bay, 2 = Cape Anguille to Cape Ray, 3 = Cape Ray to Bay de Loup, 4 = Bay de Loup to Hermitage Bay, 5 = Fortune Bay, 6 = Placentia Bay, 7 = Avalon Peninsula, 8 = Trinity Bay to Bonavista Bay, 9 = Cape Freels to Dog Bay, 10 = Notre Dame Bay, 11 = White Bay to the Strait of Belle Isle, 12 = Strait of Belle Isle to Bonne Bay, 13 = Bonne Bay to Port au Port Bay, 14 = Port au Port Bay.

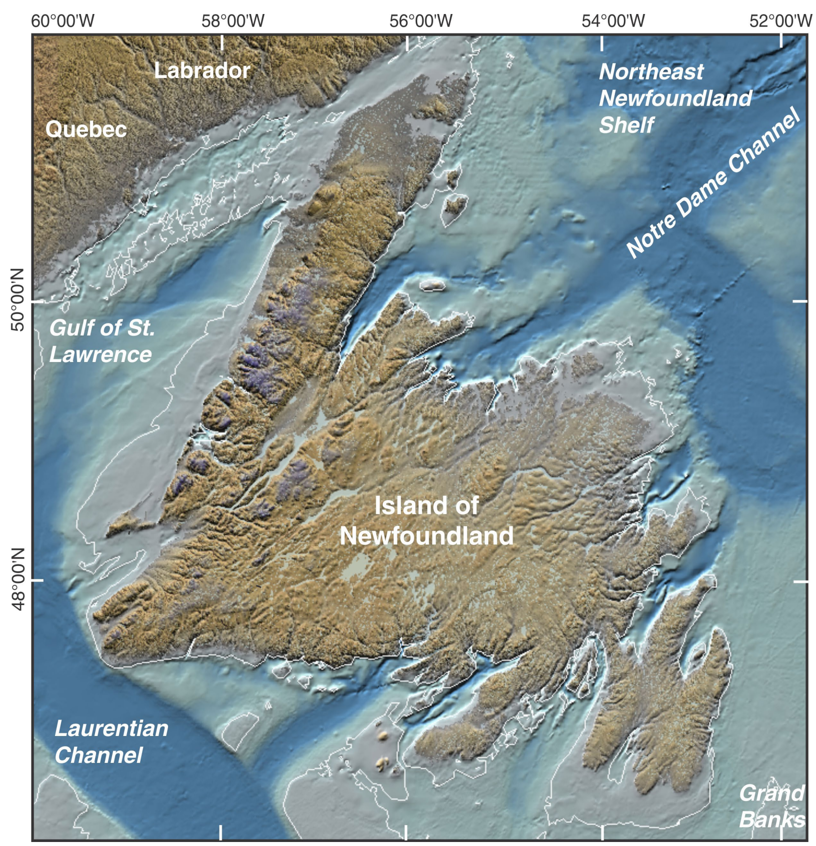


Figure 2. Map of Island of Newfoundland and the adjacent shelf, with the 100 m isobath shown.

Peninsula) was formerly part of the African Plate. The island is primarily composed of relatively old 'basement' rock, comprising altered volcanic, metasedimentary, and granitic intrusive rocks. Only in the west, from St. George's Bay northward, have relatively undisturbed Carboniferous and Ordovician sediments been observed (Colman-Sadd et al., 1990). In coastal areas, the basement rocks generally extend only a short distance offshore, and drop steeply into deeper water where more recent rocks, ranging from Carboniferous to Tertiary are noted (Fader et al., 1989). The boundary is commonly fault-bounded, for example, on the linear coast of the Baie Verte Peninsula. On the Northern Peninsula, however, the basement boundary lies inland, at the foot of the Long Range Mountains.

Physiography

The island (Fig. 2) is surrounded by the Gulf of St. Lawrence in the west, the deep Northeast Newfoundland Shelf in the northeast, the shallow part of The Grand Banks of Newfoundland in the southeast, and the deep Laurentian Channel in the southwest. The littoral fringe of shallow (<100 m) water is limited in most areas, but is notably extensive off the west coast (Port au Port to Bonne Bay) and the northeast coast. Slopes offshore are gentle in comparison with those on land. The steepest slopes occur where basement rocks plunge straight into deep water, as along the south coast and in Notre Dame Bay, particularly in fiords. The northeast coast — the 'Straight shore' — is remarkable in that gentle slopes on land continue offshore. On the western Northern Peninsula the steepest slopes are inland, at the foot of the Long Range Mountains, and gentle slopes at the coast continue offshore. There are many local variations, for example, along the southwest coast the steep basement boundary slope lies offshore, creating a coastal platform 5–10 km wide; however, east of Burgeo the boundary lies at the coast, and a platform is absent. In the extreme southwest of the Island of Newfoundland (region 2) the shallow littoral platform is narrow, and is bounded by the steep slopes flanking the Laurentian Channel.

The numerous fiords radiating out from the interior of the Island of Newfoundland display regional variations. West coast fiords are moderately deep (230 m maximum at Bonne Bay), are surrounded by high plateaus, and bordered by a shallow inner shelf. Southwest coast fiords are extremely narrow (1–2 km) and are commonly several hundred metres deep; Bay d'Espoir has a maximum depth of 770 m, making it the deepest fiord on the island (and indeed, the deepest area on Atlantic Canada's continental shelves). The plateau into which these fiords are incised has moderate elevation (300 m) and the adjacent inner shelf is comparatively deep (>200 m). Maximum depths in northeast coast fiords exceed 600 m. These fiords are wide, and local terrestrial relief (100–200 m) is relatively low. Troughs extending northeast from Trinity Bay and Notre Dame Channel reach the shelf edge (Warren, 1976).

Glacial history

Shaw et al. (2006d) published a conceptual model of ice extent and patterns of retreat in Atlantic Canada for the last glaciation, the Late Wisconsinan. They depicted an ice sheet that completely covered the Island of Newfoundland, drained by ice streams that connected with the Laurentian Channel in the west and were located in Trinity and Bonavista troughs and Notre Dame Bay (*see also* Dyke et al., 2002). A lobe of ice shown extending onto The Grand Banks of Newfoundland was, it was argued, less dynamic than the ice draining ice streams elsewhere. Subsequently Cameron and King (2011) argued that the lobe on The Grand Banks of Newfoundland was slightly less extensive that Shaw et al. (2006d) depicted, so that morainal banks farther out on The Grand Banks of Newfoundland (Shaw et al., 2006d) are thus older than the last glacial maximum. Figure 3 shows the last glacial maximum situation as depicted by Shaw et al. (2006d). It was argued that Notre Dame Bay deglaciated earliest, and that subsequent retreat was concentrated along troughs. Retreating ice margins halted near present coastal areas in west and south Island of Newfoundland. The patterns of divides in the series of maps in Shaw et al. (2006d) appear to be broadly consistent with those interpreted on land by analysis of recent shuttle radar topography mission (STRM) data (Liverman et al., 2006).

Coastal areas of the island record the effects of overdeepening by glaciers in fiords, the formation of streamlined glacial landforms by ice moving offshore, and the formation of moraines after the retreating ice margins reached modern coasts ca. 14 ka BP. Many examples will be given in subsequent sections. The primary examples of moraines are: 1) large submarine moraines on the west coast at Bonne Bay, Bay of Islands, and St. George's Bay; 2) a suite of fiord-mouth moraines along the south coast; 3) transverse moraines in the deep fiords of Notre Dame Bay. Streamlined landforms that are best developed in Placentia Bay (Brushett et al., 2007; Shaw et al., 2009), and to a lesser degree in Bonavista Bay and Notre Dame Bay, testify to the movement of glacier ice offshore, beyond the limits of earlier models (e.g. Dyke and Prest, 1987).

Postglacial relative sea levels

As a result of glacio-isostasy and eustasy, postglacial relative sea-level changes were registered in coastal areas around the island (Flint, 1940; Grant, 1972, 1980, 1987, 1989, 1991, 1994; Tucker, 1974; Brookes, 1977; Tucker et al., 1982; Brookes and Stevens, 1985; Brookes et al., 1985; Forbes et al., 1993; Liverman, 1994; Shaw and Forbes, 1995; Bell et al., 2003). Across most of the island, higher than present relative sea level was recorded immediately following deglaciation, the exception being a peripheral zone in the south in which relative sea level was never above the present (*see* marine limit map in Grant (1989)). Relative sea level fell continuously on the Northern Peninsula, with the maximum levels being registered in the extreme north (+120 m

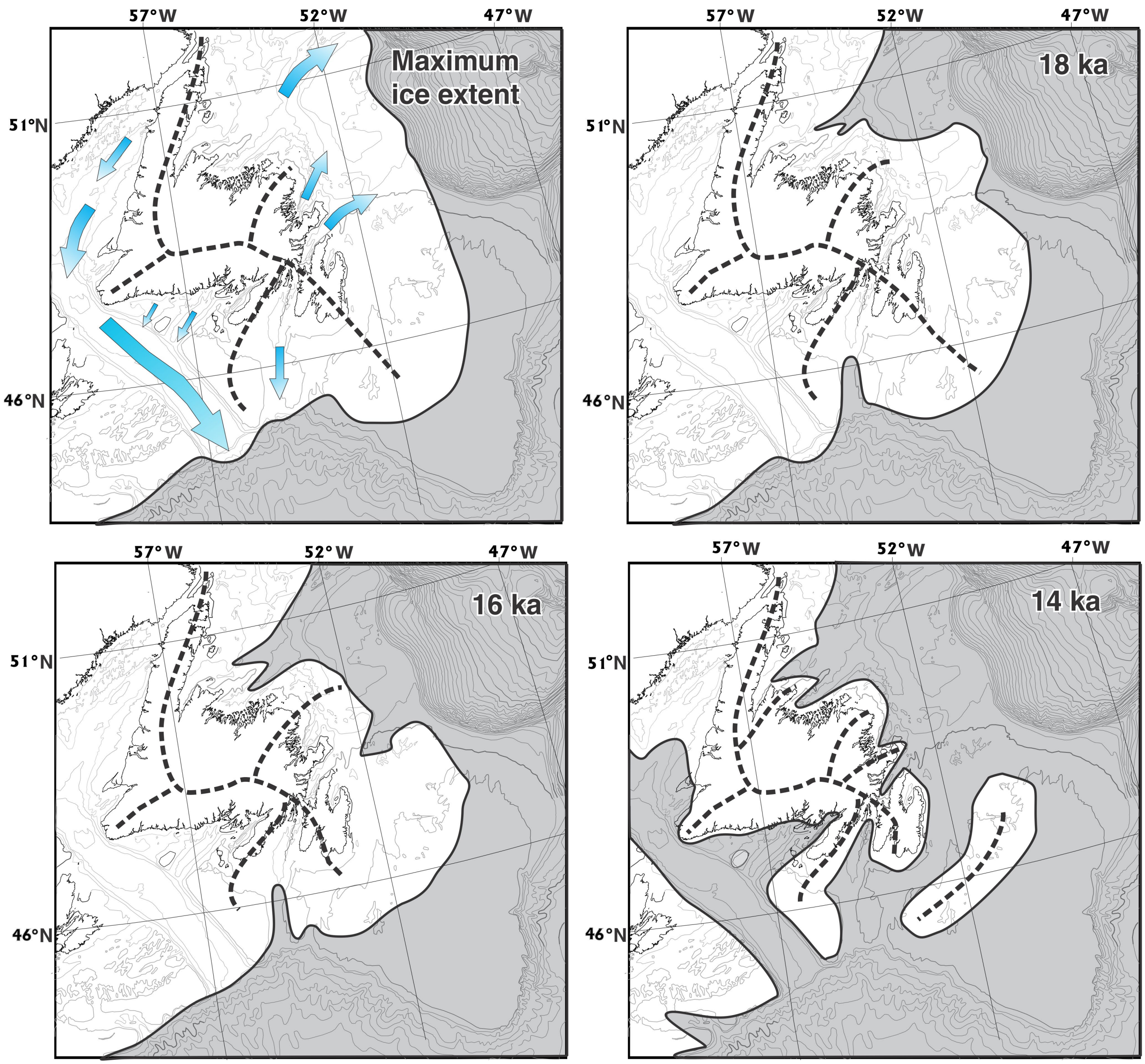


Figure 3. Ice extent at the last glacial maximum (upper left) and at several stages of deglaciation (*modified from* Shaw, 2003).

according to Grant (1994)). Across most of the island's coastal regions, relative sea level dropped to a lowstand, and subsequently rose again. The lowstand depth was spatially variable (Fig. 4) and also diachronous, i.e. the lowstand was at differing times in various areas. North of the zero isopleth (Fig. 4), relative sea level has not fallen below the present level since deglaciation.

The resultant geographic changes in Atlantic Canada are shown in a series of graphics published by Shaw et al. (2002b). Because of the deep water adjacent to most coastal regions, geographic changes in the Island Newfoundland were small compared with those on the offshore banks (where postglacial relative sea level also reached lower depths). The largest emergent area during the relative sea-level lowstand was off the island's northeast coast (region 12). By contrast, the equally shallow shelf at the northwestern tip of the Northern Peninsula was not subaerially exposed because relative sea level was above modern sea level in that region. The sill across inner St. George's Bay (region 1) was mostly emergent ca. 9 ka (Shaw and Forbes, 1992). A notable geographic change that is not evident on this map series was the formation of numerous lowstand deltas now submerged in southwest, south, and eastern Island of Newfoundland (Shaw and Forbes, 1995).

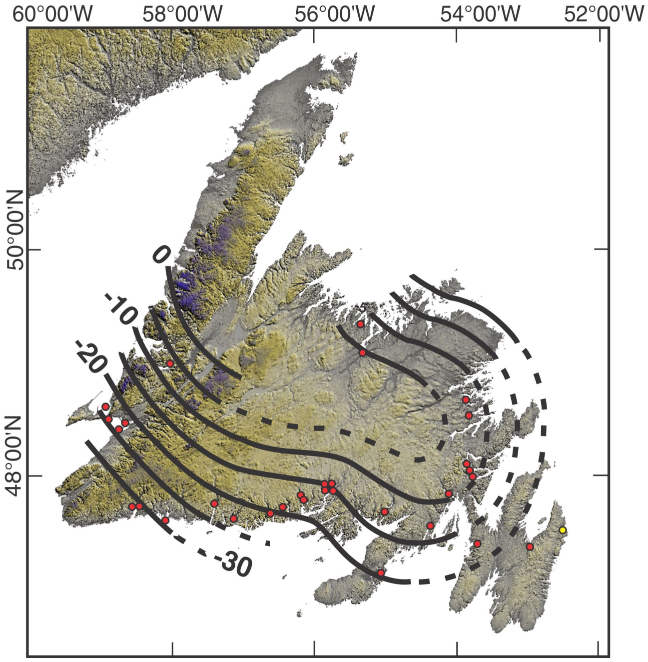


Figure 4. Isopleths (in metres) of the postglacial sea-level lowstand. Red dots = lowstand deltas; yellow dot = sill-controlled delta.

Modern rates of sea-level rise as determined by tide gauges at Channel-Port aux Basques, Argentia, St. John's, and elsewhere, indicate ongoing submergence in the south of the island (Shaw and Forbes, 1990b), but not in the northern Gulf of St. Lawrence. The implications of an acceleration in the rates of submergence for the coasts of the island are discussed by Shaw et al. (1998, 2001).

Surficial geology

Generally the seismo- and lithostratigraphic units are equivalent to the formations defined by Fader and Miller (1986). Syvitski (1991, 1994) considered the question of stratigraphic sequences on glaciated continental shelves, and argued that a complete deglacial sequence consists of some or all of the following: 1) ice-contact (ice-deposited and/or ice-loaded) sediments; 2) ice-proximal sediments; 3) ice-distal sediments; 4) paraglacial coastal sediments; and 5) postglacial sediments. For the study area the following lithostratigraphic units apply (equivalent formation name in parentheses).

Bedrock forms acoustic basement and may display parallel internal reflections or (more usually) a dense tone and incoherent reflections. On sidescan sonograms bedrock is typically strongly reflective, and the positive relief casts sharp acoustic shadows. On multibeam-sonar imagery, joints, bedding planes, and faults are commonly observed, and backscatter strength is high.

Glacial diamicton (equivalent to the Newfoundland Shelf Drift Formation of Fader et al. (1982)) was deposited at ice margins or under the ice, and comprises randomly oriented subangular pebbles and cobbles in matrices of sandy silty clay. On seismic-reflection profiles this unit commonly has a dense grey tone formed by incoherent reflections, and occasionally contains irregular, quasihorizontal reflections separating wedges of the unit. Commonly it has a positive relief, and may occur as mounds, flow-transverse morainal ridges, De Geer moraines, fiord-mouth moraines, and flow-parallel streamlined landforms (drumlins, crag-and-tail features, megascale glacial lineations). It has high backscatter on multibeam-sonar imagery.

Glaciomarine sediment (Downing Silt Formation) was deposited by meltwater plumes, close to ice margins or at greater distance, and usually comprises gravelly sandy mud with moderate- to high-intensity continuous coherent reflections that are highly conformable to the underlying terrain, and contains scattered point-source reflections. In fiords this unit contains basin-fill style acoustically transparent intervals commonly up to 20 m thick, but 150 m thick in Green Bay (Northeast Newfoundland Shelf). Such intervals are interpreted as mass-transport deposits formed close to former ice margins. Where exposed at the seafloor, glaciomarine sediments have veneers of sandy muddy gravel with high backscatter on multibeam-sonar imagery, and are commonly imprinted by iceberg furrows and pits.

Postglacial mud (Placentia Clay Formation) comprises mud or silty mud produced by reworking of glacial sediments. It has conformable, closely spaced internal reflections the intensity of which weakens upward, producing an acoustically transparent appearance on seismic profiles. Thickest deposits occur either in deep, offshore basins or in fiords, harbours, and estuaries. This unit has a smooth surface on multibeam imagery and low backscatter.

Postglacial sand and gravel (Grand Banks Sand and Gravel Formation) was also formed by reworking of glacial sediments, and includes: 1) well sorted sand with shell hash and scattered pebbles, with wave- and current-generated bedforms including ripples and dunes; 2) thin sheets of mobile, poorly sorted sandy gravel, with ripples; 3) veneers of largely immobile (*Lithothamnion*-coated) gravel overlying glacial diamicton; and 4) shoreface prisms of sand or muddy sand that pass laterally into beaches (as seen in region 2). Postglacial sand shows weak internal reflections or is acoustically incoherent on seismic profiles, whereas surficial gravel inhibits acoustic penetration. The sand has low backscatter and the gravel has high backscatter on multibeam-sonar data.

Modern processes

Waves and currents

Figure 2 demonstrates the relative paucity of wide, shallow inner shelves around the island, compared with areas farther offshore, i.e. The Grand Banks of Newfoundland. In coastal waters the chief areas of intense seafloor disturbance by waves are: 1) off the island's northeast coast (region 10); 2) off the southwest coast (region 2); 3) off the west coast (outside the fiords and inlets); 4) off the tip of Avalon Peninsula; and 5) off the tip of Northern Peninsula. In comparison with waves, oceanographic currents have minimal effect at the seafloor, but play a major role in the advection of icebergs in east and northeast offshore areas. The recent assessment of circulation off eastern Canada (Wu et al., 2012) shows that the south-moving Labrador Current is the dominant feature of the circulation, but it has maximum strength at the shelf break. An inshore branch circulates around the tip of the Avalon Peninsula and into Placentia Bay (Ma et al., 2012). The circulation in the Gulf of St. Lawrence adjacent to coastal Island of Newfoundland is weak and variable in comparison with areas to the west (the Gaspe Current, Ma et al. (2012)).

Icebergs and sea ice

The effects of iceberg scouring and grounding (Woodworth-Lynas et al., 1985, 1991) as observed at the seafloor can be classified into two zones. In a zone extending down the east side of Northern Peninsula, into Notre Dame Bay, and south to Avalon Peninsula (regions 7–12 in this study), scours are created by icebergs advecting southward in the Labrador Current. These modern iceberg scours coexist with relict iceberg scours formed during deglaciation (ca. 14 000 ^{14}C years BP). Todd et al. (1988) reported that modern icebergs ground on the Labrador Shelf, to the north of the study area, to depths of 220 m, and locally to 280 m; relict scours occurred in the 220–300 m range. In Notre Dame Bay (regions 9 and 10), Shaw et al. (1999b) claimed that modern iceberg scouring activity reaches depths of 200 m. Large areas of seafloor are imprinted by scours and grounding pits, with the pits tending to predominate close to the coasts. The imprinting is commonly in the glaciomarine unit, in which intense iceberg turbation can destroy acoustic stratigraphy, but furrows are also imprinted into glacial diamict in Notre Dame Bay, in depths of 200–300 m.

In a second zone, comprising the remainder of continental shelf around the island (regions 13, 14, and 1–6 in this study), almost all iceberg scours are relict, and were created ca. 14 000 ^{14}C years BP. Nevertheless, some areas experience modern iceberg grounding. For example, Centre for Cold Ocean Research and Engineering (unpub. report, 2004) argued that about 10–15% of the icebergs that pass the latitude of the Strait of Belle Isle drift into the strait. Similarly, a small number of icebergs pass round the southern tip of Avalon Peninsula and reach Placentia Bay.

Most of the island is subject to pack ice in coastal waters. The least pack ice occurs on the south coast, from Cape Ray to the southern tip of the Avalon Peninsula (Canadian Ice Service, 2011). The highest frequency is in a region extending from the Strait of Belle Isle to Bonavista Bay.

DATA SOURCES

Coastal surveys

Beginning in 1981, the Geological Survey of Canada established coastal survey sites in the Island of Newfoundland. The network expanded in the late 1980s, and today comprises about 140 sites. These sites have provided information on coastal processes, including, for example, rates of erosion and beach retreat; data collected near some of the sites were also the basis for an understanding of postglacial sea-level changes. Another activity of the Geological Survey of Canada has been the acquisition of aerial video of the coastline of the island (e.g. Forbes and Frobel, 1986). This video imagery was usually collected using a helicopter flying at low altitudes a short distance from the coast. The lead author, for example, participated in a survey of the southwest coast, from Channel-Port aux Basques to Terrenceville (Shaw and Frobel, 1992). Video surveys for other coastal areas were commissioned by nongovernment agencies and were not used for this study.

Geophysical surveys

Surveys of areas just offshore the coast were driven by a series of factors, including: 1) a desire to understand the offshore extent of coastal systems; 2) a need for information on past relative sea-level changes; and 3) a quest to ascertain the occurrence of marine gold placers (northeast Island of Newfoundland inner shelf). These surveys extended far offshore in some instances, a good example being Notre Dame Bay (Shaw et al., 1999b). They permitted exploration of shelf areas that had not been examined during regional surveys of earlier years, and facilitated the creation of models of glacial and sea-level history. The principal surveys are summarized in Table 1 and their distribution is shown on Figure 5, together with other surveys that were consulted on occasion.

Multibeam-sonar mapping: bathymetry

The initial multibeam-sonar mapping by GSC researchers was undertaken in 1995 off western Island of Newfoundland using CCGS *Frederick G. Creed*, a SWATH vessel (Small Water Plane Twin-Hull), equipped with a Simrad EM-1000 system. All multibeam sounders work using essentially the

Table 1. Ground-truthing surveys.

Cruise	Reference	Area and vessel
88018 E	Forbes and Shaw, 1989	Port au Port, St. George's Bay, La Poile to Barasway; CSS *Navicula*
89008	Josenhans et al., 1989	Western Newfoundland; CSS *Baffin*
89026	Shaw et al., 1990b	Placentia Bay; CSS *Navicula*
90013	Shaw and Wile, 1990	Notre Dame Bay; CSS *Hudson*
90035	Shaw et al., 1990a	Notre Dame Bay; CSS *Navicula*
91026	Shaw et al., 1992	Southwest coast (La Poile to Bay d' Espoir) and Notre Dame Bay; CSS *Dawson*
91031	Edwardson et al., 1992	Notre Dame Bay; CSS *Navicula*
92042	D.L. Forbes and J. Shaw, unpub. cruise report, 1993	Notre Dame Bay; CSS *Parizeau*
92054	J. Shaw, unpub. cruise report, 1992	Bay d'Espoir; CSS *Matthew*
92301	Edwardson et al., 1993	Notre Dame Bay; CSS *Nicholas and Paul*
94138	Shaw et al., 1995b	Bay of Islands; CSS *Hart*
97060	Shaw et al., 1999a	Western Nfld. Bonne Bay to St. George's Bay; CCGS *Matthew*
2000030B	J. Shaw and P.R.G. Girouard, unpub. cruise report, 2000	Southwest Newfoundland; CCGS *Hudson*
2005051	Shaw et al., 2006e	Placentia Bay; CCGS *Matthew*
2006039	Shaw et al., 2007	Placentia Bay; CCGS *Hudson*
2008052	J. Shaw, unpub report, 2008	Placentia Bay; CCGS *Shamook*

same basic principles. Sound is projected toward the seabed in cone-shaped beams formed using an array of piezoelectric transducers mounted on the hull of the survey vessel. These beams span a fan-shaped sector below and to the sides of the survey vessel. The width and number of beams depends on the multibeam manufacturer and model. The Simrad EM-1000 system projected 60 beams over a 150° arc with an effective beam width of around 2.5° in the along-track direction. The beam width determines the footprint of the mapping system and the footprint expands with the distance between the transducer array and the seabed. The Simrad EM-1000 system has a footprint between 2 m and 10 m in water depths less than 100 m, dependent on the angle of the beam off nadir (nadir being directly below the vessel).

By measuring the transit time of a pulse projected from the transducer array and reflected from the seabed back to the array, an estimate of the distance to the seabed could be calculated separately for each beam. Since the beams were projected at variable angles to nadir, the sound-velocity profile of the water column was needed to account for ray-bending effects. As the survey vessel moved along its track, estimates of depth were measured below and to the sides of the vessel. The along-track spacing between successive soundings was determined by the time for sound to return from the outermost beams, the speed of the ship and the computational speed of the multibeam processing unit. By running closely spaced survey lines, complete-coverage, high-resolution, digital-terrain models (DTMs) of the seabed were assembled.

When Placentia Bay was systematically mapped using Canada's 'Oceans Action Plan' funding, systems in the vessels were being upgraded. The CCGS *Frederick G. Creed* was initially equipped with the Simrad EM-1000 multibeam-sonar system, with the transducer mounted in the starboard pontoon. This was replaced by a Simrad EM-1002 system from 2005 onward. The CCGS *Matthew* was equipped with a Simrad EM-1002 multibeam-sonar system during the 2004 surveys; this was replaced by a Simrad EM-710 system in 2005. Shallow areas were surveyed using the hydrographic launch *Plover*, which deployed a hull-mounted Simrad EM-3000 multibeam-sonar system. (Note: Simrad was eventually incorporated into Kongsberg, and multibeam sounders are now manufactured under the company name Kongsberg Maritime.)

For most surveys, Differential Global Positioning System (GPS) receivers were used for navigation, providing positional accuracy of about 3 m. Survey speeds for most vessels averaged 10 knots. Data were adjusted for tidal variations using output from tide gauges. Data were cleaned and gridded in 5 m (horizontal) bins using the CARIS Hydrographic Information Processing System, exported and subsequently imported into GIS systems, including GRASS (initially

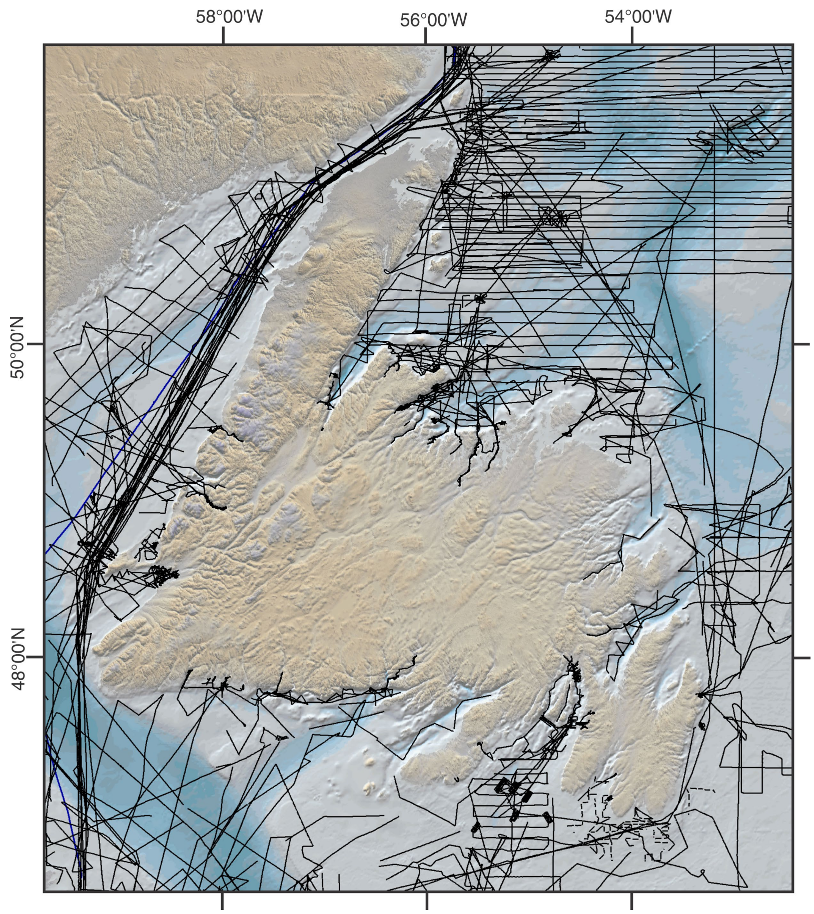


Figure 5. Ships' tracks for the principal geophysical surveys used in the study, together with additional surveys consulted on occasion.

developed by the U.S. Army Corps of Engineers); however, almost all images used in this report were developed in the Global Mapper GIS program. A variety of colour schemes have been used, and these have been selected to cope with the depth ranges in the specific areas. Artificial illumination has been varied to create the optimum views of landforms.

Multibeam-sonar mapping: backscatter strength

Multibeam systems generally record the mean energy and the time series of amplitudes returned in each beam. This mean energy is commonly called the backscatter amplitude. It is tempting to use this backscatter amplitude as a direct proxy for surficial sediment type, but a careful consideration of the physics of the problem presents a more convoluted picture. Whereas there exists no direct and simple relationship between the backscatter amplitude and surficial sediment type (Courtney and Shaw, 2000), for angles outside the specular range there is a general correspondence between backscatter amplitudes and surficial sediment roughness that can be used for cursory mapping and sediment identification. Coarse gravels and cobbles tend to be locally rough and return high-amplitude, wide-angle, backscatter signals, whereas sand and fine-grained materials can be locally smooth with a much lower backscatter. For this publication backscatter data have been extracted in many areas, but in some instances there was access only to gridded bathymetry, and inferences had to be made based on groundtruthing and also on the known seismostratigraphic framework.

Comparison of multibeam-sonar data with conventional data

Multibeam sonar, originally created as a tool for hydrographers, is now a standard seafloor mapping tool used by marine geologists. The potential was realized by GSC researchers as soon as the first of the new images derived from this methodology became available (Loncarevic et al., 1994; Shaw and Courtney, 1997; Shaw et al., 1997). The high-resolution imagery, supplemented by backscatter data, showed the true complexity of the seafloor, not revealed by previous methods. This can be illustrated by comparing part of the surficial geology map for Notre Dame Bay, Island of Newfoundland (Shaw et al., 1999b), with the multibeam image for the same area (Fig. 6). The surficial geology map is based on a limited number of transit lines, and shows three units: bedrock, glaciomarine mud, and postglacial mud; however, the multibeam image shows much greater complexity. It reveals the distribution of bedrock and permits differentiation of several lithologies. For example, the area in the upper part of the multibeam image (A in Fig. 6)

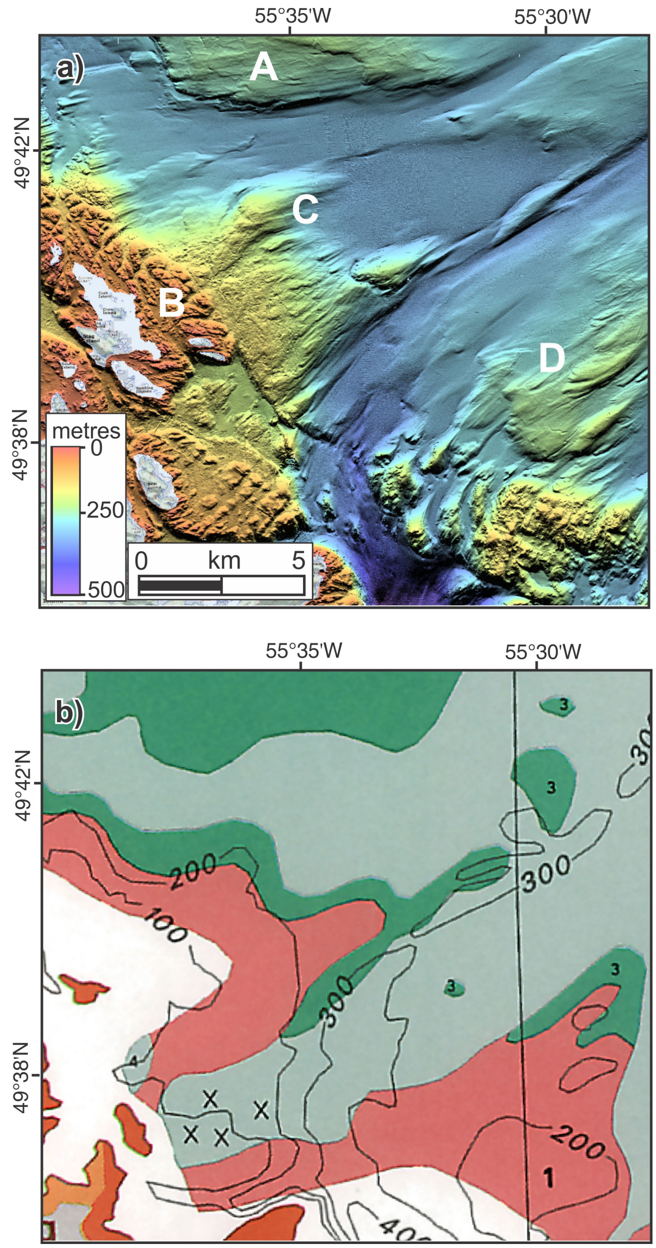


Figure 6. Comparison of **a)** an image derived from multibeam sonar with **b)** the surficial geology map by Shaw et al. (1999b). In Figure 6b red = bedrock predominant, green = glaciomarine mud, grey = postglacial mud, x = furrows. Letters in Figure 6a represent features discussed in the text.

is likely Carboniferous bedrock, and its gentle, low relief contrasts with the irregular relief of basement bedrock near the coast (B). Most significantly, the surficial geology map fails to even hint at the existence of the streamlined glacial landforms seen on the multibeam-sonar image (C) and (D).

There have been several approaches to presenting the interpreted imagery. The GSC adopted the approach of standard 1:50 000 scale maps showing shaded relief, backscatter strength, and surficial geology. Relevant maps in this series are the maps of shaded-relief, backscatter strength, and surficial geology for St. George's Bay (Shaw et al., 2006a, b, c), the maps of shaded-relief and backscatter strength for Placentia Bay (Potter and Shaw, 2009a, b, c, d, e; Potter and Shaw, 2010a, b, c, d, e), the maps of shaded-relief and backscatter strength for Bonavista Bay (Patton and Shaw, 2011a, b), and the map of shaded-relief for Bay d'Espoir (Shaw and Hayward, 2010).

More recently a 'seascape' approach has been explored. The definition of a 'seascape' is based on the Australian Land-System approach, developed to manage agricultural land (Christian and Stewart, 1953; Commonwealth Scientific and Industrial Research Organization, 1967). Seascapes are defined as "underwater landscapes characterized by unique combinations of geomorphology, texture, and biota". A relevant example is the seascape map for Placentia Bay (Shaw et al., 2011).

In this bulletin, a pragmatic approach is taken to presenting and interpreting multibeam imagery. The illustrations consist primarily of shaded-relief images derived from multibeam sonar, and are interpreted using available ground-truthing data: subbottom profiler data, sidescan sonograms, bottom photographs, grab samples, and cores. Where possible, backscatter-strength data have been used to facilitate interpretation, although few images derived from this source are presented in order to reduce the number of illustrations.

Radiocarbon dating

Radiocarbon dates are cited as conventional, i.e. $\delta^{13}C = 25‰$. Thus, 15 ka implies 15 000 ^{14}C years BP. In previous publications cited herein a marine reservoir correction of 400–410 years has commonly been applied. For calibration purposes McNeely et al. (2006) showed ΔR values of 180 ± 30 years and 130 ± 50 years on the shelf south of the island.

COASTAL AND INNER SHELF REGIONS

For purposes of description fourteen coastal–inner shelf regions are delineated (Fig. 1).

REGION 1: ST. GEORGE'S BAY

Setting

St. George's Bay (Fig. 7) lies within the Humber tectonic zone, is bounded to the south by post-Ordovician overlap sequences, to the north by Ordovician carbonate rocks on Port au Port Peninsula, and is floored by Carboniferous rocks. Eroding coastal bluffs composed of glaciogenic sediments stretch from the Anguille Mountains to the isthmus

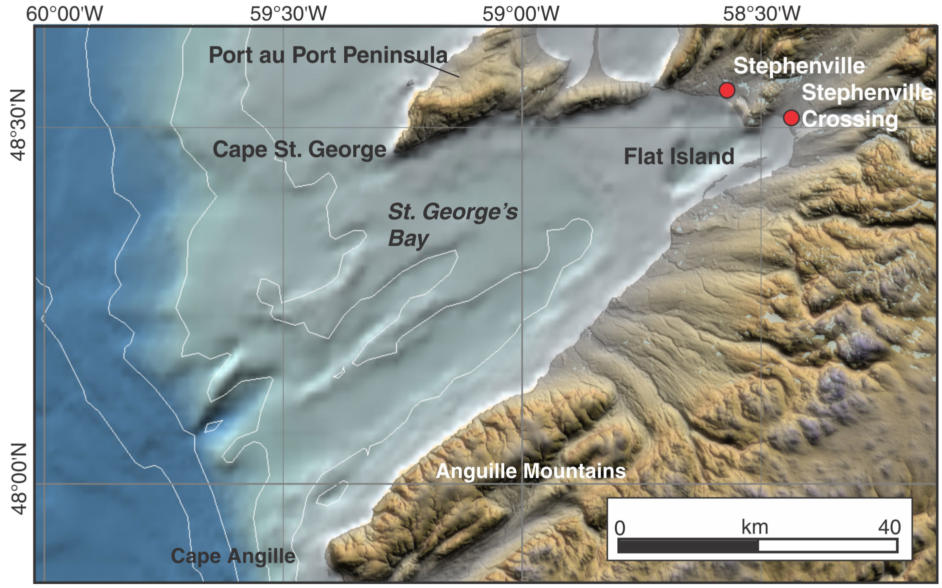


Figure 7. The St. George's Bay region, showing locations cited.

at Port au Port. They supply fine-grained sediment to the basins of inner St. George's Bay, and coarser sediments to large coastal systems at the head of the bay: Stephenville, Stephenville Crossing, and Flat Island. In the inner bay, Late Wisconsinan and early postglacial sediments attain great thicknesses, whereas in the outer bay Quaternary sediments are relatively thin and are restricted to several southwest-trending basins separated by bedrock ridges displaying complex structures in Carboniferous rocks.

Coastlines

Morphology and evolution

Coastlines along the north side of the bay are generally steep and rocky, but from the Port au Port isthmus, around the head of the bay, and southwest to the Anguille Mountains, thick Quaternary deposits are exposed in high coastal bluffs containing heterogeneous sediments deposited at former ice margins (MacClintock and Twenhofel, 1940; Brookes, 1969, 1974, 1977; Grant, 1987, 1991; Shaw and Forbes, 1987; Batterson and Janes, 1997; Bell et al., 2001, 2003). High erosion rates (up to 1.25 m per year) have been documented in the eroding bluffs between the isthmus at Port au Port and the beach at Stephenville, an area which is heavily developed for housing (Forbes et al., 1995a).

Rapid and ongoing erosion of the coastal bluffs has supplied sediment to beaches and spits that have evolved in their present positions over the past few thousand years (Shaw and Forbes, 1992) as relative sea-level continued its rise from the early Holocene lowstand of –25 m (Forbes et al., 1993, 1995b, c; Shaw and Forbes, 1995; Bell et al., 2003). The wide strand plain at Stephenville (Fig. 8) consists of gravel beach ridges that extend from the eroding bluffs in the northwest, across the mouth of a coastal compartment, to a rocky headland (Indian Head). The beach ridges decrease in elevation landward, so that the oldest beach ridges disappear below the waters of the back-barrier lagoon. Figure 9 shows a profile across part of the strand plain, and also a cross-section across one of the swales located between beach ridges. The swales contain organic-rich sediments that accumulated in brackish environments. A core located here intersected 0.45 m of herbaceous peat overlying organic-rich clay. Conventional ^{14}C dates ($\delta^{13}C$ = 25‰) on bulk samples were 640 ± 70 ^{14}C years BP at 30–35 cm (TO-3161), 1050 ± 60 ^{14}C years BP at 40–45 cm (TO-3160), and 2640 ± 60 ^{14}C years BP at 270–270 cm (TO-2159).

The gravel-beach ridges have been emplaced overlying a submerged sandy prism that extends offshore. This is not well seen in the multibeam-sonar imagery (*see* below) because of gaps in the survey, but is very clear in subbottom profiler and echo-sounder data, from which Figure 8b is derived. The seaward-thinning prism overlies glaciomarine sediments and grades into mud in the deeper water (Shaw and Forbes, 1992).

The barrier system at Stephenville Crossing at the head of the bay (Fig. 7) is sand dominated. Part of the sediment in this system originated from erosion of coastal bluffs west of Stephenville. The sand is transported along the front of the Stephenville strand plain on the submerged barrier platform, and passes round Indian Head. The transport of this bypassed sand to the northeast, and toward the sand-dominated beach at Stephenville Crossing, can be seen on Figure 8a where a small spit and submerged bar (Wolf Shoal) extend to the northeast (white arrow) toward Stephenville Crossing. This transport was only initiated once the compartment at Stephenville had become replete with sediment.

Flat Island spit (Fig. 10) is a 12 km long structure built from the glacial materials eroded from the bluffs that extend a further 40 km to the southwest (Shaw and Forbes, 1987, 1992). It has accumulated in the past several thousand years

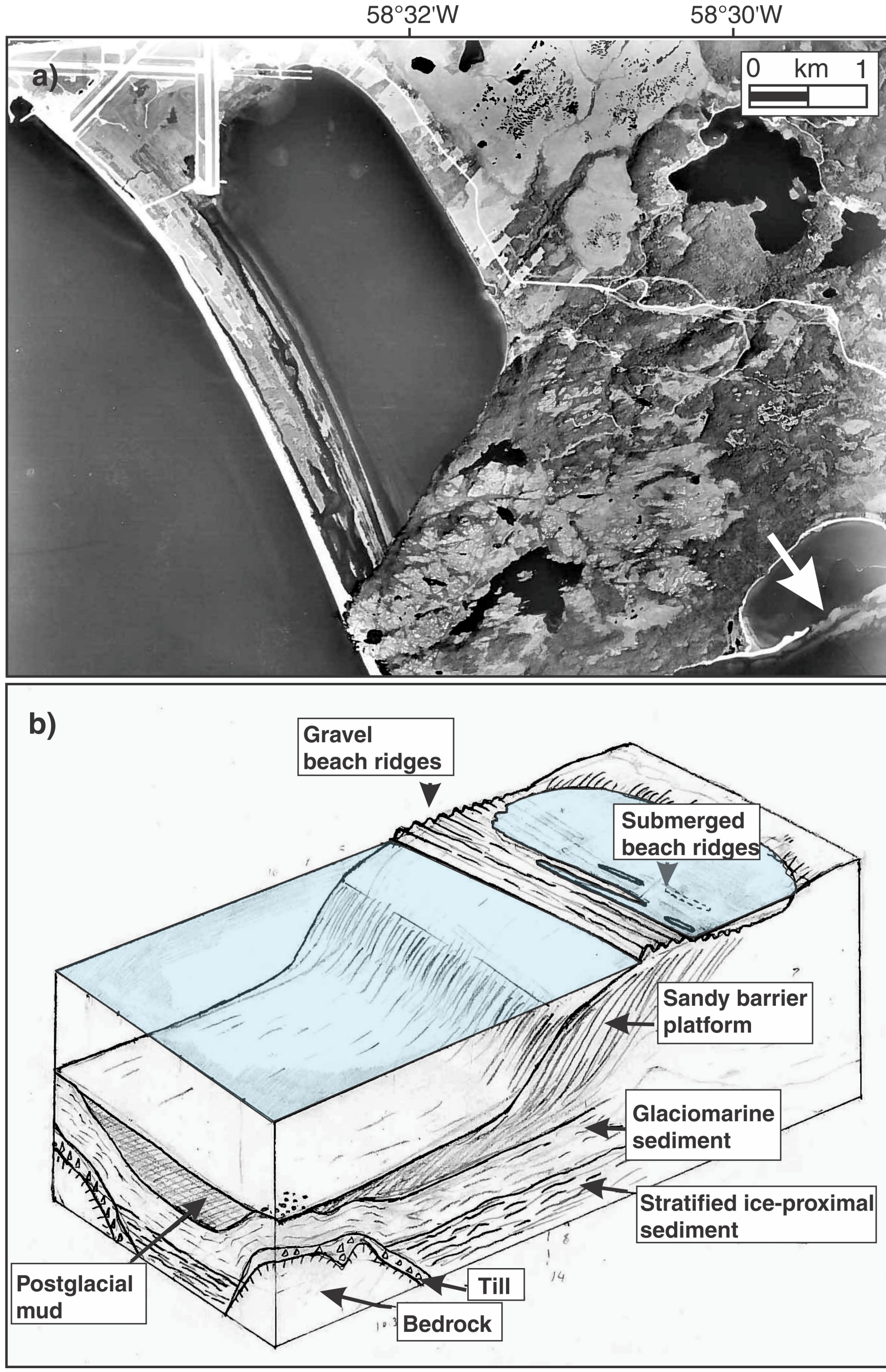


Figure 8. a) Aerial photograph (NAPL A-12156-166) of the gravel strand plain at Stephenville in 1949. This photograph predates the creation of a navigable channel at the south end of the barrier. The white arrow indicates where sediment that has bypassed the rocky headland is making its way toward Stephenville Crossing, at the head of the bay; **b)** illustration of the relationship between the gravel strand plain and an underlying sandy, barrier platform.

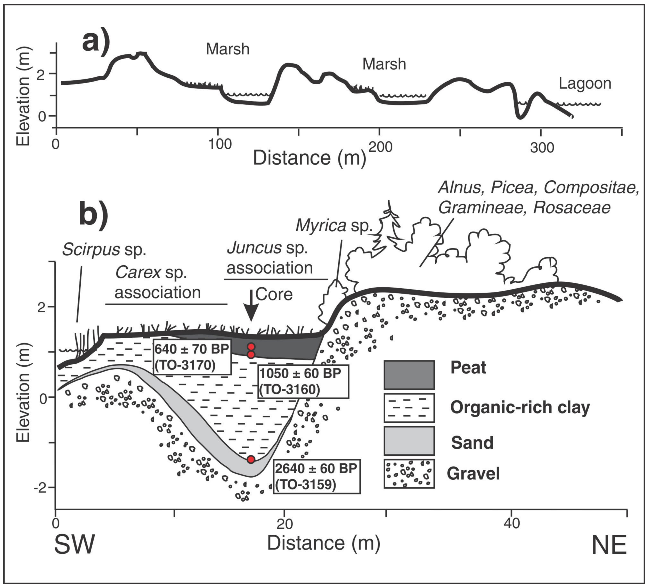


Figure 9. a) A short profile in the Stephenville strand plain, illustrating the landward decrease in beach-ridge elevation that reflects the formation of the strand plain in a regime of rising relative sea level; **b)** cross-section across one of the swales showing the accumulation of organic-rich clay and peat, and several radiocarbon dates that constrain the chronology of beach-ridge progradation.

under a regime of rising relative sea level. It comprises gravel beach ridges (*see* Fig. 11 in Shaw et al., 1990d), with some minor eolian decoration, and a nearshore bar system extending from the proximal attachment point of the spit to its end. The spit has two principal portions, one in the southwest and a wider and larger one in the northeast (once the site of Sandy Point, formerly one of the largest settlements on the west coast of the island).

The emergent parts of Flat Island spit comprise gravel beach ridges with an eolian veneer. Cross-sections contained in Shaw and Forbes (1992) show the very limited vertical extent of the spit (<3 m above mean water level (m.w.l.)) and the equally limited development of eolian dunes. The beach ridges are lowest to the rear, i.e. bordering Flat Bay, and numerous swales contain both freshwater peat and salt-marsh peat up to 1 m thick. The distal end of the spit has numerous recurves, and is evidently extending rapidly. Shaw and Forbes (1992) listed radiocarbon dates on freshwater and salt-marsh peat extending back to 1350 ± 70 ^{14}C years BP (Beta-19583), and argue that the spit has developed in a regime of rising relative sea level.

Flat Island spit is highly dynamic. A central portion of the spit was overwashed and breached in the 1950s, but the breach was subsequently progressively healed by longshore transport of gravel from the southwest. Shaw et al. (1990d) showed (their Fig. 11) that the west side of the northeast part of the spit was badly eroded between 1976 and 1986. There is no information on breaching events, but unpublished wave-refraction modelling shows how wave period is critical in focusing energy upon differing parts of the spit. With short-period (<7 s) waves approaching from the southwest, wave energy was estimated to be at a constant level along much of the spit, and to decrease at the distal end; the longshore component of wave power indicated eastward sediment transport; however, at wave periods of 7 s and upward, strong refraction caused by the topography of the shallow and morphologically complex sill that lies just offshore from Flat Island spit (*see* 'The sill — a modified moraine') focused waves at several locations on the spit, including the area of the former breach. This suggests an interaction between coastal processes and the evolution of the morphology in the deeper water.

The coastal systems in inner St. George's Bay are anomalous in comparison with coastal systems elsewhere on the Island of Newfoundland because they have formed adjacent to copious glaciogenic sediment sources, and are in close proximity to deep basins with large accommodation space (Shaw and Forbes, 1992; Forbes and Syvitski, 1995; Forbes, 2012). This results in barriers with relatively small subaerial portions and relatively large subaqueous components.

Future coastal evolution

It is interesting to speculate on the future of coastlines and coastal systems at the head of St. George's Bay under a scenario of global sea-level rise. The region is undergoing submergence. Whereas Shaw and Forbes (1990b) did not have confidence in their rates of sea-level rise derived from the Port aux Basques tide gauge, a value of 30.4 cm/century may be their best estimate. Carrera and coworkers (G. Carrera, P. Vanìček, and M.R. Craymer, unpub. report to Geodetic Survey of Canada, 1990) estimated a value of 37.5 cm/century. When factoring in predicted rates of sea-level rise from International Panel on Climate Change reports (18–59 cm/century), it is evident that, in a century's time, rates of sea-level rise will be well in excess of those at Port-aux-Basques today. Cliff recession and episodic retrogressive failure in the coastal bluffs west of Stephenville will supply copious sediment to the Stephenville strand plain; however, the coastal compartment at Stephenville has long been infilled to capacity. The wide gravel strand plain at the rear of the barrier will continue to submerge, eventually resulting in a single gravel storm beach. Sand will continue to bypass Indian Head, and will nourish beaches at Stephenville Crossing. The future of Flat Island spit is uncertain. Whereas it is supplied by copious amounts of sediment from the 40 km of eroding coastal bluffs to the southwest, the two large widest portions of the spit will continue to be submerged from the rear, resulting in a much narrower barrier. The central narrow portion will be overwashed repeatedly, but healing of the breaches will be aided by sediment supply from the southwest. The attachment point of the barrier will migrate to the northeast.

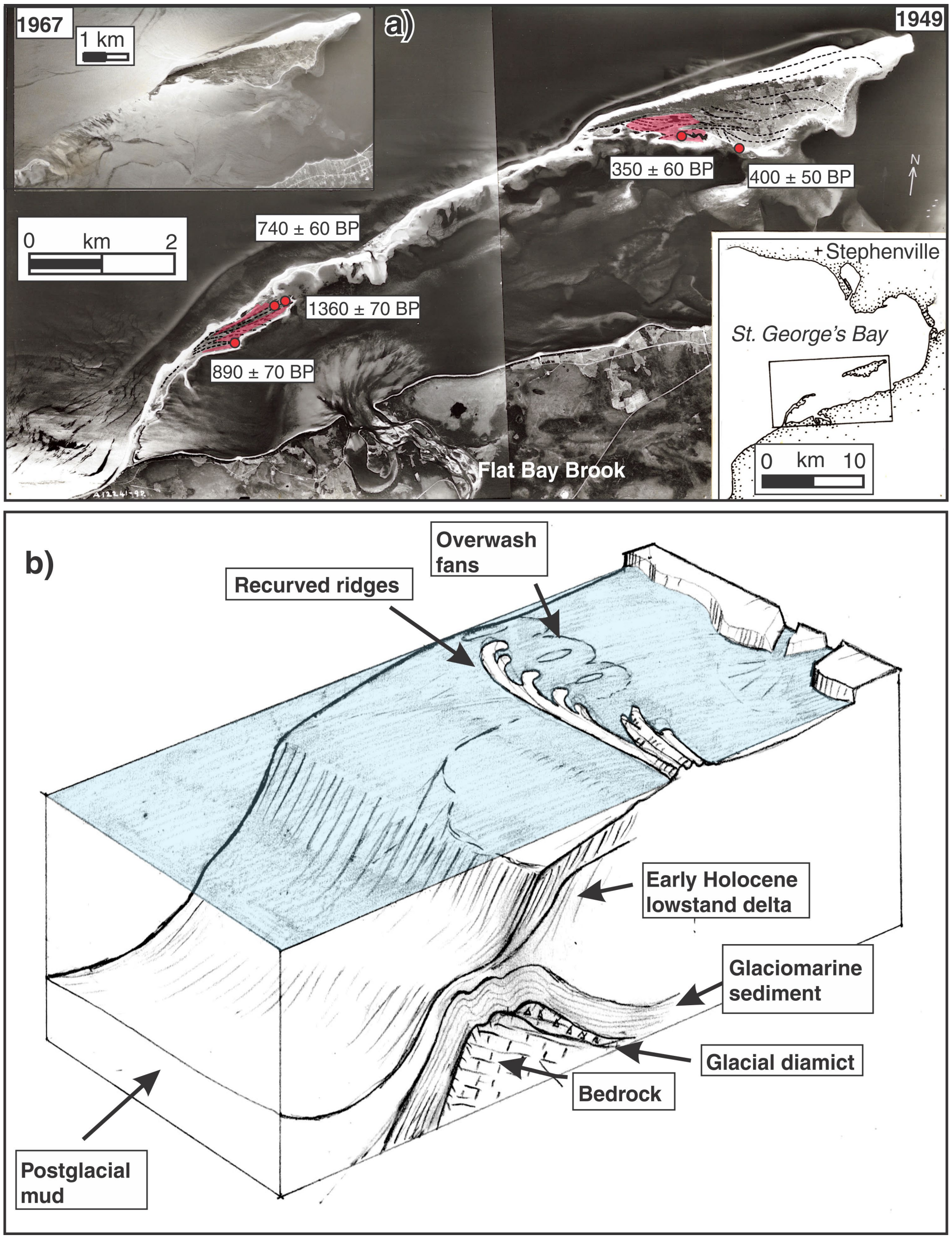


Figure 10. a) Airphoto mosaic of Flat Island spit compiled from 1949 photographs (NAPL A12241-92, A12241-93, A12241-95). The nearshore bar that feeds sand to the distal end of the system is clearly seen, as is the delta of Flat Bay Brook. The narrow central section of the spit was overwashed in the 1950s, initiating erosion at the west end of the wide, eastern barrier section. The inset photograph (1967) shows the breach, which has subsequently partly healed by spit extension from the south-west; **b)** illustration of the relationship between the spit, the submerged spit platform, and the early Holocene lowstand delta of Flat Bay Brook.

Inner St. George's Bay

Inner St. George's Bay was the first area of the Island of Newfoundland to be surveyed by GSC scientists using multibeam sonar. The new imagery, shown on Figure 11, was described and discussed in several papers (Shaw and Courtney, 1997; Shaw et al., 1997), and preliminary papers (Shaw et al., 1996a), and was subsequently used to produce three GSC A-series maps: Shaw et al., 2006a (shaded-seafloor relief), Shaw et al., 2006b (backscatter strength), and Shaw et al., 2006c (interpreted surficial geology). The interpretation of the multibeam data was done using groundtruthing data previously summarized by Forbes and Shaw (1989), Shaw and Forbes (1990a, 1992), and Forbes et al. (1995a).

The sill — a modified moraine

The sill across inner St. George's Bay (A in Fig. 11) is a submarine moraine that extended across the bay after ca. 14 ka (Shaw, 2003; Shaw et al., 2006d). An airgun seismic-reflection profile (Fig. 12) shows it to comprise acoustically incoherent sediment (glacial diamict) that passes laterally (seaward) into stratified sediment (glaciomarine mud). The moraine surface has been heavily modified by waves during and after the early Holocene relative sea-level lowstand, so that highs were eroded and depressions were infilled. The highly variegated surface includes areas of bouldery glacial diamict, sheets of rippled sand, sand waves, and fields of gravel ripples, resulting in complex areas of high and low backscatter (Shaw and Courtney, 1997; Shaw et al., 2006b). Several large sediment lobes (e.g. Fig. 11, B) have formed under the influence of the prevailing westerly waves and advanced eastward into the basins of inner St. George's Bay. The steep landward sides of the sill have been formed by spillover of gravel and sand into the basins. One of the migrating lobes apparently collapsed into the deeper water to form a large area of irregular topography (C).

Figure 13a shows the principal spillover lobe at a larger scale. In one area the migrating lobe overrode glaciomarine sediments, forming a series of compression ridges (A). Figure 13b shows a simplified version of the relationship between the prograding spillover deposits and the underlying sediments. Whereas parts of the sill area consists of sand bodies interspersed with sheets of rippled gravel, places where the moraine is close to the seabed are dominated by boulder gravel — for example, the rough terrain at B on Figure 13a.

To the west of the sill, a prism of sand thickens seaward, and becomes finer with depth (Fig. 11, D). The submarine moraine extends to the southwest, reaching the coast at the Anguille Mountains. Part of this extension of the moraine (E) was surveyed in 2011 by the Marine Institute, Memorial University. This survey revealed a very complex morphology (*see* inset on Fig. 11) that is difficult to decipher in the absence of backscatter and groundtruthing. There appear to be complex sand waves extending toward the deeper water, and patches of till.

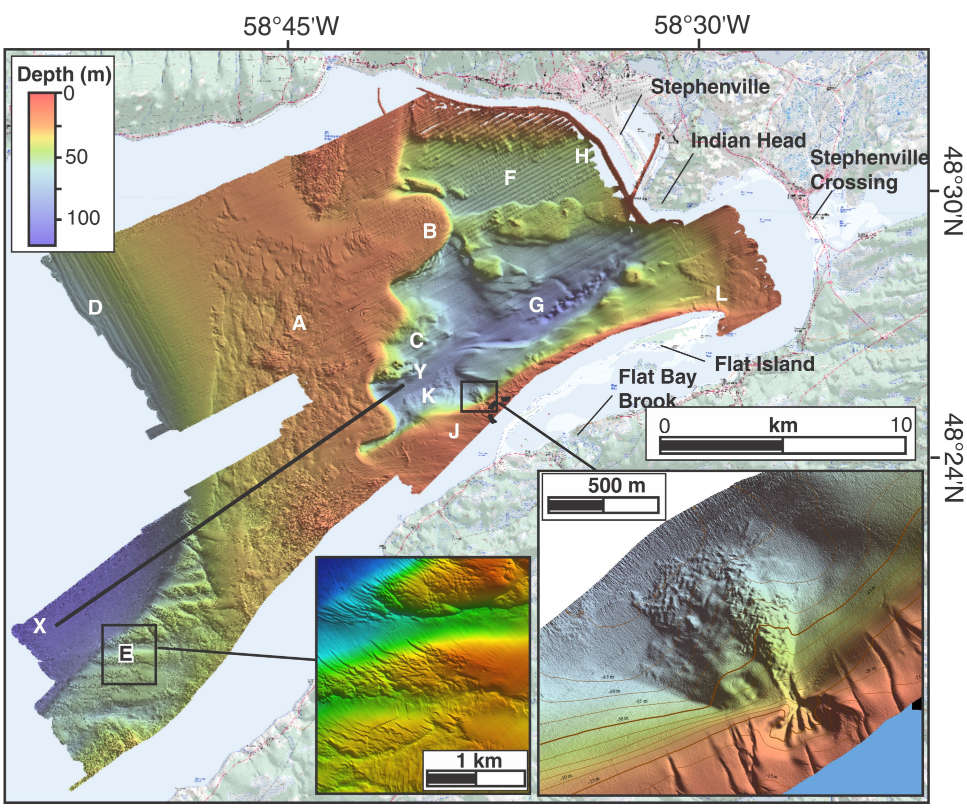


Figure 11. Multibeam imagery of inner St. George's Bay. Principal features are the sill (A), spillover deposits on the sill (B), and an area of sill collapse (C). To the west the glacial diamict on the moraine is buried by a seaward-thickening prism of sand that grades to sandy mud at depth (D). The continuation of the moraine to the southwest is buried by sandy bedforms (E). The basin off Stephenville (F) contains postglacial mud overlying thick glaciomarine deposits, as does the deep basin extending southwest from Stephenville Crossing (G); the latter also hosts a series of large pockmarks. The barrier platform (J) is flanked by submarine fans, one of which is shown in the inset map. Other features include a large area of mass transport (K) and fans at distal end of Flat Island spit (L).

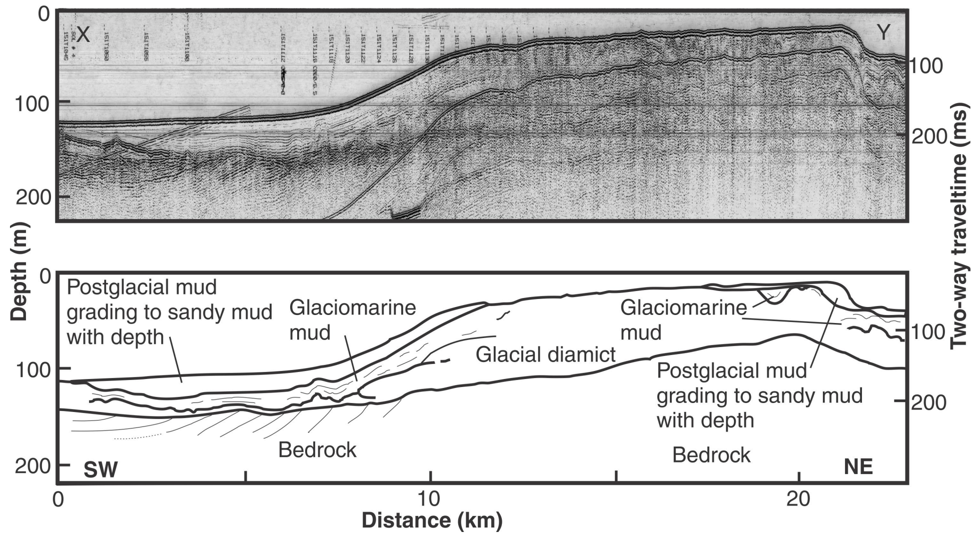


Figure 12. Airgun seismic-reflection profile of the moraine across St. George's Bay. Location shown on Figure 11.

Basins of the inner bay

The basins of inner St. George's Bay (Fig. 11, F, G) contain acoustically stratified deposits of glaciomarine mud up to 190 m thick in the deep channel extending southwest from the Stephenville Crossing area; thinner glaciomarine mud is draped over the interbasin area. The basins also contain postglacial mud up to 25 m thick, with faint acoustic stratification except where gas masking occurs (Shaw and Forbes, 1990a). Large pockmarks occur in the deep basin off Flat Island spit (G).

The subaqueous extensions of the barriers and spits described above are a dominant component of the very complex topography of the inner bay. A large barrier platform off Stephenville (Fig. 11, H), described above, comprises a seaward-fining prism of sand with a clinoform structure on acoustic records. The large sandy barrier platform off Flat Island spit (Fig. 11, J) similarly has a clinoform structure on acoustic records, and a break of slope that is –25 m in the southwest but –20 m at the distal end. Large shore-normal sand waves decorate the platform surface. Sand transport off the platform and into deeper water, perhaps during major storms when large amounts of sediment are suspended on the platform, is evidenced by a large area of mass transport (K) (*see* Shaw et al., 2006a, b, c) and several submarine fans with levéed channels. The inset on Figure 11 shows one of the fans. A resurvey in 2010 revealed no observed morphological change since the original 1995 survey, suggesting that fan evolution is episodic. A series of small submarine fans (Fig. 11, L) at the distal end of the spit is associated with spit extension. What cannot be seen on the multibeam imagery is the large early Holocene delta graded to –25 m off Flat Bay Brook. This feature, dated at 9820 ± 50 ^{14}C years BP (Bell at al., 2003), is entirely enveloped in barrier platform deposits.

Shaw and Forbes (1992) and Forbes et al. (1995c) showed that several deep bedrock valleys underlie the moraine and other Quaternary deposits. The 'Stephenville valley' extends southwest from Stephenville, and reaches 180 m below sea level. The 'St. George valley' runs southwest from beneath the Stephenville Crossing barrier, attains a depth of 195 m, and continues to the southwest below the moraine that comprises the sill.

Central and outer St. George's Bay

As seen on Figure 12, the thick Quaternary sediment deposits of the submarine moraine at the head of the bay thin seaward, so that in the central and outer bay, underlying bedrock structure determines gross seafloor morphology (Fig. 14). Southwest-trending bedrock ridges (A) in the central bay separate elongated basins (B) containing thin glaciomarine mud overlain by postglacial mud. In most places the bedrock ridges have a veneer of Quaternary sediments, heavily imprinted by relict iceberg furrows and pits, although outcrops do occur. The inset at bottom left on Figure 14 shows the high backscatter (dark tone) of a ridge and the patterned backscatter of its flank, where relict iceberg furrows and pits have infills of postglacial mud (light tone). Several small moraines (C) register halts in the retreat of late Wisconsinan ice up the bay, whereas the shallow area (D) near the coast is the southwest extension of the St. George's Bay moraine that forms the sill of the upper bay.

The bedrock structure evident at the seafloor is complex and is indicative of a series of anticlines and synclines, interrupted by northwest-trending faults. It includes a series of depressions, notably a 10 m deep trench and a circular depression (Fig. 14, inset at bottom right) that is 350 m across and 40 m deep. This depression is analogous to the submarine sinkholes of Great Bras d'Or, Cape Breton (Shaw and Potter, 2007). The terrain is part of the Mesozoic basin off western Island of Newfoundland (Lavoie et al., 2009). The Carboniferous rocks have been intruded by salt diapers, and belong to the Magdalen Basin salt-diaper zone (*see* Fig. 72 *in* Lavoie et al., 2009). In some areas there is strong

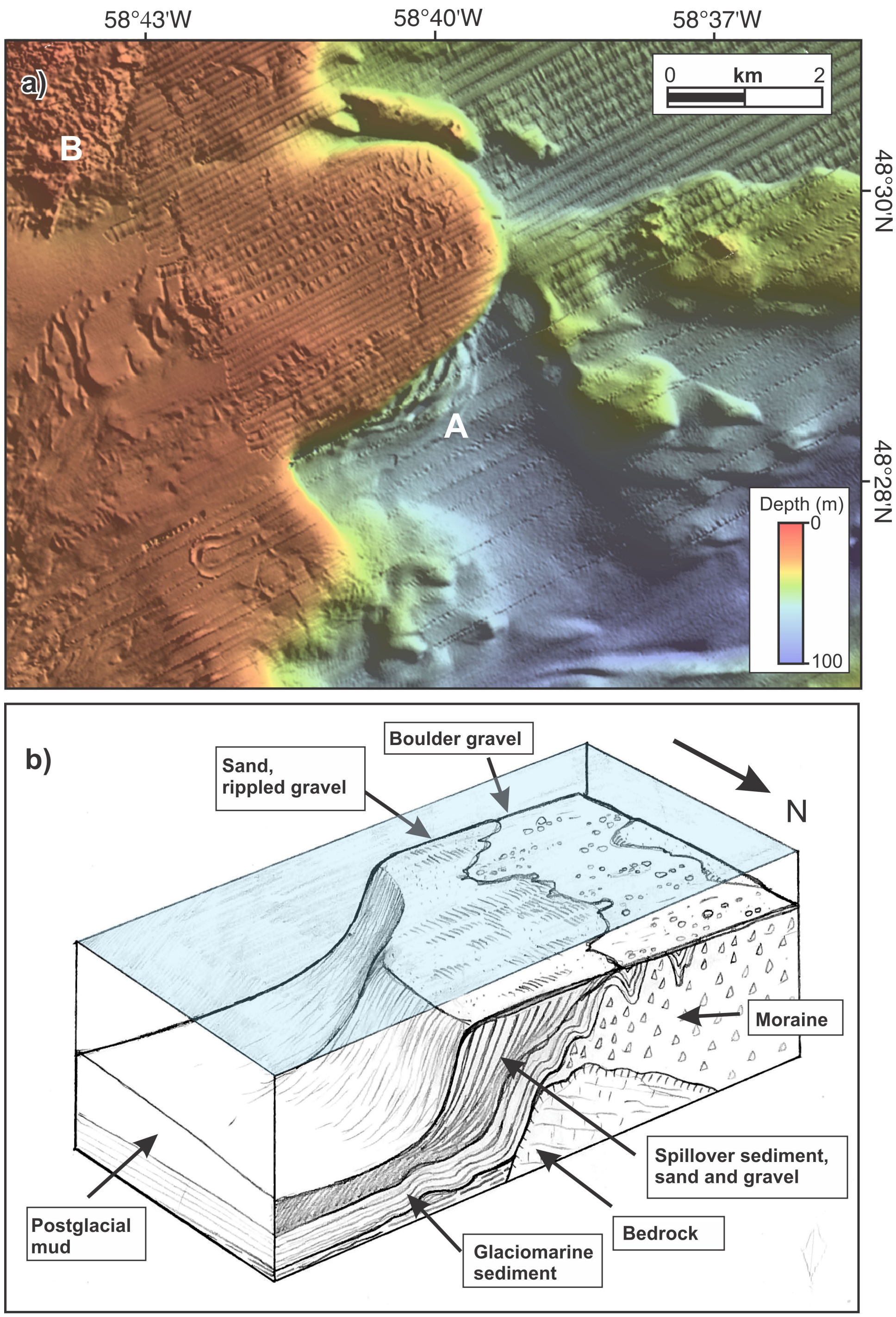


Figure 13. a) Enlarged view of the principal spillover lobe on the sill that extends across the bay. A series of ridges (A) are interpreted as resulting from compression of glaciomarine sediments by the advancing sediment lobe; **b)** illustration of the relationship between the spillover lobe and the moraine that extends across the bay. Much of the surface is sand and gravel (*see* Shaw et al., 2006b), but areas of glacial diamict are exposed at the seafloor (B).

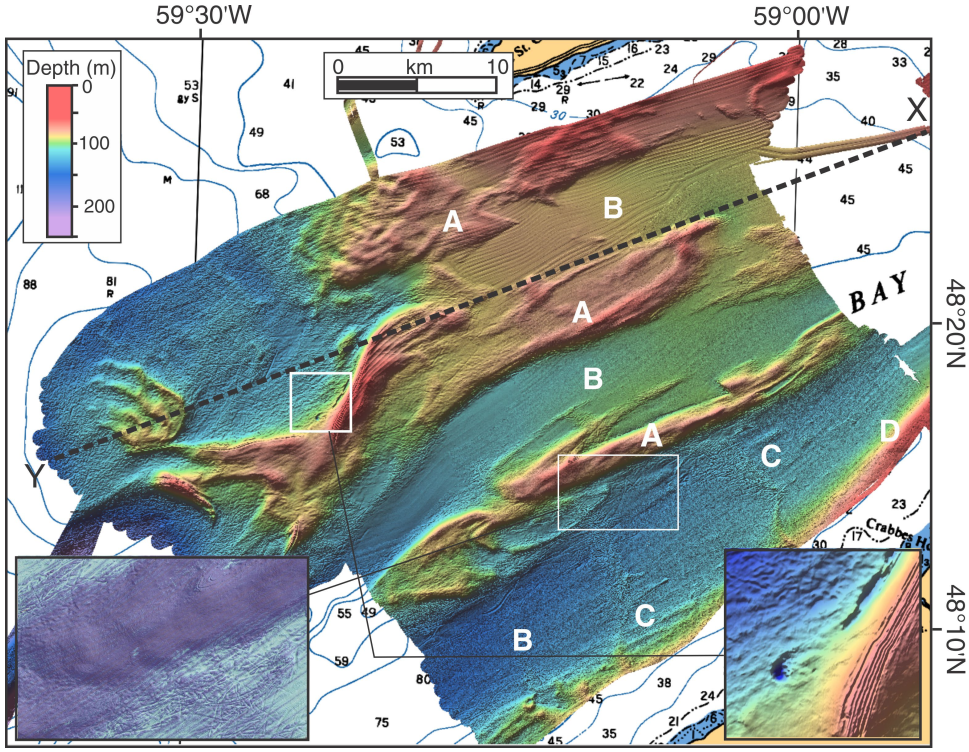


Figure 14. Multibeam image of the seafloor in outer St. George's Bay, showing bedrock ridges (A) separated by muddy basins (B). Several small moraines (C) record the retreat of an ice margin up the bay. The inset at bottom left shows backscatter strength, with high backscatter (purple) on the ridge and variegated backscatter in deeper water (where relict pits are partially filled with postglacial mud). The insert at bottom right shows the deep hole on the seafloor, interpreted as a solution-collapse feature. The submarine moraine in inner St. George's Bay extends to the southwest, and reaches the coast at 'D'. Base is from Canadian Hydrographic Service (2003b).

correspondence between bedrock structure in the multibeam imagery and the VG2 component of the recently collected aeromagnetic data (Dumont and Jones, 2013).

Figure 15 is an airgun seismic-reflection profile down the bay (location on Fig. 14) showing the presence of internal reflections in the bedrock. At the left side of the image the seaward edge of the prism of postglacial sediment is seen; this prism is also shown on Figure 12. Several morainal ridges overlying bedrock testify to halts in the retreat of ice up the bay, before the margin became established at the head of the bay.

REGION 2: CAPE ANGUILLE TO CAPE RAY

Setting

Encompassing the Humber and Dunnage tectonic zones, and several major northeast-trending structural elements (Long Range and Cape Ray faults), the coast of region 2 (Fig. 16) is nevertheless remarkably linear in a northerly direction. The shelf bordering the Laurentian Channel narrows southward: the 200 m isobath is 12 km offshore in the north but only 3 km offshore in the south. Eroding coastal bluffs composed of glacial sediments have supplied sediment to sandy barrier beaches off the Codroy River valley in the north, and to sandy beaches and coastal dunes between Cape Ray and Channel-Port aux Basques in the south. The inner shelf is characterized by widespread bedrock exposures and thin sheets of mobile sand. A significant feature in this region is a previously unrecognized submarine canyon, informally named the Cape Ray submarine canyon.

For descriptive purposes the region is divided into four sections. The northernmost section is offshore from the Grand Codroy River, and was mapped with multibeam sonar in the 1990s (grey shaded-relief image on Fig. 17). The remaining three sections (coloured image, Fig. 17) comprise a zone that was mapped more recently with multibeam-sonar and airborne-LiDAR ('Light Detection And Ranging') techniques in connection with a planned submarine electricity cable, resulting in complete coverage from land to sea. The

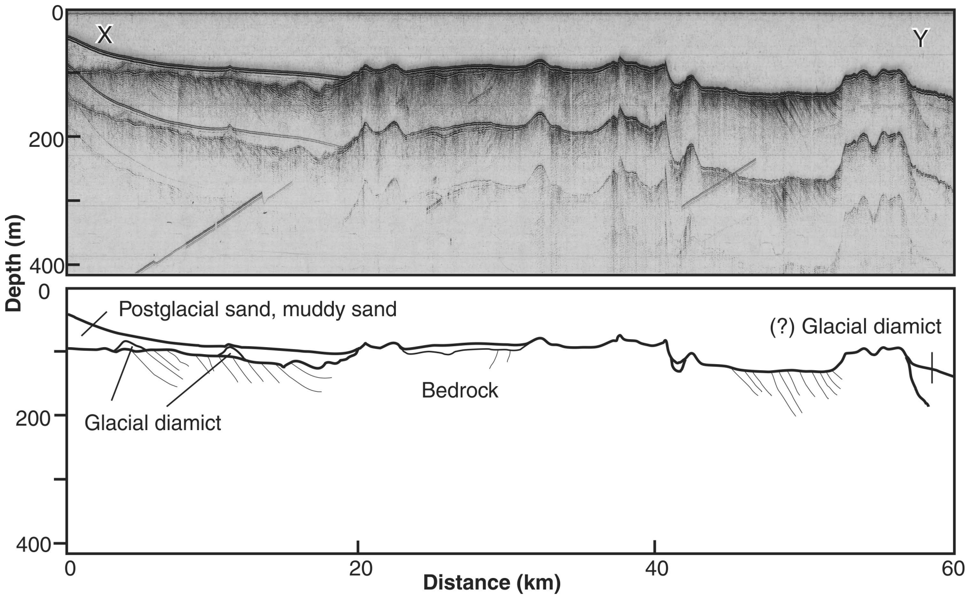


Figure 15. Airgun seismic-reflection profile down the bay, location on Figure 14.

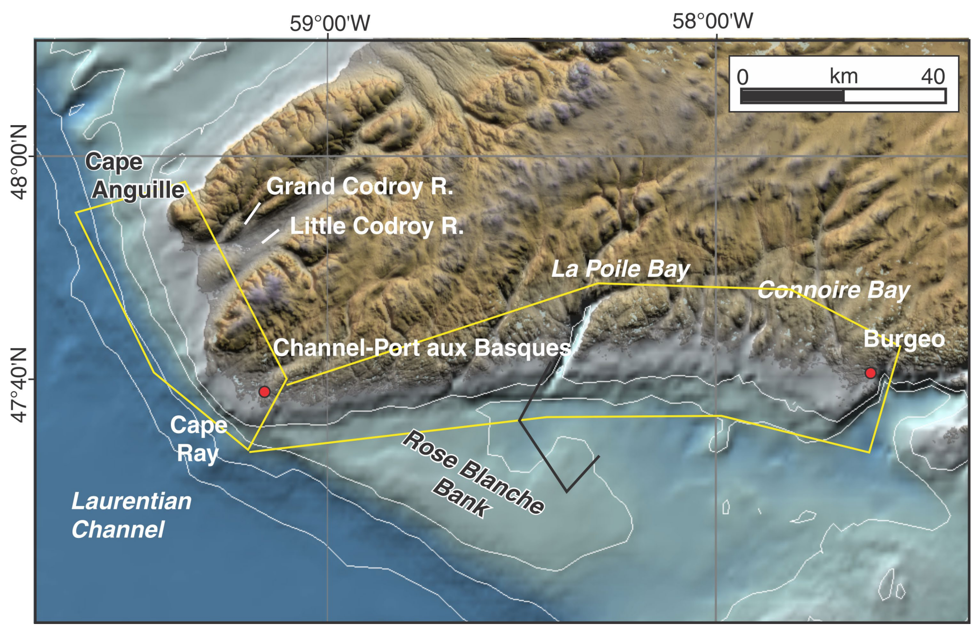


Figure 16. Location of regions 2 and 3.

resulting comprehensive data set offers a unique opportunity to glimpse the submarine components of terrestrial coastal systems such as barrier beaches and coastal dunes.

Coastlines

At the mouths of the Grand Codroy and Little Codroy rivers (Fig. 16), moraines noted by Grant (1987) intersect the coast, and supply sediment to sandy spit barriers. The northern Grand Codroy River barrier has well developed flood- and ebb-tidal deltas, whereas the Little Codroy River barrier has a well developed flood-tidal delta. Farther south, coastal bluffs contain glacial sediments, and bedrock is exposed at or just below sea level. In the south of the region, between Cape Ray and Channel-Port aux Basques, the coast is characterized by low bedrock headlands, sandy beaches, back-barrier lagoons, and the most extensive coastal dunes on the Island of Newfoundland (Catto, 2002, *see* his Fig. 2).

Grand Codroy River area

The area offshore from the estuary of Grand Codroy River (Fig. 18) was mapped in the 1990s to facilitate route planning for a fibre-optic cable between Island of Newfoundland and Cape Breton, Nova Scotia. Unfortunately, only images of shaded relief and backscatter strength were available for this study (i.e. no gridded data). Nevertheless, these images provide some insight into the offshore component of the

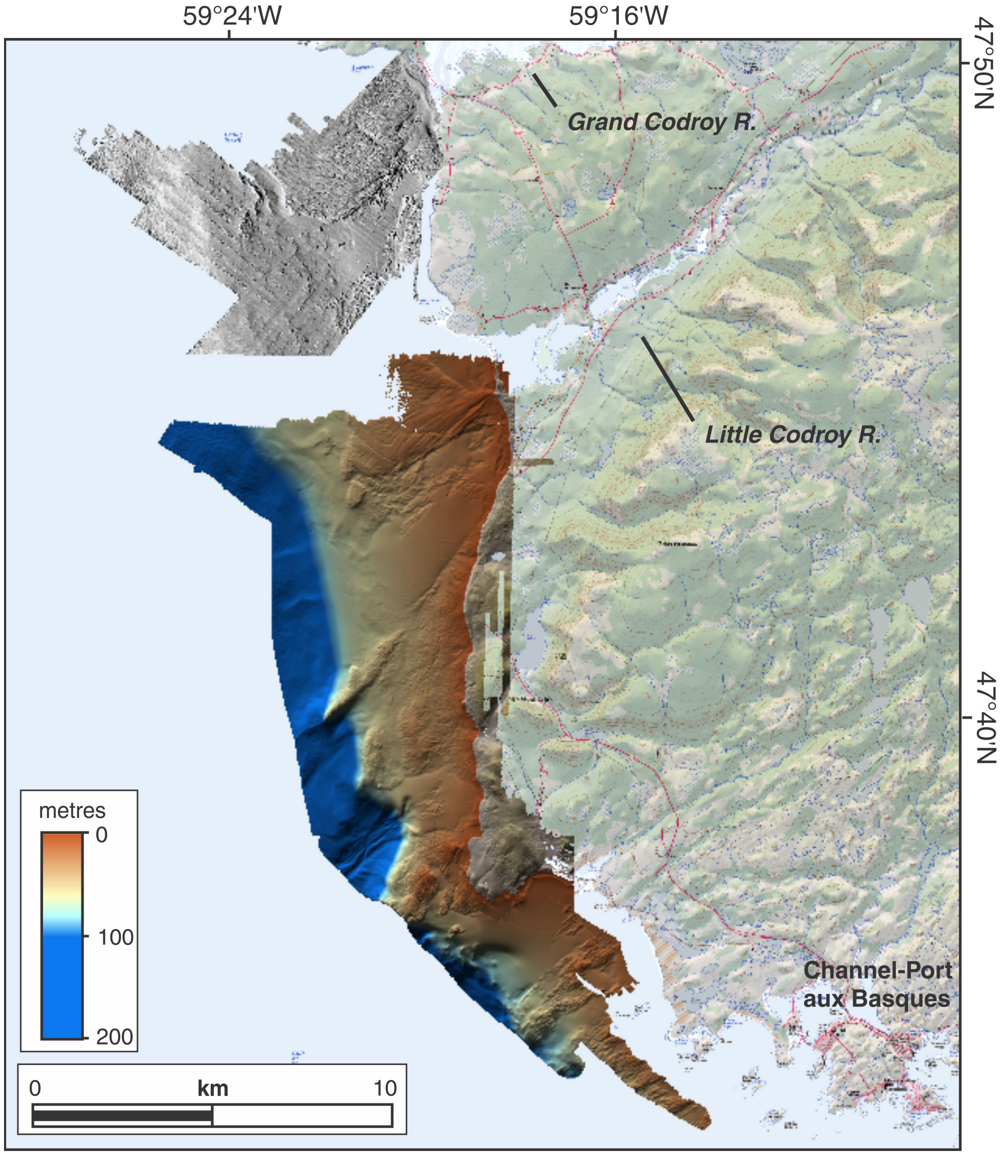
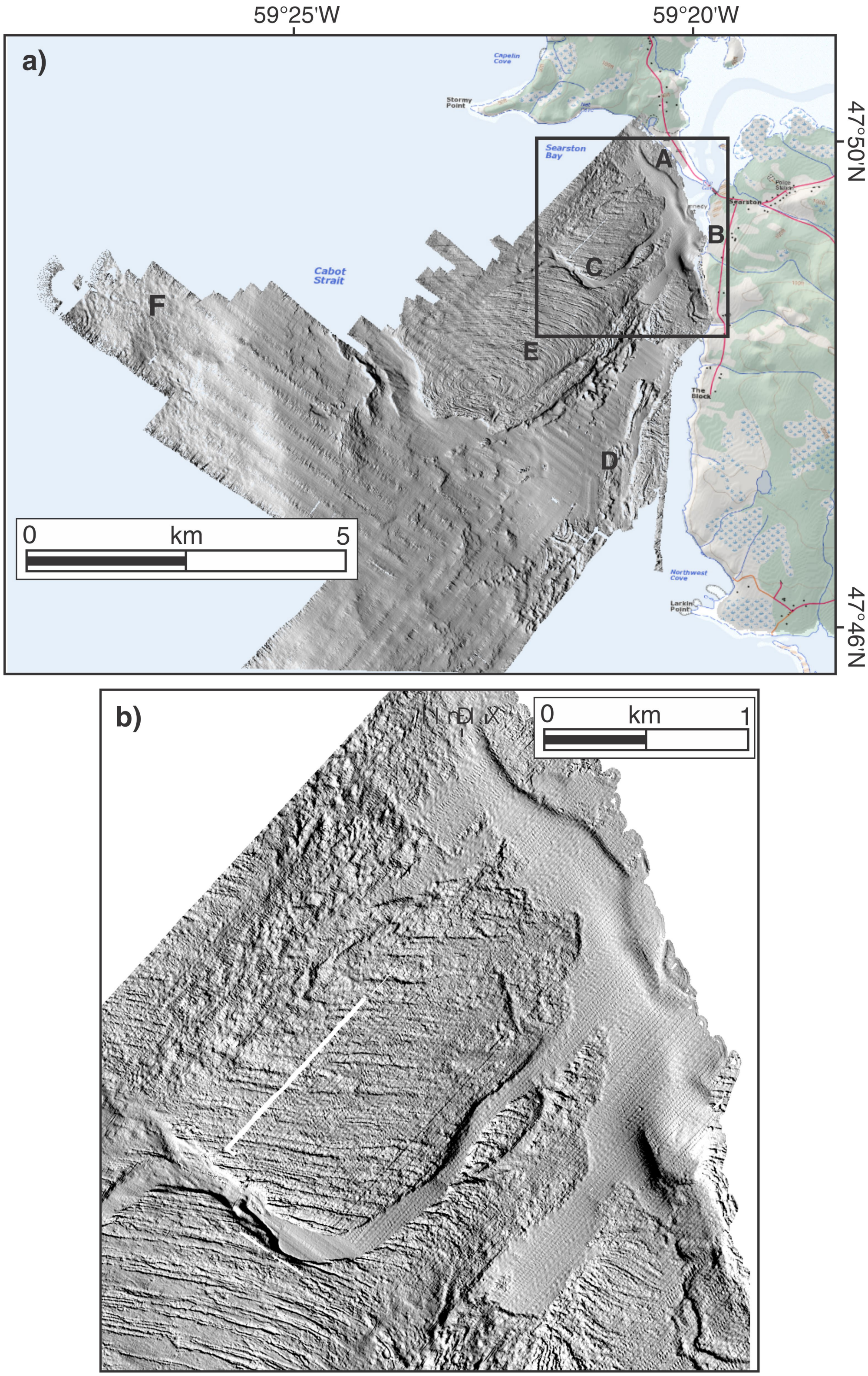


Figure 17. Shaded-relief images of terrain in zone 2. The grey image (top) shows the area surveyed in the 1990s in connection with a fibre-optic cable route. The coloured imagery shows an area mapped in 2011 in connection with a planned electricity cable crossing from Island of Newfoundland to Nova Scotia. Areas deeper than 20 m were surveyed with multibeam sonar, whereas shallower areas and adjacent land (brown) were surveyed with LiDAR.

Figure 18. a) Seafloor offshore from the estuary of Grand Codroy River, showing a meandering channel incised into bedrock, with more muted terrain farther offshore. Notable features include a nearshore sand prism separated from the coast except in the north where it may connect with the barrier beach offshore of Codroy (A). The connection with the ebb-tidal delta of Grand Codroy River (B) suggests a possible clockwise sediment transport pathway. The sand extends offshore in depressions and also on the floor of a valley (C) incised into folded Carboniferous bedrock (E) with well developed bedding planes and no Quaternary sediment cover. The meandering valley appears to be an extension of the Grand Codroy River valley onshore. Relict terrain turbated by icebergs occurs at F. **b)** Enlargement of the area immediately offshore from the estuary, showing the sandy offshore bar and the meandering sand-filled bedrock channel that extends offshore to depths of about –30 m. These data were not available for analysis in a grid format.


59°25'W
59°20'W
47°50'N
47°46'N
a)
Capelin Cove
Stormy Point
Searston Bay
Searston
Cabot Strait
The Block
Northwest Cove
Larkin Point
A
B
C
D
E
F
0
km
5
b)
0
km
1

nearby barrier beach. Unlike inner St. George's Bay, where Flat Island spit and the Stephenville strand plain have voluminous submarine counterparts, very thin unconsolidated sediment deposits lie offshore from the barrier beach at the mouth of Grand Codroy River. A nearshore sand bar (A in Fig. 18a) is separated from the adjacent beach by a zone of bedrock and thin sediment. This sand bar may connect with the beach at the north end and link with the ebb-tidal delta of Grand Codroy River at its south end (B), thus perhaps functioning as a sediment-transport pathway that cycles sand from the ebb delta to the beach and thence back to the ebb delta. Sand extends offshore, filling a meandering bedrock valley (C) that reaches depths of −30 m. A large area of smooth terrain (D) is sand. It is unknown whether the bedrock valley was formed during the early postglacial relative sea-level lowering that extended to about −30 m in this region (Shaw and Forbes, 1995) or was formed under glacial ice as a meltwater conduit.

A distinctive feature in this area is the large exposure of Carboniferous bedrock (E) with a strongly marked syncline or anticline. Farther offshore the seafloor has a smoother appearance, reflecting the thickening of Quaternary sediments. In about 150 m depth the Quaternary cover — probably glacial diamict (till) — is iceberg furrowed (F). As the water deepens toward the Laurentian Channel, the glacial diamict is progressively mantled by a seaward-thickening blanket of glaciomarine mud, and in the deepest water both units lie beneath thick postglacial mud. Backscatter data show that sandy sediments (low backscatter) are confined to the nearshore bar and ebb delta, the channel infill, and the smooth area (D); elsewhere high backscatter shows the presence of gravel or bedrock at the seafloor. Figure 18b is an enlargement of the area off the Grand Codroy River estuary, and shows the nearshore bar, ebb delta, and incised channel.

Northern section of the multibeam-LiDAR coverage

The inner shelf and adjacent coastal areas were surveyed with multibeam-sonar and LiDAR methods in 2011 (Shaw, 2012). The LiDAR coverage extends from land to approximately 30 m water depth, overlapping with the multibeam coverage that commonly extends landward to depths of 20 m. The resolution of the two data types is comparable, but for the purposes of this study backscatter information was extracted only from the multibeam data.

This area is depicted in three figures, showing colour-shaded relief (Fig. 19), backscatter strength (Fig. 20), and an interpretation (Fig. 21). On adjacent land areas a major structural lineation, the Long Range Fault (or Cabot Fault), separates Carboniferous rock in the north from granitoid intrusions to the south (Colman-Sadd et al., 1990). The feature is strongly marked on the VG2 component of the aeromagnetic data (Dumont and Jones, 2013). In depths shallower than about 50 m, much of the area in the north of the imagery is sedimentary bedrock, with well developed ridges that have relief of 1–2 m (A in Fig. 19), and sand and gravel between ridges. This bedrock has a crisp southern boundary that probably represents the offshore extension of the Long Range Fault; however, this lineation is not identifiable farther offshore, although Fader et al. (1989) thought that it was. The apparent fault running east-west at B is an artifact where the multibeam data overlie the LiDAR data. Nevertheless, many small displacement faults are evident in the data. A series of ridges (C) 40 m apart with relief of 1.5 m is interpreted as De Geer moraines, formed during ice retreat toward the coast and preserved because they lie below the estimated postglacial lowstand of relative sea level (−30 m according to Shaw and Forbes (1995)).

Several large sand bodies (Fig. 19, D, E, F) that extend offshore have low backscatter and are relatively thin (1–2 m). Sand sheet 'F' is separated from the coast by a bedrock zone with a minimum width of 200 m, but sand sheets 'D' and 'E' connect with systems of inner and outer nearshore bars (G, H). Nearshore bars (G) are located immediately offshore from beaches at the foot of coastal bluffs, with the seafloor between the bars and the shore probably comprising gravel. Farther north, the nearshore bars (H) are crescentic in form, and are located offshore a barrier beach. The anchor (J) for the south end of this barrier appears to be the bedrock that extends farther offshore.

In depths over about 50 m, the sand sheets are absent and the bedrock is buried by Quaternary unconsolidated sediment with high backscatter, probably glacial diamict (Fig. 19, K). In the extreme northwest, the Quaternary sediments are imprinted by relict iceberg scours (M). A large area of smooth seafloor in deep water with low backscatter (N) is interpreted as postglacial sandy mud that overlies the glacial deposits.

In the backscatter-strength imagery (Fig. 20) high backscatter strength has a green-blue tone, and low backscatter strength is pink. Note that backscatter values are not available for the LiDAR data. The crisply defined edge of the low-backscatter strength thin sand sheets is very evident. All other terrain types have high backscatter strength. Figure 21 provides a summary of the surficial geology, and also shows the location of a submarine channel that is partly filled with sand or sandy gravel, and likely continues seaward (dashed line) under the sand sheet. This channel either formed under glacial ice, or during the postglacial lowering of relative sea level.

Central section of the multibeam-LiDAR coverage

In this area (Fig. 22, 23, 24, 25) bedrock extends offshore for distances of 2–3 km, except where a large bedrock ridge extends much farther. In the southern half of this area, however, a large offshore sand body (A in Fig. 22) approaches to within 100 m of the coast. This sand sheet is estimated to be generally 2 m thick. About 1 km offshore it thins, and forms

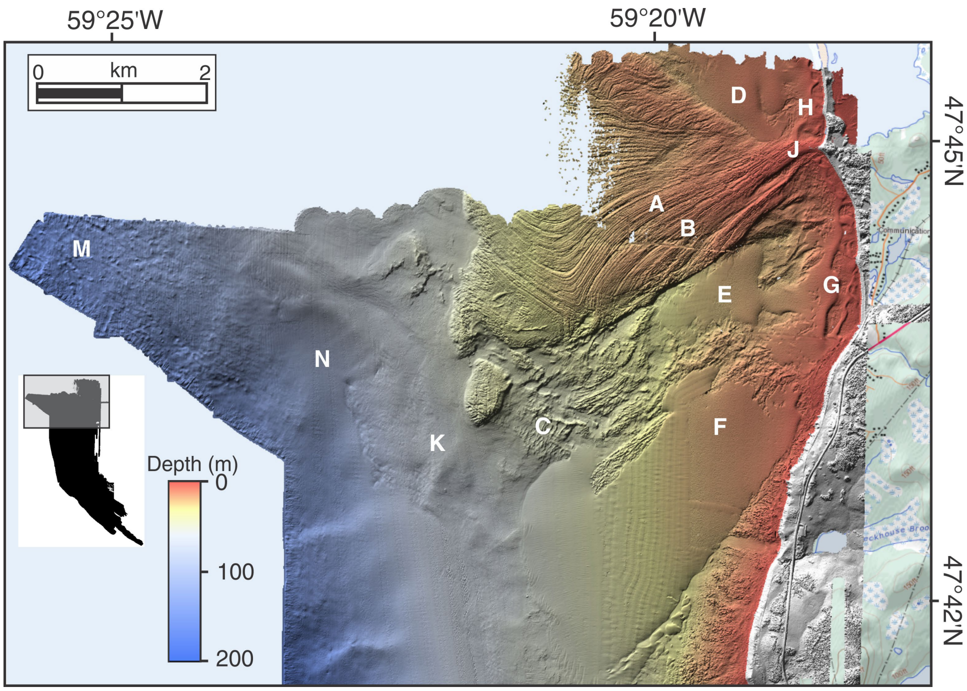


Figure 19. The northern part of the multibeam coverage in zone 2, with land areas coloured grey. Features include bedrock (A), an apparent fault (B) that is an artifact of the boundary between multibeam and LiDAR data, possible De Geer moraines (C), sand bodies (D, E, F), nearshore bars (G, H), and a bedrock ridge (J) that anchors the adjacent barrier beach.

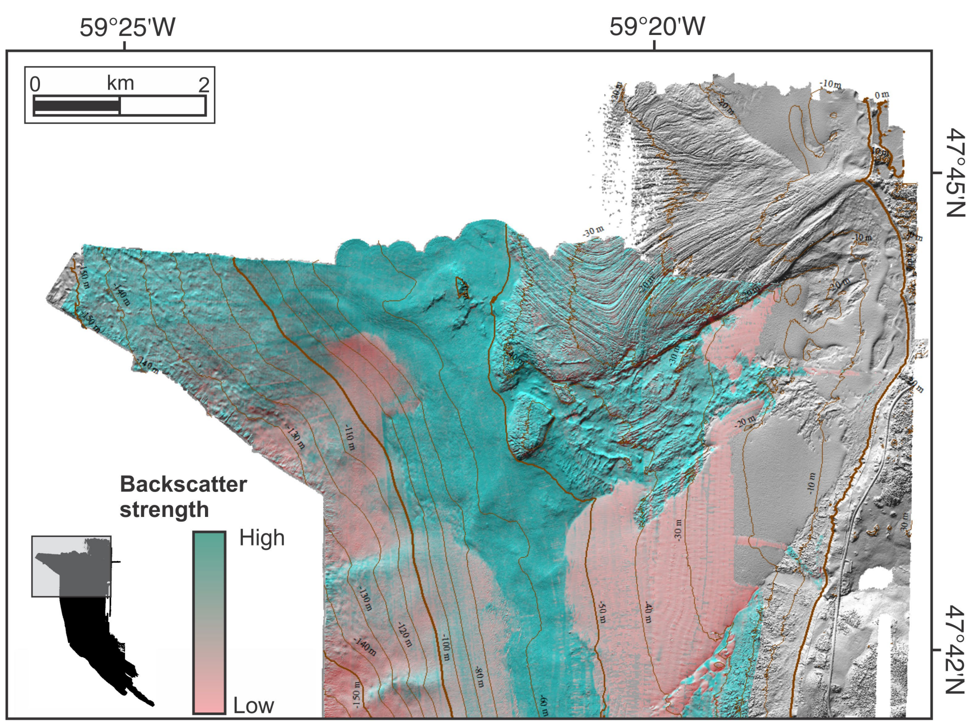


Figure 20. Backscatter of the northern area. Backscatter is not available for the LiDAR area (right).

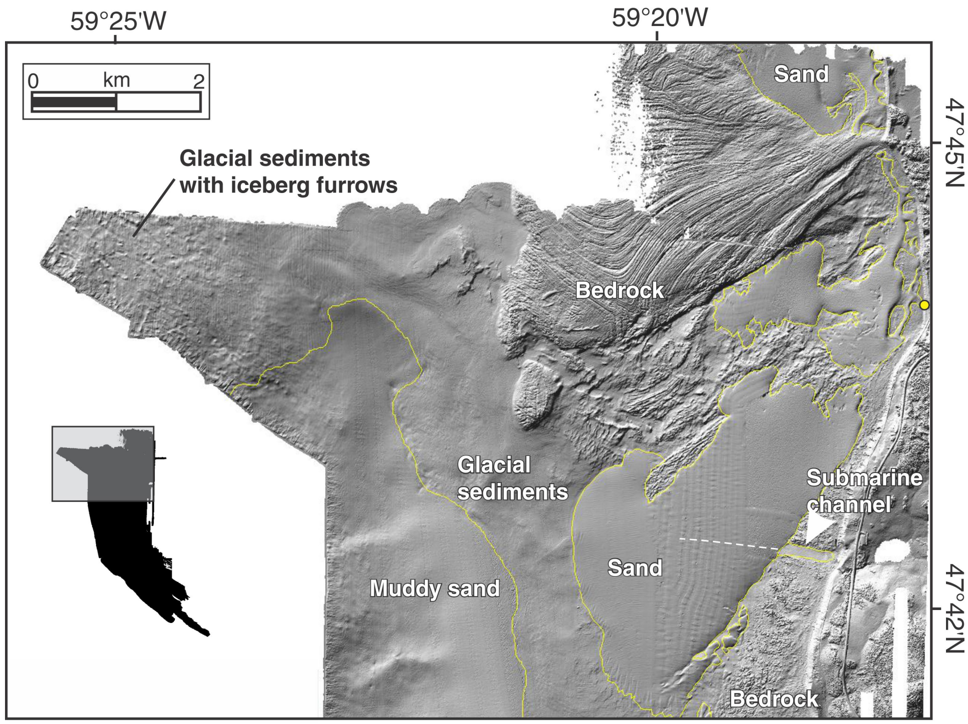


Figure 21. Interpretation of the northern area. The white dashed line suggests the westward extension (under the sand sheet) of the submarine channel that is present near the coast.

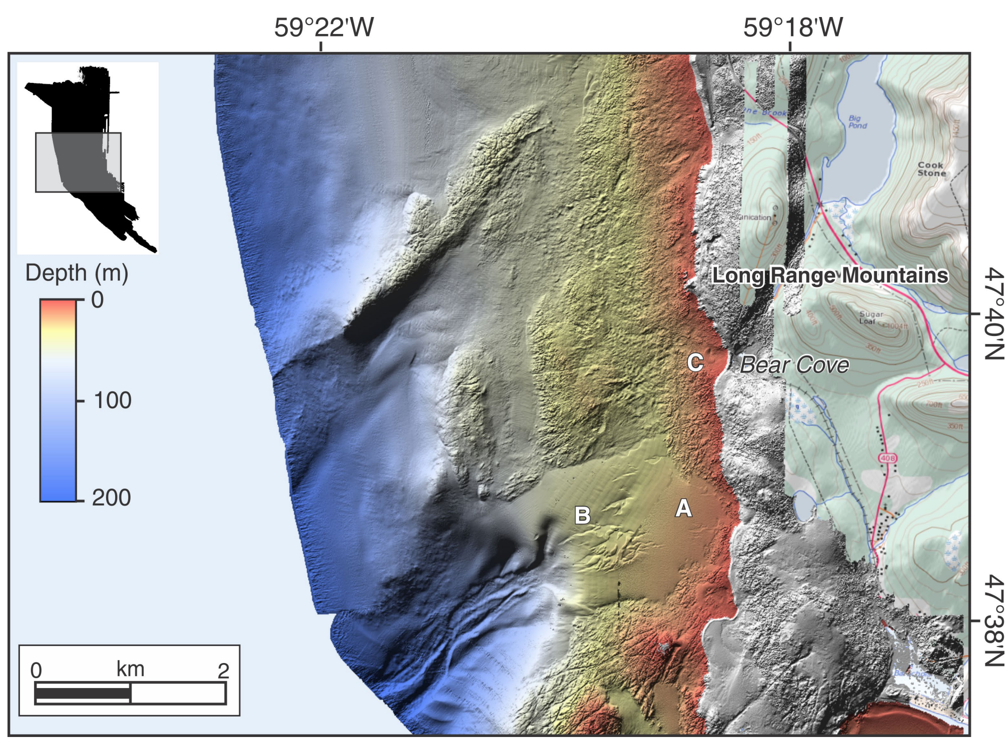


Figure 22. Shaded-relief image of the central area in zone 2, showing the large offshore sand body (A), depletion of the sand body (B) by loss of sand into the submarine canyon, and small, nearshore sand body (C) off Bear Cove.

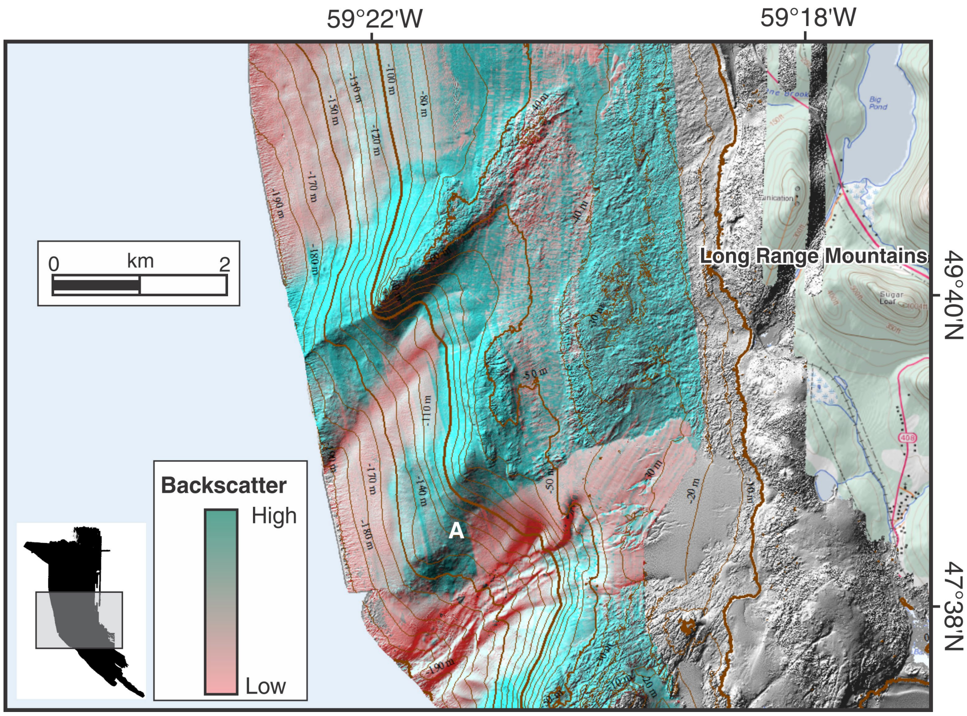


Figure 23. Backscatter of the central area. The sharp boundary (A) is a processing artifact.

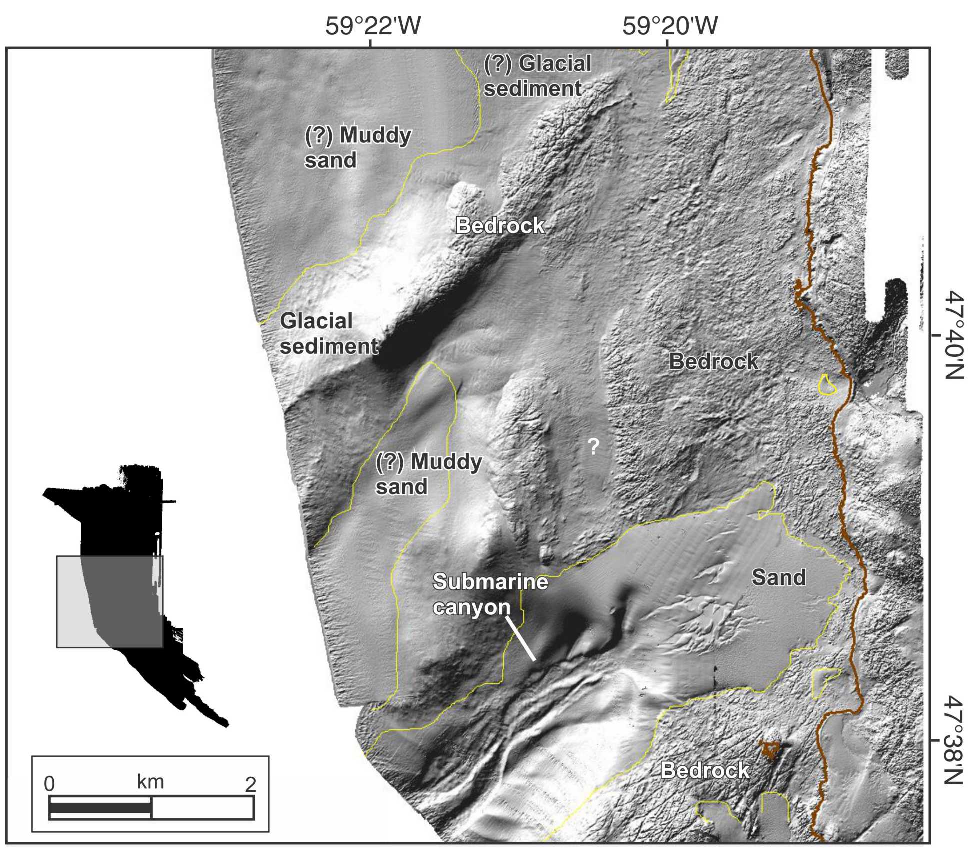


Figure 24. Interpretation of area 2. Coastline indicated by brown line.

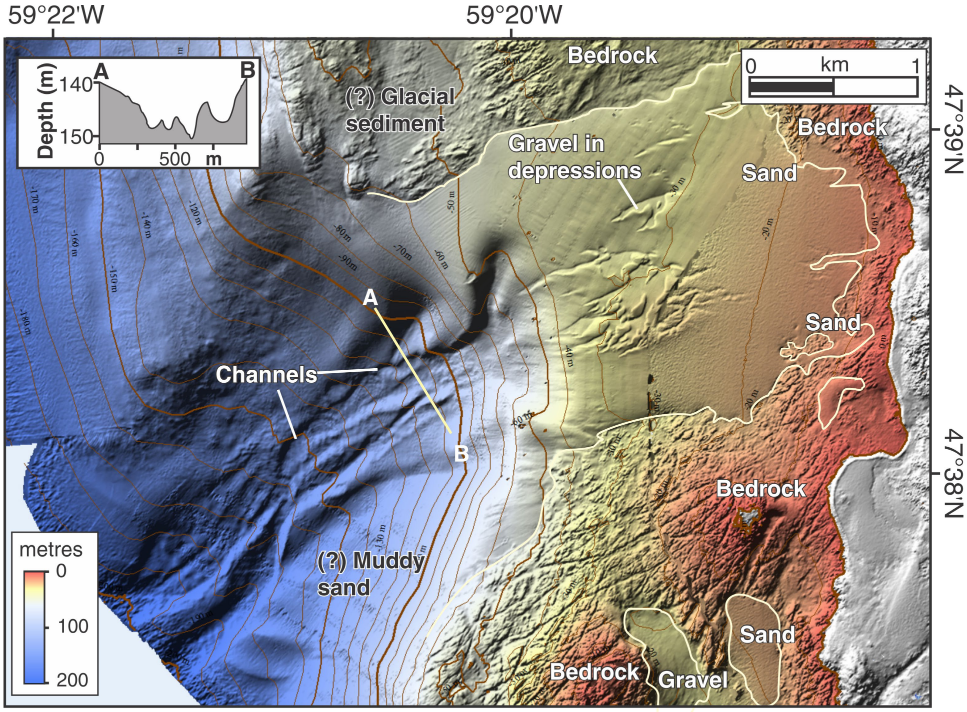


Figure 25. Colour shaded-relief map of the submarine canyon near Cape Ray.

a series of 'fingers' of sand separated by a gravel seafloor with high backscatter (B). This area of uneven topography extends into deeper water, and connects with a submarine canyon, hereby informally called the Cape Ray submarine canyon. This feature is discussed in more detail below. Close to the coast, a very small nearshore sand bar (C) is located in the coastal embayment of Bear Cove, where a small barrier beach encloses a lagoon. Farther offshore the smooth seafloor indicates either glacial sediment (high backscatter) or postglacial sandy mud and/or muddy sand (low backscatter).

The backscatter-strength image (Fig. 23) shows the distribution of sand and finer sediments (pink) verses harder substrates (dark tone). The sharp boundary (A) is an artifact of data processing. No backscatter strength data are available in the area of LiDAR coverage, but it is nevertheless clear that the large sand body in the lower part of the coverage extends close to the coast. On the surficial geology summary (Fig. 24) the extent of this sand is shown. In most areas, however, bedrock areas extend right up to the coast.

The Cape Ray submarine canyon (Fig. 25) extends across the inner shelf, with a break of slope at –40 m. The canyon has low backscatter strength (*see* Fig. 23) indicative of sand or mud. It is comprised of several channels with slopes of 3°, incised to depths of about 5 m and containing transverse ridges 40 m apart with relief of about 1 m. The canyon appears to connect with the large, thin (2 m) sand sheet that extends to near the coast. In the vicinity of the canyon headwall the sand sheet thins into a series of 'fingers' of sand, and isolated small sand bodies, that are separated by a lag gravel. This topography is suggestive of sediment starvation. A large headwall area connects with a series of channels that extend downslope. A protrusion of the 335 m isobath on CHS chart 4015 (Canadian Hydrographic Service, 2003a) suggests that a submarine fan is developed in the deeper water, beyond the coverage.

The discovery of this canyon on the continental shelf is of interest because its only analogues occur on the edge of the continental shelves of Atlantic Canada. Piper and Normark (2009) recognized three major initiation processes on submarine canyons: transformation of failed sediment, hyperpycnal flows from rivers or ice margins, and resuspension of sediment near the shelf edge by oceanographic processes. The third mechanism seems most likely here (Piper and Normark, 2009, p. 353). The recurrence interval for sand moving down the canyon is unknown. Possibly the sand body adjacent to the canyon head has been the source of sediment moving down the canyon. This sand body is not presently connected to any sediment source (e.g. eroding coastal bluffs), and may have been depleted.

Southern section of the multibeam-LiDAR coverage

This region is shown on Figures 26, 27, and 28. The Cape Ray fault zone, a Late Silurian to early Devonian reverse-oblique fault zone, occupies part of an embayment (Colman-Sadd and Scott, 1994) (Fig. 29). To the north (exposed at Cape Ray) are rocks of the Ordovician Cape Ray igneous complex, to the south is the Port aux Basques gneiss (Silurian and older), and in between a variety of lithologies of the Windsor Point Group in the fault zone (Dubé and Lauzière, 1995).

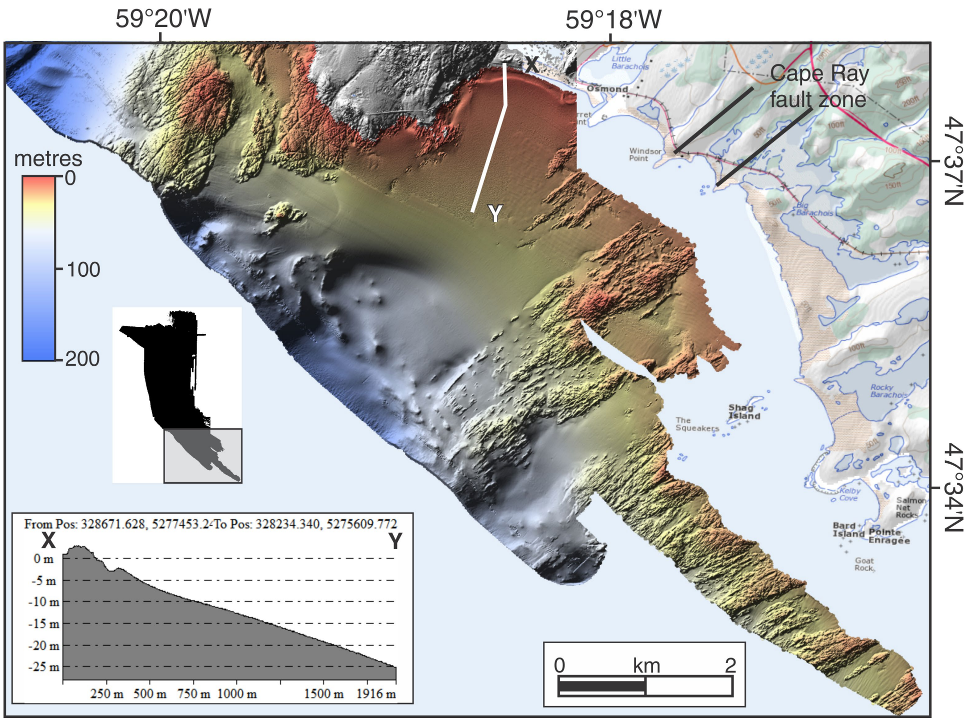


Figure 26. Southern part of the coverage in zone 2.

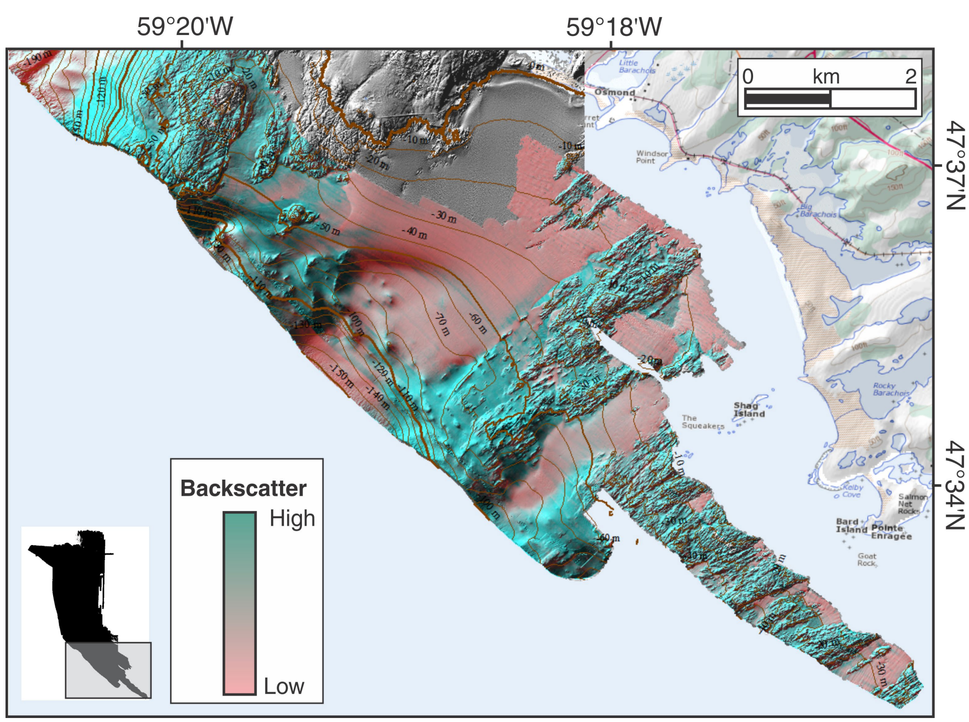


Figure 27. Backscatter of the southern part of the coverage in zone 2.

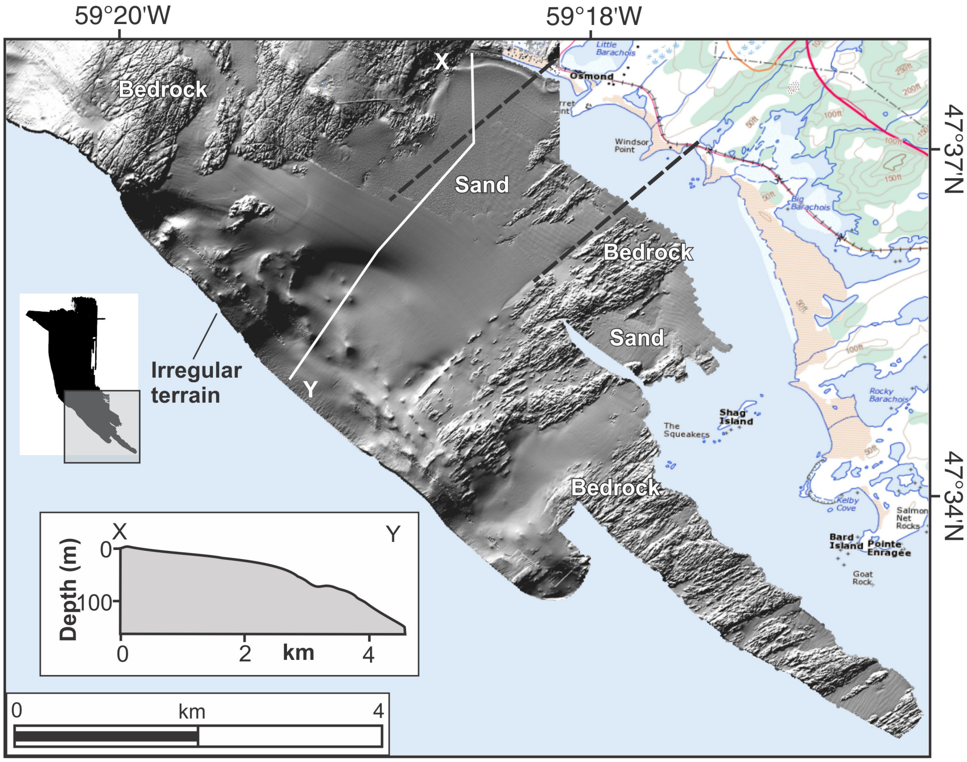


Figure 28. Interpretation of the southern area.

In this region large volumes of sediment are present onshore, in coastal dunes, beaches, and inlets (Catto, 2002). Multibeam and LiDAR imagery shows large volumes of sand offshore also. Immediately east of Cape Ray, at Cape Ray Cove, a sandy barrier beach encloses a large lagoon that connects to the ocean via a channel at the west end of the beach. The beach appears to be overwashed regularly, most likely when major storm waves approach from the south, and over the long term is probably migrating landward. Offshore from this barrier beach is a sandy nearshore bar that is 1.5 m high. The sand continues offshore as a prism perhaps more than 10 m thick, and may be as much as 20 m. The sand in the offshore in this area, therefore, appears to be greater in volume and thickness than in areas farther north. The relatively large amounts of sand offshore from the beaches and coastal dunes in this subregion appear to challenge the assertion by Catto (2002, p. 30) that sand from the beaches is "insufficient to replace that removed through natural and anthropogenic causes". On the other hand, there are no eroding coastal bluffs supplying material to the beaches south and east of Cape Ray, nor is there evidence of a pathway offshore, around the cape. Thus, Catto's (2002, p. 17) conclusion that under present conditions development of eolian dunes is hindered may be correct.

Bedrock predominates offshore just west of Cape Ray and also to the southeast of the study area. An offshore sand body comes within 250 m of the coast, but is limited in extent. Two kilometres south of Cape Ray the nearshore sand prism thins over a short distance. The seafloor steepens, with a break of slope at 45 m depth. Seaward of here the seafloor is irregular: numerous bedrock pinnacles come close to the surface, and are separated by a smooth seafloor probably comprising sand overlying glacial sediments.

REGION 3: CAPE RAY TO BAY DE LOUP

Setting

Region 3 (Fig. 16) lies in the Gander tectonic zone (Colman-Sadd et al., 1990), and includes the boundary between basement rocks and Carboniferous sediments that predominate farther offshore. Offshore from the steep, rocky, east-west oriented coast, basement rocks extend up to 8 km offshore, forming a shallow (mostly <100 m deep), irregular, wave-dominated rocky platform with pockets of sand and gravel between outcrops. Several fiords intersect the coastline, the largest of which is La Poile Bay. They contain glaciomarine and postglacial mud deposits. The outer margin of the platform has a continuous, shore-parallel moraine with typical crest depths of 150 m. Fluted till ridges with relief up to 20 m and oriented normal to the moraine extend to the north for several kilometres. Beyond the edge of the littoral platform the seafloor drops off steeply, and in deeper water relatively thick Quaternary sediments are found — glacial diamict overlain by glaciomarine and postglacial mud.

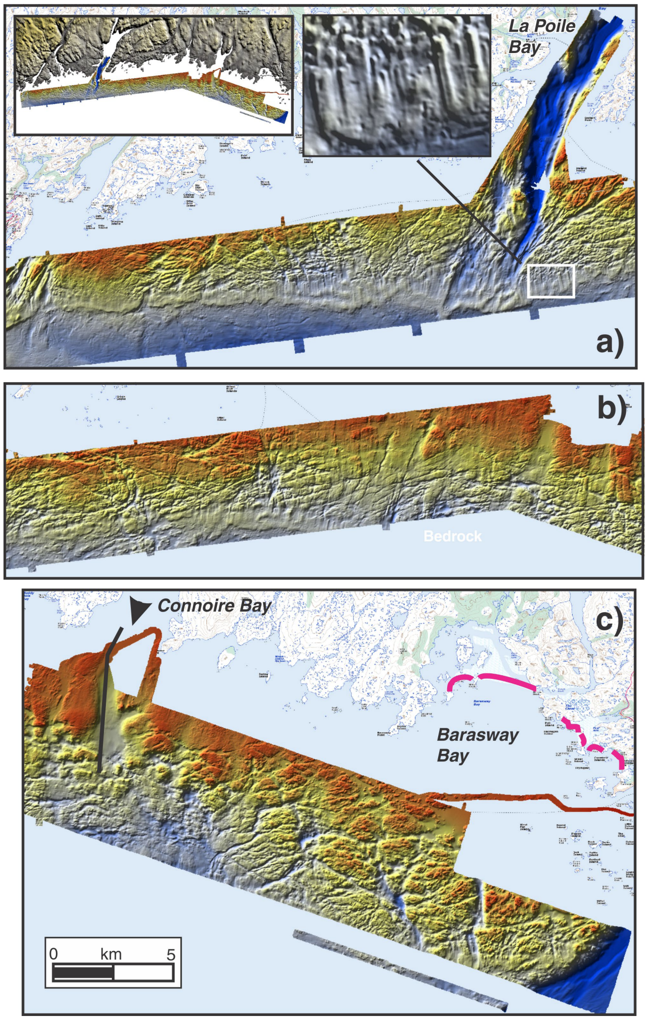


Figure 29. a), b), c) Multibeam coverage in zone 3, separated into three adjoining sections. The inset at left on Figure 29a shows the complete coverage, and the other inset shows an enlarged view of fluted till in the area of the white box. The northern part of the sandur plain off Connoire Bay (arrow in Fig. 29c) was not surveyed with multibeam sonar, but the southern part shows up clearly in the multibeam image—the black line indicates the profile on Figure 31. Red lines on Figure 29c indicate the numerous sandy beaches off Barasway Bay and the Burgeo area.

Coastlines

Coastlines in this region were imaged in the aerial video survey of Shaw and Frobel (1992). The coast is low and rocky, with little vegetation except in major inlets. A few gravel pocket beaches are found in the west of this region, and a small spit occurs in Connoire Bay. Large volumes of unconsolidated sediment at the shore and immediately offshore are found in the 10 km long stretch of coast just west of the settlement of Burgeo, where sandy barrier beaches and sets of dune ridges connect low rocky headlands (Catto, 2002). The largest barrier system is at Barasway Bay. Gaps in the beaches link shallow back-barrier lagoons with the ocean.

Inner shelf

The inner shelf (Fig. 29) is divided into coast-parallel zones of strongly contrasting character. The innermost zone is underlain by basement rocks, extends 7 km seaward on average, to a depth of about 100 m, and is distinguished by rugged bedrock outcrops with relief of 20 m separated by pockets of sand and gravel (Fig. 29). Forbes and Shaw (1989) reported that this terrain prevailed off Barasway Bay, despite the presence nearby of the largest sandy barrier beach on the south coast of the Island of Newfoundland. Evidently, as off Codroy, the sediment prism associated with the barrier does not extend far offshore.

At the seaward margin of the rugged bedrock topography is a belt of moraines that is between 1.0 km and 1.5 km wide. The acoustically incoherent glacial diamict averages 20 m thick, and has been grooved by the grounded ice, producing parallel flute-like ridges commonly 150 m apart with relief up to 20 m (*see* inset on Fig. 29a). The outermost part of the moraine belt is a 10 m high shore-parallel ridge. Shaw et al. (2000b, 2009) and Shaw (2003) argued that this moraine, located at the margin of basements rocks, was coeval with arcuate fiord-mouth moraines east of Bay de Loup (where the basement margin was at the coast), and thus dates to ca. 14 ka. It formed when glacier ice formerly far out on the shelf retreated to a temporary margin in shallow water.

Beyond the morainal belt the steep slope that is the boundary of the basement rocks has a veneer of Quaternary sediments, and is imprinted by relict iceberg scours and grounding pits, visible on the multibeam imagery. In depths more than 200 m the seafloor contains glaciomarine mud with a thin cover of postglacial mud. The cross-section shown on Figure 30 extends from Rose Blanche Bank into La Poile Bay. It shows acoustically stratified bedrock, likely Carboniferous in age. The bank has a veneer of iceberg-turbated Quaternary sediments.

The inner zone is crossed by several fiords, the largest of which, La Poile Bay, attains a depth of 358 m near its mouth. La Poile Bay contains Quaternary sediments up to 160 m thick, primarily glaciomarine mud overlain by postglacial mud, gas charged in places. Piston cores 020 and 022 collected on survey CCGS *Hudson* 2000-030B (J. Shaw and P.R.G. Girouard, unpub. cruise report, 2000) targeted the glaciomarine sediments using seismic data from cruise 91026 (Shaw et al., 1992). Core 22 penetrated 7.61 m into thick deposits of acoustically stratified glaciomarine sediment in a perched basin on the flanks of the fiord. The core intersected silty clay with sandy layers and scattered angular clasts. Radiocarbon dates have not been obtained on the many gastropods and bivalves noted in the core.

La Poile Bay also contains one of the series of submerged deltas along the southwest coast described by Shaw and Forbes (1995). These increase in depth westward, indicating an increase in the depth of the postglacial relative sea-level lowstand. Shaw and Forbes (1995, their Fig. 8) shows the

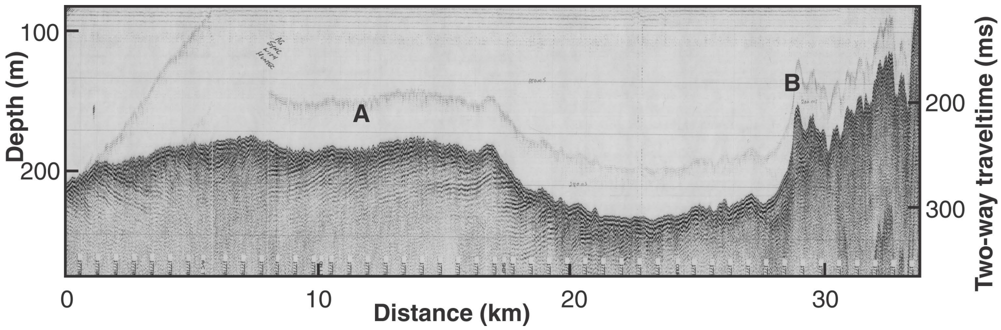


Figure 30. Airgun seismic-reflection profile across Rose Blanche Bank (A), where carboniferous rocks are overlain by a veneer of turbated sediment, into a basin with thicker Quaternary sediments, then up onto the basement platform with a moraine (B) at its outer limit.

acoustic structure of the delta at the head of La Poile Bay. A date of 6320 ± 60 ^{14}C years BP was obtained from organic detritus topset beds, cored with a vibracore (TO-3174).

Connoire Bay, the second largest embayment on this stretch of coast, does not have the fiord characteristics of La Poile Bay. By contrast, the seafloor in Connoire Bay and immediately offshore contains thick deposits of sand and gravel, described by Shaw and Forbes (1995). A prism of sediment extends offshore for 4 km from the mouth of the bay, occupying a bedrock trough 175 m deep. Cross-profiles inside the bay show Quaternary sediments up to 60 m thick. The floor of the bay is smooth and sandy. As shown on Figure 10 of Shaw and Forbes (1995) the prism slopes seaward to a break of slope at –44 m, and pinches out at 5 km from the mouth of the bay. Figure 31 shows this feature. Surficial sediments are either sand or rippled gravel. This sediment body was interpreted as a sandur plain reworked by falling sea level in the early Holocene. Shell material in a vibracore at 84 m depth was dated at 8650 ± 70 BP (TO-3175). This date includes a 410 year reservoir correction applied by the laboratory; the conventional age is 9060 ± 70 ^{14}C years BP.

REGION 4: BAY DE LOUP TO HERMITAGE BAY

Setting

Post-Ordovician intrusive rocks predominate onshore in region 4 (Colman-Sadd et al., 1990). The linear coastline (Fig. 32) trends east-west, as in region 3, but is intersected by numerous fiords incised into a plateau with relief of 200–300 m at the coast. The region includes Bay d'Espoir, a large fiord with the deepest water on the Island of Newfoundland continental shelf. Several aspects distinguish this region from the preceding Port aux Basques to Bay de Loup region. First, the boundary between basement and younger rocks is located at the coast rather than about 7 km offshore. This results in a very bold, rocky coast with deep water just offshore, and the absence of a wide shallow zone with mobile sand and rocky outcrops. This is seen on the surficial geology map (Fader et al., 1982) that shows a strip of Sable Island Sand and Gravel Formation west of the Burgeo area, but not to the east. Thick deposits of glaciomarine and postglacial mud lie offshore, the latter heavily pockmarked off White Bear Bay.

Second, after a major calving episode (Shaw et al., 2006d), glaciers stood at the fiord mouths in this zone ca. 14 ka, whereas to the west, the margin stabilized at the edge of the platform formed by the basement rocks (Shaw et al., 2000b, 2009; Shaw, 2003). This stillstand created the many fiord-mouth moraines in the region.

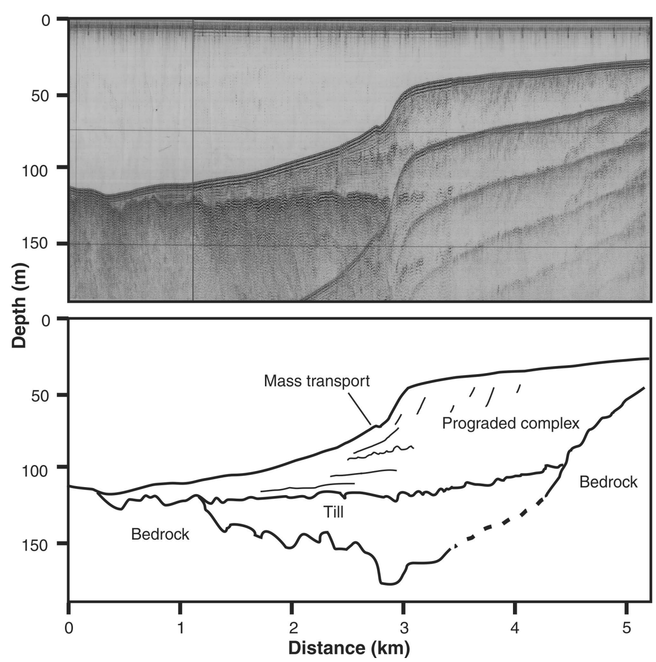


Figure 31. Huntec DTS seismic-reflection profile and interpretation across a submerged sandur plain off the mouth of Connoire Bay.

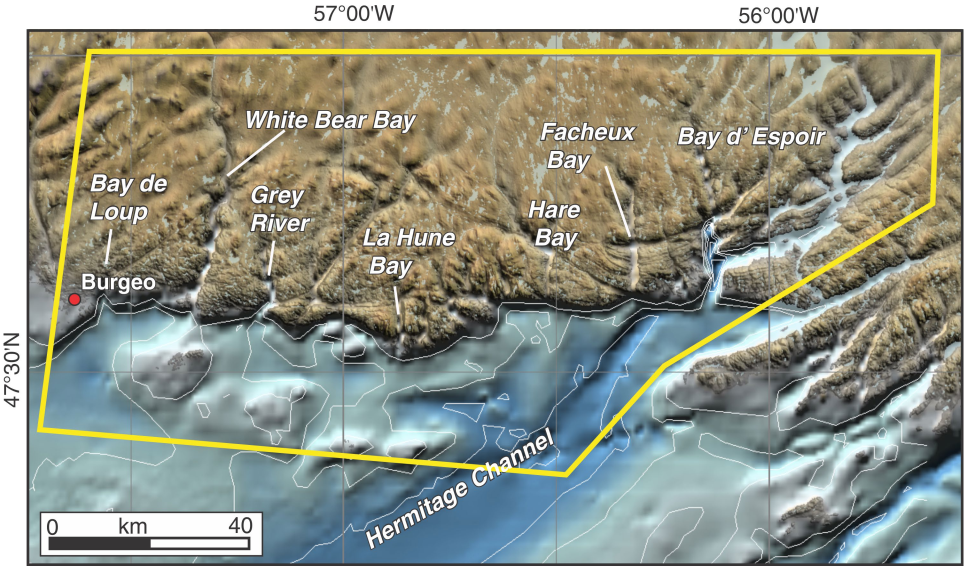


Figure 32. Location map for zone 4, which extends from Bay de Loup to Bay d'Espoir.

Coastlines

As can be seen in the GSC aerial video (Shaw and Frobel, 1992), this region has steep rocky coasts devoid of vegetation, with few beaches. Raised marine terraces at the fiord heads have been incised by rivers, and have supplied sediment for the construction of early Holocene deltas that are now submerged (Shaw and Forbes, 1995). One characteristic of the outer coasts is the presence of a vegetation-free zone some considerable distance above sea level, presumably because of sea-spray effects.

Inner shelf: the fiords

From west to east, the principal fiords (maximum depths in parentheses) are: Bay de Loup (421 m), White Bear Bay (272 m), Bay de Vieux (368 m), Grey River (68 m), La Hune Bay (152 m), Aviron Bay (157 m), Francois Bay (125 m), Chaleur Bay (208 m), Rencontre Bay (256 m), Devil Bay (123 m), Hare Bay (181 m), Facheux Bay (390 m), and Bay d'Espoir (790 m). The fiords are commonly very narrow (White Bear Bay, for example, averages 1 km in width) and are relatively shallow compared with those of Notre Dame Bay (cf. Shaw et al., 1999b). The exception is Bay d'Espoir, the deepest fiord in the Island of Newfoundland. Grey River is unusual in that it is remarkably shallow, having been largely infilled with Quaternary sediments.

Fiord-mouth submarine moraines

A majority of the fiords have submarine moraines at their mouths (Shaw et al., 2000b, 2009; Shaw, 2003). These arcuate, multilobed ridges of glacial diamict formed when glacial ice margins stabilized for a short period at fiord mouths. The moraines at the mouths of White Bear Bay (Fig. 33) and Facheux Bay (Fig. 34) display some of the characteristic morphology of these moraines. The Facheux Bay moraine has multiple lobes. An acoustic profile (Fig. 35) across this feature shows how the moraine interfingers with acoustically stratified glaciomarine sediments, resulting in till-tongue architecture (King and Fader, 1986). Piston cores targeting the glaciomarine sediments yielded numerous shell samples the dates of which constrain moraine formation. The oldest date at the Facheux Bay moraine is 13 770 ± 90 ^{14}C years BP (TO-3190, conventional age, i.e. δ^{13}C = 25‰). At this time, the ice flows emanating from the fiords were fed by a series of individual catchments. Figure 36 shows these catchments, and also demonstrates how the margin to the west lay along the basement rock boundary, and formed a continuous moraine.

Fiord stratigraphy

The fiords contain thick infills of Quaternary sediments, mostly acoustically stratified glaciomarine mud deposited by meltwater plumes, interfingered with acoustically transparent debris-flow deposits. A cross-section of White Bear Bay illustrates this aspect (Fig. 37). A thick sequence of acoustically stratified glaciomarine sediments partly fills the fiord. An uppermost unit that is acoustically transparent is postglacial mud, most likely deposited following the early Holocene lowstand of relative sea level. This unit appears to be banked, suggestive of current activity in the fiord, and as the inset shows it is separated from the underlying sediment by a strong unconformity. Near the head of the bay the upper part of the lower sequence thickens and internal reflections trend upward into the early Holocene delta (*see* 'Submerged deltas in fiords' section), so that the lower sequence is both glaciomarine and paraglacial (Syvitski, 1991, 1994). In the much larger Bay d'Espoir, the Quaternary sediments occur in a series of separate basins rather than a single prism as in White Bear Bay.

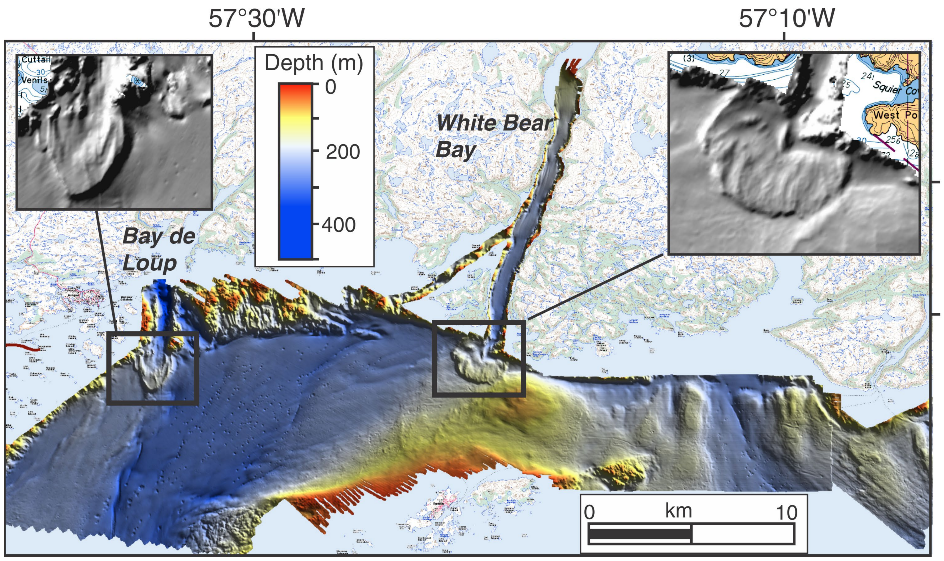


Figure 33. Seafloor terrain from Bay de Loup to just east of Grey River. The insets show enlarged views of the fiord-mouth moraines off Bay de Loup (left) and White Bear Bay (right). The relatively deep seafloor to the southwest of the entrance to White Bear Bay contains acoustically stratified glaciomarine sediment overlain by acoustically transparent postglacial mud with pockmarks. Background from Canadian Hydrographic Service (2002b).

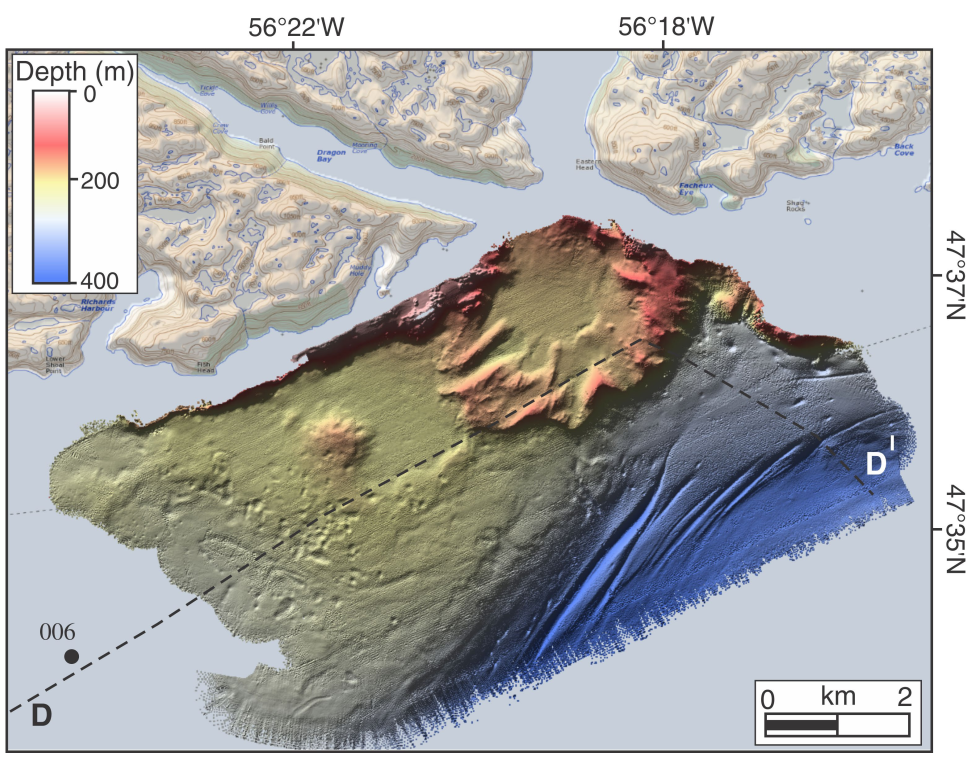


Figure 34. Arcuate submarine moraine at the entrance to Facheux Bay, also showing the survey line from cruise 2003030B that is illustrated in Figure 35, and the location of piston core 91026-006.

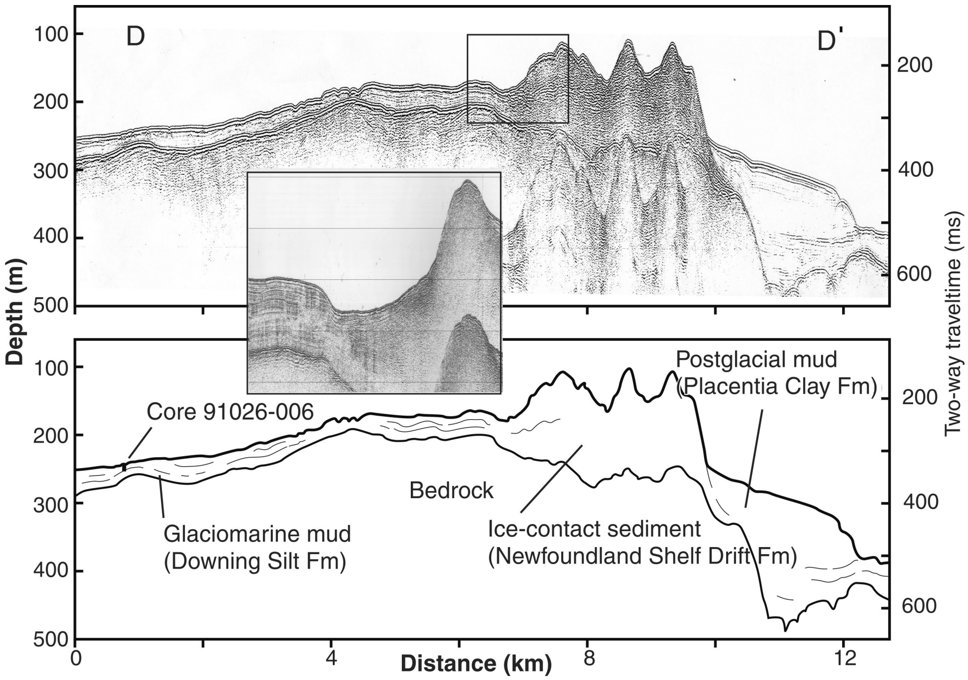


Figure 35. Airgun seismic-reflection record of the arcuate submarine moraine at the entrance to Facheux Bay, also showing the location of piston core 91026-006. The insert show part of the moraine in higher resolution (Huntec DTS profile from cruise 2003030B), and illustrates the interfingering of the acoustically incoherent moraine sediment (till) with the acoustically stratified glaciomarine mud.

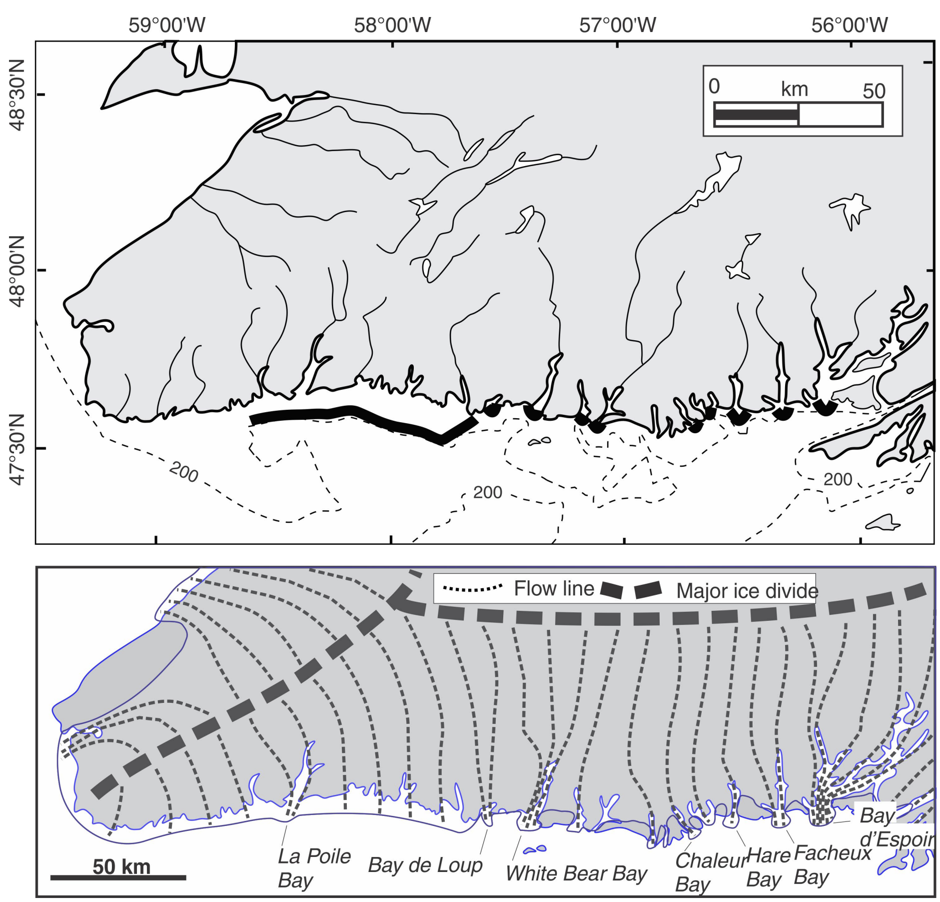


Figure 36. Moraines along the southwest coast of Island of Newfoundland.

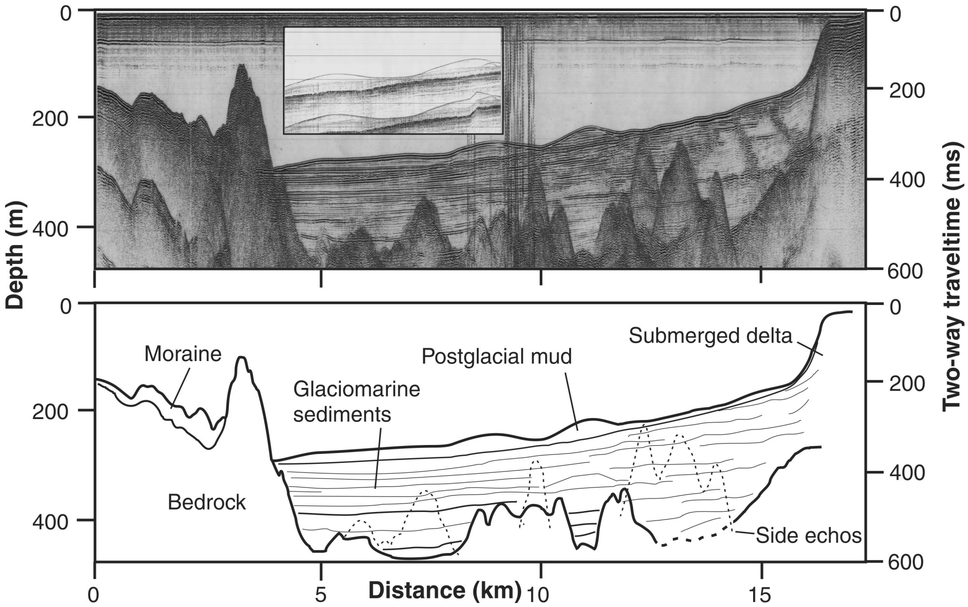


Figure 37. Airgun seismic-reflection profile, White Bear Bay. The inset is a Huntec DTS profile showing the inconformity separating the uppermost postglacial mud unit (acoustically transparent) from the underlying glaciomarine and paraglacial sediments (acoustically stratified).

Submerged deltas in fiords

Submerged early Holocene deltas in regions 3 and 4 record the spatially and temporally varying postglacial relative sea-level lowstand along the south coast of Island of Newfoundland (Shaw and Forbes, 1995). Relative sea level at La Poile Bay fell to a −30 m lowstand ca. 10 ka (Grant, 1989; Shaw et al., 2000b). The deltas comprise wedges of sediment with prograded, seaward-dipping internal reflections (*see* Fig. 8, Shaw and Forbes, 1995). In White Bear Bay, piston core 2003030B-005 targeted postglacial mud immediately overlying bottomset beds of the lowstand delta shown on Figure 37. A conventional date of 9750 ± 40 ^{14}C years BP was obtained on a marine gastropod *Solariella varicosa* from a depth of 0.30 m in this core. In a vibracore on the delta top (core 2003030B-03) a sample of organic debris at a depth of 1.54 m returned a date of 5730 ± 40 ^{14}C years BP (Beta-15063).

Several lowstand deltas are well developed at the head of Bay d'Espoir (Fig. 38), where relative sea level fell from +50 m at ca. 12.5 ka to a well constrained lowstand of −16 m at 8.5 ka that is based on a series of radiocarbon dates from samples in gravity cores — *see* seismic profiles and other supporting data in Shaw and Forbes (1995).

Bay d'Espoir

Bay d'Espoir (Fig. 39) is a large fiord, with several arms, including North Bay and Lampidoes Passage. The sidewalls are extremely steep in places. Off Goblin Head, for example, a slope analysis of the multibeam sonar data shows that rock-wall slopes reach a maximum angle of 84°. Haedrich and Gagnon (1991) and Gagnon and Haedrich (2003) reported a rich fauna at this location, including the first record of the European giant file clam, at depths of 550 m to 775 m. Bay d'Espoir is constricted in several areas, resulting in a series of flat-floored basins containing glaciomarine sediment overlain by postglacial mud. The principal east-west-oriented channel has a thin Quaternary sediment cover, and bedrock structure is strongly evident on the seafloor. The shoal at the entrance to the bay (Fig. 39, bottom inset) is an arcuate ridge of till 60 m high and 75 m thick, comprised of seaward-projecting lobate ramps ornamented with grooves expressive of basal sliding by glacier ice. The adjacent deep trough has a maximum depth of 790 m. The sidewalls are steep and rocky, with submarine fans on the east side. The flat floor is underlain by nearly 300 m of postglacial and glaciomarine mud, so that the bedrock base of the trough is approximately 1000 m below sea level (Shaw et al., 1992).

Submarine fans (A in Fig. 39) occur on either side of North Bay. Immediately south of the fans, numerous blocks up to 20 m high sit on the seafloor, suggestive of collapse of the fiord sidewalls (*see* inset upper left). Submarine fans up to 40 m high originating from the west side of the bay (Shaw et al., 2006e) are superimposed on the failed material, showing that the failure is not recent. East Bay contains a submerged delta at its head and also a jumble of large blocks on the seafloor (B). A moraine (C) is located at the entrance to Lampidoes Passage. The irregular seafloor (*see* inset upper right) in the narrows was caused by massive rock falls from vertical and overhanging fiord sidewalls. A series of submarine fans originating from the north shore are superimposed on the blocky slump debris, again suggesting that sidewall failure was not recent. Roti Bay (D) is a shallow area extensively used for aquaculture (Tlusty et al., 2000). North of Little Crow Head (E) the basin floor is flat, whereas the valley walls have a channellized appearance.

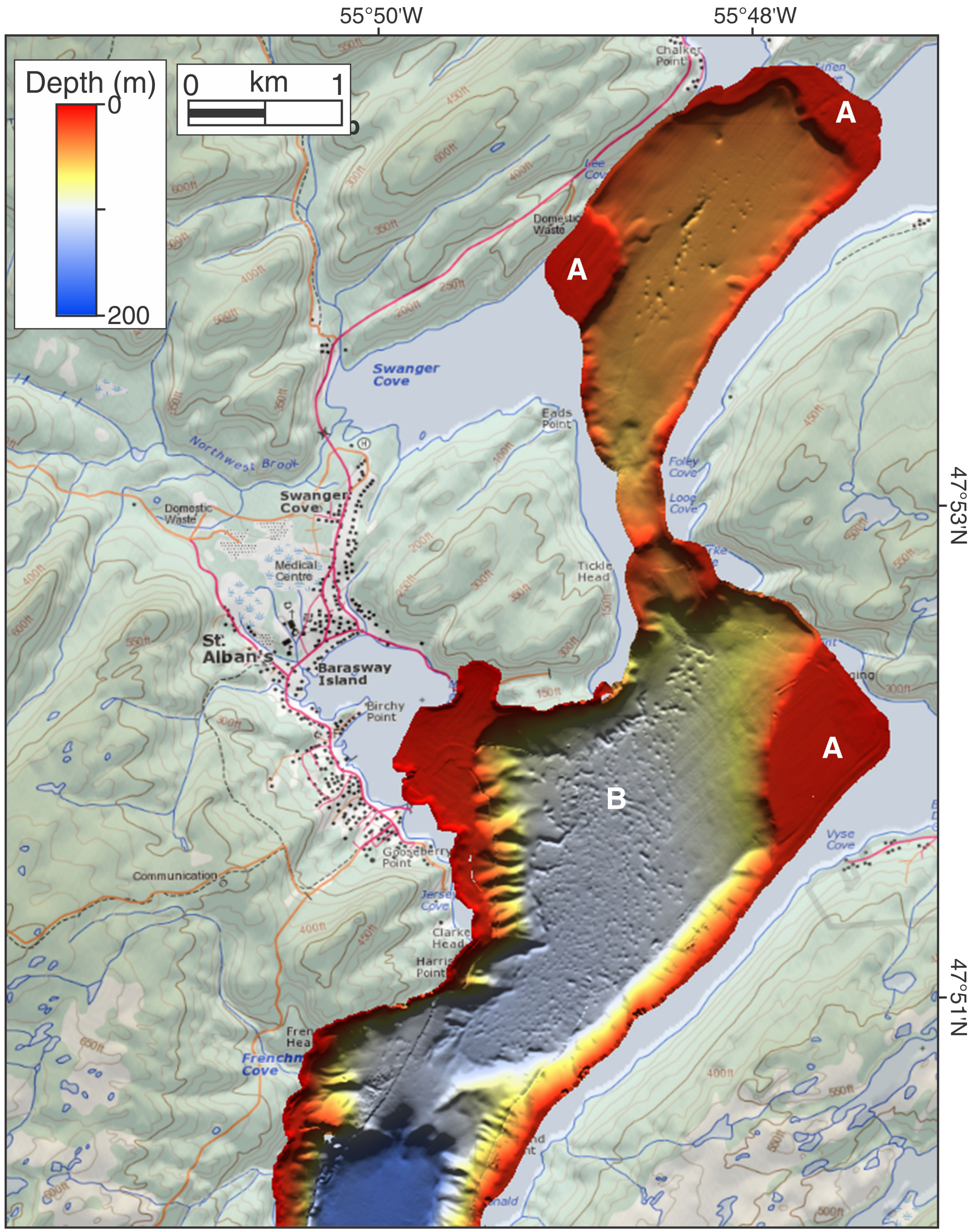


Figure 38. Three prominent lowstand deltas (A) are seen in this multibeam image of the head of Bay d'Espoir. The seismic profiles and core information are illustrated in Figures 15 and 16 of Shaw and Forbes (1995). The deeper areas contain postglacial mud with numerous pockmarks (B).

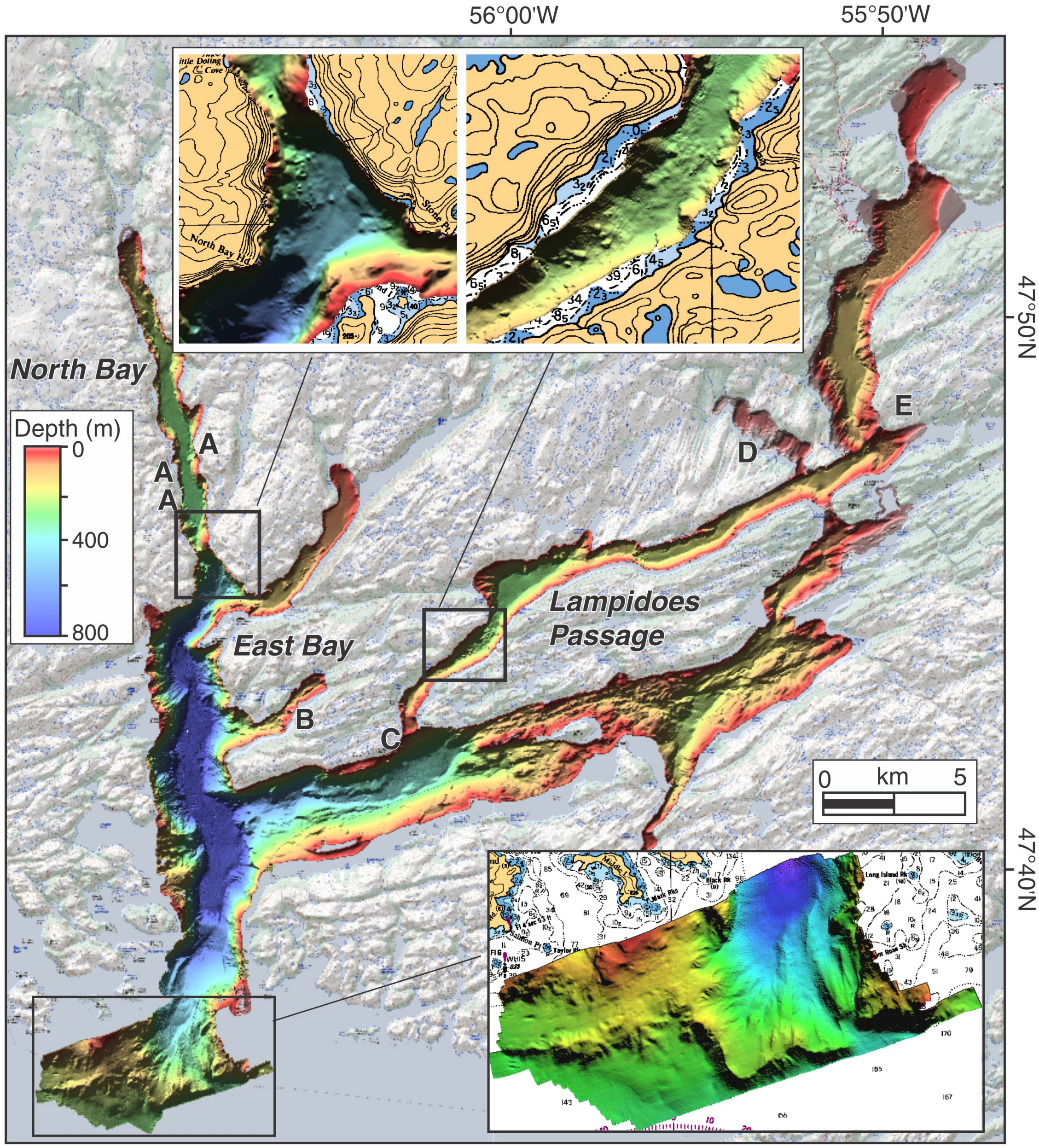


Figure 39. Bay d'Espoir, with insets show areas of rockfall in North Bay (left), Lampidoes Passage (right), and the submarine moraine at the fiord mouth (bottom right). Letters A to E are discussed in text. Base is from Canadian Hydrographic Service (1999).

Seafloor offshore from the fiords

The relatively deep (>200 m) water encountered immediately offshore in region 4 is characterized by thick glaciomarine mud overlain by heavily pockmarked postglacial mud (Fig. 40). The source of the gas may be the underlying Carboniferous bedrock: the map by Fader et al. (1989) shows Carboniferous bedrock far offshore, but the evidence of this multibeam imagery suggests the Carboniferous lies very close to the coast (E. King, pers. comm., 2010). Conventional radiocarbon dates ($\delta^{13}C$ = 25‰) on glaciomarine sediments in piston cores 91026-003, 91026-004, 91026-005, and 91026-006 from this area (Shaw et al., 2000b) range from 12 220 ± 90 ^{14}C years BP (TO-3187) to 14 580 ± 90 ^{14}C years BP (Beta-88458), and show that open-water conditions prevailed from ca. 14.6 ka onward. Reservoir corrections were applied to these dates in Shaw et al. (2000b).

Thick Quaternary sediments are found in the deep Hermitage Channel that terminates just southeast of Facheux Bay. In this area, immediately southeast of the Facheux moraine (Fig. 34), thick deposits of postglacial mud with gas masking in places overlie an acoustically transparent ponded unit interpreted as a debris flow, overlying thick glaciomarine sediments. Possibly there was instability on the front

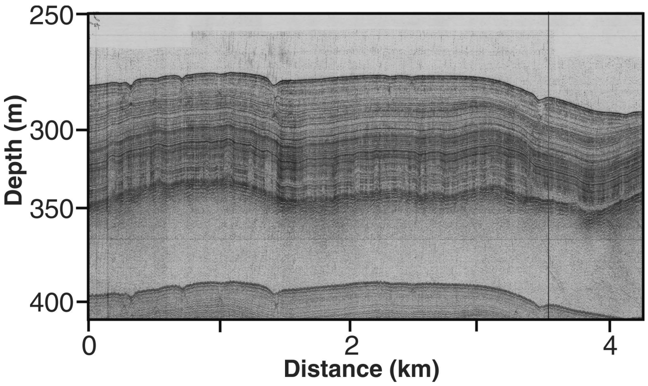


Figure 40. Quaternary sediments in deep water west of Ramea Islands.

of the moraine: the multibeam image shows signs of possible submarine slide scars. The Quaternary sediments thin in shallower areas around the Ramea Islands, where there is an outer zone of glaciomarine sediment imprinted by iceberg furrows and pits, and an inner zone of bedrock and mobile sand and gravel.

REGION 5: FORTUNE BAY

Setting

Region 5 (Fig. 41) encompasses the 'Hermitage peninsula', the indented, steep rocky coasts to the east, the long, low coastlines of the west side of the Burin Peninsula, and the deep trough of Fortune Bay. Hermitage Bay lies west of the 'Hermitage peninsula', and is included within this region. The 'Hermitage peninsula' is indented by three deep arms: Connaigre Bay, Northeast Arm (maximum depth 183 m), and Great Bay de l'Eau (316 m). The western boundary of the region is the northeast-trending Dover Fault (Colman-Sadd et al., 1990), east of which the region lies entirely within the Avalon tectonic zone. The bedrock geology is much more complex than in region 4, and includes rocks of various ages, from Cambrian to Carboniferous. With one exception, the region has not been mapped with multibeam sonar. Raised marine terraces supply sediment to the numerous substantial sand and gravel beaches scattered across the region.

Coastlines

On the north side of Fortune Bay, the many arms of the 'Hermitage peninsula' (Fig. 41) are characterized by raised marine terraces (Jenness, 1960; Tucker et al., 1982), and small gravel pocket beaches are common in embayments. A particularly large glacial deposit at Great Harbour Bight — perhaps an ice-contact delta (Shaw and Frobel, 1992) — has supplied sediment to a gravel barrier that spans the bay. The barrier is flood dominated, and consists of gravel beach ridges mantled by freshwater and salt-marsh peat. A sample of heath peat from 45–50 cm in a 50 cm thick peat layer on this barrier returned a conventional age of 730 ± 90 ^{14}C years BP (GSC-4681). At Coombs Cove barrier beach a sample from the basal 5 cm of an 85 cm thick freshwater peat layer overlying beach sand is dated at 720 ± 80 ^{14}C years BP (GSC-4789). A large sandy barrier beach at Terrenceville spans the head of Fortune Bay, nourished by adjacent raised marine terraces; a sample of peat from the barrier was dated at 430 ± 50 ^{14}C years BP (GSC-5715).

By contrast with the section of coast on the north side of Fortune Bay, the coast of the Burin Peninsula has low relief and few indentations; however, raised marine terraces are also typical of this region, and the supply of sediment from these has given rise to large coastal deposits in several areas. The Frenchmans Cove strand plain is a complex structure (Shaw and Forbes, 1987) composed of gravel beach ridges. The earliest and oldest beach ridges have an elevation of 1 m above mean water level, whereas the western part of the strand plain is occupied by modern high gravel that attains maximum elevation of 3.5 m above mean water level. This shows that, like the barriers at the head of St. George's Bay, it grew during a period of rising relative sea level. It connects with a bedrock island, and links to the coast to the east via a narrow gravel barrier, thus enclosing a brackish lagoon. Shaw and Forbes (1987) discussed the evolution of this strandplain. Pollen analysis of a core that intersected 1.2 m of peat in a shallow lagoon revealed brackish conditions at the base followed by freshwater conditions in most of the core, and colonization of the beach ridges by *Alnus* sp. The oldest of three radiocarbon dates on peat in the core was 1210 ± 70 ^{14}C years BP (Beta-19581). Subsequent to the GSC investigations in the 1980s, the previously unoccupied portions of the strand plain — the oldest areas with lowest elevations — have been converted for use as a golf course.

South of here, the barrier at Grand Beach presents a striking contrast (*see* Fig. 2.7 in Forbes, 1984). Sheets of overwash landward of the single gravel beach testify to rapid beach migration. The beach enclosed a tidal lagoon in which a rooted stump of *Picea* sp. 0.1 m below mean water level at the base of a 0.7 m thick peat layer overlying gravel was dated at 4310 ± 70 ^{14}C years BP (GSC-4903). Peat at this level was dated at 4720 ± 80 ^{14}C years BP (GSC-4913). Whereas these dates give a general indication of rising sea level, they do not permit more exact estimate of rates.

Hermitage Bay

Data from survey CCGS *Matthew* 99020 (Fig. 42) provide an instructive cross-section of Hermitage Bay from the mouth of Bay d'Espoir to Eastern Cove on the 'Hermitage peninsula'. This line shows thick, gas-charged postglacial mud in the deepest part of the trough, overlying glaciomarine sediments. Toward the south, as the depth decreases, the glaciomarine unit is exposed at the seafloor. Airgun seismic-reflection profiles from cruise CCGS *Hudson* 73006,

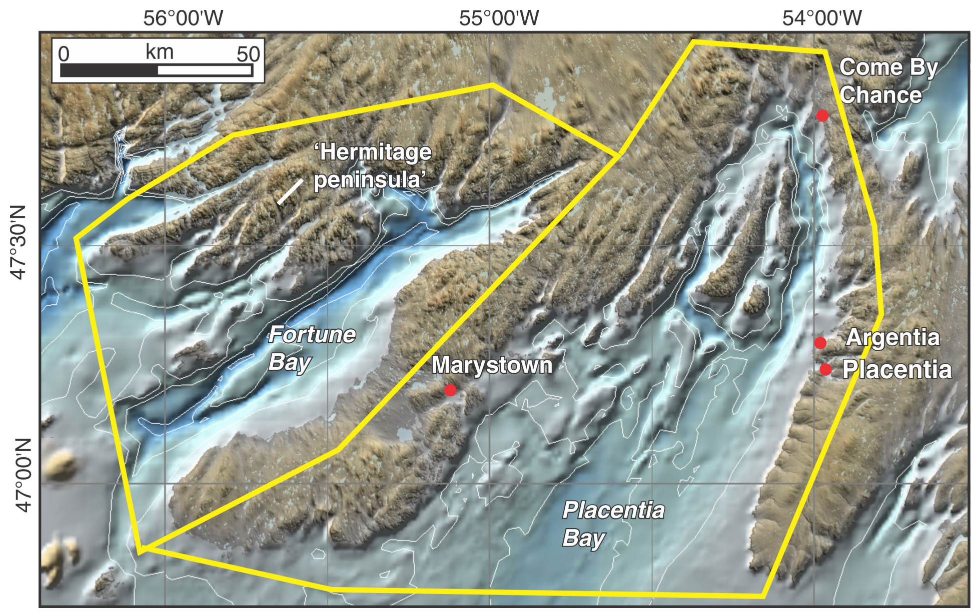


Figure 41. Location of zones 5 and 6.

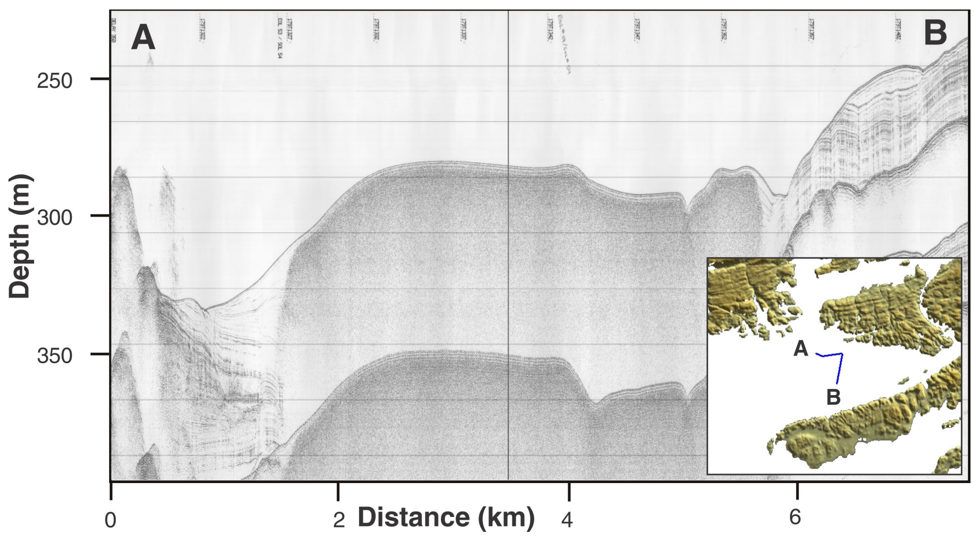


Figure 42. Huntec DTS section in Hermitage Bay, showing thick, gas-charged postglacial mud overlying glaciomarine sediments. Cruise CCGS *Matthew* 99020 daytimes 1791317–1791407.

and also information on the A-series map of the area (Fader et al., 1982) show that outer Fortune Bay has thin Quaternary sediments that thicken in the inner bay.

Connaigre Bay and Northeast Arm

Whereas proper interpretation of multibeam data requires 'interrogating' gridded bathymetry and backscatter data, it is nevertheless necessary on occasion to make do with raster imagery. Such is the situation with regard to Connaigre Bay (Fig. 43). The bay has three basins and a broad sill at a depth of about 98 m according to CHS chart 4827 (Canadian Hydrographic Service, 2004a). Amphitheatres in several areas (A in Fig. 43) similar to those in Placentia Bay are suggestive of mass transport, perhaps where thick glaciomarine sediments have failed, leaving scars and forming depositional lobes in the deeper water. Toward the head of the bay, submarine moraines occur in several areas. The moraine at 'B' is located at a constriction in the channel, the favoured location for stabilization of a retreating ice front (Syvitski and Shaw, 1995). Up to four transverse moraines occur at the entrance to Great Harbour (C) and a fifth moraine to the northeast may underlie the gravel barrier that stretches across the bight (D). The inset map (upper left) shows these moraines and the barrier. On CHS chart 4832 (Canadian Hydrographic Service, 1987) a transverse shoal is evident near the head of Northeast Arm (Fig. 43, E). It is located at a constriction in the channel, and is also likely a submarine moraine.

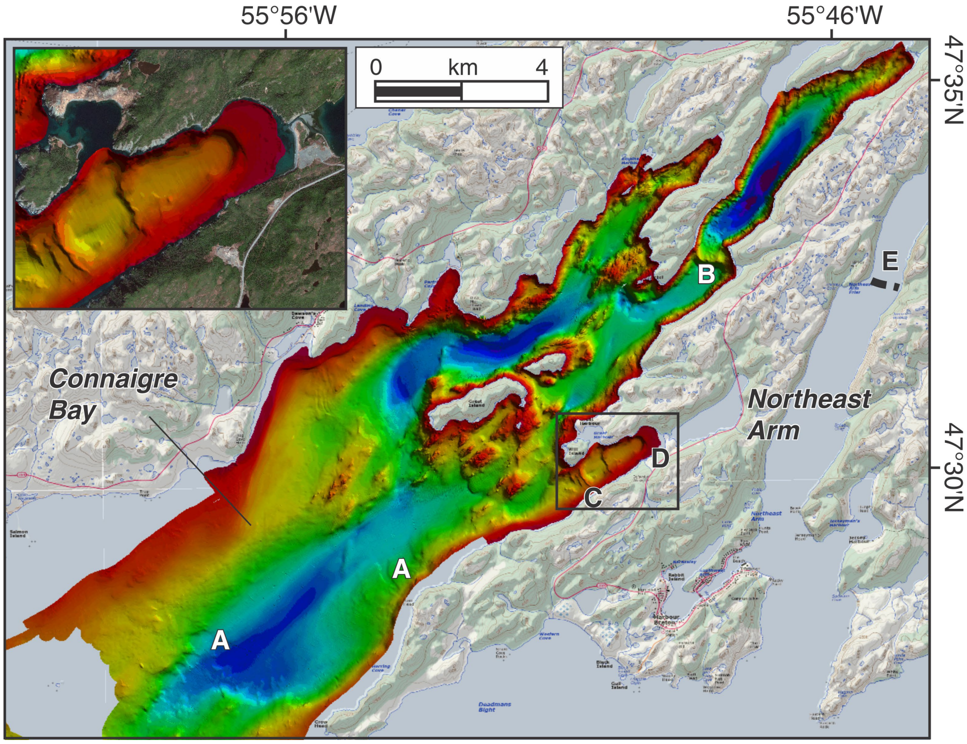


Figure 43. Shaded-relief image of Connaigre Bay, derived from a poor-resolution geotiff rather than gridded data. Letters A to E are discussed in text.

Fortune Bay

Fortune Bay exceeds 400 m depth along its thalweg. On GSC A-series map 4015 (Fader et al., 1982) the surficial geology of the deeper portions of the bay is depicted. As might be expected, postglacial mud occupies the deepest areas of the channel, overlying glaciomarine mud that is exposed on the channel flanks.

REGION 6: PLACENTIA BAY

Setting

Placentia Bay (Fig. 41) lies within the Avalon tectonic zone. It is approximately 100 km wide at its mouth, and extends northeast for about 130 km. It is bounded to the west by Burin Peninsula and to the east by the Avalon Peninsula. In the upper bay, three deep, north-trending channels are separated by islands and contain thick Quaternary sediments, including strongly banked postglacial mud overlying glaciomarine sediment. The outer bay divides into three zones, from west to east. In the west the topography is dominated by fields of drumlins and megascale glacial lineations, with De Geer moraines superimposed. In the central area thick postglacial mud with pockmarks overlies glaciomarine sediments; on the eastern edge of this zone strong currents have eroded the seafloor, forming megaflutes. The eastern zone is characterized by glacial landforms that have been truncated above −40 m to form an erosional platform. Inlets at Argentia, Long Harbour, and Marystown contain accumulations of postglacial mud, and are generally separated from the deep outer basins by shallow, wave-dominated sills. Submerged early Holocene deltas are found in several locations. A major beach-ridge plain has formed at Placentia over the past three thousand years, against a backdrop of rising relative sea level.

Placentia Bay has been entirely mapped with multibeam sonar, extensively surveyed with subbottom profilers, and sampled. Results from these activities are contained in a series of five 1:50 000 scale shaded-relief maps (e.g. Potter and Shaw, 2009a, b, c, d, e), five maps of backscatter strength (e.g. Potter and Shaw, 2010a, b, c, d, e), a seascape map (Shaw et al., 2011), various papers (Batterson et al., 2006; Brushett et al., 2007; Shaw et al., 2009), and cruise reports (e.g. Shaw et al., 1990b). Consequently a reasonable comprehensive picture exists of glacial history, relative sea-level changes, coastal and submarine geomorphology, surficial geology, and modern processes.

Coastlines

The west side of the bay has low relief, with scattered sand and gravel beaches. Relief increases northward toward the head of the bay, where coasts are steep and rocky. The large islands separated by Eastern Channel and Western Channel are characterized by steep rocky shores and scattered pocket beaches composed of gravel. Rocky shores predominate down the east coast as far as the Argentia area (Fig. 41).

Placentia strand plain

South of Argentia, glacial deposits are exposed in coastal bluffs. One of the largest deposits is at Pointe Verde (Fig. 44a). Longshore drift from prevailing southwesterly winds has contributed vast quantities of sand and gravel to the strand plain at Placentia (Forbes, 1984; Forbes et al., 1995b). The strand plain (Fig. 44b, c) consists of a series of prograded pebble-cobble gravel beach ridges (Shaw and Forbes, 1987) that record the progressive infilling of an embayment. Barrier construction was initiated by extension of a spit across Southeast Arm ca. 2500 years ago. Subsequently (after 2.1 ka) the inlet at the distal end of the spit was closed, and the barrier infilled against the high ground. Twelve sets of beach ridges are separated by truncation events (Forbes et al., 1995b). Crest elevation increases from the oldest ridges at the rear of the system, adjacent to the high ground of a former island, to the highest, at the modern storm beach.

The swales and beach ridges (Fig. 44c) are overlain by both freshwater and saltmarsh peat (Shaw and Forbes, 1987). Radiocarbon dates on basal peat in swales extend back to 530 ± 70 ^{14}C years BP (Beta-19578). Basal peat in a former tidal channel is dated at 2100 ± 120 ^{14}C years BP (Beta-19570) and shell material in a more recent part of the strand plain at 1570 ± 60 ^{14}C years BP (TO 2380) (locations shown on Fig. 8 *in* Forbes et al. 1995b). (The TO- date was from a depth of 6 m in gravelly sand. Forbes et al. (1995b) reported it as 1160 ± 60 ^{14}C years BP, i.e. with a 410 year reservoir correction applied, but it is possible that the date already had a correction applied). Another date from the former tidal channel is 2590 ± 120 ^{14}C years BP (GSC-5879). Thus, the beach-ridge plain has formed over the past 3000 years, against a backdrop of rising relative sea level. Only the reduced tidal levels in the rear of the barrier (ShawMont Martec Limited, unpub. report for Department of the Environment (Newfoundland and Labrador) and Environment Canada, 1984) prevent the submergence of the lowest-lying beach ridges. A modern seawall on the west-facing storm beach also provides protection. The lower and older parts of the strand plain are nevertheless subject to occasional inundation due to high water levels in the back-barrier tidal lagoon (ShawMont Martec Limited, unpub. report for Department of the Environment (Newfoundland and Labrador) and Environment Canada, 1984) whereas the high modern storm beach was, until recently, subject to overflow during storm surges (ShawMont Martec Limited, unpub. report for Department of the Environment (Newfoundland and Labrador) and Environment Canada, 1984); a new seawall has eased this problem. The problem of storm-surge flooding has been exacerbated by extensive building on the plain in the past few decades.

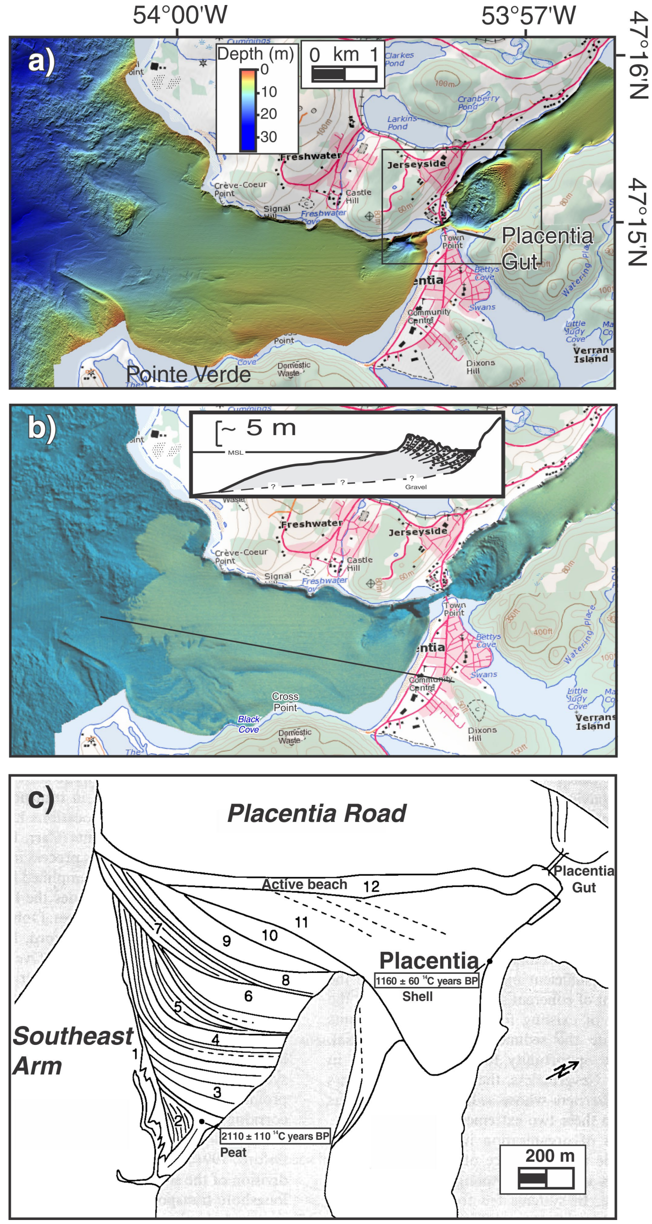


Figure 44. a) Shaded-relief image of Placentia Road, the embayment offshore from Placentia; the box shows the location of Figure 45; **b)** backscatter strength for the same area, dark tones = high backscatter strength (coarse substrate) and light tones = low backscatter strength (sand). The insert shows an estimate of the subaqueous sand body lying offshore from the gravel strand plain. MSL = mean sea level; **c)** illustration showing the pattern of gravel beach ridges on the Placentia strand plain (base from Canadian Hydrographic Service, 1989).

Subaqueous components of the Placentia strand plain

Data from multibeam surveys permit one to consider the subaerial Placentia strand plain as part of a larger system. Offshore from the strand plain, Placentia Road contains a prism of sand (Fig. 44a, b) that extends 4.5 km offshore and overlies older Quaternary sediments. The thickness is uncertain, but a maximum of about 5 m seems likely (*see* inset, Fig. 44b). The seaward edge of the sand prism is clearly delineated on the backscatter-strength image (Fig. 44b). The existence of this submerged sandy sediment prism contrasts with the dominance of gravel-sized clasts on the strand plain, and reflects the segregation of sediments according to grain size during infilling of the embayment, with gravel forming the subaerial beach ridges, and sand contributing to the subaqueous prism. This segregation has strong parallels with Stephenville strand plain and Flat Island spit (region 1), although in the last cases larger accommodation space offshore has resulted in much thicker subaqueous sediment prisms.

Northward progradation of the strand plain has created the narrow tidal channel at Placentia Gut (Fig. 45), that connects Placentia Road with the tidal areas to the rear of the Placentia strand plain (ShawMont Martec Limited, unpub. report for Department of the Environment (Newfoundland and Labrador) and Environment Canada, 1984). Seaward of Placentia Gut the ebb system includes a trough (A) extending to a depth of 19 m. Backscatter values indicate that the shoal immediately south of Placentia Gut is composed of gravel, probably cycled northward along the beach by longshore drift. The flood tidal system is more complex. A relatively deep scoured area lies adjacent to Placentia Gut. Ridges in the trough (B) may be De Geer moraines excavated by tidal scour. To the northeast is a crescentic flood-tidal delta (C) consisting of sand up to 7 m thick, the mobility of which is attested by the presence of numerous dunes that are commonly 1 m high. Yet farther north irregular mounds on the seafloor (D) are likely dredge spoil.

Placentia Bay: overview

Figure 46 shows the shaded-relief image of the bay derived from multibeam-sonar mapping. The upper bay includes three narrow, glacially overdeepened channels: Western Channel (maximum depth 450 m), Central Channel (265 m), and Eastern Channel (333 m). The outer bay has lower relief, and encompasses a rugged western part, with bedrock ridges and glacial deposits, a central section that is relatively deep and has gentle slopes except for the shallow bedrock ridge of Merasheen Bank, and a shallow section fringing the Avalon Peninsula.

Figure 47 shows backscatter strength, a proxy for sediment texture, for the same area as Figure 46. Large areas of high backscatter strength (bedrock) occur in the upper bay, with narrow, discontinuous strips of low backscatter extending down the deep channels. The three channels contain thick deposits of postglacial mud with acoustic stratigraphy masked by gas in some areas. In Eastern and Western channels the postglacial mud is thickest in banks, to either side of the thalweg. Elsewhere postglacial sedimentation has been precluded by currents, and glaciomarine sediments are exposed at the seafloor. In the south of Placentia Bay,

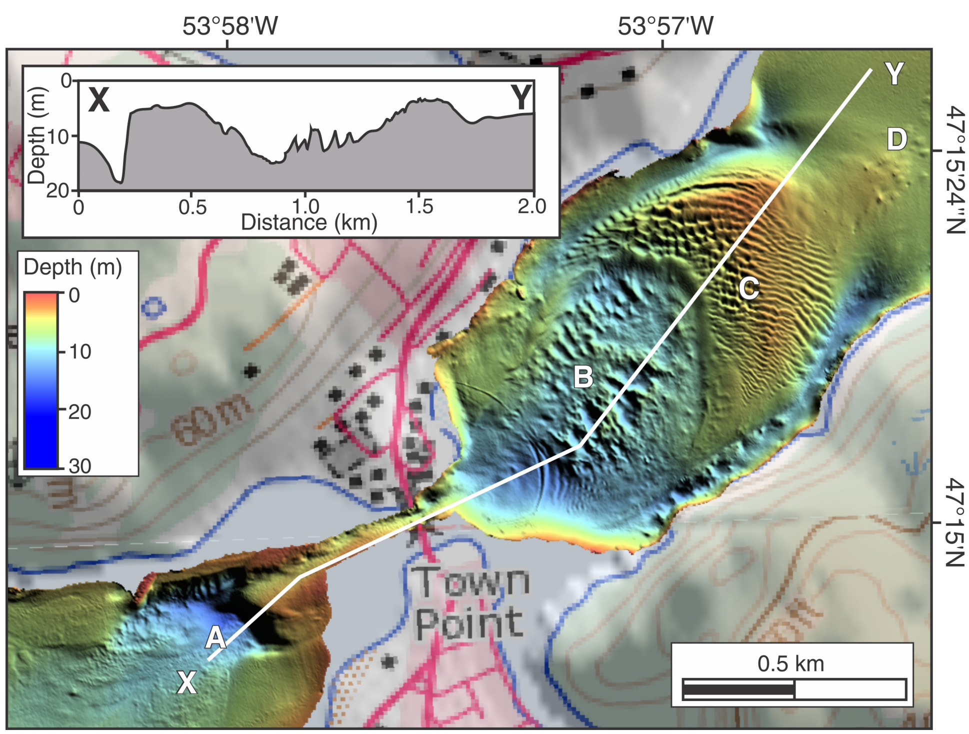


Figure 45. Placentia Gut area showing ebb trough (A), possible De Geer moraines (B), sand dunes (C), and dredge spoil (D). The cross-section (inset) shows from left to right the ebb scour, the relatively shallow gut, the deep flood trough, and the sand bank. The background is CHS chart 4841 (Canadian Hydrographic Service, 1989). Location shown on Figure 44.

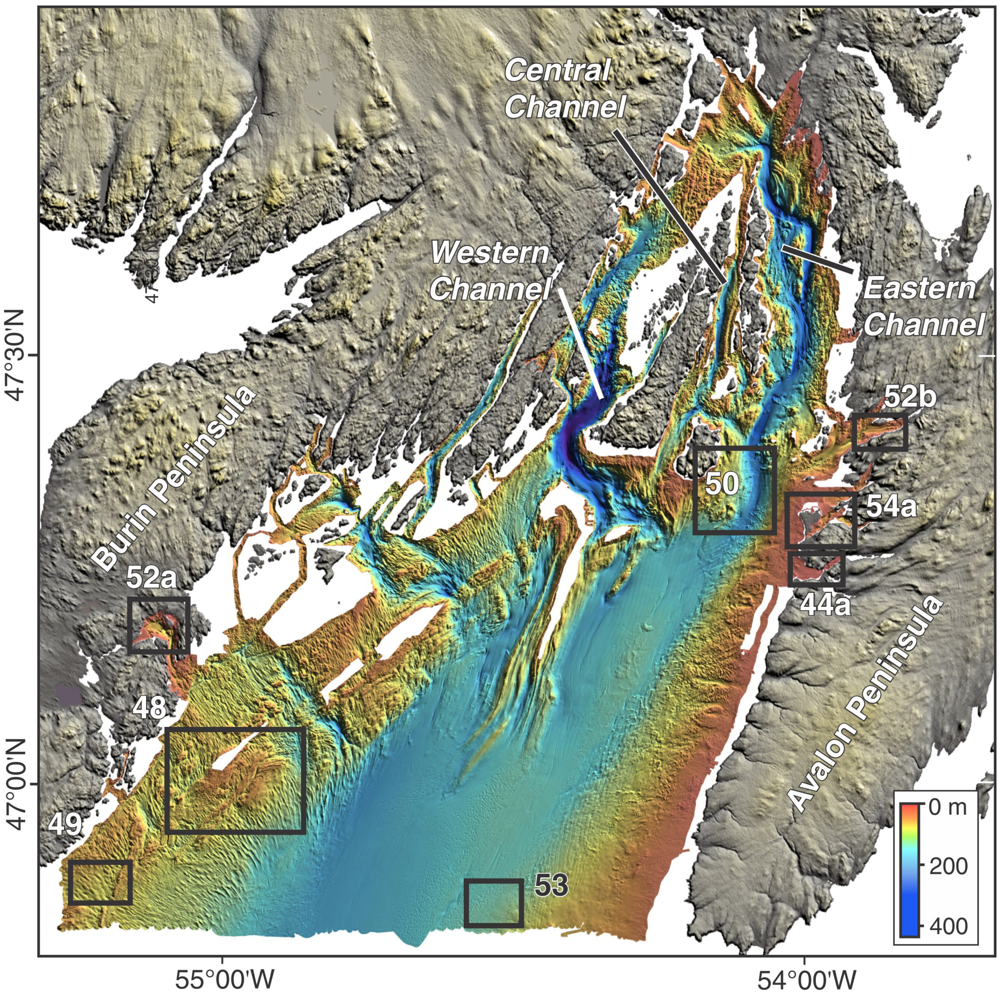


Figure 46. Shaded-seafloor relief of Placentia Bay, also showing the locations of Figures 44a, 48, 49, 50, 52a, b, 53, and 54a.

high backscatter strength predominates in the west and east where glacial sediments and bedrock occur at the seafloor, but low backscatter strength typifies the central area where thick deposits of postglacial mud marked by pockmarks and sedimentary furrows overlie glaciomarine mud.

Detailed descriptions of the seafloor geomorphology are contained in the A-series maps of shaded-seafloor relief (Potter and Shaw, 2009a, b, c, d, e), backscatter strength (Potter and Shaw, 2010a, b, c, d, e), and the seascape map (Shaw et al., 2011). Here the present authors focus on several aspects of this interesting bay, namely: glacial landforms and the style of deglaciation; late-glacial mass-transport events; the chronology of deglaciation; effects of postglacial relative sea-level changes; widespread erosion of the seafloor in the outer bay (megaflutes); and anthropogenic effects.

Glacial landforms and the style of deglaciation

The west side of the outer bay (Fig. 48) contains fields of 20 m high drumlin ridges (elongation ratio E<7:1, Stokes and Clark (1999, 2001)). Toward the south, in deeper water, elongation ratio increases to more than 7:1, and the drumlins transition into a convergent field of megaflutes. These streamlined landforms in southwest Placentia Bay occur in depths of 50–240 m. Lengths range from 0.5 km to 3.5 km, with an average of 1.5 km; widths range from 350 m to 50 m, and heights average 10 m, ranging from 2 m to 15 m. They tend to be ‘comet’ shaped, with a broad upstream section and a narrower, tapering downstream section.

Placentia Bay is considered to be the onset area for ice that formerly flowed south out of the bay toward Halibut Channel (Brushett et al., 2007). Assuming that increasing elongation is a function of former ice velocity (Stokes and Clark, 2001; Evans, 2003), then the elongation of drumlins and their transformation into megaflutes, together with the

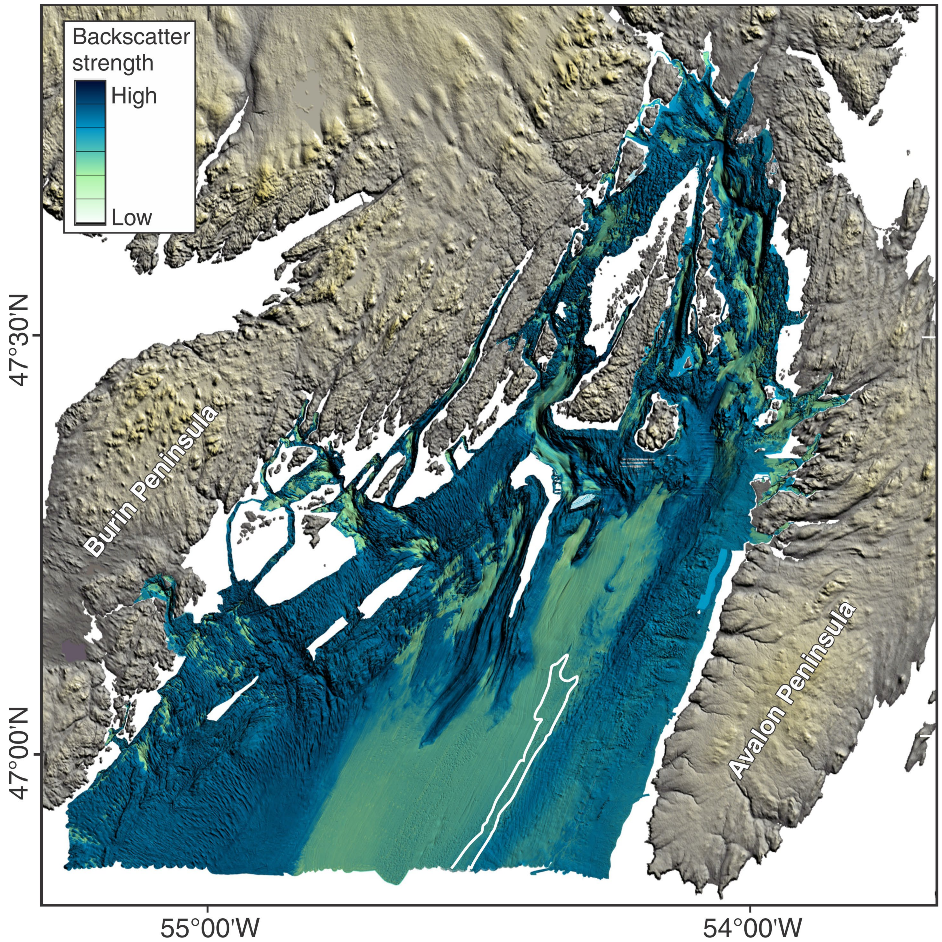


Figure 47. Backscatter strength of Placentia Bay. White line shows position of megaflute zone. Much of the outer bay has low backscatter (mud, sandy mud).

pattern of convergence, reflects acceleration of ice into the flow down the bay. Swarms of 3–4 m high De Geer moraines are superimposed transversely on the streamlined landforms in places (Shaw et al., 2009), indicating that the ice retreated up the bay by calving, with the ice margin being roughly parallel to the isobaths.

In the extreme southwest of the multibeam-sonar coverage, the drumlins are overprinted by northeast-southwest flutings (Fig. 49), demonstrative of an ice readvance from the southwest, from offshore. This is interpreted as a readvance of an ice cap in shallow water south of the Burin Peninsula, in the vicinity of the Burin Moraine mapped by Fader et al. (1982). Shaw et al. (2009) proposed that this readvance was consistent with the conceptual model proposed by Shaw et al. (2006d), and that ice marooned south of the Burin Peninsula by rapid ice retreat up Placentia Bay advanced northward a short distance.

Late-glacial mass-transport events

In Eastern Channel a series of escarpments in the 180 m to 200 m depth range are commonly 15 m high, U-shaped, and open to the east (Fig. 50). They are interpreted as scars formed by submarine mass failure. Acoustic profiles show that the unit that failed was the glaciomarine facies, the Downing Silt Formation. The failure occurred as a series of slabs became detached and moved into Eastern Channel, where stacked wedges of acoustically incoherent sediment are interpreted as mass-transport deposits. The timing of mass transport is indicated by radiocarbon dates from core 2006039-032 that targeted the slide scar (Fig. 50, bottom). In the scar, acoustically transparent sediment (postglacial mud) overlies acoustically stratified sediments (ice-distal glaciomarine mud). The core penetrated to near the base of the glaciomarine unit, from which samples are dated at 11.6 ka and 12.6 ka (Table 2). The interpretation is that mass transport followed soon after deglaciation, and occurred before 12.6 ka. The trigger may have been earthquake activity related to glacioisostatic adjustment. There are some parallels with the mass-transport events about this time in Bay

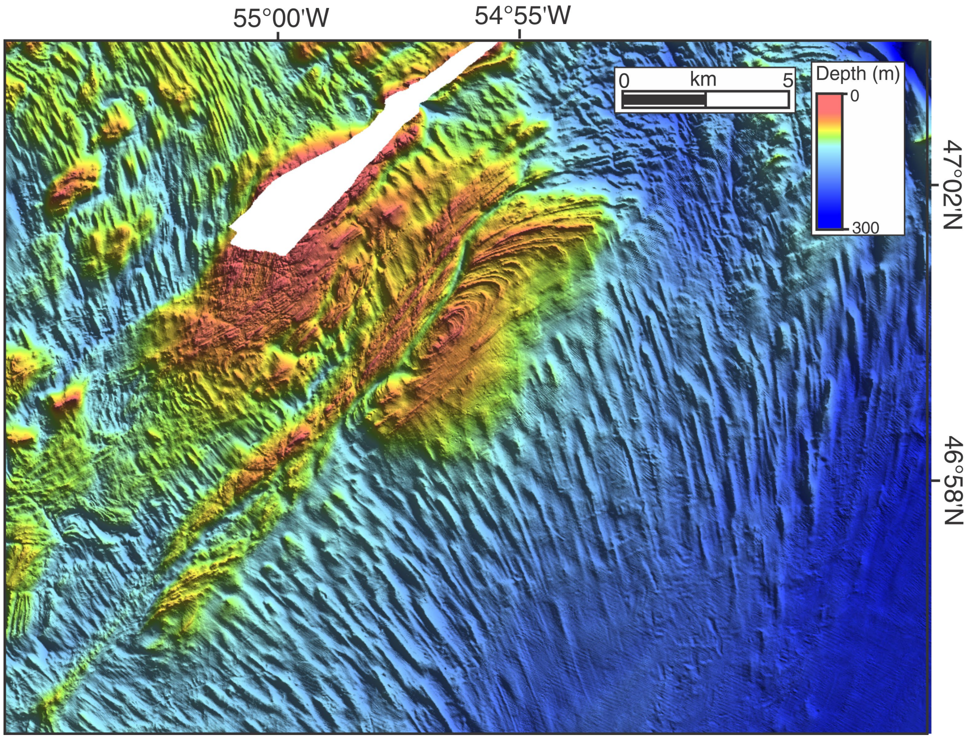


Figure 48. Glacial landforms, drumlins and megaflutes, overlain by glaciomarine mud imprinted by furrows (bottom right). De Geer moraines are superimposed on the streamlines landforms, but are not seen with this illumination angle (from the northeast).

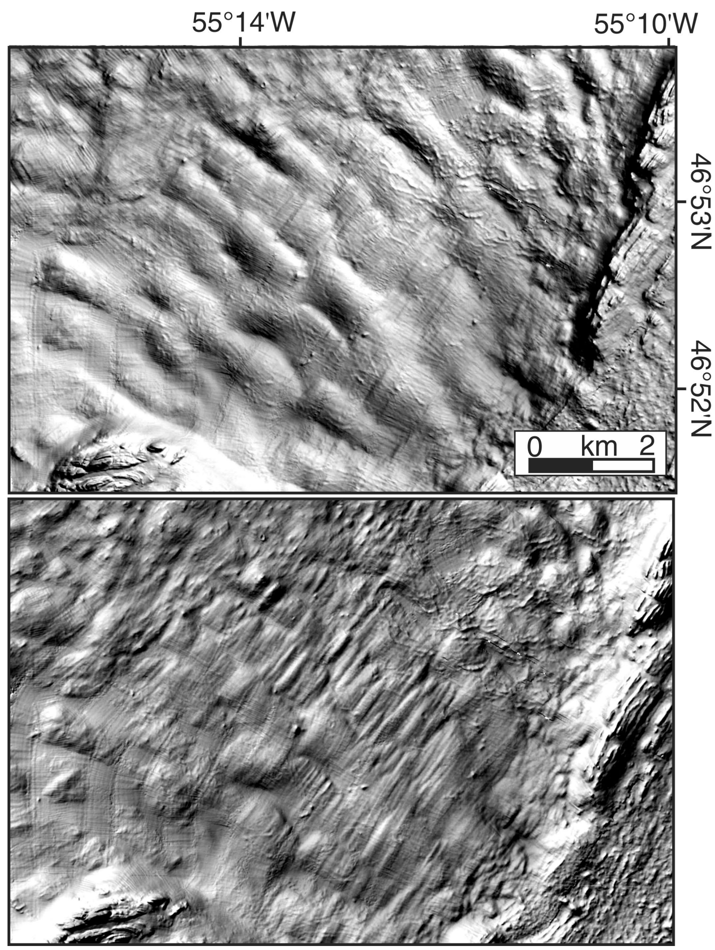


Figure 49. Evidence of a glacial readvance. The upper image, created with artificial illumination form the north-east, shows a series of drumlin ridges with relief of about 10 m. When illumination is from the northwest, however, the drumlins are overprinted a series of ridges with relief of 2–3 m, indicating a glacial readvance from the southwest (Shaw et al., 2009).

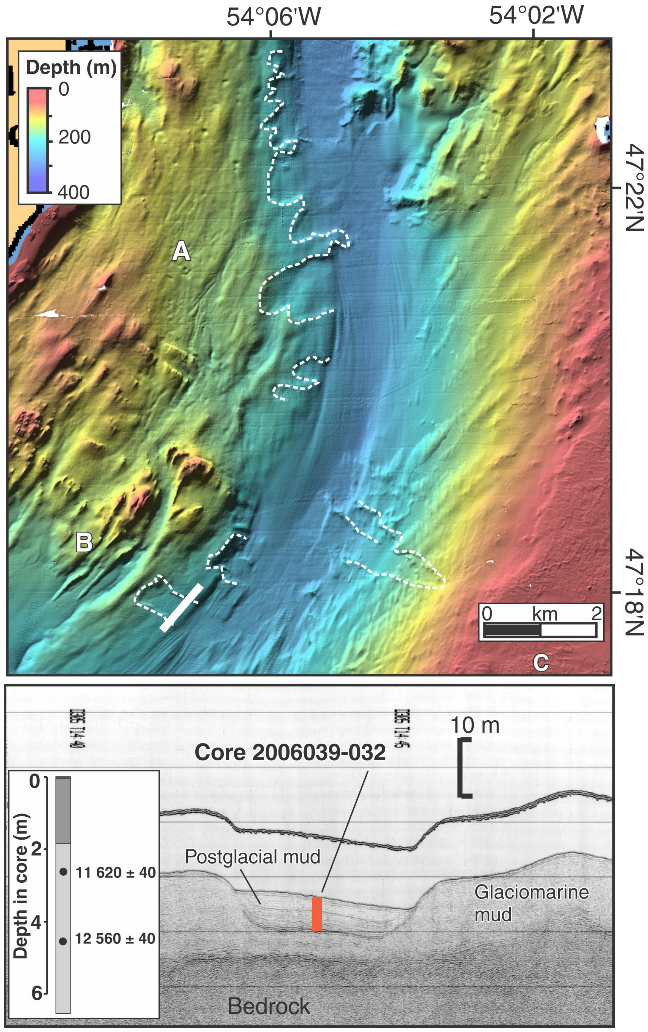


Figure 50. A series of 10 m high escarpments on the flanks of Eastern Channel (top) are interpreted as scars formed by failure of glaciomarine sediments and mass transport into the adjacent deep water, where stacked lobes of acoustically incoherent sediment are found. At bottom, a Huntec DTS record shows a piston core in the slide scar. Radiocarbon dates (conventional) constrain the age of failure to before 12.6 ka BP. Other features visible include relict furrows (A) and crag-and-tail landforms (B), indicating ice-sheet movement to the southwest.

of Islands, in region 13, where glacioisostatic adjustment was also proposed as a causative mechanism (Shaw et al., 2000a).

Chronology of deglaciation and postglacial sedimentation

A series of radiocarbon dates (Table 2) on marine pelecypod samples contained in glaciomarine and postglacial sediments constrain the timing of glacial events in Placentia Bay and the nature of postglacial sedimentation. The dated samples were from cores (Fig. 51) that intersected the acoustically stratified glaciomarine unit (the Downing Silt Formation of Fader et al., 1982) that is found throughout the southern part of the bay, and the overlying postglacial mud (Placentia Clay Formation). The earliest date on the glaciomarine facies is 14 090 ± 40 ^{14}C years BP ($\delta^{13}C = 25‰$), which when calibrated is equivalent to 15 739–16 491 cal. BP (two sigma range). The glaciomarine facies is dated over a range from 14.1–11 ka (Table 2). The present authors infer therefore that ice pulled back from the outer bay after 14.1 ka. Extrapolation in core 007, south of the entrance to the bay, suggests open water at that location a little earlier, perhaps by ca. 14 500 ^{14}C years BP.

Radiocarbon dates from the postglacial sediments (Table 2) are suggestive of large changes in sedimentation rates over time, and of contrasts between sedimentation in the inner and outer bays. Core 2006039-013, near the western limit of the postglacial mud in the bay sediments, contained 1.7 m of coarsening-upward silty clay, and a sample at the base of this unit is dated at 11.0 ka. This suggests a cessation of deposition in the early Holocene. In core 2006039-019, located just west of the megaflute zone, the transition from glaciomarine mud to postglacial mud is dated ca. 10.6 ka. Dates in the postglacial mud suggest initial rapid sedimentation (250 cm/ka), but some time after ca. 9 ka sedimentation stopped and erosion commenced, resulting in a surficial sand layer containing inclusions of the underlying silt.

In the inner bay, by contrast, muddy sedimentation was sustained in numerous deep depocentres throughout the Holocene. In Western Channel, core 2006039-027 penetrated the uppermost part of thick postglacial mud and contained three dates on marine pelecypods dating back to 2.4 ka, suggesting a late Holocene sedimentation rate of about 370 cm/ka. In Eastern Channel, core 2006039-029 contained two dates, showing a sedimentation rate of about 90 cm/ka, and core 2006039-035 had a single date of 8.3 ka at 4.72 m (~60 cm/ka). Core A107-12G (location in Shaw et al., 2013) showed sedimentation at rates of 69 cm/ka to 98 cm/ka in the middle-late Holocene, with a sample at 4.5 m dating back to ca. 6 cal. a BP (ca. 5.6 ka BP) (Solignac et al., 2011).

Effects of postglacial relative sea-level changes

The postglacial relative sea-level lowstand depth as measured at lowstand deltas varies from −12 m at Pipers Hole River, at the head of the bay, to −16 m in Mortier Bay and Long Harbour (Shaw and Forbes, 1995). The delta in Mortier Bay (Fig. 52a) is similar to that at the head of Long Harbour (Fig. 52b) and comprises a flat-topped wedge of sediment (B) that has prograded into an adjacent muddy basin (maximum depth 128 m). As seen on Figure 52a, the entrance to Mortier Bay is deep (sill depth 55 m), so the delta has been graded to the former relative sea level, unlike at St. John's, in region 7, where a shallow sill created a lake to which the delta was graded. Similarly, the submerged delta

Table 2. Radiocarbon dates on marine pelecypod samples from cruise CCGS *Hudson* 2006039. The calibrated values are intercepts on the calibration curves, as supplied by Beta Analytic. These calibrations have not taken into account Delta-R, which is uncertain (McNeely et al., 2006) (although Solignac et al. (2011) used a value of 139 ± 61 a).

Core	Water depth (m)	Depth (cm)	Age ^{14}C years BP ($\delta^{13}C$ = 25 ‰)	Age calculated BP (no local reservoir correction)	Lab. #	Material
007	231	109	13860 ± 40	15730	Beta-227605	*Nucula tenuis*
007	231	300	13710 ± 40	15780	Beta-227606	*Yoldia* sp.
007	231	558	14090 ± 40	16280	Beta-227607	*Portlandia arctica*
009	184	177	13860 ± 60	15990	Beta-238551	*Nuculana tenuissulcata*
013	243	169	11000 ± 40	12680	Beta-227608	*Nuculana pernula*
013	243	245	12650 ± 40	14100	Beta-227609	*Nuculana pernula*
013	243	248	12540 ± 40	14000	Beta-227610	*Buccimun tenue*
013	243	271	12710 ± 40	14160	Beta-227611	*Nuculana pernula*
013	243	495	13740 ± 40	15820	Beta-227612	*Nuculana tenuissulcata*
015	189	22	12200 ± 50	13680	Beta-238552	*Nuculana* sp.
015	189	115	12990 ± 40	14890	Beta-238554	*Nucula expansa*
015	189	167	12870 ± 40	14610	Beta-227613	*Nuculana tenuis*
015	189	332	13610 ± 40	15630	Beta-227614	*Portlandia arctica*
015	189	514	13930 ± 40	16080	Beta-227615	*Portlandia arctica*
019	198	41	9020 ± 40	9690	Beta-227616	Bivalve fragments
019	198	125	10010 ± 40	11080	Beta-227617	*Nuculana pernula*
019	198	270	10620 ± 40	11940	Beta-227618	*Nuculana pernula*
019	198	302	10670 ± 40	12020	Beta-227619	*Nuculana expansa*
019	198	495	12010 ± 40	13420	Beta-227620	*Nuculana* sp.
019	198	644	12890 ± 40	14650	Beta-227621	*Mytelis trossulis*
027	403	146	1050 ± 50	630	Beta-227622	*Yoldia* sp.
027	403	391	1840 ± 40	1370	Beta-227623	*Macoma* sp. fragments
027	403	657	2410 ± 40	2040	Beta-227624	*Nuculana pernula*
029	224	218	4950 ± 40	5290	Beta-238555	*Nuculana tenuisulcata*
029	224	379	6470 ± 40	6960	Beta-227625	*Nuculana pernula*
031	297	25	9910 ± 40	10790	Beta-238556	Sea urchin fragments
031	297	574	10720 ± 40	12100	Beta-238553	*Nucula expansa*
032	211	267	11620 ± 40	13110	Beta-227626	*Nuculana pernula*
032	211	458	12560 ± 40	14010	Beta-227628	*Nucula tenuis*
035	212	472	8250 ± 40	8770	Beta-227629	*Macoma calcerea*

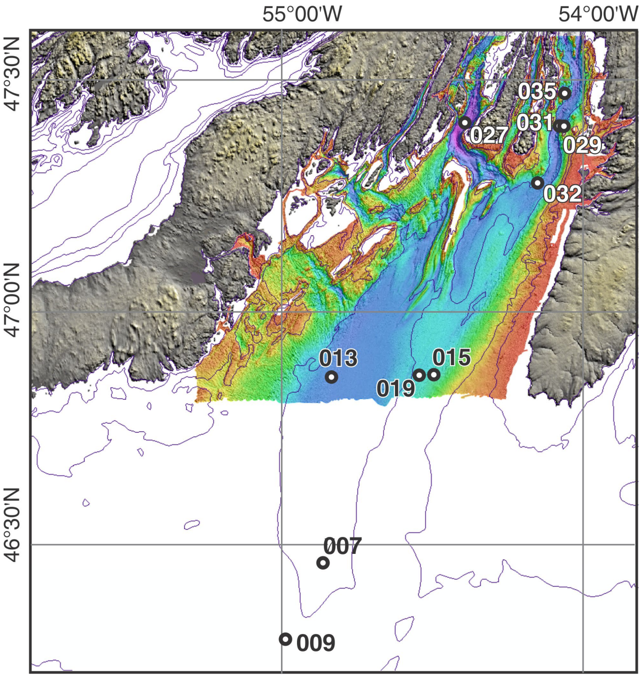


Figure 51. Location of piston cores (numbered) from cruise CCGS Hudson 2006-039. The radiocarbon dates from these cores are detailed on Table 2. The 200 m isobath is shown.

in Long Harbour (Fig. 52b) formed without a sill, and provides a reasonable estimate of maximum relative sea-level lowering.

The recently collected multibeam-sonar data show that south of Argentia, offshore from the Avalon Peninsula, glacial deposits have been eroded to create an erosional platform that extends to depths of about 40 m; below this depth glacial landforms are well preserved (*see* Potter and Shaw, 2009e). It is difficult to determine whether the relative sea-level estimates for Placentia Bay need to be re-evaluated, because wave-reworking of the glacial sediments would extend some way below the low-tide level. Furthermore, the regional isobase pattern suggests that a lower relative sea-level value would be predicted south of Long Harbour anyway.

Megaflutes

Megaflutes — erosional scours normally found in deep water on continental slopes — were identified in 1978 on sidescan sonograms and seismic-reflection profiles from Placentia Bay. (Confusingly, the term 'megaflutes' is also used to connote glacial landforms, cf. Stokes and Clark (1999, 2001)). They occur on the east side of the outer bay, at a depth of about 200 m, in a 2–3 km wide swath that continues to the south into Halibut Channel, over a total distance of about 100 km. The megaflutes have been formed by removal of a layer of postglacial mud, exposing underlying glaciomarine sediments and releasing a volume of 4.5 km^3. They occur in a range of forms, including single, multiple, and coalescent types, and in some areas at least their inception was related to pre-existing pockmarks (Fig. 53). Radiocarbon dates from piston cores were used by Shaw et al. (2013) to demonstrate that megaflute formation postdated ca. 9 ka BP. Whereas megaflute formation in Placentia Bay had previously been attributed to a 'reverse flow' from the tsunami generated by the 1929 Grand Banks earthquake, Shaw et al. (2013) proposed that they are formed by south-flowing density currents generated when volumes of cold saline water stored in the deep (>250 m) basins at the head of Placentia Bay are intermittently displaced and spilled south.

Anthropogenic effects

The large inlets in the bay have been the scene of human activities that have modified the seafloor. Modifications at Mortier Bay (Fig. 52a) include anchor-drag marks, and spud marks created by large drill rigs. At Long Harbour, off the former phosphorous reduction plant, dredge spoil lies on the floor of the submerged delta and also in the adjacent muddy basin, where its presence is indicated by patches of high backscatter strength (Fig. 52b).

The primary example of human impact on the seafloor is in the Argentia area (Parrott et al., 1995; Shaw et al., 1995a, c, 1996b). The Argentia Peninsula lies seaward of a series of sheltered basins (Fig. 54a). The entrance area (A) is shallow and wave dominated, with mobile sand and gravel at the seafloor, interspersed with areas of boulder gravel where glacial diamict is exposed. The sand forms a spillover wedge (B) on the flank of a basin. A series of shallow sills (C) characterized by mobile sand and gravel are transverse moraines modified during the postglacial relative sea-level lowstand. The basins (D) contain muddy sediments, gas-charged in places.

During World War II, and as part of agreements between the colonial power, the United Kingdom, and the United States of America, Argentia was selected as the site of a large military base for aviation and naval vessels. The transformation of the area involved among other things the removal of an existing community, the construction of an airfield on the peninsula, the excavation of storage areas for petroleum, and the construction of docks. The area was used intensely throughout the war, and less intensively for some years subsequently, until 1995 when the U.S. Naval Facility Argentia finally closed and the area was subject to remediation.

Evidence of the intense activity of the war and early post-war years can be seen on the seafloor in Figure 54a. Several areas were dredged to improve navigation. The spoil from dredging of the shoal (E) was dumped in the adjacent deep basin, where it is distinguished by higher backscatter strength than the surrounding mud. Large concrete blocks laid to anchor submarine nets are easily distinguished on the multibeam sonar imagery (F). The largest dredged area (Fig. 54b) is a submerged early-mid-Holocene gravel spit that has been dredged to produce a grooved pattern on the seafloor (A). The linear escarpment (B) marks the edge of

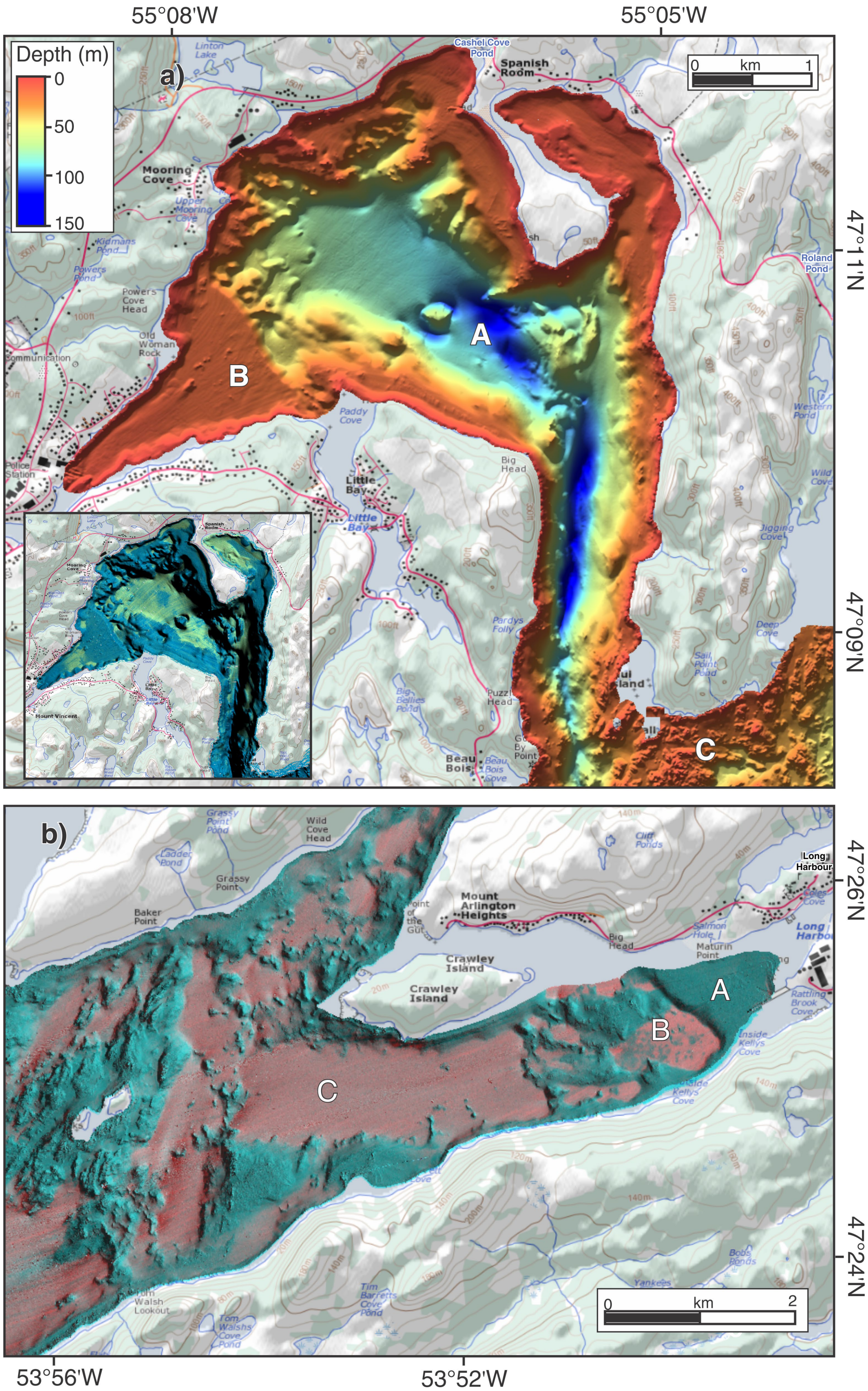


Figure 52. Shaded relief of **a)** Mortier Bay. The bay is relatively deep, with deposits of postglacial mud up to 35 m thick (A) overlying 0–5 m of glaciomarine mud in the deepest areas. The acoustic stratigraphy is mostly obscured by gas masking. High-resolution multibeam and sidescan-sonar data show that the seafloor is strongly imprinted by anchor-drag marks, and that mounds of dredge spoil occur on the shallow platform (B), which is an early Holocene submerged delta. Outside the entrance the terrain is rocky (C) with a few small sediment pockets. **b**) Long Harbour, with backscatter strength draped over the terrain. Pink areas are mud and greyish-green areas are areas of gravel and rock with high backscatter. Mounds on the submerged delta platform (A) are dredge spoil. The spoil can be seen in the adjacent muddy basin as patches of greyish-green (B). A series of muddy basins (e.g. C) contain gas-charged postglacial mud and are separated by irregular terrain with high backscatter.

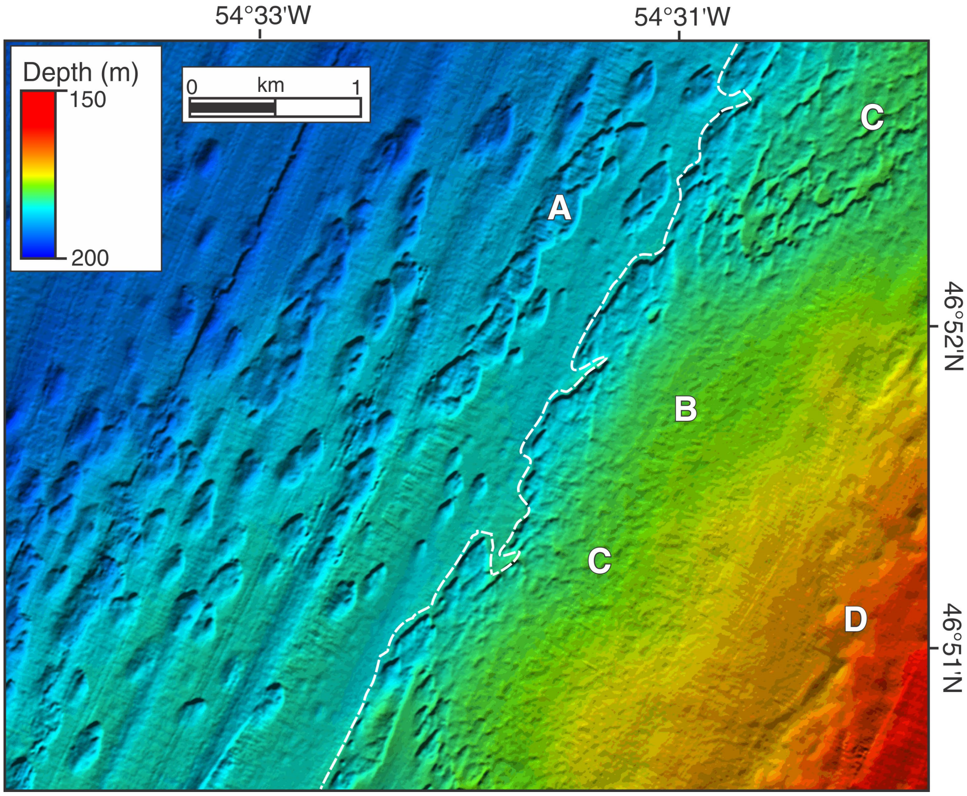


Figure 53. Field of megaflutes incised into the upper, postglacial mud unit. In places megaflutes overlap or have coalesced (A). The 2 m high escarpment (dashed line) is the boundary between the megaflute zone and the area of rough seafloor to the east (B) in which small residuals remain. Area C contains larger residual ridges. Yet farther east, all the postglacial unit has been removed, leaving a smooth seafloor (D). *See* full discussion in Shaw et al. (2013).

the dredging. Spoil was dumped just across the bay (C), where it resulted in a small submarine slide. A large pit (D) was dredged to create room for a floating dry dock. Systems of anchors and chains used to anchor seaplanes remained on the seafloor in 1995 (*see* inset on Fig. 54b). Finally, the muddy seafloor of Argentia Harbour is intensely turbated as a result of anchor dragging, particularly near wharves.

A deep-water dump site was established in the 450 m deep area at the south entrance to Western Channel. Multibeam-sonar surveys at the time of base closure targeted this area, and resulted in the first multibeam-sonar collection in the region; however, at this depth the resolution of the Simrad EM-1000 system was insufficient to detect targets. Sidescan-sonar surveys did, however, detect targets, many of them away from this deepwater site. Anecdotal evidence reveals the practice of dumping of material destined for the deepwater site at locations closer to Argentia!

Finally, a fleet of five trawlers was scuttled at the west side of the outer bay some years ago. These can be clearly seen on the multibeam imagery (*see* Potter and Shaw, 2010d); high-resolution sidescan-sonar images were published by Shaw et al. (2007).

REGION 7: AVALON PENINSULA

Setting

This region (Fig. 55) encompasses the areas offshore from the Avalon Peninsula, excluding Placentia Bay. It includes the southwest-facing St. Mary's Bay, the northeast facing Conception Bay, and a series of peninsulas. It lies in the Avalon tectonic zone, characterized by Late Proterozoic rocks of varying lithology overlain unconformably by a Late Proterozoic and Early Paleozoic shallow-marine succession. The steep rocky coasts and the shallow inner shelf are exposed not only to high-energy Atlantic wave processes, but also to impact by icebergs drifting south in the Labrador Current. The latter is only moderately strong near coasts compared with farther offshore, near the shelf break (Wu et al., 2011). Multibeam-bathymetry data are available only for two small areas: St. John's, and Holyrood, at the head of Conception Bay.

Coastlines

This region has predominantly bold, rocky coastlines, one exception being the bayhead barriers in St. Mary's Bay and Trepassey Bay (Forbes, 1984; Shaw and Forbes, 1987, 1988). In an early phase these were prograded strand plains; subsequently, the oldest (>1.2 ka), lowest beach ridges have been submerged in lagoons. A high barrier spans Holyrood Pond, a deep fiord basin. The sand and gravel barrier sits

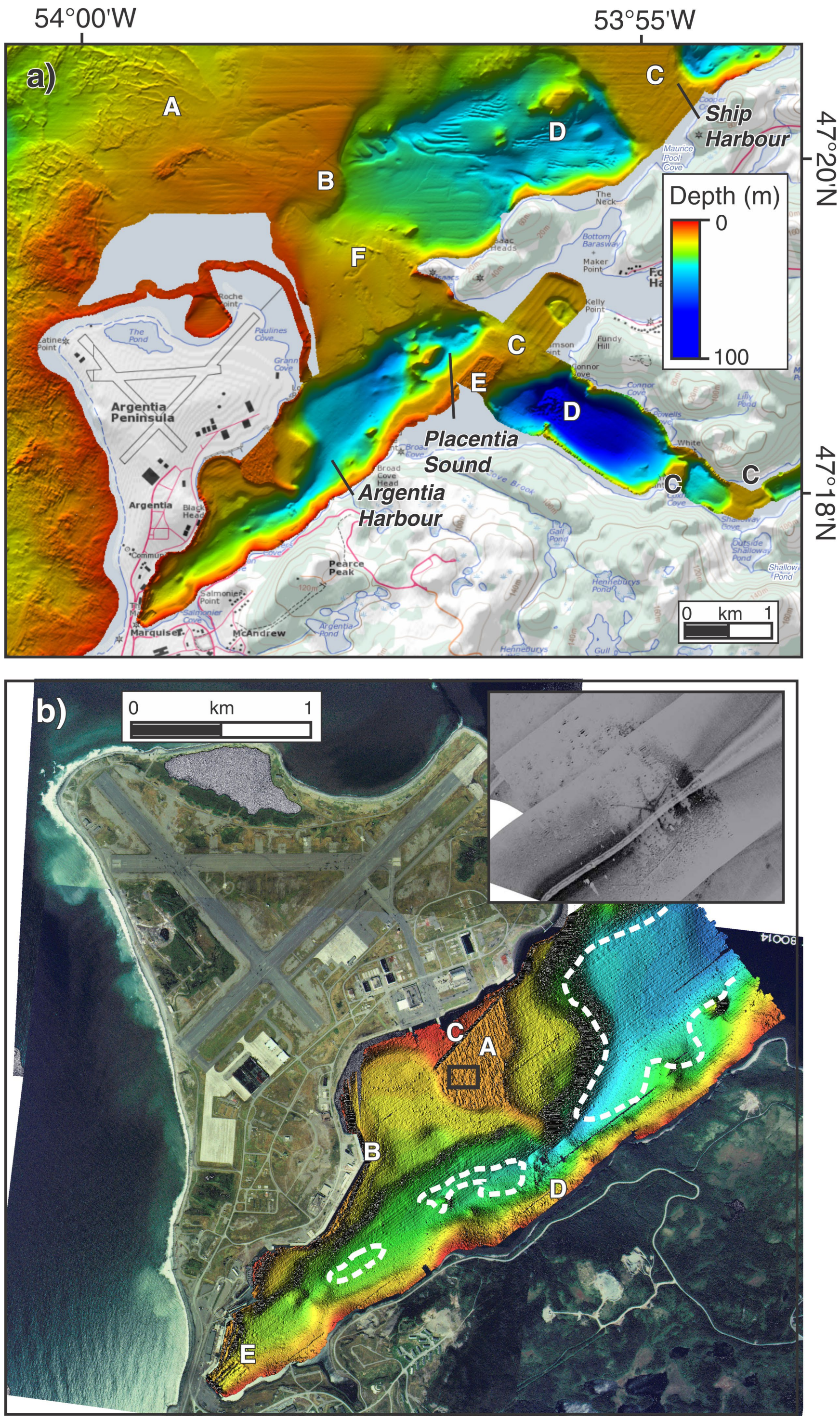


Figure 54. Geology of Argentia, Placentia Sound, and Ship Harbour. **a)** Colour shaded-relief image of Argentia Harbour and environs showing the sill (A), where glacial diamict is exposed at the seafloor, and sandy, spillover lobes (B). Flat-topped shoals (C) separate muddy basins (D). Area E was dredged during World War II. **b)** Image of Argentia Harbour derived from a sweep-style multibeam sonar operated by Public Works Canada. Area A, dredged in WW 2 to permit the approach of shipping to the new dock (B), is bounded by a sharp escarpment (C). Spoil was dumped just across the harbor (D). The excavation for a floating dry dock is seen at area E. The sidescan-sonar image (inset; location shown by box) shows debris not evident on multibeam images (anchor blocks, chains formerly used for anchoring seaplanes). Dashed line indicates extent of gas masking. The airphoto in Figure 54b is a composite image from aerial photographs taken on September 11, 1980, Newfoundland and Labrador Department of Forest Resources, rolls 80014 and 80015.

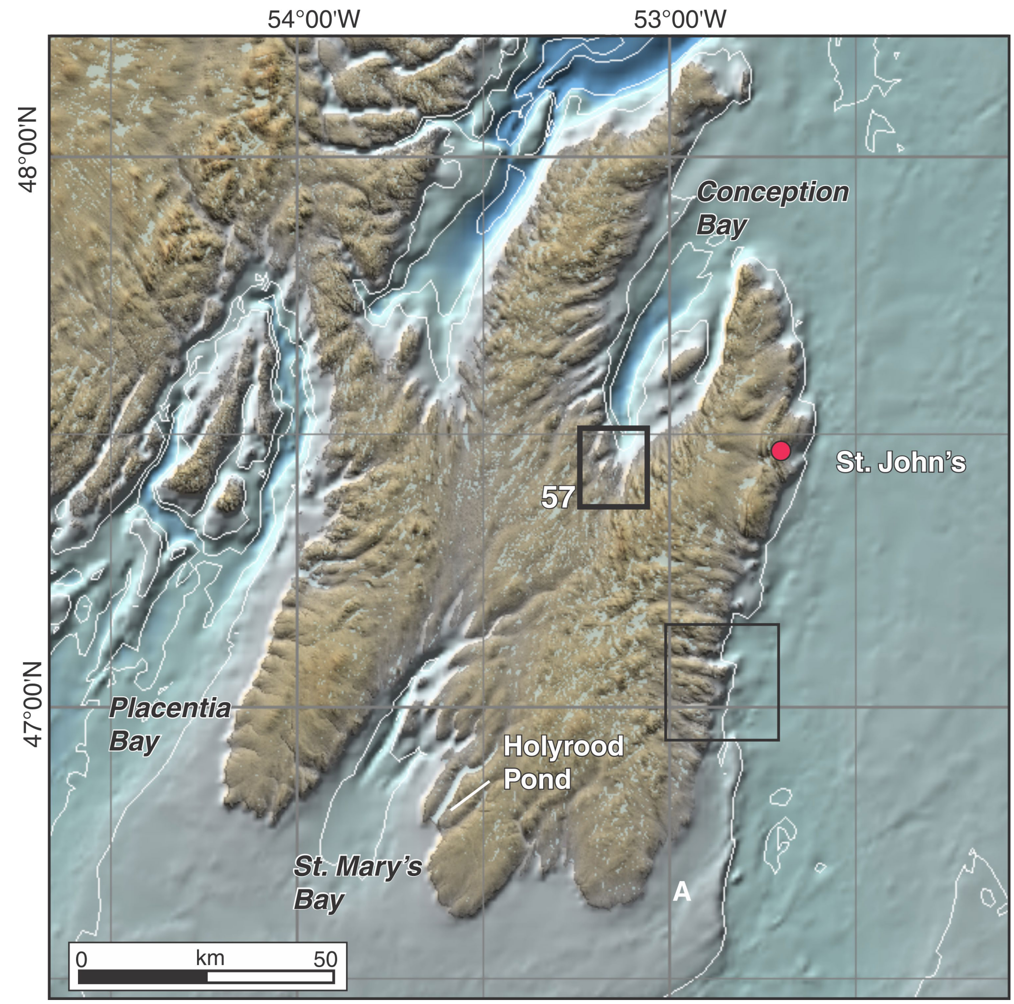


Figure 55. Region 7; the inset box shows coverage of the surficial geology map of King (2013). A submarine platform (A) lies off the southeast coast of the Avalon Peninsula. Figure also shows location of Figure 57.

on glacial diamict overlying the bedrock that forms the sill of Holyrood Pond. Sediment is supplied to the barrier from extensive eroding glacial bluffs (Forbes, 1984, Forbes and Taylor, 1987; Shaw and Forbes, 1988). The barrier is occasionally overwashed when high, long-period, storm waves approach from the south-southeast (Forbes, 1984). Parts of the settlement are then flooded, and traffic on the coastal highway is disrupted. Coastal protection works have been erected.

Nakashima and Taggart (2002) examined the pocket beaches and barrier beaches that are scattered around Conception Bay, and linked beach orientation and other factors to concentrations of capelin eggs.

Inner shelf: overview

Relatively little multibeam-sonar mapping has taken place in this region. The marine geology of St. Mary's Bay is relatively unknown. Hydrographic chart 4817 (Canadian Hydrographic Service, 2002a) shows that the bay is separated from the continental shelf by a sill at depths between 60 m and 70 m, north of which is a deep, narrow trough extending to −130 m. Troughs at the head of the bay extend to maximum depths of 226 m and 195 m. Limited survey data are available to reveal the complexity of the bay that is apparent on hydrographic charts. Huntec DTS surveys offshore from Holyrood Pond (Forbes, 1984) show no penetration of the seafloor, where surficial gravel overlies sand.

The low-resolution elevation model (Fig. 55) shows a platform offshore from the southern Avalon Peninsula. It is generally shallower than 100 m, and is bounded at its seaward edge by a ridge with shallowest depths of 30 m to 60 m. This ridge appears to comprise a series of bedrock pinnacles (G. Cameron, pers. comm., 2012), based on a line from GSC survey 2007-016 that traversed the ridge just east of Cape Race.

East Avalon Peninsula: Seal Cove to Motion Bay

This area (*see* box on Fig. 55) was mapped by King (2013) who demonstrated that topography is governed by Precambrian metasediments with a strong north-south structural trend, manifested as ridges and valleys, locally outcropping from a thin glacial diamict mantle, across a 2–4 km wide zone offshore the headlands. East of this, low-relief, Cambrian and Ordovician slightly metamorphosed shale and siltstone are overlain by a glacial diamict blanket of several metres or more thickness. This has scattered drumlins oriented transverse to the coast, i.e. an orientation similar to flow-parallel glacial indicators (Catto, 1998) and landforms seen on Shuttle Radar Topography Mission (SRTM) imagery of the adjacent Avalon Peninsula.

St. John's Harbour and approaches

Multibeam imagery of the St John's area (Fig. 56a) demonstrates the contrast between an inner-shelf environment and a sheltered harbour with anthropogenic influences. Outside the entrance to the harbour (Fig. 56b), the seafloor shelves steeply. Bedrock ridges trending northeast are veneered with glaciomarine sediment. Deeper areas have an overlay of postglacial mud heavily impacted by iceberg pits, and a few iceberg furrows, most likely modern rather than relict. At the harbour entrance the sediment cover is absent, and bedrock is exposed near the seabed.

On Figure 56c a notable feature is the −14 m bedrock sill that divides the harbour into two basins. Lewis et al. (1987) argued that the Holocene transgression over the rock sill at ca. 10 ka was registered in massive silty marine sediment overlying laminated lacustrine sediment. Prior to this time, the inner harbour was a lake. The multibeam data show clear evidence of the lake phase in the form of a submerged delta at the left in Figure 56c. The deltaic platform has a veneer of postglacial mud, is heavily marked by anchor-drag marks, and has been dredged in the vicinity of the shipyard. East of the frontal slope of the delta the basin contains soft postglacial mud, heavily marked by anchor-drag marks (Fig. 56d), and a mound of sediment deposited from a sewage outfall.

Conception Bay

Conception Bay (Fig. 55) has a deep (>275 m) central trough that shallows seaward to a sill at 150 m depth. Slatt (1974) showed that seafloor sediments in Conception Bay are heterogenous, with a strip of mud down the axis of the bay and very complex distributions of gravel and sand elsewhere, giving a total of 11 textural types. This complex distribution is partly explained with reference to the bathymetric imagery from Olex Ltd. that shows a series of narrow, curving bedrock ridges that converge at the outer bay (arrowed in inset at upper right, Fig. 57).

The multibeam-sonar imagery from the head of Conception Bay (Fig. 57) shows submerged platforms at the head of Holyrood Bay (A) and Gasters Bay (B), with breaks of slope at −22 m. These are interpreted as submerged early Holocene deltas. Seaward of the submerged delta in Holyrood Bay the hydrographic chart 4847 (Canadian Hydrographic Service, 2001) shows a sill shallower than −30 m, and also shows a spot sounding of −23.2 m. This raises the possibility that the deltas do not record maximum relative sea-level lowering, but formed in lakes behind a sill, as at St. John's Harbour. Unfortunately the multibeam coverage does not extend onto the sill to any great extent; however, the fact that there appears to be no sill for the Gasters Bay delta (B) suggests that these deltas do indeed record a postglacial relative sea-level lowstand of about −22 m.

A submerged platform (C) extending north from the peninsula separating Holyrood Bay and Harbour Main has a break of slope at −25 m. Shallower parts of the platform appear to consist of bedrock. It is interpreted as an erosional landform the formation of which was coeval with that of the deltas. The hydrographic chart (Canadian Hydrographic Service, 2001) shows that this platform is part of an arcuate ridge (dashed line). Similar arcuate ridges lie offshore from the other two bays. These ridges are interpreted as submarine moraines, formed by ice margins located at the bay mouths. Mounds and ridges in the bays (D) may be glaciofluvial deposits, and the anastomosing ridge at E may be an esker.

The seafloor in Holyrood Bay has several intriguing features (lower left inset, Fig. 57). The Holyrood thermal generating station burns fuel oil and is served by a wharf, offshore of which the seafloor is marked by anchor-drag marks that radiate away from the wharf to depths of 90 m (F). The presence of the marks suggests minimal disturbance of the seafloor sediments by modern processes. Just to the south of here, regularly spaced channels (G) several metres deep terminate at submarine fans 4 m above the surrounding terrain, very similar to those off Flat Island spit in St. George's Bay. The causative mechanism for these channels is unknown.

Lastly, the floor of Harbour Main is intensely impacted by fresh-appearing iceberg-grounding pits and some linear iceberg scours (upper left inset, Fig. 57). The sills outside Holyrood Bay and Gasters Bay protect them from impact by modern icebergs, whereas the −80 m sill outside Harbour Main allows them to enter. The seafloor in the deep area (H) outside all these bays is intensely turbated by iceberg pits, mostly old, but also some fresh in appearance.

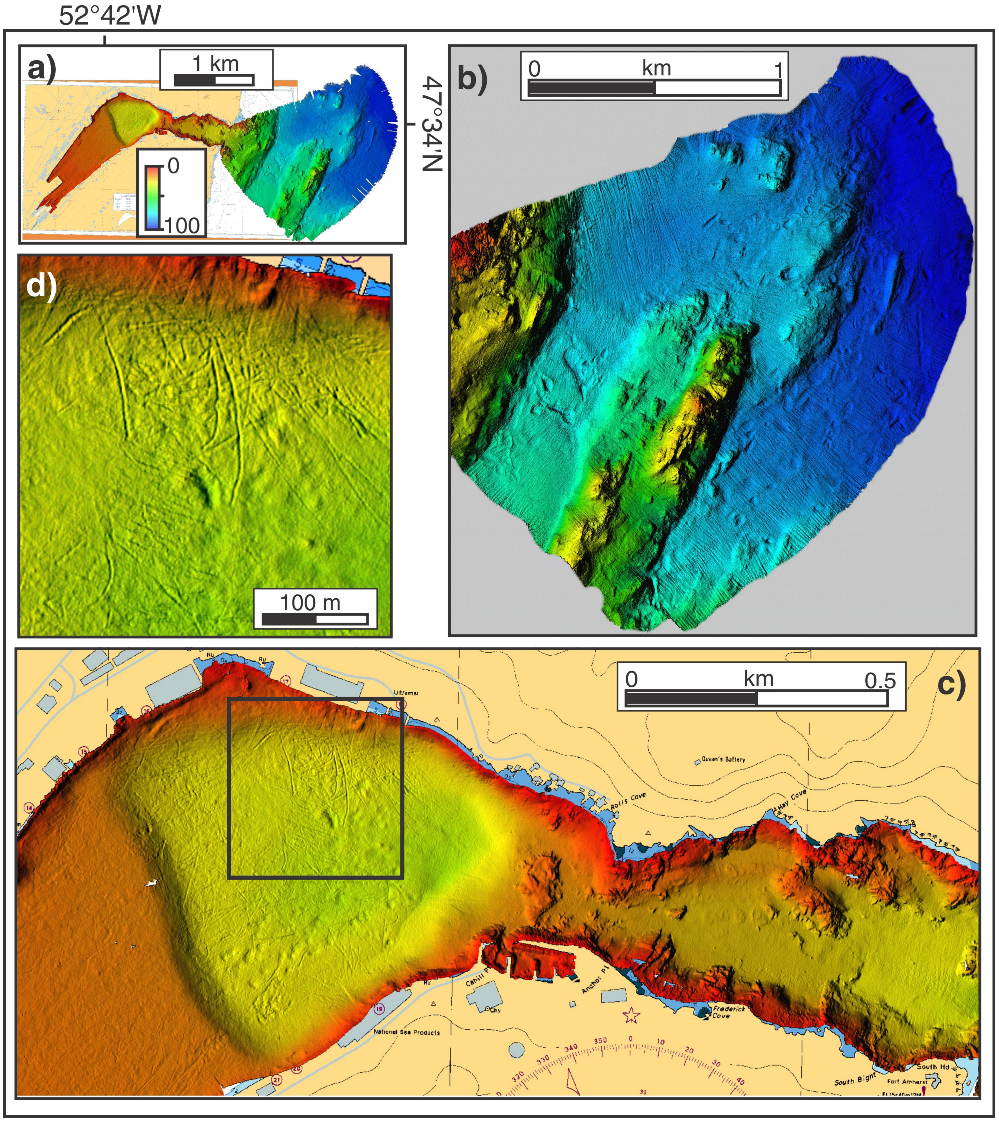


Figure 56. Multibeam bathymetry of St. John's Harbour and approaches; **a)** complete coverage, **b)** approaches to the harbour, **c)** part of the inner harbour, **d)** enlargement of part of the inner harbour (box in Fig. 56c) showing anchor-drag topography. Background is from Canadian Hydrographic Service (1995).

REGION 8: TRINITY BAY TO BONAVISTA BAY

Region 8 (Fig. 58) lies mainly in the Avalon tectonic zone, and partly in the Gander Zone (Colman-Sadd et al., 1990). The dividing line between the two — the Dover Fault — extends in a northeast direction, reaching Bonavista Bay at Freshwater Bay. The Gander Zone rocks are primarily Cambrian to Ordovician stratified sediments with Devonian and Carboniferous granitic intrusions. The Avalon Zone rocks are more varied, and comprise Late Proterozoic to Early Ordovician sedimentary and volcanic rocks, with Later Proterozoic to Cambrian intrusive rocks. Most geological boundaries extend in a northeast direction, and are intersected by a series of fiords that contain thick deposits of glaciomarine and postglacial sediments. Whereas the region is fully exposed to Atlantic Ocean, the inner shelf is deep and the area of seafloor subject to wave mobilization is relatively small.

Coastlines

The coastlines in this region are bold and rocky, and beaches and other unconsolidated coastal landforms are rare. Forbes's (1984) map of coastal survey sites shows only two locations in this region. The Geological Survey of Canada conducted limited investigations at Eastport (Fig. 58), located on the peninsula that is the northern boundary of Newman

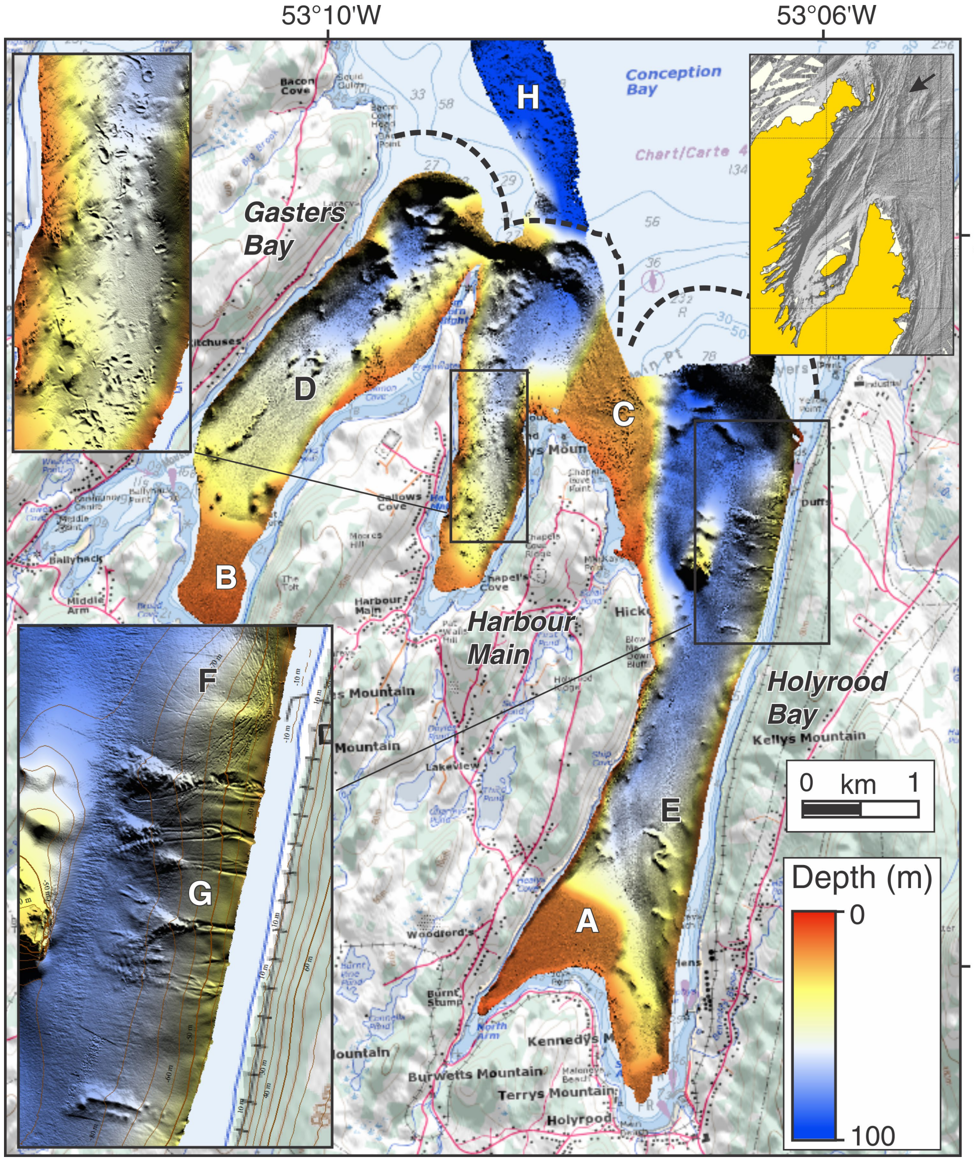


Figure 57. Multibeam bathymetry of innermost Conception Bay: Holyrood Bay, Harbour Main, and Gasters Bay. Location shown on Figure 55. Letters explained in text.

Sound. A conventional radiocarbon age of 5490 ± 120 ^{14}C years BP (Beta-27231) was obtained on a bulk sample of freshwater peat about 3 m below modern mean water level. The site was on the banks of a stream behind the beach, and the corer penetrated 4.5 m of peat, muddy sand, and gyttja. The implications of this date are unclear.

Inner shelf: overview

The region is dominated by two deep bays: Trinity Bay and Bonavista Bay. Trinity Bay deepens to more than 450 m, and shallows on the shelf to a sill at about 250 m. Bonavista Bay deepens to more than 325 m and has a sill at about 225 m depth. About 130 km northeast of the Bay de Verde Peninsula, a large arcuate submarine ridge (depth ~185 m) marks a major ice margin dating to ca. 16 ka that formed during ice retreat across the continental shelf (King and Sonnichsen, 2000; Cameron and King, 2011). The bathymetric complexity of the region is increased by the presence of a series of fiords that drain into the two principal bays. Random Sound and Smith Sound drain into Trinity Bay, whereas farther north, Newman Sound and Clode Sound are the principal fiords in the head of Bonavista Bay, with depths in excess of 300 m (chart 4855, Canadian Hydrographic Service, 1997). The last two converge to form a trough with a maximum depth of 455 m.

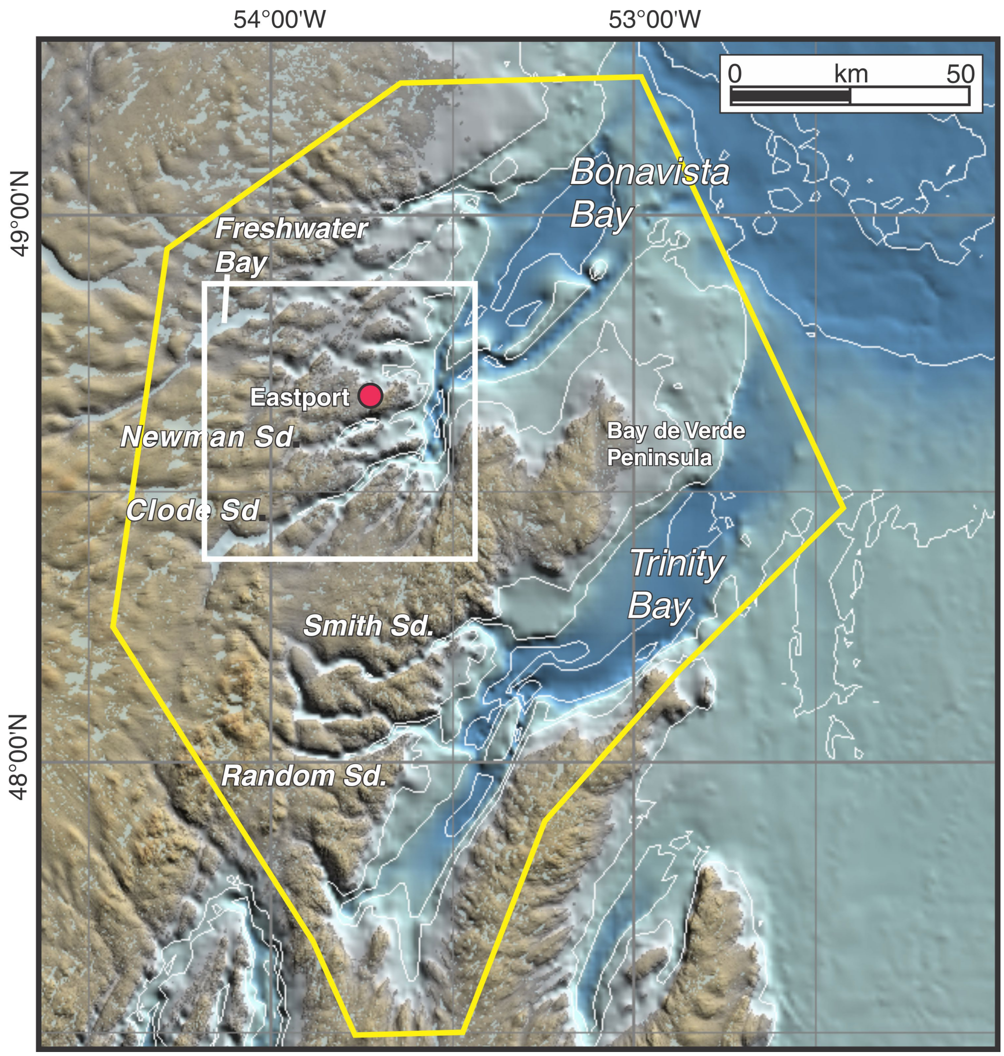


Figure 58. Region 8 encompasses the inner parts of Trinity Bay and Bonavista Bay, each of which contains a series of fiords. The white box outlines the coverage of GSC maps of seafloor topography (Patton and Shaw, 2011b) and backscatter strength (Patton and Shaw, 2011a).

Trinity Bay

Multibeam data are not available for this area and this brief description is based mainly on sparse geophysical survey data and bathymetric imagery data from Olex Ltd. Due east of the entrance to Smith Sound, data from GSC cruise 89006 clearly show the boundary between Proterozoic volcanic rocks near the coast and well stratified Cambrian-Ordovician rocks in deeper water. Relatively thick postglacial and glaciomarine deposits are found in the deep, narrow troughs of Trinity Bay, whereas the shallow, flanking shoulders have a thin Quaternary cover and are heavily imprinted by iceberg furrows and pits in water depths above about 200 m. The bathymetric data from Olex Ltd. show a smooth central bay, with possible traces of large, streamlined glacial landforms (megascale glacial lineations), and shallow (~100 m) bedrock platforms on either side, particularly on the north side off the Bay de Verde Peninsula.

Little is known of the marine geology of Random Sound (maximum depth 347 m) and Smith Sound (365 m), fiords that tributary to Trinity Bay. Shaw and Forbes (1995) showed that the head of Random Sound contained three submerged early Holocene deltas at depths 9.7 m below modern mean water level.

Bonavista Bay

The trough of Bonavista Bay attains a maximum depth of 347 m, and is separated from the shelf beyond by a sill at –220 m. Bathymetric imagery data from Olex Ltd. show that the sill is largely bedrock, but the western part, adjacent to the trough, has a smooth appearance, and may comprise Quaternary sediments. Cumming et al. (1992) showed that the sedimentary package above acoustic basement in Bonavista Bay was deposited during the deglaciation of the late Wisconsinan ice sheet. During the late Wisconsinan maximum (~20 ka), grounded ice in Bonavista Bay extended farther offshore. Deglaciation of the bay was rapid and occurred prior to about 13.5 ka. A basal till was deposited beneath the grounded ice, and after lift-off an ice shelf developed over the outer basins where diamicton was deposited. The inner bay gradually deglaciated as the ice margin retreated to the present-day shoreline by about 13 ka, and fine-grained outwash sediments transported by interflows were rapidly deposited. The inner bay remained under the influence of one or more remnant ice centres until ca. 10 ka, with ice positioned on the areas informally named 'Bonavista peninsula' and 'Gander peninsula'. Normal marine conditions were established in the outer bay by about 13.5 ka and in the inner bay and fiords by about 10 ka.

Figure 59 shows the multibeam-sonar imagery of the inner part of Bonavista Bay, mainly the fiords of Clode Sound–Chandler Reach, and Newman Sound. Figure 60 shows the backscatter strength for the same area. A detailed analysis of this region is contained on the pair of maps by Patton and Shaw (2011a, b).

Newman Sound

A sill at a depth of about 270 m (*see* arrow on inset profile C-D, Fig. 59) separates the fiord from the deeper waters to the east. This sill is mostly bedrock, possibly with some morainal material superimposed. As shown on the profile, Newman Sound can be divided into two parts in terms of bathymetry. In the outer portion, inland of the sill, the sound has steep sidewalls and a flat floor with a maximum depth of 313 m. Acoustically transparent postglacial mud with low backscatter strength (*see* Fig. 60) overlies thick, acoustically stratified glaciomarine sediments.

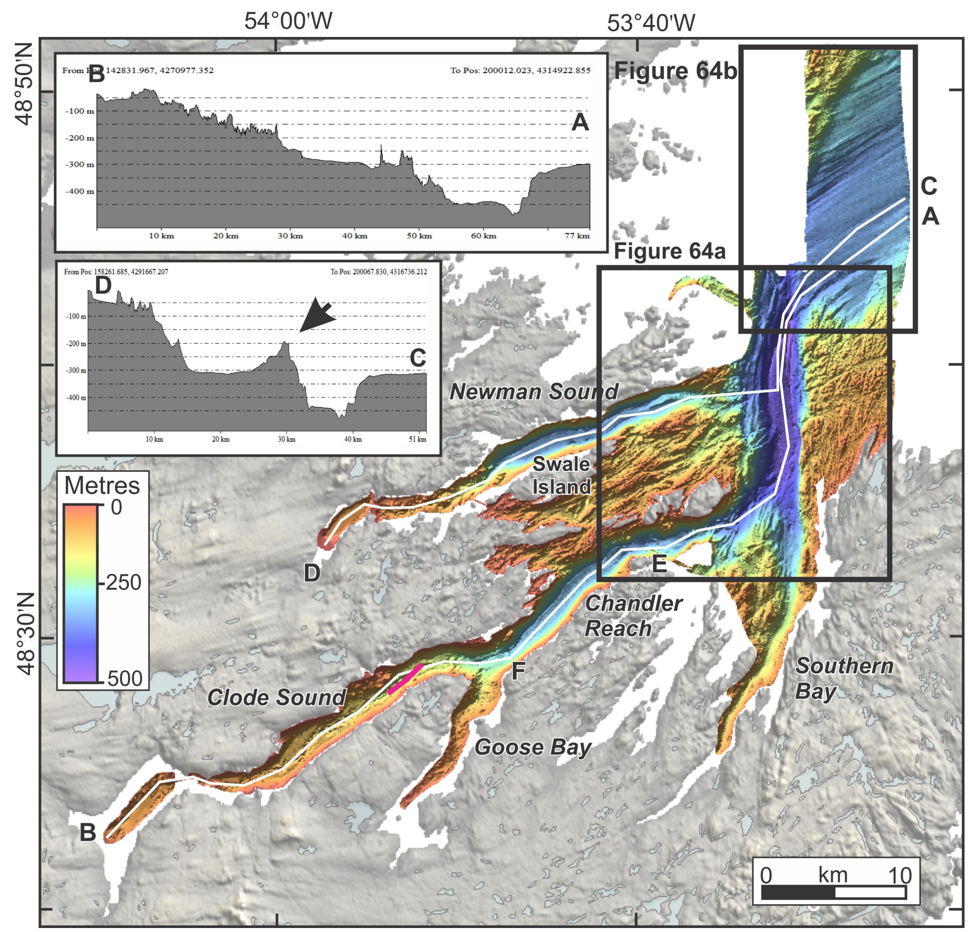


Figure 59. Multibeam imagery of inner Bonavista Bay, with insets showing profiles of Newman Sound (top) and Clode Sound–Chandler Reach (below). White lines show profile locations on the insets. E-F marks the location of Huntec DTS seismic-reflection profile of Figure 62. Red line is the Huntec DTS record shown as an inset on Figure 61. Boxes indication locations of Figures 64a and b.

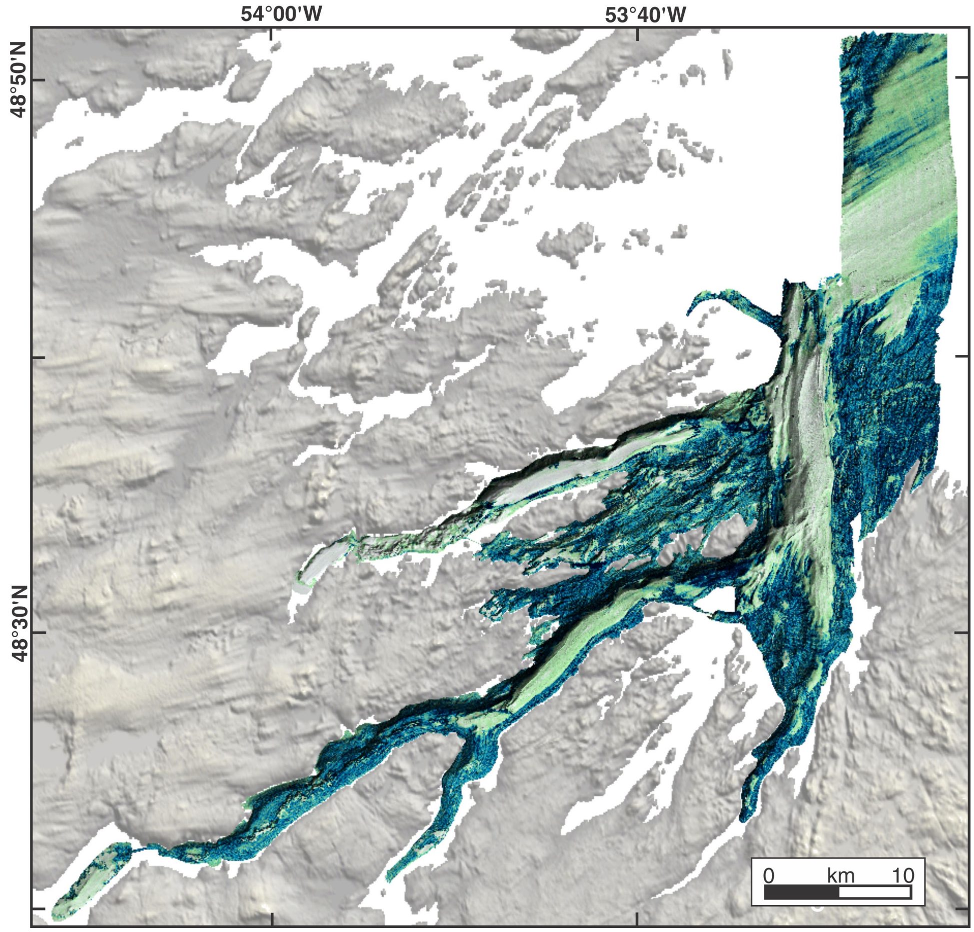


Figure 60. Backscatter strength of inner Bonavista Bay. The backscatter is bimodal, so that light tones correspond with areas of postglacial mud and dark tones with bedrock and surficial gravel.

In the second, inner portion, the sound is shallow (30–40 m) and topographically irregular, with patchy backscatter strength (Fig. 60). The seafloor is indented by a series of depressions up to 40 m deep, as shown on Figure 61a (A), similar to the topography in the upper reaches of Clode Sound. A ridge (B) stretching across the sound at Buckley Point is likely a moraine. Inland of Buckley Point the depressions are covered by thick postglacial mud, resulting in a smooth seafloor (C). A shallow terrace (D) at the head of the bay is an early-Holocene submerged delta. A brief outline of the connections between the geomorphic subunits of the sound and marine habitats is provided by Copeland et al. (2007).

Clode Sound and Chandler Reach

Like Newman Sound, this fiord trough is composed of several sections with differing morphology. The outer limit of the fiord is a 270 m deep sill (inset profile A-B, Fig. 59) just north of Chance Harbour Ledge. This sill may be a moraine. Southwest of here, as far at the point where it bifurcates into two arms, Chandler Reach is a classic fiord, with steep sidewalls and a smooth deep floor marked only by sedimentary furrows. Depth shallows from 320 m in the outer areas to 230 m in the inner. As in Newman Sound, postglacial mud overlies thick glaciomarine deposits. Profile E–F (Fig. 62) shows that Chandler Reach contains up to 50 m of an acoustically incoherent ponded unit with an irregular upper surface, interpreted as a mass-transport deposit (debris flow) formed close to a former ice margin in the fiord. It is overlain by a drape of acoustically stratified glaciomarine mud capped by acoustically transparent postglacial mud.

West of the bifurcation into two arms, Clode Sound (*see* profile, Fig. 59) is considerably shallower. It also has the characteristics of inner Newman Sound. As seen on Figure 61b, a terrace at a depth of about 90 m (A) is incised by a series of irregular depressions (B) up to 40 m deep. Backscatter data show that the flat-lying, terrace-like areas have high backscatter, and are probably floored by fine, angular gravel, whereas the depressed areas contain pockets of postglacial mud. Huntec™ DTS data (*see* inserted profile) show that the remnant areas between pits contain closely spaced, parallel, horizontal internal reflections similar to the acoustic character of glaciomarine sediments. In the wider section of the

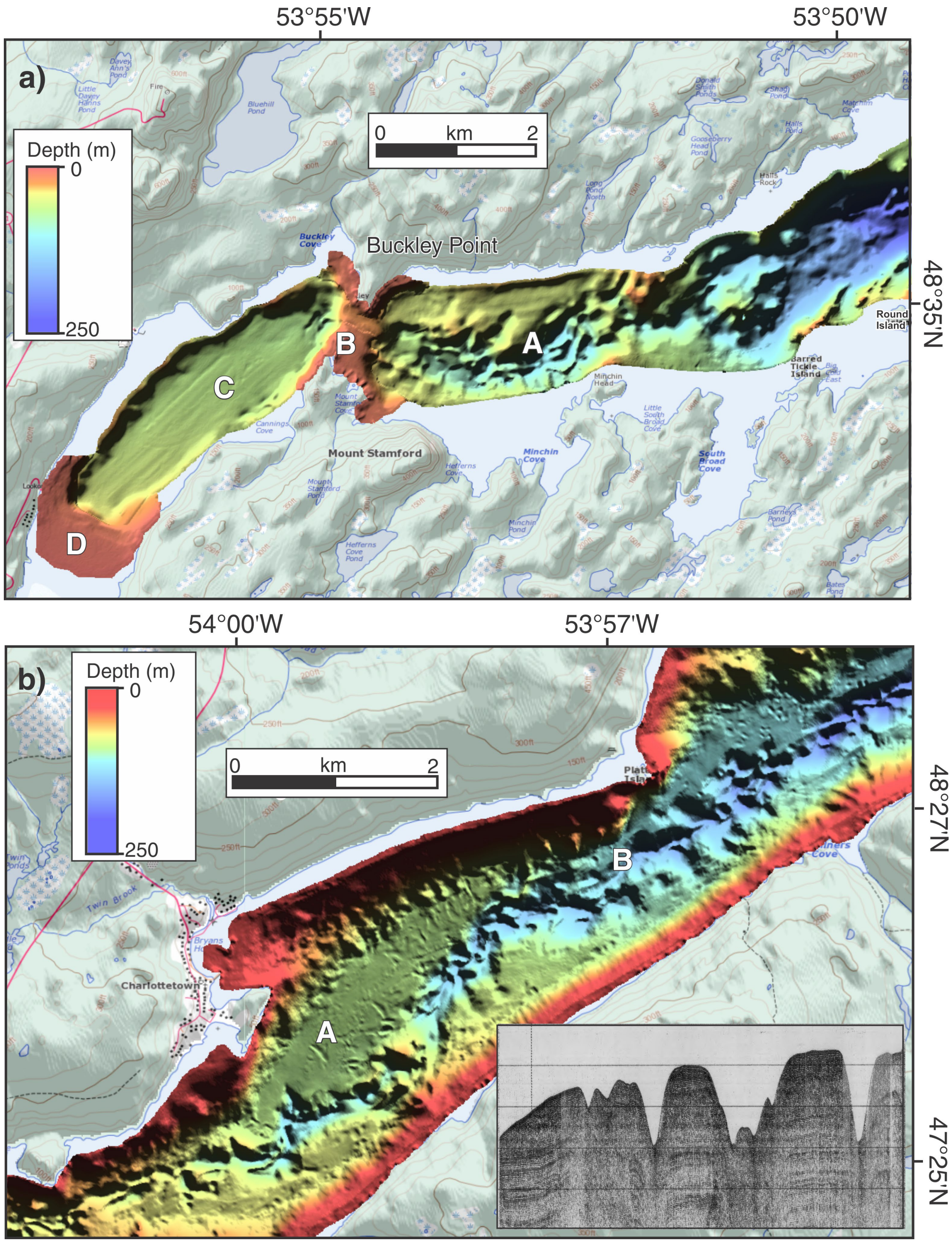


Figure 61. a) Inner part of Newman Sound; **b)** inner Clode Sound with inset showing part of a Huntec DTS record across the pitted terrain. The location of the Huntec record is outside this image, and is shown as a red line on Figure 59.

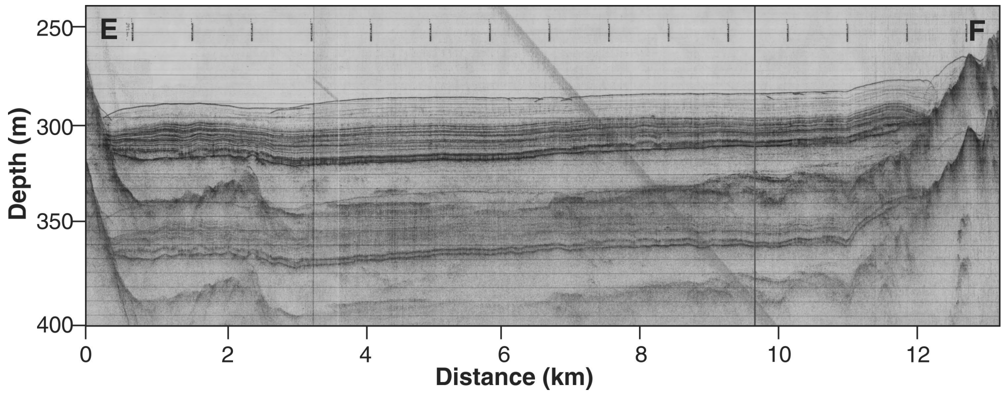


Figure 62. Part of a Huntec DTS record (cruise 87033) on a line running along the thalweg of Chandler Reach (E-F profile location marked on Fig. 59).

fiord, inland of the narrows and just west of Bunyan's Cove, the irregular terrain is also present, but is mostly masked by thick and expansive deposits of postglacial mud.

Swale Tickle

The great complexity of the seafloor in the region precludes a detailed discussion in this bulletin. This complexity is illustrated by reference to Swale Tickle (Fig. 63a, b), a shallow area south of Swale Island (Fig. 59). In the west, Swale Tickle connects with Newman Sound via several narrow channels at the White Islets (arrow on Fig. 63a), whereas in the east the seafloor drops off steeply into the deep trench. The maximum water depth in the tickle is about 120 m. Swale Tickle is floored by folded and faulted northeast-trending bedrock ridges. Quaternary sediments occur only in the deeper areas (Fig. 63a, A), where the smooth seafloor probably consists of sand or sandy mud, depending on depth and exposure. A notable feature in the tickle is a smooth, 3 km long ridge with high backscatter strength located just south of Swale Island (B). It is interpreted as an elongated drumlin that formed under the influence of glacier ice moving northeast. As shown on the higher resolution image (Fig. 63b), the ridge is imprinted by several iceberg pits, and indeed, iceberg pits are common throughout the tickle, at depths below 100 m. They are found only where Quaternary sediments occur (the icebergs leave no trace on bedrock). By analogy with the iceberg-pitted coastal waters of adjacent Notre Dame Bay (Shaw et al., 1999b), they most likely were formed by modern icebergs.

Seafloor outside the fiords

This area immediately outside the fiords (Fig. 64a, b) comprises several geomorphic subprovinces, primarily distinguished on the basis of depth. The north-south trough is mostly deeper than 450 m and attains a maximum depth of 499 m at its north end (Fig. 64a), beyond which it widens and shallows into a broad sill at depths of about 320 m. The trough is floored by thick postglacial mud, marked by sedimentary furrows and fluid-escape structures, overlying thick glaciomarine mud deposits. Areas of rugged bedrock occur in shallow water on either side of the trough. Seaward of the trough (Fig. 64b) the seafloor is relatively smooth, and consists of postglacial mud overlying terrain heavily imprinted by northeast-trending megaflutes. The limited extent of the multibeam and the thick postglacial cover make it difficult to provide dimensions of these landforms. Relief ranges up to 14 m, width up to 400 m, and length up to 2.6 km. Elongation ratios (Stokes and Clark, 1999, 2001) exceed 7:1, placing these features in the megaflute class. In places the glacial lineations are in the form of crag-and-tail features, that is, tapering ridges formed in the lee of bedrock outcrops. Bathymetric data from Olex Ltd. show that the lineations continue offshore for a distance of about 50 km.

Glacial processes in inner Bonavista Bay

Both Bonavista Bay and Trinity Bay are believed to be feeder areas for ice streams that continued offshore at the last glacial maximum (Shaw et al., 2006d). The orientation of megaflutes just outside the deep basin, and of the large drumlin in Swale Tickle, are consistent with a cover of eastward-flowing ice derived from an ice-dispersal centre to the west, as argued by Batterson and Taylor (2001). A provisional interpretation of the pitted terrain (Fig. 61) in the upper reaches of Newman Sound and Clode Sound is that it results from the decay of stagnant glacial ice that occupied the fiord during deglaciation, ca. 14 ka. It has strong similarities with the terrain produced on land (dead-ice sinks and dead-ice moats) in which wasting ice lobes became isolated and buried by outwash (*see* Fleisher, 1986; Fig. 11.75 *in* Benn and Evans, 1998). Why this terrain is not observed in fiords elsewhere on the island is unknown at present.

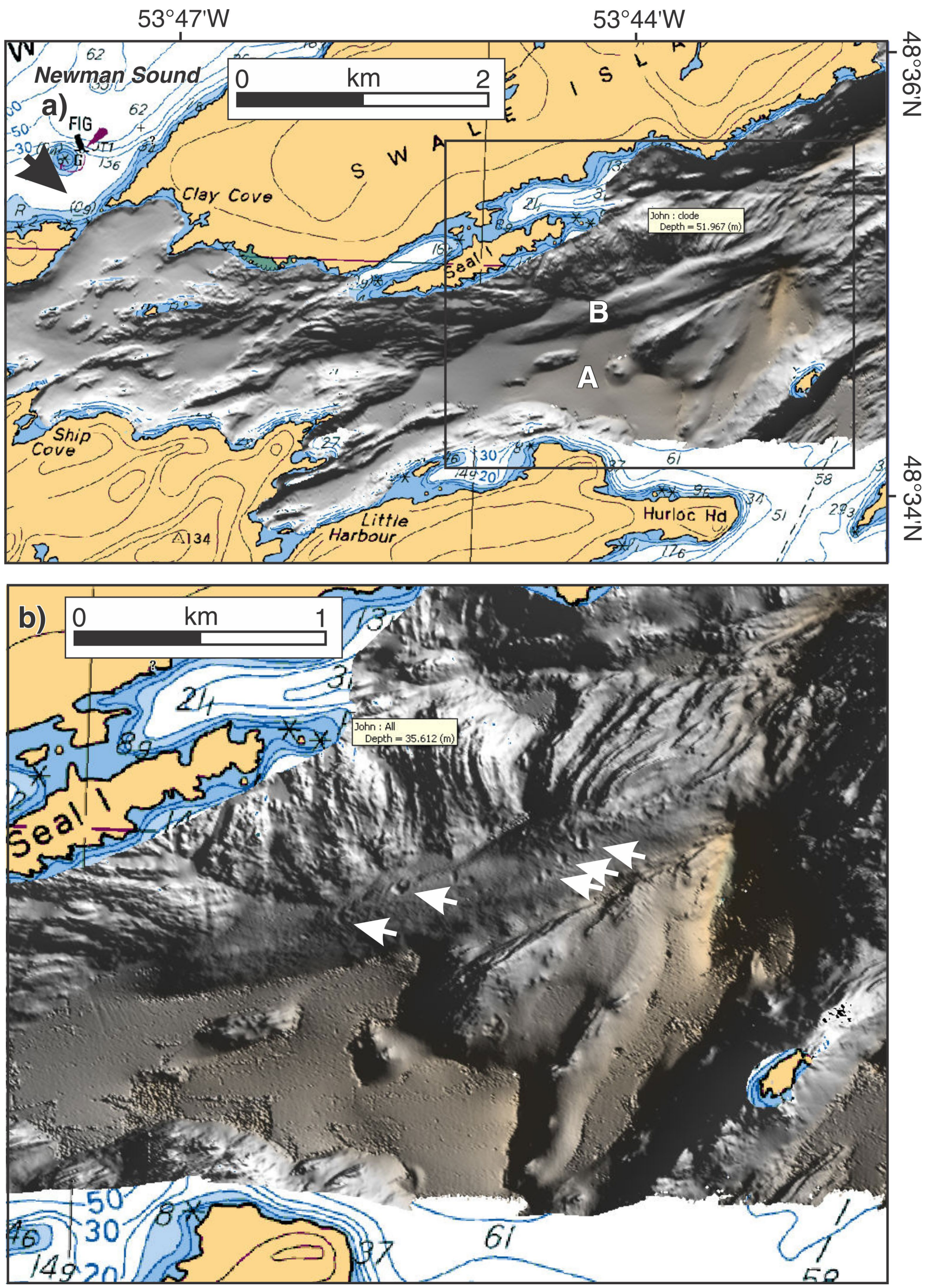


Figure 63. **a)** Grayscale image of part of Swale Tickle, illustrating the complex terrain away from the principal fiord troughs, including folded bedrock, a large drumlin marked by iceberg pits, and smooth sandy seafloor in depressions; **b)** enlargement of part of Figure 63a showing modern pits (arrowed) on the flanks of the drumlin. Background from Canadian Hydrographic Service (1997).

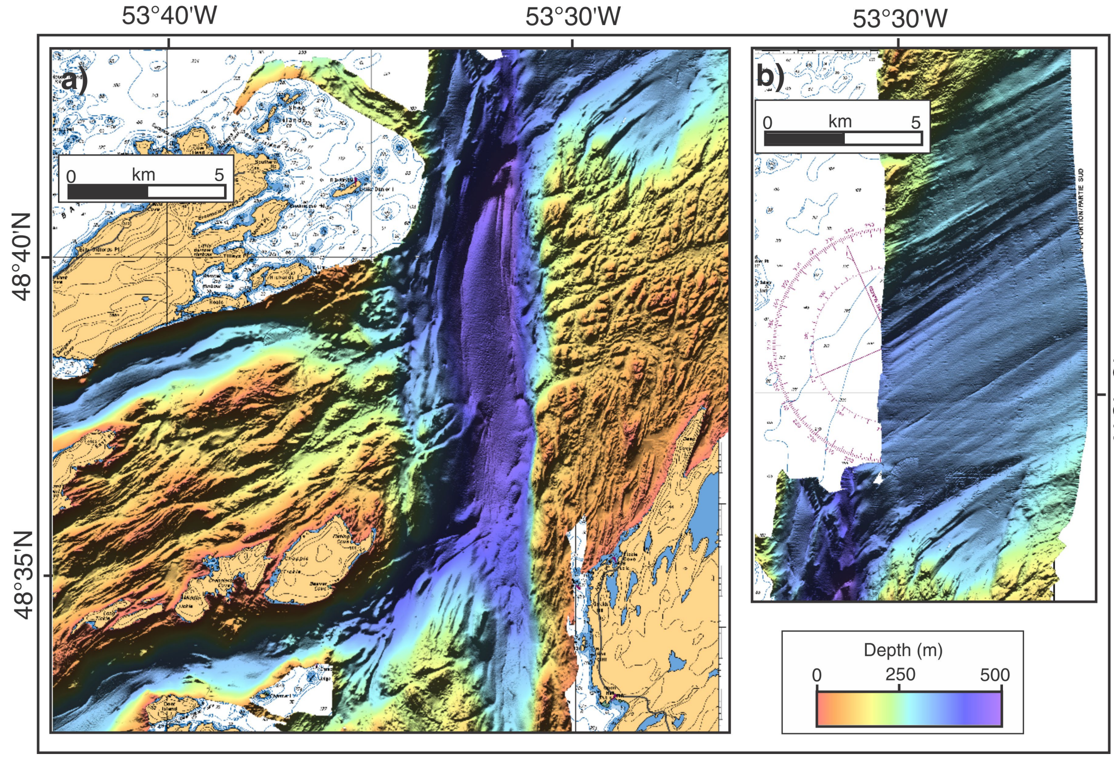


Figure 64. Outer parts of the Bonavista Bay multibeam-sonar coverage: **a)** a deep trough at the entrance to Newman Sound and Chandler Reach, flanked by areas of bedrock to the east, and to the west, north of Swale Island; **b)** seafloor fluting indicative of glacial ice movement out of the bay. Locations shown by boxes on Figure 59. Background is by Canadian Hydrographic Service (1997).

REGION 9: CAPE FREELS TO DOG BAY

Setting

Region 9 (Fig. 65) lies within the Gander tectonic zone, and is characterized by Devonian and Carboniferous granitic intrusions, with Cambrian-Ordovician stratified rocks west of Ragged Harbour. This region is distinctive because of low coastal relief and the extensive sandy beaches and dune systems of the 'Straight shore', the informal appellation for the coast between Cape Freels and Hamilton Sound. It encompasses the estuary of the Gander River, the principal river of northeast Island of Newfoundland. The inner shelf is relatively wide and shallow — the 100 m isobath lies an average of 25 km offshore. Coasts and shallow areas are fully exposed to Atlantic Ocean wave processes. The entire region is subject to seafloor impact by icebergs moving south in the Labrador Current, as well as sea ice.

Region 9 was surveyed in the early 1990s as part of an effort to evaluate the gold placer potential of the Notre Dame Bay region. This work is described in various cruise reports (Shaw et al., 1990a, 1992; Jenner and Shaw, 1992; Edwardson et al., 1993) and a synthesis report (Shaw et al., 1999b). Other relevant publications address the marine geology of Hamilton Sound (Shaw and Edwardson, 1994) and the postglacial sea-level history of the region (Shaw and Forbes, 1990c). Shaw et al. (1999b) described two models for marine processes in Notre Dame Bay, one of which was specific to region 9, primarily because the wave-impacted zone is so wide.

Coastlines

The coastlines in this region present the largest area of sandy coastal landforms on the Island of Newfoundland. Along the so-called 'Straight shore' (Fig. 65, 66) coastal dunes lie behind sandy beaches (Shaw et al., 1999b). Sandy tombolos, back-barrier lagoons, and extensive sand dunes occur at Cape Freels, where a broad active washover sheet extends into the backbarrier lagoon. The modern dunes overlie eroded remnants of older dunes that are capped by a layer of peat with a basal age of 1630 ± 50 ^{14}C years BP (Shaw and Forbes, 1990c). Another sandy tombolo is located at Lumsden. Farther north the barrier beach in Deadman's Bay encloses a brackish lagoon connected to the ocean at its south end. This barrier has a complex structure that includes moribund flood-tidal deltas, tidal channels, and washover fans that attest to an arrested transgressive phase. Radiocarbon

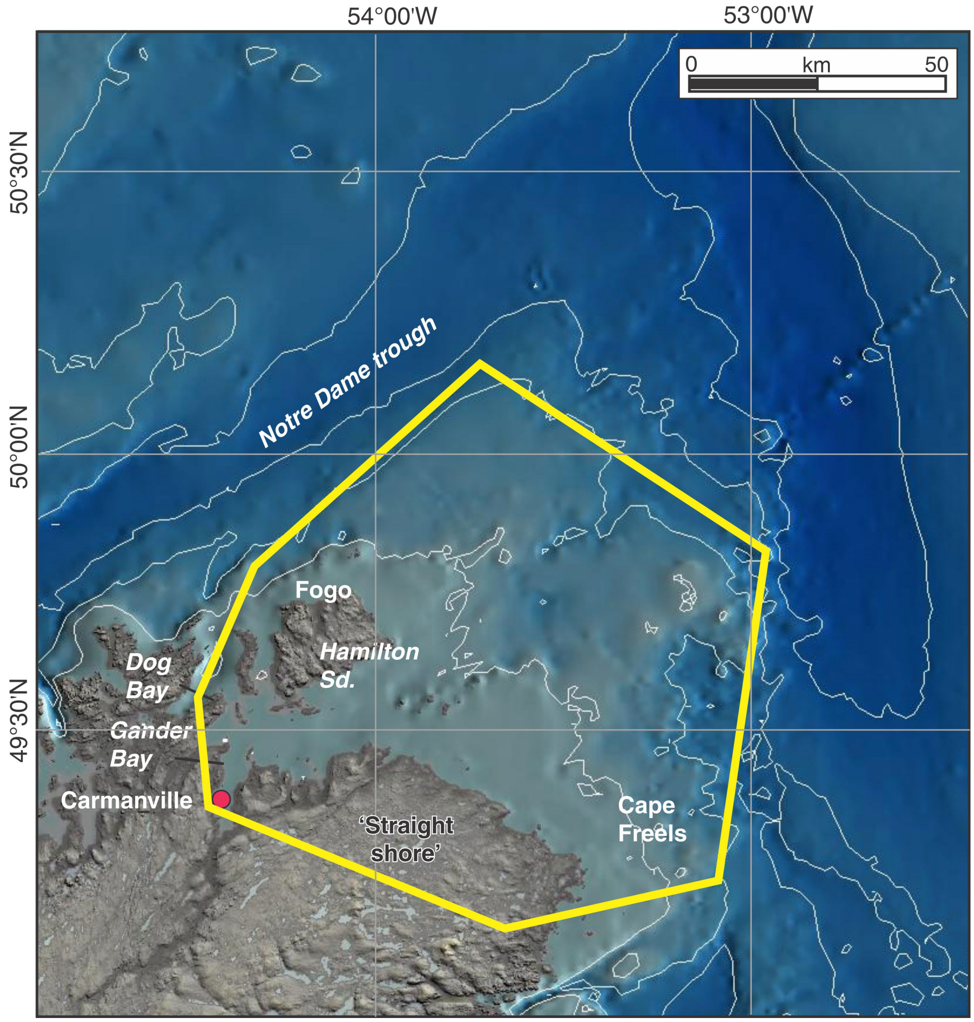


Figure 65. Region 9.

dates on organic deposits overlying the moribund landforms range back to 1780 ± 80 ^{14}C years BP (Shaw and Forbes, 1990c). Yet farther north, the large foreland at Man Point (Fig. 67) consists of beach and dune ridges mantled by up to 2.9 m of freshwater peat with basal dates ranging back to 3150 ± 90 ^{14}C years BP. The foreland is being trimmed by erosion, exposing the peat-mantled dune ridges and swales.

Shaw and Forbes (1990c) reached several conclusions regarding these shorelines. Relative sea level approached the modern level ca. 3000 years ago, and formerly transgressive coastal systems such as that at Deadman's Bay became stabilized in their present positions at an early date. Similarly, the large dune-ridge foreland at Man Point, which must have been emplaced very rapidly (perhaps as a system farther offshore disintegrated), ceased growth at that time, and became blanketed by freshwater peat. The foreland is now in disequilibrium, is being rapidly eroded, and sediment is moving farther north on the nearshore bar system, bypassing headlands to reach Ragged Harbour, where prograded beach ridges have formed over the past century. The coastal compartment at Ragged Harbour will become the recipient of sediment released by the destruction of Man Point, and when it fills entirely, sediment will bypass it and reach Hamilton Sound.

By contrast with the 'Straight shore', coastlines of Hamilton Sound (*see* map *in* Shaw et al., 1999b) are mostly low and rocky, with boulder-strewn intertidal zones. The east coast of Dog Bay is predominantly rocky, with gravel-pocket beaches, but near the head of the bay, pebble and cobble beaches lie behind wide intertidal flats, and a delta has extensive flats of pebbly sandy mud strewn with boulders organized into polygonal 'garlands' (cf. Forbes and Taylor, 1987, 1994). In Gander Bay, the west coast as far south as Fox Island has numerous pebble-cobble–pocket beaches. Farther south, the coast is predominantly low and rocky with mixed sand and gravel beaches and wide, sandy intertidal flats in places. Along the east side of the bay, near the causeway, boulder-strewn tidal flats front a low, rocky shoreline.

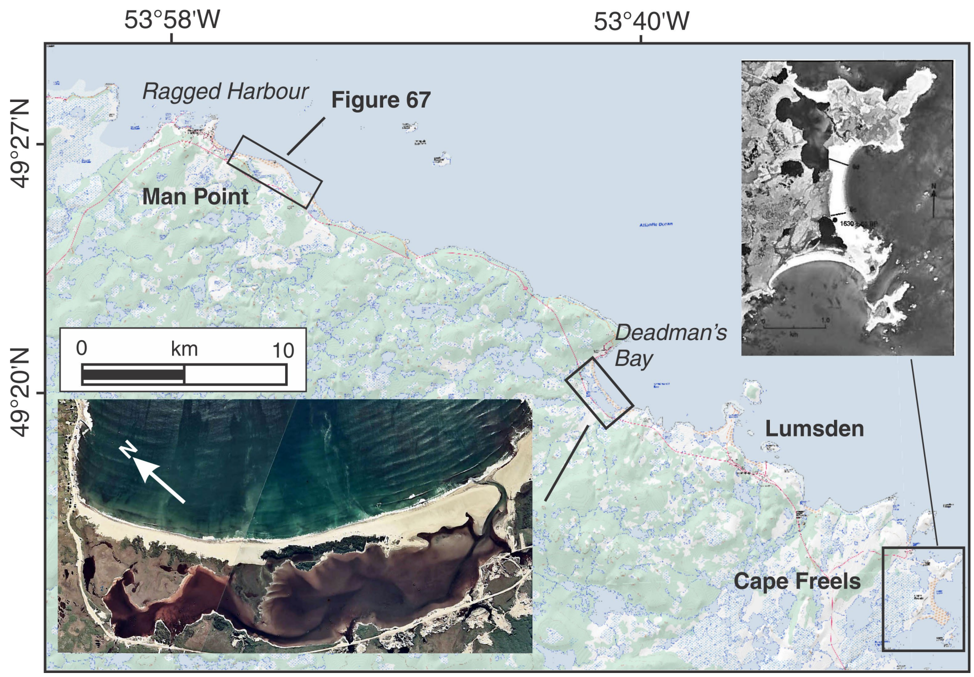


Figure 66. Sandy coastal systems along 'Straight shore'. The inset of the Cape Freels area at upper right shows sandy tombolas, with bedrock shoals visible just offshore. The image of Deadman's Bay (lower left) shows an overwashed barrier beach that appears to be relatively stable. The complex history of this beach is discussed in detail by Shaw and Forbes (1990c).

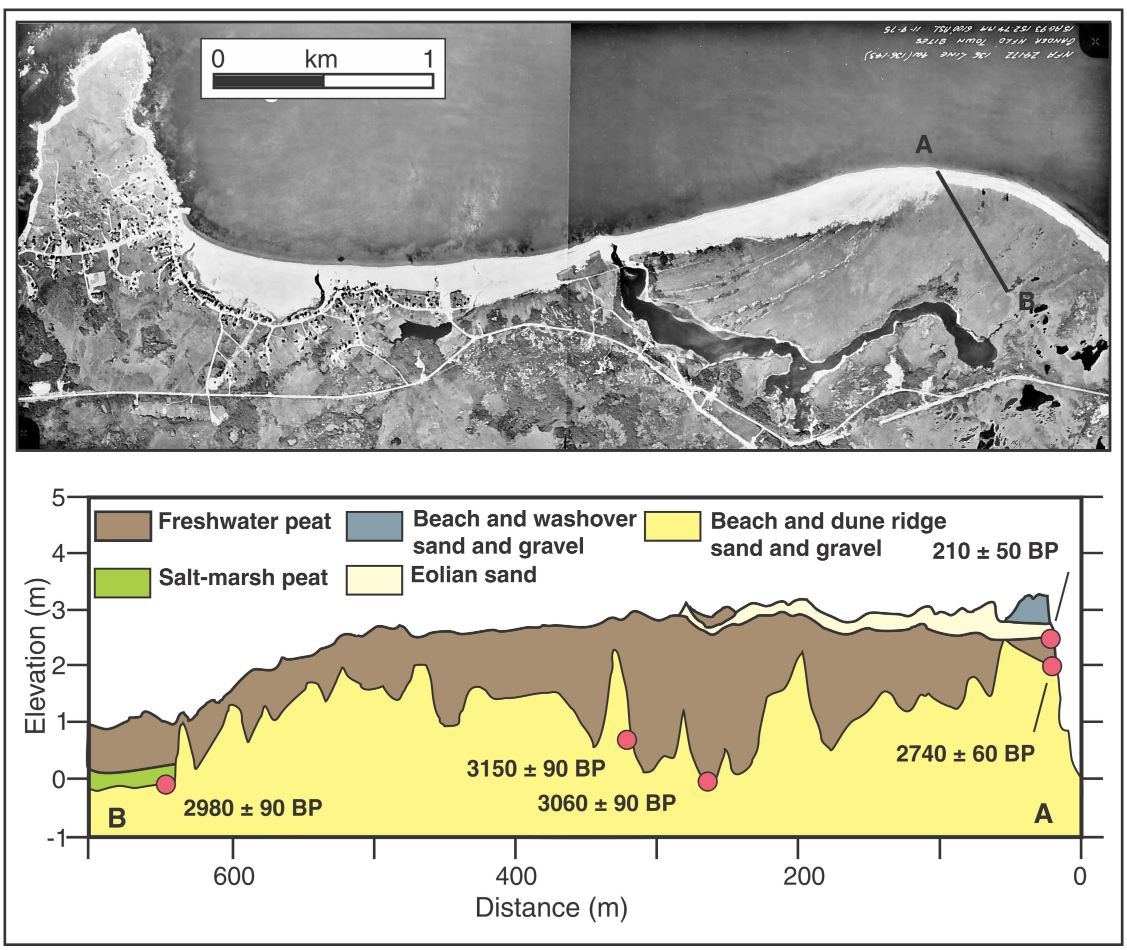


Figure 67. Man Point dune-ridge foreland is overlain by up to 3 m of freshwater peat. Excellent sections of the peat-capped dune ridges are to be seen on the modern rapidly eroding beach.

The inner shelf: overview

The inner shelf is underlain by Devonian granite of the Gander Zone (Fader et al., 1989). In an inner zone shallower than 75 m, relief averages a few metres. Scattered bedrock outcrops are surrounded by gravel and sand deposits up to 9 m thick. Extensive sheets of gravel ripples are overlain in places by sand. Numerous grounded icebergs were observed within this inner zone during the surveys on 5th July 1991, and several very fresh multiple-keel iceberg-grounding pits were observed on the sidescan sonograms. The high mobility of the sand and gravel deposits due to waves and currents precludes the preservation of such records of grounding. Below a depth of 75 m relief is rugged (up to 70 m), and bedrock outcrops over about 20% of the seabed. The surrounding sediments are highly furrowed and pitted by icebergs.

Shaw et al. (1999b) described a conceptual model for the inner shelf of northeast Island of Newfoundland, with five zones: 1) deep offshore muddy basins; 2) basin flanks; 3) an iceberg-impacted zone; 4) a shallow-water (<60 m) wave- and current-dominated zone; and 5) an intertidal-supratidal zone. The variant of this model applicable to this region had a particularly wide zone 4 — *see* Figure 71 *in* Shaw et al. (1999b).

Cape Freels to Fogo Island

The multibeam-sonar imagery shown in Figure 68a illustrates the geomorphology of the inner shelf off Cape Freels. (This is a tiff image supplied by CHS, so appropriate colouration and shading could not be applied.) The joint patterns typical of granitic rocks are very evident, but, as noted above, actual outcrop probably accounts for a small percentage of the seafloor. Large areas of smooth seafloor (sand and gravel) occur at depths of 50–60 m, and fingers of sand and gravel extend inshore, occupying the many depressions that delineate joints. Smooth terrain off Lumsden may indicate sheets of sand that thicken toward the sandy tombolos at the coast, but it is notable that bedrock occurs at the seafloor a short distance offshore from the extensive sandy

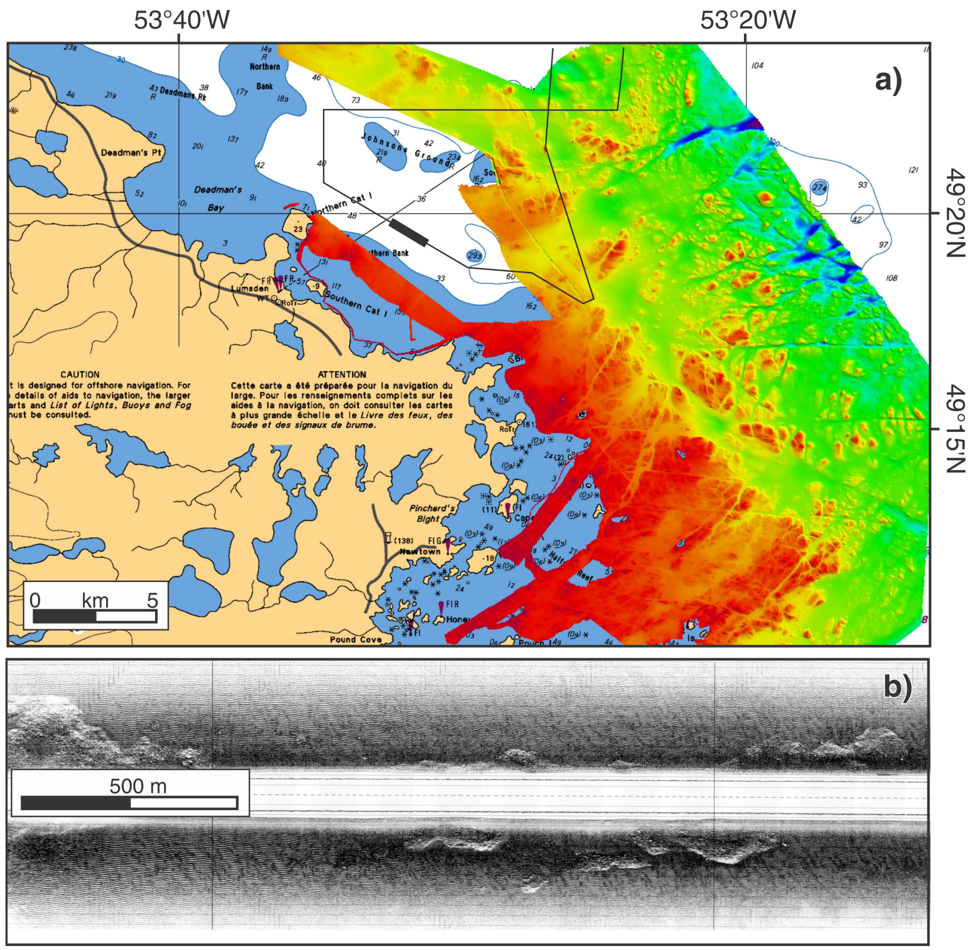


Figure 68. **a)** Multibeam imagery offshore from Cape Freels, also showing the track of survey Dawson 91026 (thin black line). The thick black line shows the location of the **b)** sidescan-sonar record that shows scattered bedrock outcrops and sheets of mobile rippled gravel. Note: Figure 68a is based on a raster image supplied by CHS, and does not have the most appropriate colour bar. Background used is by Canadian Hydrographic Service (2005).

deposits at Cape Freels. Figure 68b is a sidescan-sonar record (cruise CSS *Dawson* 91026) showing terrain off the Lumsden area consisting of rippled gravel and scattered bedrock outcrops.

Pinnacles on the inner shelf

The seafloor northeast of the 'Straight shore' is characterized by pinnacles that typically rise from depths of 150 m to attain depths less than 50 m. Two such pinnacles located east of Fogo Island are illustrated on Figure 69 (another raster image from CHS, for which it was not possible to apply superior colour bars). The easternmost of the two is in an area mapped as Devonian granite, whereas the westernmost lies at the boundary between Devonian granite and mid-upper Ordovician rock. On cruise CSS *Dawson* 91026 (Shaw et al., 1992) the sidescan-sonar towfish narrowly missed grounding on the first pinnacle, and indeed had the ship's course been slightly to the north, it would have probably hit the seafloor at the shallowest part of this feature (Fig. 69b). The sidescan-sonar towfish was lost on the first pinnacle on Figure 69c, which rose to about 47 m below sea level.

Figures 70a and b illustrate the terrain in the pinnacled area. Figure 70a illustrates the terrain in the vicinity of the first pinnacle encountered on the survey (Fig. 69b). The upper part of the record is the subbottom profiler data, whereas the lower part is one part of the sidescan-sonar swath. On either side of the pinnacle the terrain has high backscatter (dark tone), indicative of gravel at the seafloor; however, there are many light-toned patches, indicative of iceberg furrows or pits that have been infilled with mud or sandy mud. Figure 70b shows the terrain in the area where a sidescan-sonar towfish was lost through collision (Fig. 69c). In the deeper water surrounding the pinnacles the seafloor is also predominantly gravelly, with patches of finer sediment and numerous infilled iceberg furrows and pits. The sudden termination of the sidescan-sonar record documents the collision.

Fogo Island area

Figure 69a shows the terrain offshore Fogo Island, where rugged bedrock predominates near the coast. As seen on Figure 65, the shallow inner shelf extends far offshore in this area. Bathymetric data from Olex Ltd. show bedrock terrain extending for about 75 km northeast of Fogo Island, and indeed, most of this shallow area was mapped as 'bedrock predominant' by Shaw et al. (1999b). The inset shows that the bedrock terrain near Fogo Island forms a shallow (–50 m) platform 5–10 km wide, highly dissected near its outer margin. Beyond the sharp, very linear boundary, the relatively smooth seafloor is at depths of about 150 m, with scattered pinnacles. It is probable that lithology in the main bedrock area is granite (Colman-Sadd et al., 1990), and that the pinnacles farther offshore also consist of granite. The relatively deep seafloor offshore from the bedrock zone has been strongly pitted and furrowed by icebergs in depths of 100–300 m. What proportion of this is modern is uncertain, but it is likely that the pitted and furrowed seafloor in the deeper areas comprises glaciomarine sediment, with no postglacial cover.

Notre Dame Channel

Although Figure 69 is based on a raster image supplied by CHS, gridded data were eventually accessed for the deep area to the north of Fogo Island. This coverage straddles the Notre Dame Channel. It is a 20 km wide swath along the channel axis, where maximum depths shallow from 350 m in the southwest to 320 m in the northeast; the channel flanks shallow to 300 m in the northwest, and 250 m in the southeast. Huntec DTS surveys (cruises 75009 and 78023) in this area show acoustically stratified bedrock overlain by thin, iceberg-turbated surficial sediments. In the southwest, along the trough axis, the seafloor is smooth (A in Fig. 71a, b) where acoustically transparent postglacial mud overlies thin, acoustically stratified glaciomarine mud draped over acoustically transparent till. The largest iceberg furrows are 14 m deep from berm to trough, whereas the widest are 300 m across. The iceberg furrows occur in several forms, one notable type being ridged furrows. The large iceberg furrow at 'B' on Figure 71b contains transverse ridges, slightly convex in the assumed direction of motion, 2 m high and 200 m apart. This is similar to the 'washboard' pattern observed in furrows off Antarctica at depths of 600 m to 400 m (Meyer et al., 2000). Another type of iceberg-impact feature in this area of multibeam-sonar coverage is a series of pits several metres deep (C on Fig. 71b), perhaps created by an iceberg that ungrounded repeatedly during its movement.

This is a region of modern impact, so it is interesting to speculate how much of the furrowing is modern. Many of the larger and deeper iceberg furrows in the southwest are buried by postglacial mud, suggesting that they are relict. Davidson and Simms (1997) noted that, on The Grand Banks of Newfoundland, ice-contact features below 200 m were relict, above 110 m were modern, and between 110 m and 200 m were a mixture. It is possible that some of the irregular iceberg furrows in the shallowest areas at the southeast side of Notre Dame Channel are modern, but overall it is likely that the majority of iceberg furrows throughout this area of coverage are relict.

Hamilton Sound, Dog Bay, and Gander Bay

The remaining parts of region 9 are relatively sheltered from wave action and protected from iceberg groundings by Fogo Island. In the shallow embayment of Hamilton Sound (Shaw and Edwardson, 1994) bedrock outcrops are common. Clusters of drumlins record the egress of glacier ice out of the area in a northeasterly direction. Gravelly sand, sand,

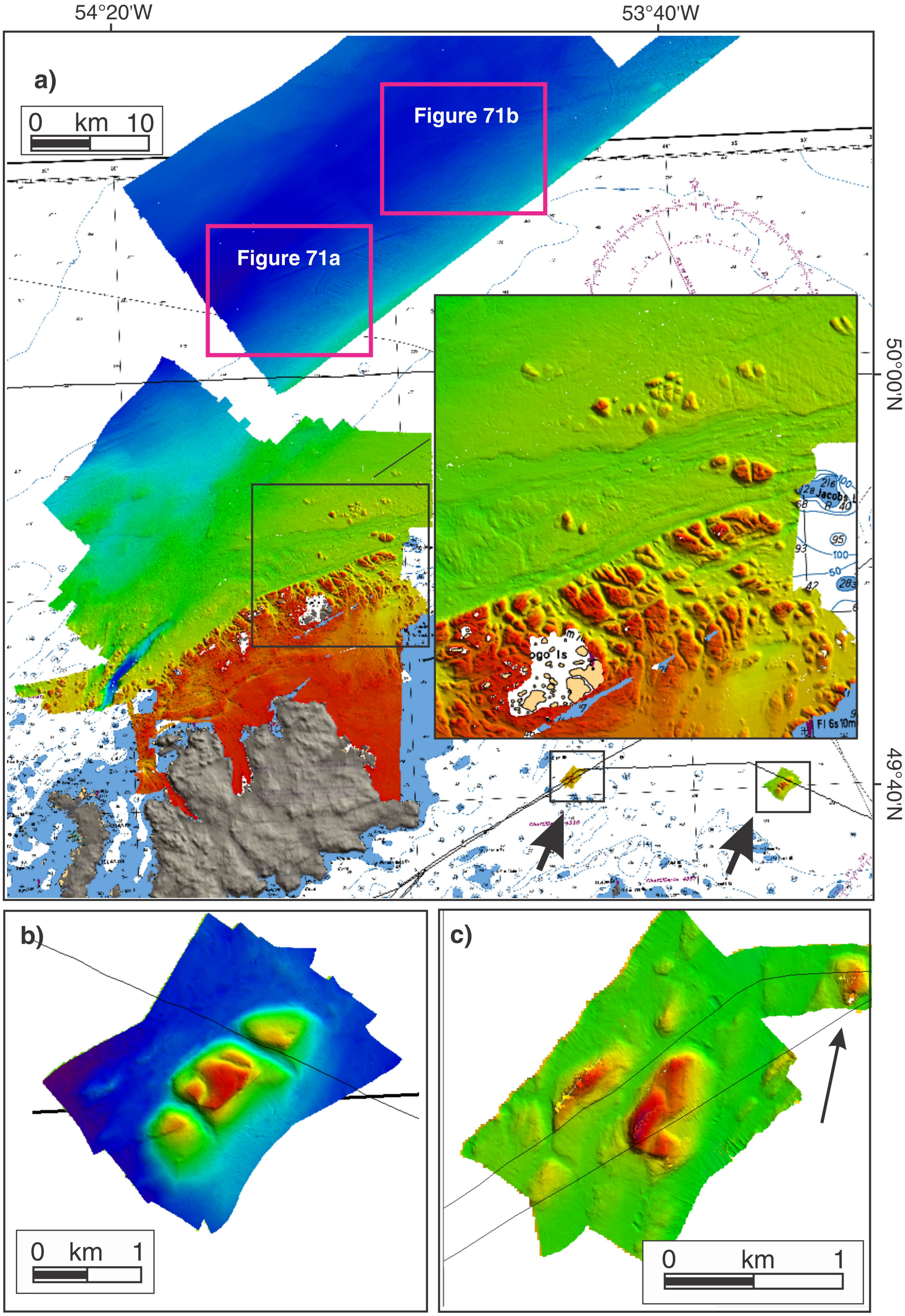


Figure 69. **a)** Multibeam-sonar imagery showing the seafloor around Fogo Island, with an enlarged view of part of the area. The pinnacles east of Fogo Island were encountered during cruise Dawson 91026. The red boxes show the location of Figures 71a and b; **b)** the most easterly group of pinnacles. Note that the ship's track fortuitously avoided the shallowest area. **c)** A second group of pinnacles. The sidescan-sonar data collided with the pinnacle at upper right (arrow) and was lost. Base includes Canadian Hydrographic Service (2005).

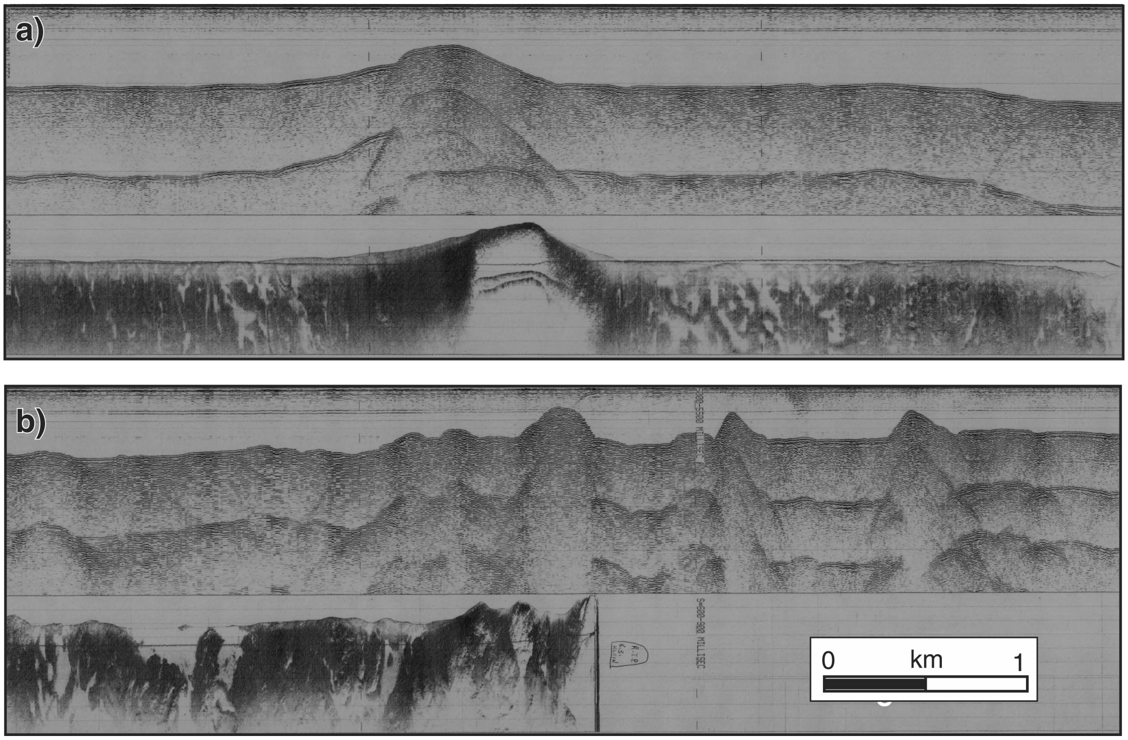


Figure 70. Combined airgun and sidescan-sonar records at the two pinnacles shown on Figure 69b, c. The deeper terrain surrounding the pinnacles has a patchy appearance. Furrows infilled with mud contrast with the surrounding gravelly seafloor. The sidescan record ends at the large pinnacle on Figure 70b.

muddy sand, or sandy mud, are found in basins. Furrows and pits show that modern icebergs make their way into the sound and ground on the seafloor.

The regional relative sea-level curve (Shaw and Forbes, 1990c) is constrained by two types of morphological evidence: rounded drumlin crests at depths below 19 m that would have been truncated if sea level had fallen below −18.5 m, and (wave-cut) terraces at depths of 17 m to 21 m. These data are indicative of a −17 m lowstand of postglacial relative sea level. Radiocarbon dates from a vibracore (Shaw and Forbes, 1990c) suggest that the lowstand occurred prior to 8.6 ka. During the lowstand, Fogo Island was connected to the mainland by a narrow isthmus.

Dog Bay and Gander Bay are shallow, funnel-shaped bays exposed to high levels of wave activity in their outer parts. Gander Bay deepens northward gradually to 33 m at its mouth. Drumlins at the mouth of Gander Bay are contiguous with those in Hamilton Sound. They have a veneer of muddy subangular to angular gravel, in contrast to the surrounding seabed which is composed of silty mud with scattered gravel. The mud is up to 8 m thick, and contains several zones of shallow gas. In the shallower water toward the head of the bay, the seabed appears more reflective on sidescan sonograms and displays point-source reflections. This indicates the presence of a hard gravelly bottom in places. Seabed photographs show angular to subangular gravel overlying a muddy substrate.

REGION 10: NOTRE DAME BAY

Setting

This region (Fig. 72) encompasses the Gander and Dunnage tectonic zones (Colman-Sadd et al., 1990), and includes a wide range of bedrock lithologies from Late Proterozoic to Silurian, with major structural elements that run in a northeasterly direction, and continue offshore in Notre Dame Bay (Fader et al., 1989). The coast has high relief and is indented by deep fiords, offshore of which the inner shelf in Notre Dame Bay has gentle topography. Outer coasts are exposed to Atlantic Ocean wave processes, but inner coasts are sheltered. In exposed areas of the innermost shelf, icebergs drifting south in the Labrador Current impact the seafloor, imprinting it with scours and grounding pits.

Together with region 9, this region was the focus of a search for marine placer gold in the early 1990s. The main outcome was GSC Bulletin 532 (Shaw et al., 1999b), which contains a 1:250 000 scale surficial geology map that includes a shoreline classification, and a series of maps and

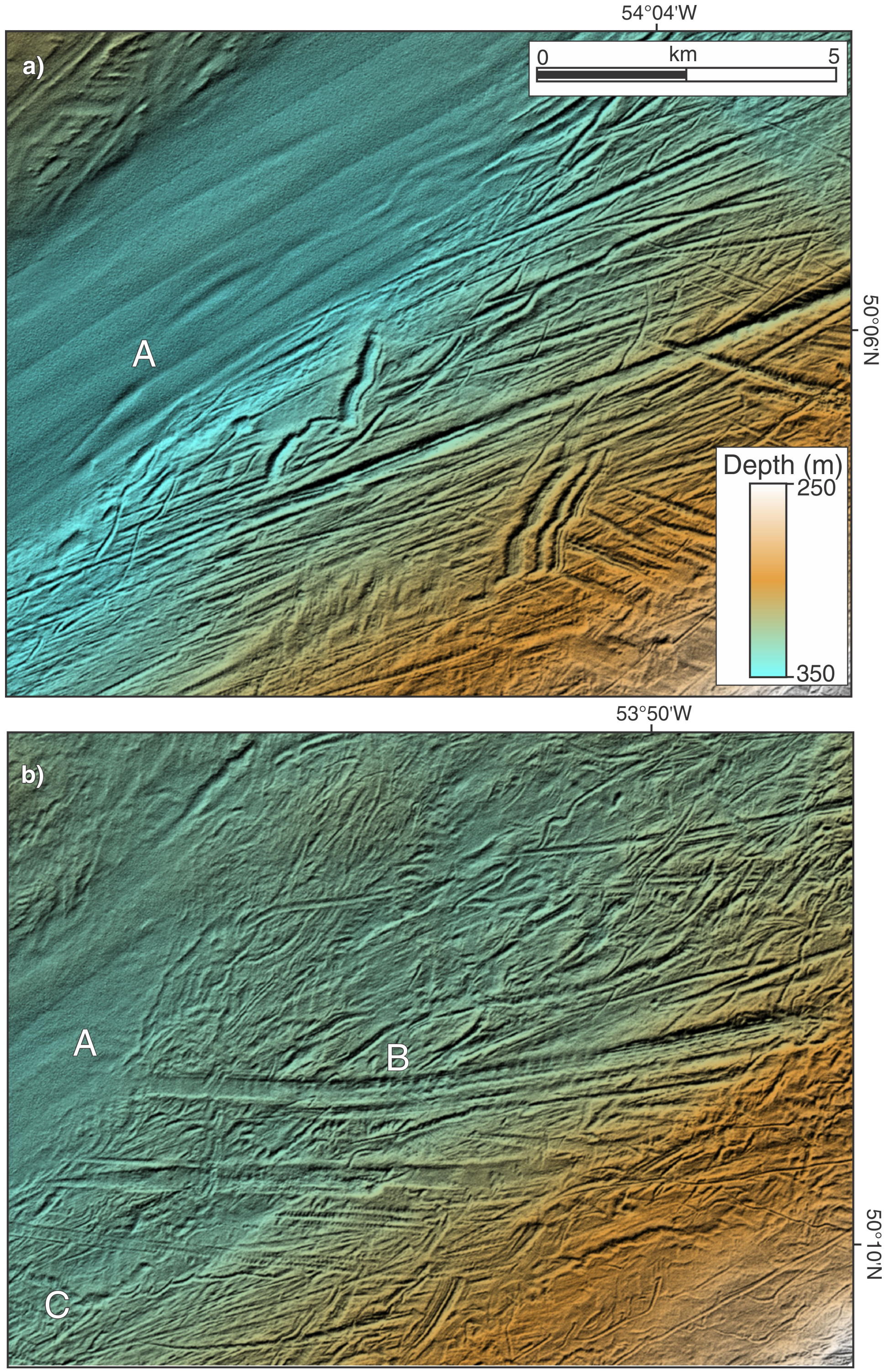


Figure 71. Terrain in Notre Dame Channel, locations shown on Figure 69a. **a)** The smooth area (A) has a cover of postglacial mud that mantles furrows. In this area the predominance of furrows indicate motion to the northeast. **b)** A small area of postglacial mud occurs at (A); otherwise the seafloor is heavily furrowed. The large furrow at (B) has a 'washboard' pattern of transverse ridges in the trough, 2 m high and 200 m apart. The furrow at (C) consists of a line of pits, suggestive of intermittent ungrounding during transit.

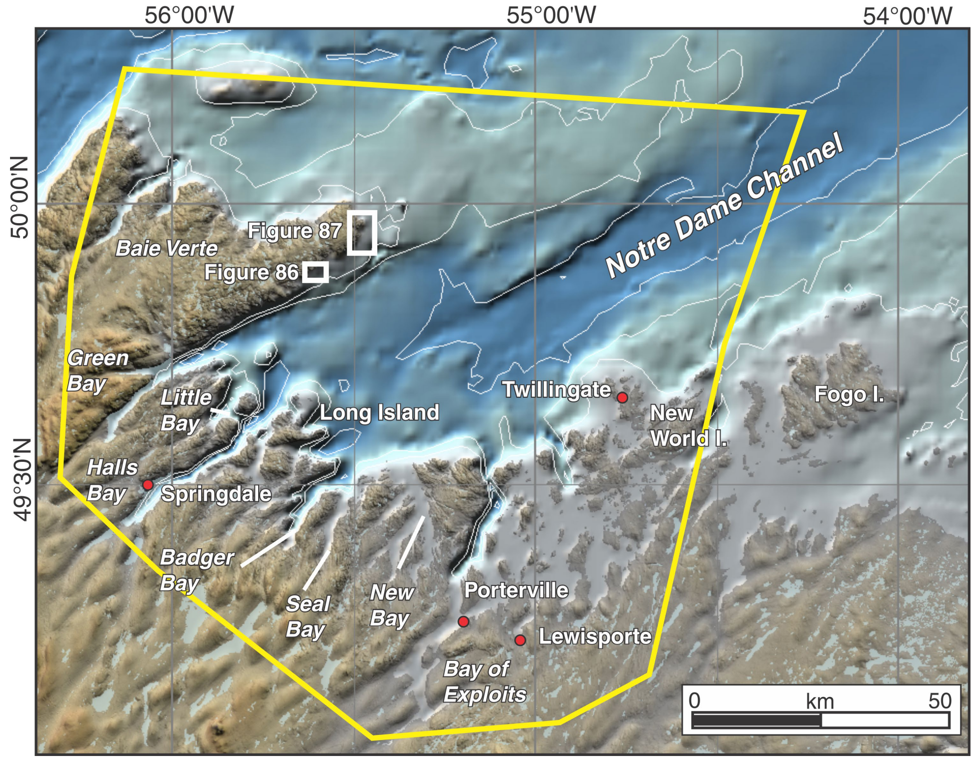


Figure 72. Region 10.

reports on individual embayments. A series of Open File reports contain information on the surveys that were the basis for the bulletin (Shaw et al., 1990c; Shaw and Wile, 1990; Edwardson et al., 1992, 1993; Jenner and Shaw, 1992). Large amounts of multibeam-sonar data collected subsequent to the 1990s mapping provide additional insights into the surficial geology of the region as understood from the 1990s mapping. Figure 73 shows the extent of coverage.

Coastlines

Coastlines in this region are bedrock dominated, with very limited development of depositional landforms, reflecting a restricted sediment supply. Erosional exposures of unlithified sediments in backshore cliffs are extremely rare and are restricted primarily to inner fiord locations. Wave exposure ranges from protected inner-bay settings to large sections of coast fully exposed to the ocean. Shaw et al. (1999b) distinguished five broad classes of coasts: rock-dominated, low- to high-relief exposed outer coasts with cliffs, wave-washed rock ramps and limited beach development; semiexposed, rock-dominated, island and outer-fiord coast with rock platforms, rock cliffs, and limited beach development; semiexposed to protected, rock-dominated, fiord coast with limited development of mixed sand and gravel beaches; partly exposed to protected, rock-dominated, low-relief coast with extensive rock and boulder platforms, boulder flats, and locally extensive mixed sand and gravel beaches; and protected, rock-dominated, low-relief, shallow embayments with extensive boulder flats locally.

Inner shelf: overview

The region is divisible into two zones. The innermost zone is underlain by basement rocks and dissected by northeast-trending fiords (Fig. 72), as follows: Baie Verte (maximum depth 324 m, sill depth 192 m); Green Bay (maximum depth 442 m, sill depth ~280 m); Halls Bay (maximum depth 472 m, sill depth 290 m); a basin east of Long Island (maximum depth 554 m, sill depth 290 m); Seal Bay–Badger Bay (maximum depth 440 m, sill depth 290 m); and New Bay (maximum depth 456 m, sill depth ~200 m). In Bay of Exploits the western trough has a maximum depth of 673 m, and a sill depth about 210 m, and the eastern trough reaches 604 m with a sill at 250–300 m. The fiord basins have steep sides and flat floors, but, ignoring the Quaternary sediment fill of glaciomarine and postglacial sediments, they actually have V-shaped crossprofiles. The adjacent outer zone on the continental shelf has low relief, and sedimentary rocks are mantled by Quaternary sediments: glacial diamict and thin glaciomarine mud in shallow areas, glacial diamict and thick deposits of glaciomarine and postglacial mud in basins.

Shaw et al. (1999b) outlined two conceptual models for this region. The fiord model had three zones: an outer zone, with deep basins separated from the adjacent shelf by shallow sills; a middle zone with steep, rocky sidewalls and thick glaciomarine and postglacial sediment; and an inner zone with thick glaciomarine sediments, thin postglacial deposits, and raised marine terraces at the coast. Outside the fiords the inner shelf model had five zones: basins in Notre Dame Bay with thick postglacial mud overlying glaciogenic sediments; basin flanks with internal acoustic reflections truncated at the

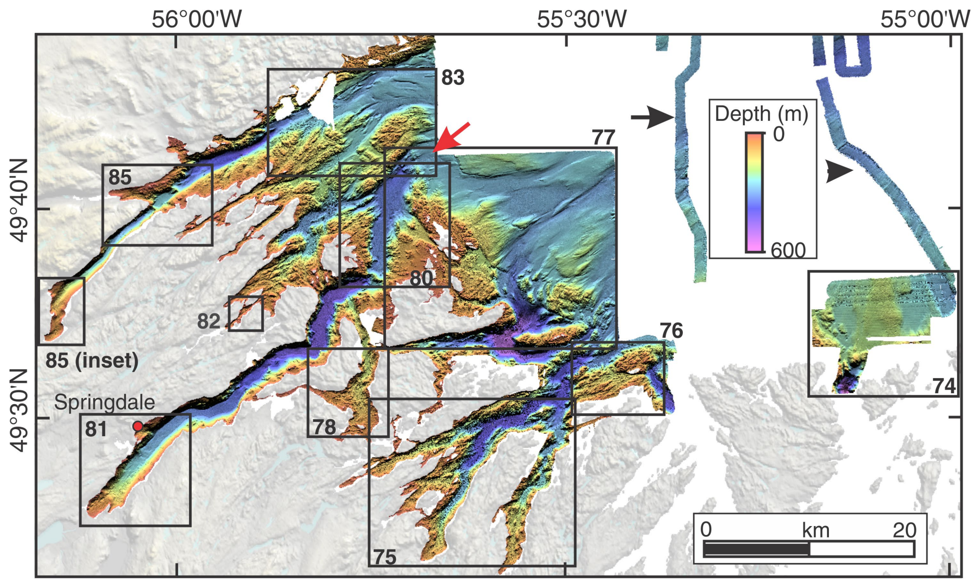


Figure 73. Extent of multibeam imagery, western and southern Notre Dame Bay, also showing location of figures. Black arrows indicate the two multibeam-sonar survey lines that traverse Notre Dame Channel.

seabed; an iceberg-impacted zone in depths under 200 m; a shallow-water (<60 m) wave- and current-dominated zone; and an intertidal-supratidal zone.

New World Island area

The New World Island area (Fig. 72) comprises a series of islands and exposed embayments bordering the eastern margin of Notre Dame Bay and opening northward to the Atlantic Ocean. Figure 59 of Shaw et al. (1999b) shows surficial sediment distribution off the outer coasts of New World Island. As befitting an exposed area much of the seafloor near the coast is basement bedrock. The most common seafloor type is mobile rippled gravel. Iceberg scours occur throughout the area.

Bay of Exploits

Multibeam-sonar data are not available for Bay of Exploits (Fig. 72), the largest and deepest fiord in the region, so the description paraphrases Shaw et al. (1999b). The bay has two principal channels. Western Channel contains ponded Quaternary sediments in a narrow (1 km) swath bounded by fiord sidewalls with slopes up to 70° (*see* Fig. 10 *in* Shaw et al., 1999b). Total sediment thickness decreases southward from 200 m just inside the sill. Several transverse moraines record halts during southward retreat of Late Wisconsinan ice. The moraine at Porterville passes laterally into acoustically stratified sediments and is postdated by radiocarbon determinations of 11 480 ± 100 ^{14}C years BP and 12 020 ± 100 ^{14}C years BP. Several basins south of Porterville contain thin postglacial sediments overlying stratified glaciomarine sediments dated at 11 440 ± 100 ^{14}C years BP and 12 020 ± 100 ^{14}C years BP (Shaw et al., 1999b). Just offshore from Botwood, up to 20 m of glaciomarine mud is draped over irregular bedrock terrain; steep slopes on the irregular bottom have creep terraces. Above 45 m depth the seabed has been impacted by iceberg pits. At the mouth of the Exploits River an extensive, gently sloping terrace with a lip at 6.1 m below mean water level is a submerged early Holocene delta (Shaw and Forbes, 1995).

Entrance to Bay of Exploits

Multibeam-sonar imagery (Fig. 74) of an area just outside the western entrance to Bay of Exploits reveals a complexity not evident on the surficial geology map (Fig. 1 *in* Shaw et al., 1999b), and also illustrates the contrast between the inner and outer zones of the inner shelf model (above). North of the bedrock fiord sill (A) the seafloor is relatively shallow. Streamlined glacial landforms (B) located in the lee of bedrock knolls indicate ice-flow directions toward the north. North of here, several survey lines (*see* lines indicated by black arrows on Fig. 73) that traverse Notre Dame Channel show streamlined landforms oriented in the east-northeast direction, consistent with convergent ice flow down Notre Dame Channel. Survey data from cruise CSS *Dawson* 91026 shows that the broad, iceberg-furrowed ridge (C) is underlain by till. This ridge extends approximately 10 km from the coast and is interpreted as an accumulation of glacial diamict in the lee of the Exploits Islands, in an area of weaker ice flow between stronger flows of ice out of the deep eastern and western channels of Bay of Exploits.

New Bay, Seal Bay, and Badger Bay

In New Bay (*see* Fig. 57 *in* Shaw et al., 1999b), steep bedrock sidewalls with a discontinuous veneer of mud or gravelly mud flank a central basin containing ponded postglacial mud overlying basin-fill stratified glaciomarine

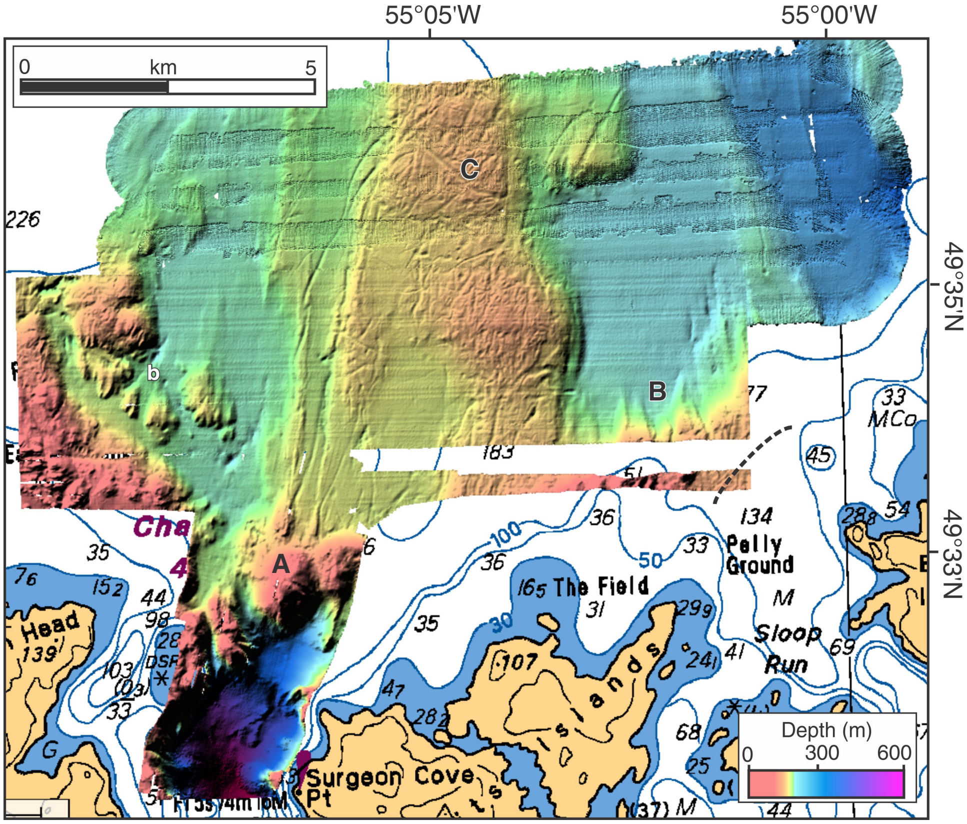


Figure 74. The seafloor at the mouth of Bay of Exploits, illustrating the transition from basement rock inner zone and the sediment-draped outer zone in Notre Dame Bay. Bedrock knolls occur at the entrance to the deep trough (A), and crag-and-tail landforms (B) record movement of glacial ice toward the north. The iceberg-furrowed terrain on the ridge (C) contrasts with inner-fiord settings, in which grounding pits predominate. Location on Figure 73. Background is Canadian Hydrographic Service (2004b).

sediments. Postglacial mud is banked to one side of the basin near the head of the bay, so that glaciomarine sediments are exposed on the seabed.

Seal Bay (Fig. 75) was not mapped by Shaw et al. (1999b). The new multibeam imagery shows a series of basins containing postglacial mud with low backscatter strength overlying glaciomarine mud, separated by bedrock sills (high backscatter strength). The structural grain of the bedrock appears to be northeasterly. At the heads of the various arms of the fiord, in depths shallower than 140 m, the seafloor is heavily imprinted by iceberg-grounding pits (*see* greyscale insert on Fig. 75). The poor resolution of the data from the survey in Seal Bay means that it is difficult to determine the 'freshness' of the iceberg-grounding pits, hence the extent to which these are modern verses relict features is unknown

The new data from Badger Bay (Fig. 75) are in reasonable agreement with the report of Shaw et al. (1999b) who described stratified glaciomarine sediments up to 110 m thick, interbedded with acoustically transparent debris flows. Postglacial mud with low backscatter, typically about 20 m thick, overlies glaciomarine sediments in the basins, and also occurs as a 2–5 m thick veneer over the irregular bedrock slopes on the basin flanks. Toward the heads of the arms, the postglacial mud tends to occur in banks, so that the underlying glaciomarine sediments are exposed at the seabed. Shaw et al. (1999b) reported that north of Gull Island, in depths of 100–130 m, the seabed has a relief of several metres and the sediments (<10 m thick) are acoustically opaque as a result of iceberg-grounding and turbate formation. As in Seal Bay, in water depths shallower than approximately 140 m the multibeam imagery shows a seafloor highly imprinted by iceberg-grounding pits, but the poor quality of the data means that the extent of recent grounding activity verses ancient is unknown.

North of Seal Bay

The area just north of Seal Bay (Fig. 76) is relatively shallow (<150 m), with many areas shallower than 100 m, and thus the seabed is exposed to sediment mobilization by waves. This area encompasses a large area of basement bedrock (A) that projects north from the coast. Flat-lying areas between the fault- and joint-bounded ridges at depths of 100–160 m are occupied by Quaternary sediment with low backscatter strength, i.e. sand. At either side of the bedrock plateau, deep troughs constitute the seaward extensions of New Bay (B) and Seal Bay–Badger Bay (C). The troughs contain glaciomarine sediment capped by postglacial mud. The New Bay trough terminates in a bedrock sill (D), perhaps with a capping of glacial diamict. The combined trough of Seal Bay–Badger Bay (C) shallows from a basin at 420 m to a perched basin at 330 m (E).

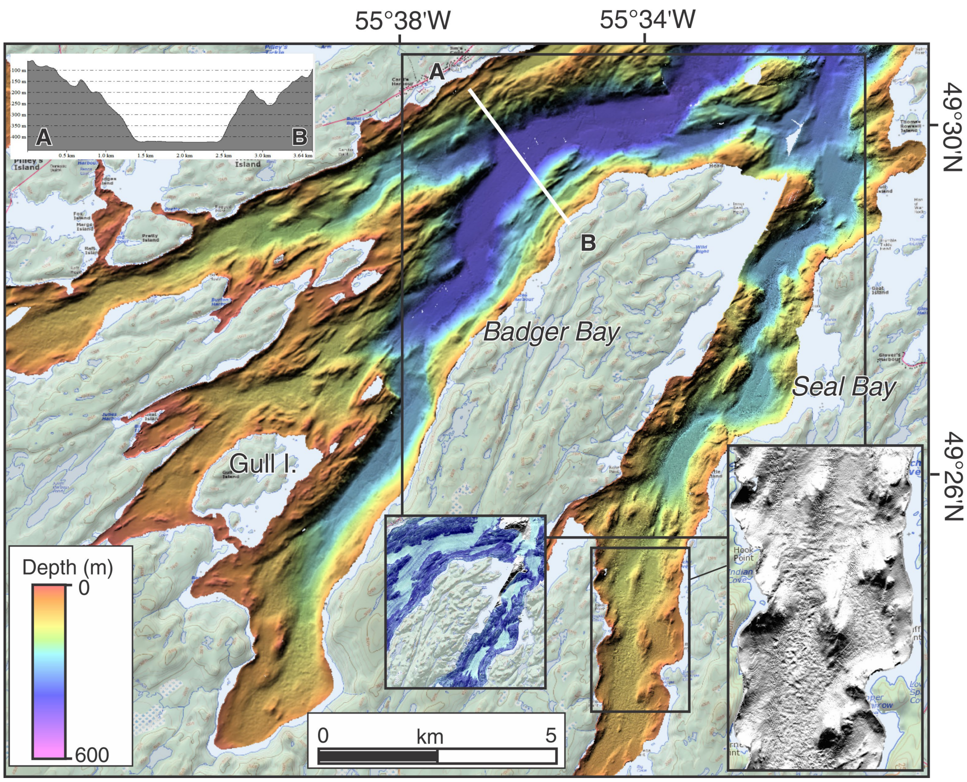


Figure 75. Shaded-relief image of Seal Bay and Badger Bay. The insert show the iceberg-pitted terrain at the head of Seal Bay in more detail. Location on Figure 73.

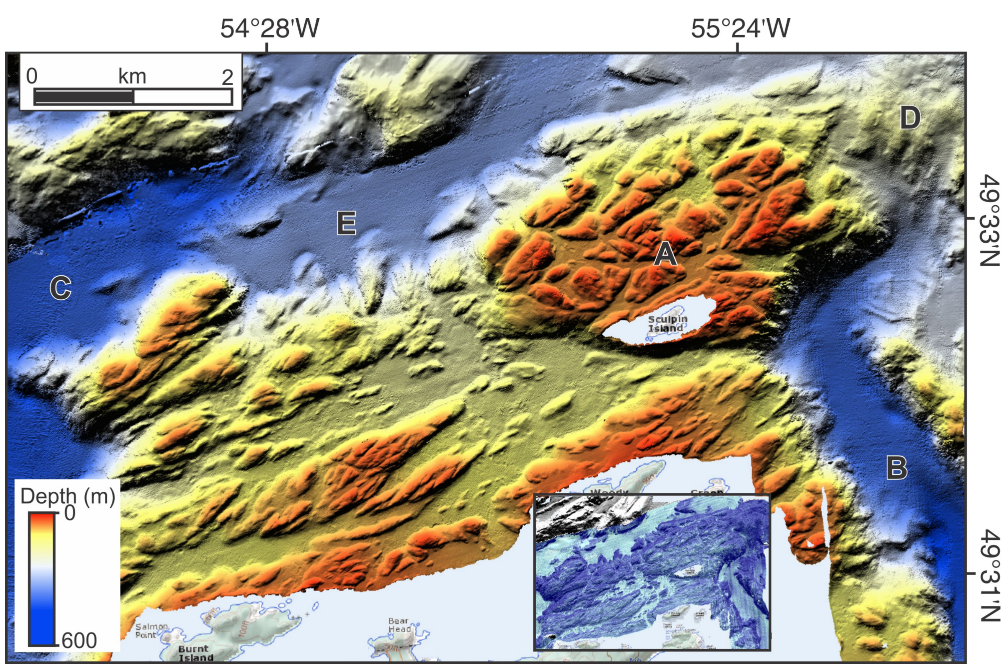


Figure 76. Seafloor north of Seal Bay showing bedrock terrain (A), and deep troughs constituting the seaward extensions of New Bay (B) and Seal Bay–Badger Bay (C). The troughs contain glaciomarine sediment capped by postglacial mud. The New Bay trough terminates in a bedrock sill (D), perhaps with a capping of till. The combined trough of Seal Bay–Badger Bay (C) shallows from a basin at 420 m to a perched basin at 330 m (E). The insert shows backscatter strength of the area, with high backscatter portrayed as a dark tone. Location on Figure 73.

Inner shelf near Long Island

This region (Fig. 77) shows the transition between the inner rugged, basement bedrock-dominated inner zone, and the low-relief floor of Notre Dame Bay, where sedimentary rocks are mantled by Quaternary sediments: discontinuous glacial diamict, a drape of glaciomarine mud, and thick postglacial mud in basins. Long Island Tickle continues east to a deep of 545 m (A). In shallower water bedrock predominates, and several major geological boundaries are evident, for example along dashed line B-C. East of this is a sharply delineated block (D) probably consisting of bedrock mantled by Quaternary sediments that are heavily marked by iceberg-grounding pits. It is uncertain whether this bedrock is Upper Carboniferous, but areas to the northeast of this are mapped such by Fader et al. (1989). The large block (E) gives the impression of gently dipping strata and indeed, an airgun seismic-reflection line (cruise 91026, day 182) shows apparent dips to the northeast. Most of this region was mapped as Lower Carboniferous by Fader et al. (1989).

An interesting aspect of the area shown on Figure 77 is the presence of streamlined glacial landforms. Length is variable from 0.2 km to 4.4 km, with a mean of 1.2 km. Width is extremely variable. The landforms nearest the coast are short and strongly tapered, and in many cases appear to have bedrock outcrops at their upstream ends, so that they resemble crag-and-tail features. Landforms farther offshore are more elongate, with an elongation ratio of less than 7:1 (Stokes and Clark, 1999, 2001). The streamlined landforms extend for a distance of 100 km down the axis of Notre Dame Channel. In the upstream end, close to the coast, they exhibit a strongly convergent pattern, indicative of accelerating ice flow to the northeast, into Notre Dame Channel. Several survey lines that traverse the channel show elongated streamlined landforms.

When compared with the extent of Figure 54 in Shaw et al. (1999b), this image shows the inability of surveys conducted before the multibeam era to portray the true complexity of the seafloor. Most of the seafloor away from bedrock highs was described as iceberg turbate, and photographs (Fig. 55 *in* Shaw et al., 1999b) show a predominance of gravelly substrates.

Sunday Cove Island area

The area shown on Figure 78 is the innermost part of a long fiord trough that connects with Halls Bay. In the Sunday Cove Island area, basins contain stratified glaciomarine sediment up to 40 m thick overlain by postglacial mud up to 12 m thick. In places the postglacial mud is banked against one flank of the basin, so that glaciomarine sediments are exposed on the basin floor (*see* Fig. 52 *in* Shaw et al., 1999b). Bedrock outcrops occur in places — one can be seen on the sidescan-sonar record (inset). Shaw et al. (1999b) noted that large areas are imprinted by iceberg-grounding depressions, often with a recent muddy infill, as shown in the inset on Figure 78. These depressions can be observed on the multibeam image. It was claimed that the depressions were mainly

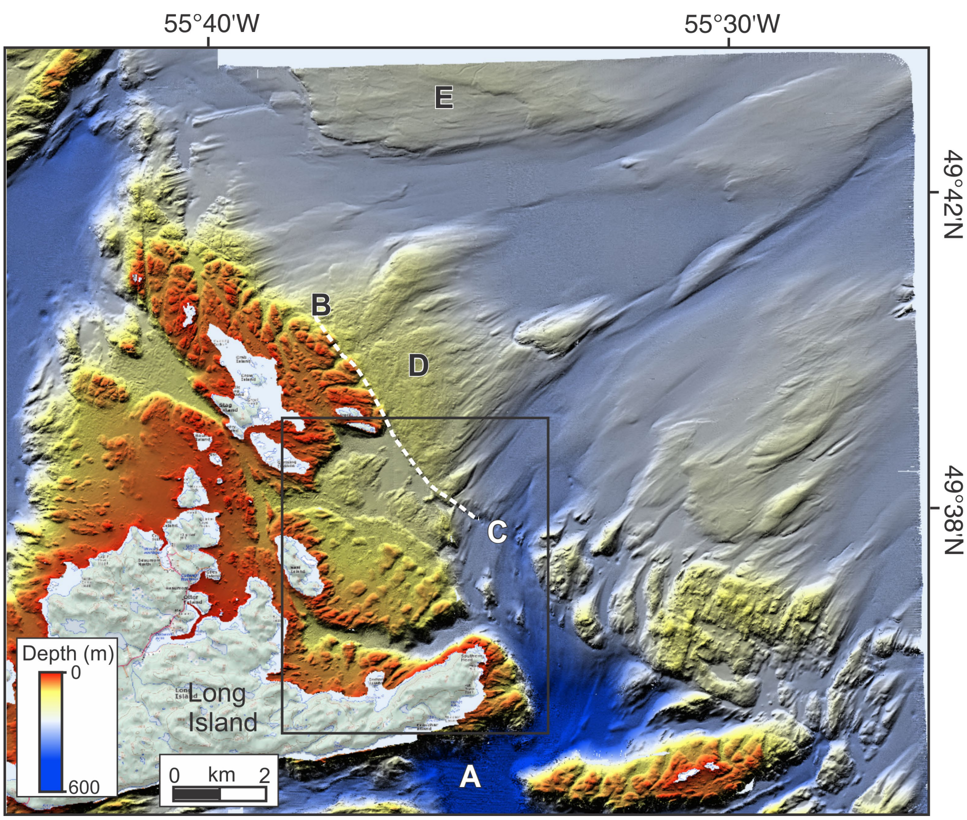


Figure 77. Shaded-relief image of the shelf near Long Island. Long Island Tickle continues east to a deep of 545 m (A). In shallower water, bedrock predominates, and several major geological boundaries are evident, for example along dashed line B-C. The sharply delineated block (D) probably consists of bedrock mantled by Quaternary sediments. On large block (E) airgun seismic-reflection data show apparent dip to the northeast. Location shown on Figure 73. Box shows location of Figure 54 of Shaw et al. (1999b).

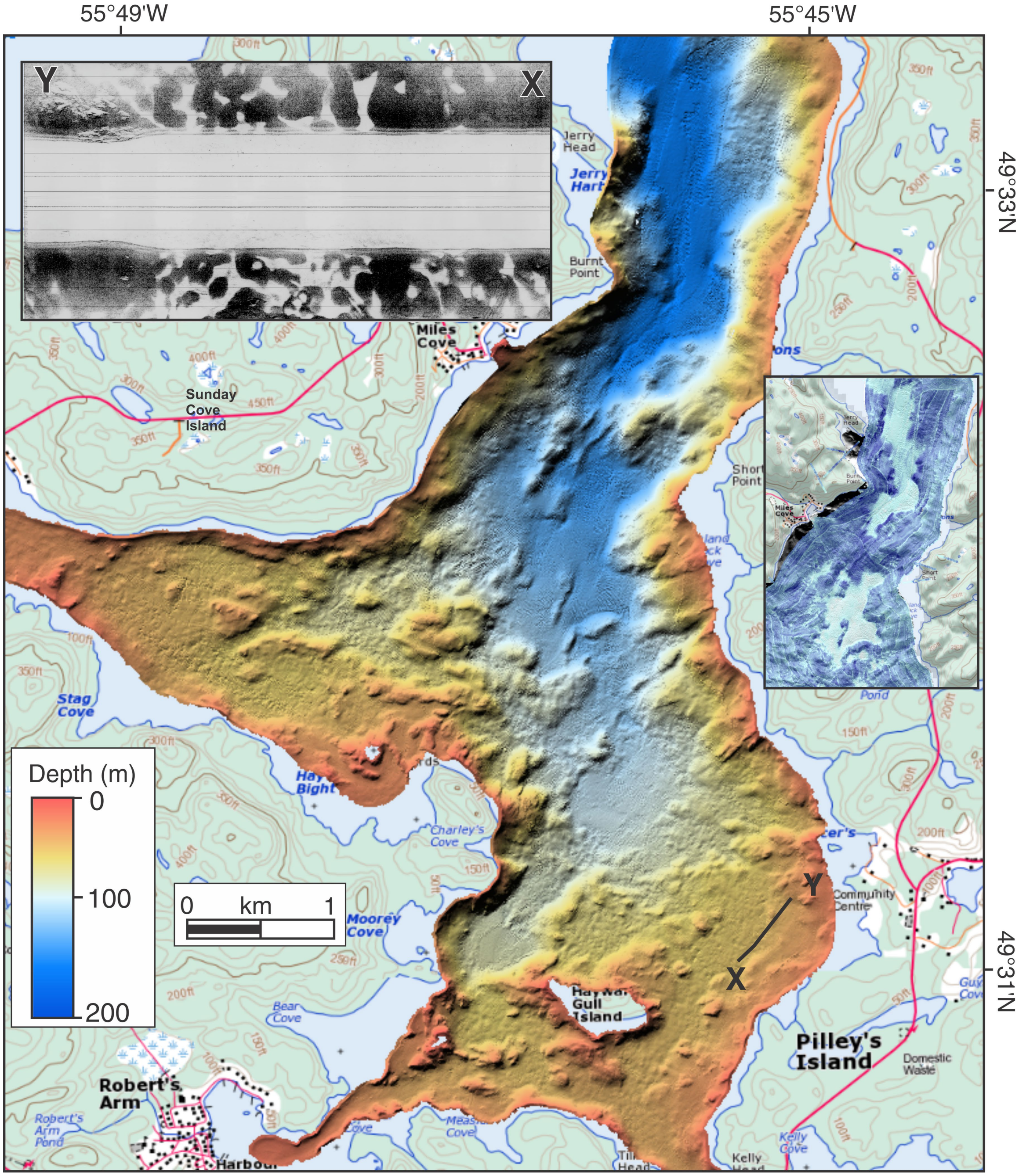


Figure 78. Shaded-relief image of Sunday Cove Island area. The inset at upper left shows a section of sidescan-sonar record, showing iceberg-impacted seabed (cruise 90-035, daytime 227/14:55–15:00), location of record shown by black line X-Y. This multibeam image can be compared with the map in Shaw et al. (1999b). Inset at right shows backscatter strength for part of the area. Lighter tone indicates low backscatter. Location shown on Figure 73.

formed by grounding of the icebergs that drift into the region in summer, but whereas some pits have a 'fresh' appearance, the degree of modern impact is unknown.

Halls Bay

The bay is divisible into several morphological elements (Fig. 79), viz: a bedrock sill; an outer basin; a large submarine moraine; the deep middle basin; and the shallow inner basin. The bedrock sill rises about 90 m above the terrain of Notre Dame Bay, to the northeast, where a series of streamlined glacial landforms (crag-and-tail type) trend east-northeast.

The basin to the south of the bedrock sill (Fig. 80) averages 350 m deep, and contains up to 25 m of postglacial mud overlying acoustically stratified glaciomarine sediments. Total sediment thickness is a maximum of 150 m. Unlike many fiords, the sidewalls do not extend above sea level, but rather the bedrock trench of the fiord is incised in a rocky platform 50–100 m below sea level (Fig. 80b). Parts of the platform have an infill of Quaternary sediment, perhaps glaciomarine mud. The surface of these flat-lying areas, at depths above 150 m, is strongly imprinted by iceberg-grounding pits. Another unusual aspect is the presence of large circular pits (Fig. 80c), up to 350 m across and 20 m deep. These are interpreted as gas-escape structures, i.e. pockmarks. Not only are they unusual in Notre Dame Bay, but pockmarks of this size are rare, and the best comparison may be to those in inner St. George's Bay (region 1).

A transverse moraine (Fig. 79, 80) separates the outer basin (depth 325 m) from the inner basin (475 m). The architecture of this moraine is unknown as the survey line (cruise CCGS *Hudson* 90013, day 170) was planned without the benefit of multibeam sonar, and crossed on a bedrock ridge to the west of the moraine, hence the absence of a moraine on the seismic profile (Fig. 79). The crossfiord moraine marks a halt in the southward recession of the former ice margin. It coincides at a constriction where the submarine platform projects across the bay, the likeliest place for a halt (Syvitski, 1991). South of the moraine, the basin averages 450 m deep and contains up to 180 m of Quaternary sediment. The low-backscatter postglacial mud unit is thinner than north of the moraine. The underlying acoustically stratified glaciomarine unit contains two transparent intervals, 15 m and 20 m thick, interpreted as debris flows.

The inner bay (Fig. 81) shallows steadily toward its head, and hosts thinner Quaternary sediments than in the fiord to seaward. The postglacial mud unit (A) thins to the southwest, and is only a few metres thick south of Springdale where core 90013-063 (location shown) contains shell at 4 m dated

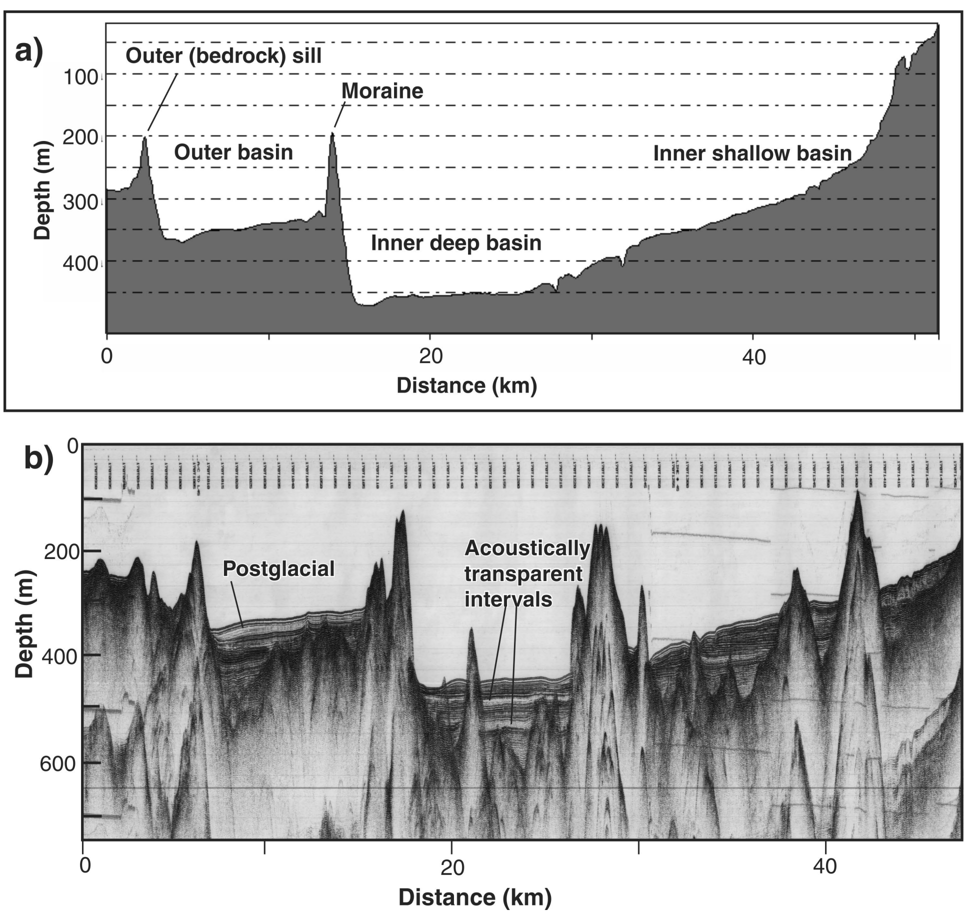


Figure 79. a) Bathymetric profile along the thalweg of Halls Bay, from beyond the sill to the head. Start of profile indicated by a red arrow on Figure 73. **b)** Airgun seismic-reflection profile down Halls Bay, from the bedrock sill (left) to the head of the bay (right), (cruise CCGS *Hudson* 90013). This profile is not exactly along the thalweg, hence the moraine was not crossed during transit, and is not shown here. The various bedrock ridges are places where the survey line departed from the thalweg and approached the sidewalls.

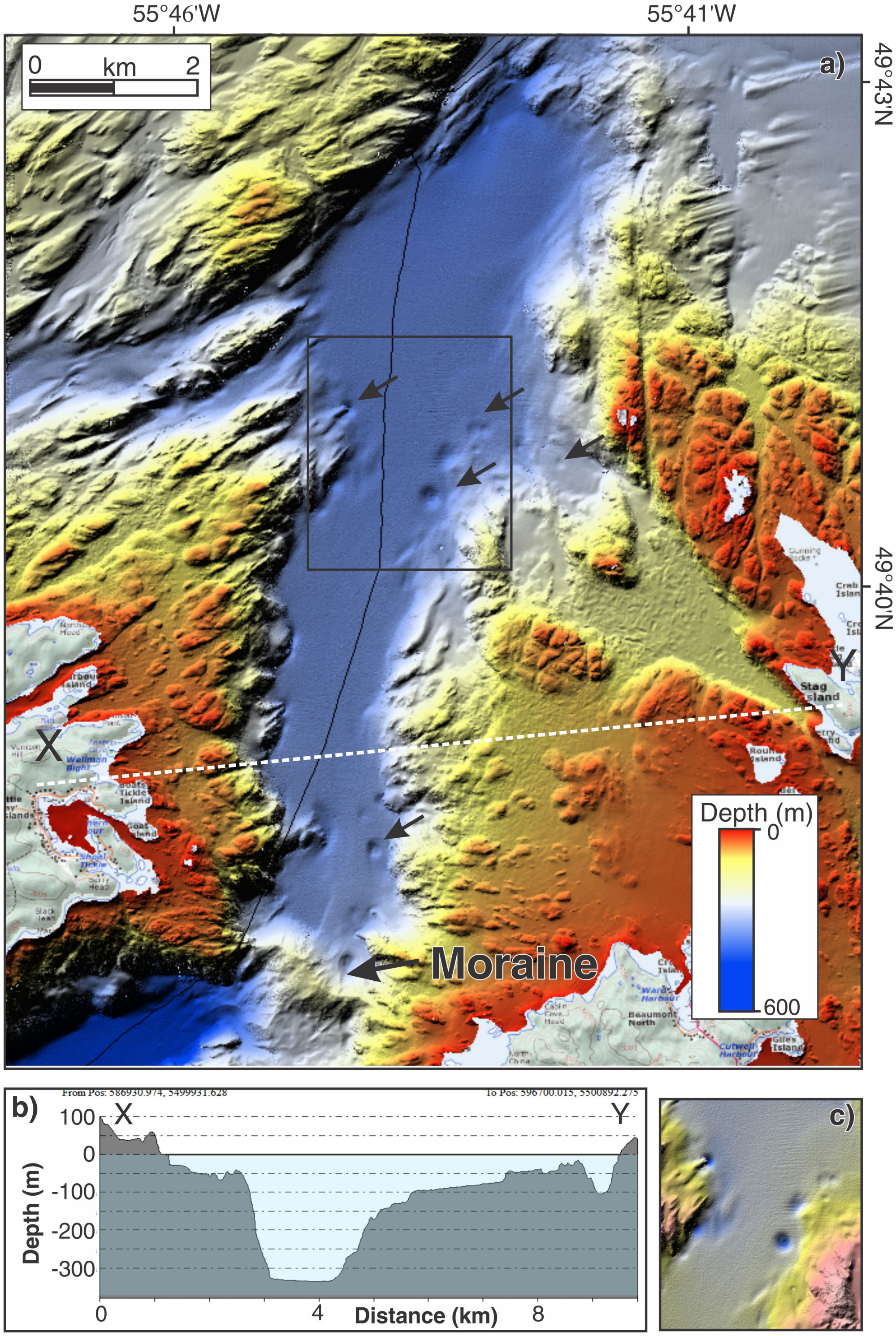


Figure 80. **a)** Shaded-relief image of the outer section of Halls Bay. The arrows indicate large pockmarks. **b)** Cross-section X-Y (Fig. 80b) shows how the fiord trough is incised into a submerged platform. An extension of the sidewalls in the cross-section gives a realistic idea of the fiord profile — V-shaped rather that U-shaped. This is also shown in profile farther in the bay on Figure 49 of Shaw et al. (1999b). **c)** Pockmarks, locations shown by arrows on Figure 80a. Location shown on Figure 73.

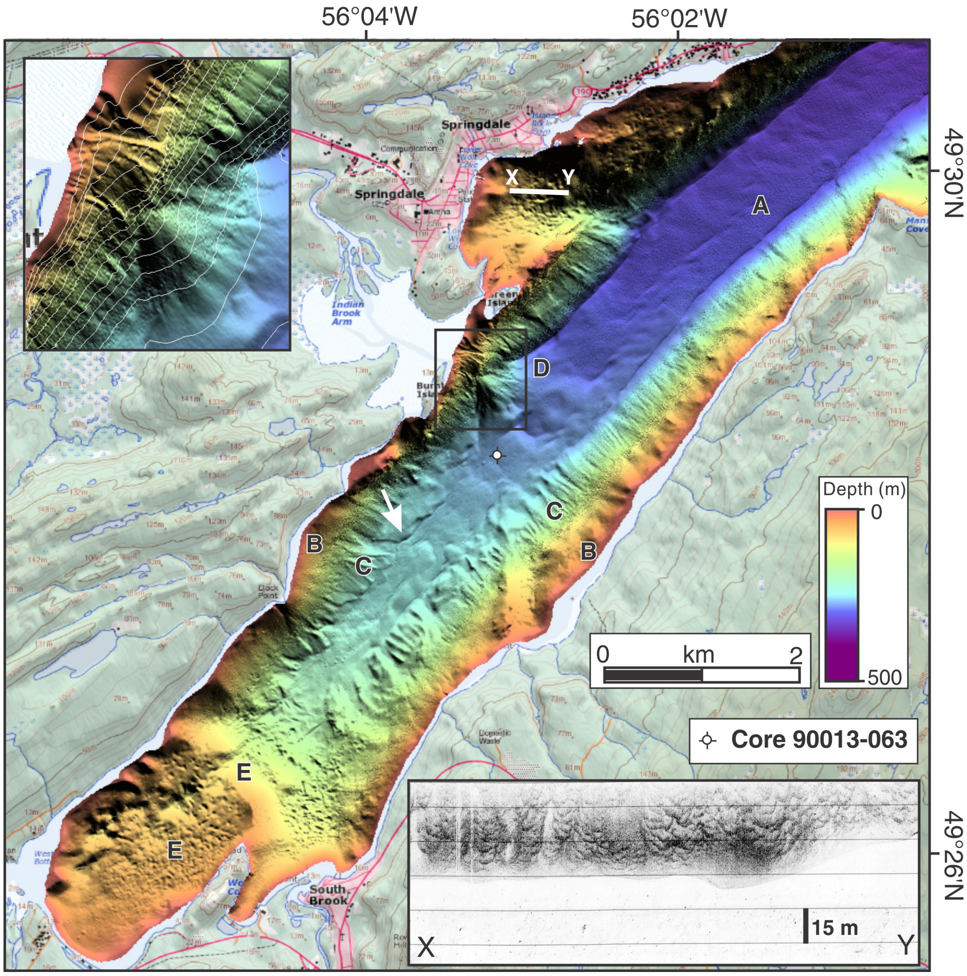


Figure 81. Shaded-relief image of inner Halls Bay. The inset (sidescan-sonar record from cruise CSS *Navicula* 90035) show a botryoidal pattern on the fiord sidewalls; location is indicated by white line X-Y off Springdale. The postglacial mud unit (A) thins toward the head of the fiord. From sea level down to −120 m the fiord sidewalls are smooth (B). Below −120 m the sidewalls are marked by a series of parallel rills (C) with relief of 2–5 m. Mass transport is indicated by a large submarine fan (D). In depths shallower than 150 m the seafloor is impacted by iceberg-grounding pits (E). Location shown on Figure 73.

11 330 ± 130 ^{14}C years BP (Shaw et al., 1999b). Inner Halls Bay contains thin postglacial mud overlying glaciomarine sediments up to 170 m thick containing acoustically transparent intervals up to about 20 m thick that commonly have upper surfaces with fine-scale (<0.5 m) roughness and are interpreted as debris-flow deposits that accumulated in close proximity to glacier ice. From sea level down to −120 m the fiord sidewalls are smooth (B), but sidescan-sonar data show a botryoidal pattern (*see* inset), interpreted as creep by Shaw et al. (1999b). Below −120 m the sidewalls are marked by a series of parallel rills (C) with relief of 2–5 m. Although their origin and relation to the upper creep zone is unknown, they are indicative of mass transport of sediment from the sidewalls into the basin. More mass transport is indicated by a large submarine fan (D) located seaward of Indian Brook (*see* inset). Shaw et al. (1999b) showed that the glaciomarine sediments at the fiord had been deformed by slumping and slipping (*see* their Fig. 14). Some evidence appears on the fiord floor where an escarpment (arrowed) may mark the edge of a dislocated sediment block. In depths shallower than 150 m the seafloor is impacted by iceberg-grounding pits (E); sidescan-sonar images show a reflective seafloor with pockets of postglacial mud (low reflectivity) in the pits.

Little Bay

Little Bay is distinguished from the large fiords of the study region by its high exposure to wave activity, and the high degree to which it is impacted by icebergs. Postglacial mud is confined to a narrow strip in the deepest parts of a central trough (Shaw et al., 1999b, their Fig. 48), and the underlying glaciomarine mud is exposed over a wider area. Bedrock predominates in shallow water. Small basins are partly or wholly filled with iceberg-turbated glaciomarine sediment. Some pits and berms have a subdued relief, whereas others have fresh, sharply defined berms. Most pits are in the glaciomarine unit, and are probably relict, but fresh pits in the postglacial mud show that the process of iceberg impact is active today.

Little Bay Arm (Fig. 82) lies southwest of Otter Island Narrows (depth 20 m) and is thus protected from the iceberg impact that is prevalent in Little Bay proper, and only two iceberg impact pits are discernable on the image. Little Bay Arm is divided into a series of basins separated by bedrock ridges with Quaternary sediment veneers. This is one of two sites in Notre Dame Bay at which mining activities on land have had an impact in the marine environment. Between approximately 1989 and 1991, the central portion of a mine

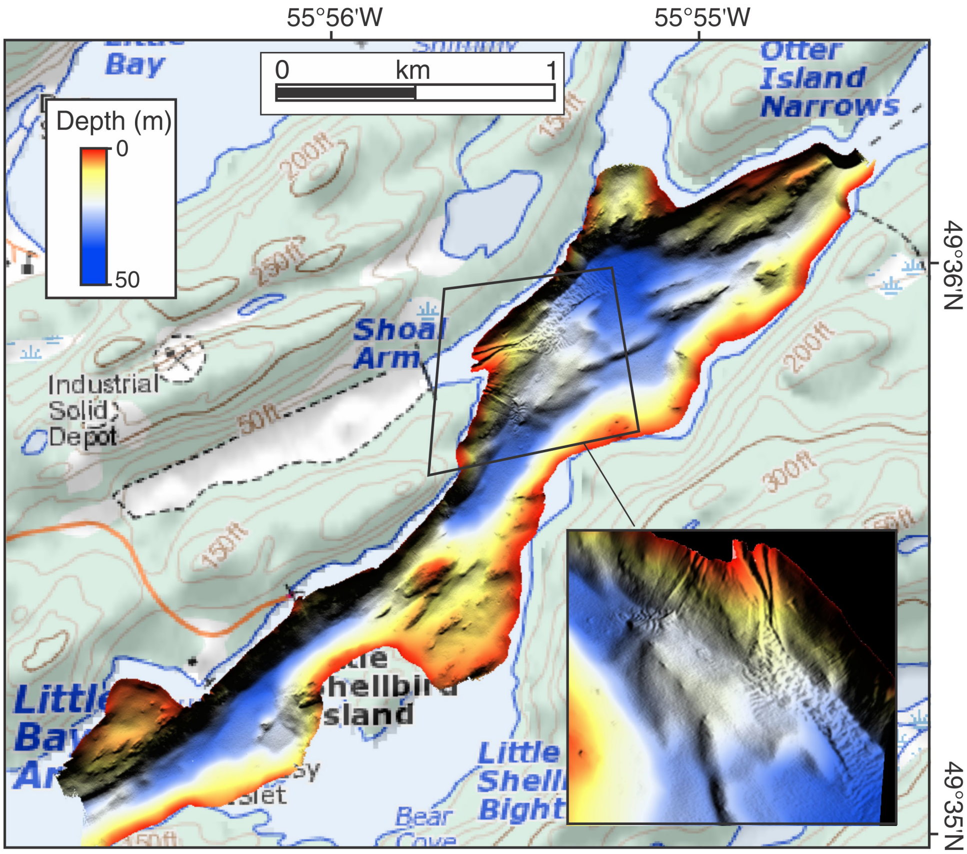


Figure 82. Multibeam image of Little Bay Arm, and an inset with a 3-D view toward the failure deposits. Compared with Little Bay proper, few icebergs enter Little Bay Arm; two furrows were detected at the extreme northeast of this image. Location shown on Figure 73.

tailings dam failed, releasing tailings into Little Bay Arm (Veinott et al., 2001, 2002, 2003; Parrott et al., 2010). Since that time, an estimated 30–50% of the tailings have flowed through the breached dam forming a tailings delta and other seafloor deposits in Little Bay Arm. The multibeam image shows submarine deposits of tailings, large bedforms, and several channels incised during the failure event.

Inner shelf outside Green Bay

This area (Fig. 83a) includes the northward continuation of the Green Bay glacial trough (A) that averages 440 m deep. The trough does not have a pronounced sill; however, unusually for this region, a series of small transverse moraines (Fig. 83b) is present, at depths between 380 m and 260 m. This may indicate a stepwise retreat of a grounded ice margin. The area is dominated by northeast-trending shallow bedrock ridges (B in Fig. 83a), some of which have streamlined glacial landforms (C) in their lee where they meet the adjoining deeper, smoother terrain. The orientation of the streamlined landforms shows that glacial ice exiting the region streamed to the east-northeast, heading down Notre Dame Channel. Above a depth of about 150 m, sediment-filled areas between bedrock outcrops (D) are intensely marked by impact features, predominately grounding pits (Fig. 83c), many with berms and a 'fresh' appearance.

Large areas of postglacial mud are evident (E in Fig. 83a) and have gentle relief. Line X-Y shows the location of a Huntec DTS profile (Fig. 84). The underlying bedrock terrain is mantled by a drape of glaciomarine mud and ponded postglacial mud, thicker in depressions than over highs. The glaciomarine unit contains several of the acoustically transparent intervals common in this region and interpreted as debris flows by Shaw et al. (1999b). The fault-controlled eastern edge of the Baie Verte Peninsula is very evident (F).

Green Bay

Green Bay (Fig. 85) is classified into two sections: a deep (450 m) outer portion flanked by shallow benches, and a shallower inner portion. In the outer portion, a gently shelving bench up to 2 km wide occurs between 50 m and 100 m water depth along the south side (A), and a narrow (1.5 km) bench is present between 100 m and 200 m water depth along the north side (B). Quaternary deposits in the fiord consist of till, glaciomarine mud, and postglacial mud, with a combined maximum thickness of 165 m. Detailed mapping of the marginal benches (Fig. 47 *in* Shaw et al., 1999b) reveals a predominance of gravel and muddy gravel, with scattered patches of sand and/or mud and numerous bedrock outcrops. Wave-formed gravel ripples are present in a number of places on both sides of the bay, attesting to

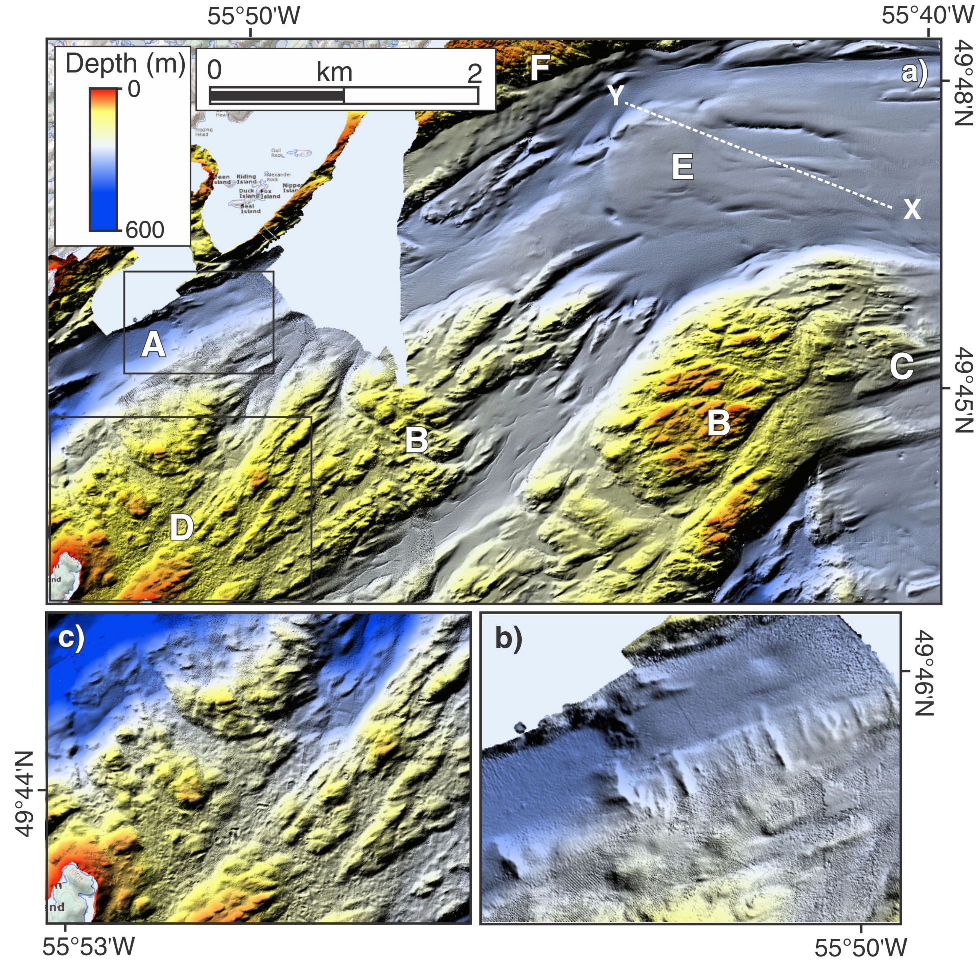


Figure 83. a) Shaded-relief image of the inner shelf outside Green Bay. This area includes the northward continuation of the Green Bay glacial trough (A). Northeast-trending bedrock ridges (B) have streamlined glacial landforms (C) in their lee. Above about 150 m, sediment-filled areas between bedrock outcrops (D) are intensely marked by iceberg-impact features. Areas of postglacial mud (E) have gentle relief. Line X-Y shows the location of a Huntec DTS profile (Fig. 84). The fault-controlled eastern edge of the Baie Verte Peninsula is very evident (F); **b)** inset showing transverse moraines at the fiord mouth; **c)** inset showing iceberg-impacted terrain at depths above 150 m. Location shown on Figure 73.

active reworking by waves propagating from Notre Dame Bay. The platforms are intensely marked by iceberg-grounding pits above about 130 m depth; some curvilinear furrows also occur.

The multibeam survey reveals the presence of two submarine moraines (C) in the fiord, missed by the survey line of cruise CCGS *Hudson* 90013. These sharp-crested landforms rising 60–75 m above the fiord floor record longer periods of ice-margin stability than the series of minor moraines observed to the northeast. In the west-trending arm of the fiord (D), above depths of 100 m, and toward the head of the bay in depths above 150 m (inset), the seafloor is imprinted by numerous iceberg-grounding pits, some with a fresh appearance.

East coast of Baie Verte Peninsula

The east coast of the Baie Verte Peninsula is fringed by a shallow (<–180 m), narrow, rocky shelf the width of which expands from less than 1 km at the mouth of Green Bay to 7 km at Cape St. John. The shelf edge marks the boundary of a major northeast-trending fault (Fader et al., 1989). The shelf drops off steeply to the floor of Notre Dame Bay, where Quaternary sediments overlie sedimentary rocks. A major moraine lies at the foot of shelf along the north side of the peninsula, and whereas overlying glaciomarine sediments have been dated (the oldest of three radiocarbon determinations in marine pelycypods in core 90013-064 was

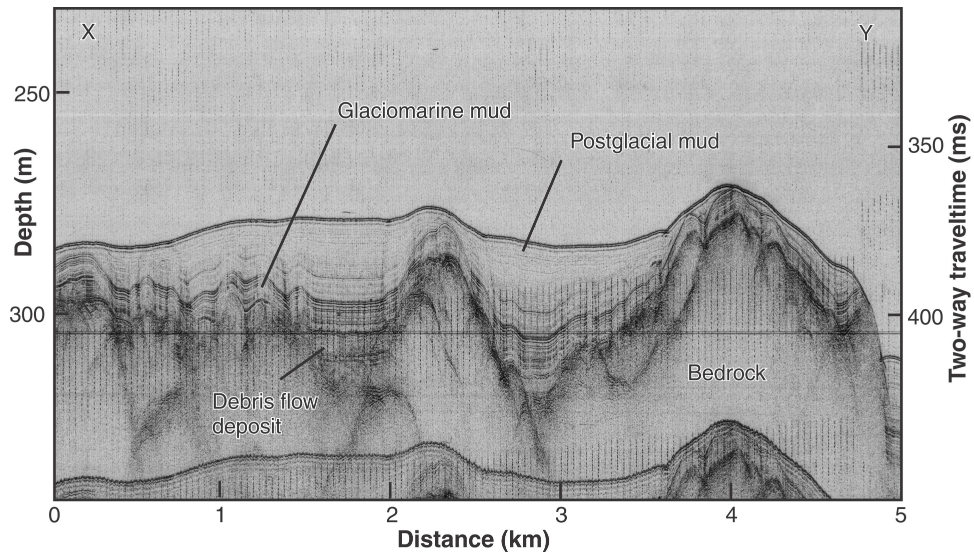


Figure 84. Huntec DTS profile outside Green Bay.

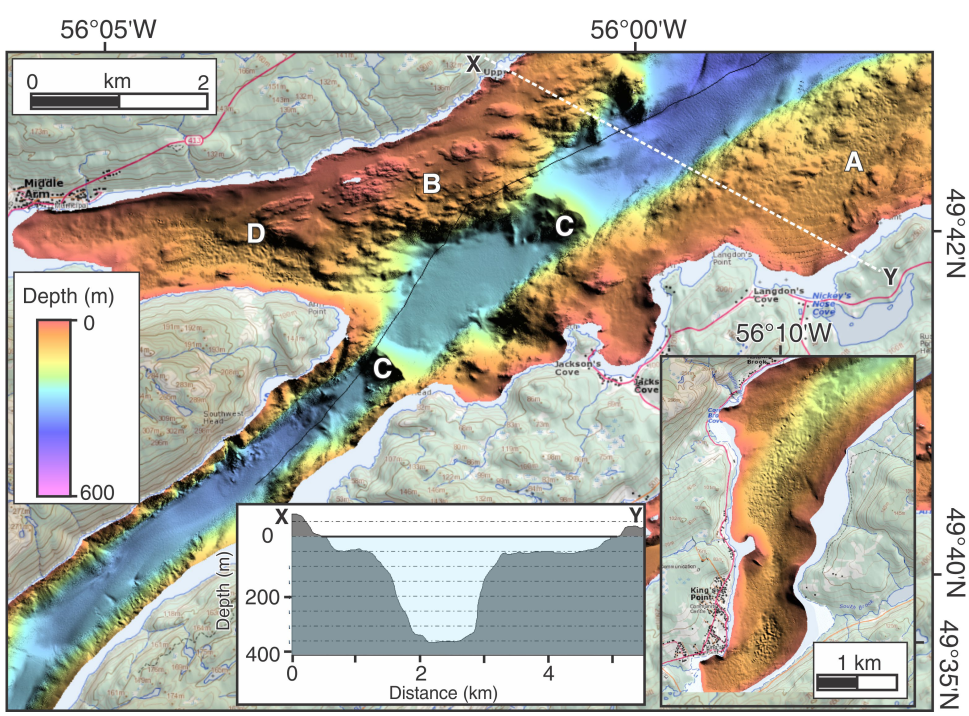


Figure 85. Shaded-relief image of Green Bay. The inset shows the head of the bay. Also showing cross-section X-Y. Location shown on Figure 73. Letters are features discussed in the text.

13 380 ± 90 ^{14}C years BP, *see* Shaw et al. (1999b)), the moraine does not pass laterally into glaciomarine sediments (till tongue relationship).

An area of the shelf (Fig. 86) was mapped in detail by Shaw et al. (1999b) and surveyed with multibeam sonar more recently, as part of a project to assess extent of mine tailings at the seafloor (Parrott et al., 2010). Copper mineralization was discovered at Tilt Cove in 1857 and mining was carried out intermittently between 1864 and 1917. No tailings were produced during this initial phase of mining. The mine was reactivated in 1957 and operated until 1967. During this time at least 5–6 million tonnes of tailings were generated (P.M. Dimmell, unpub. report for Department of Fisheries and Oceans Canada, 1999) and slurried into the ocean through a pipeline located on the rocky point on the south side of Tilt Cove bight. This site contrasts with Little Bay, in that tailings deposition was deliberate during the mining operations, whereas the tailings at Little Bay were deposited following a catastrophic dam failure around 1989 (Veinott et al., 2001, 2002, 2003).

The seafloor exhibits numerous bedrock shoals rising up to 20 m above intervening sediment-filled basins. Sharply defined furrows (A) and pits (B) represent recent iceberg impact on the seafloor. A submarine fan of mine tailings (C) is present offshore from the former mine to a depth of 45 m. The upper 18 m is smooth, possibly as a result of wave action. The deeper part of the fan has numerous small escarpments giving a botryoidal pattern as in Halls Bay, suggestive of a creep process. The fan is heavily imprinted by iceberg pits, testimony to the frequency of recent iceberg impact. Away from the fan, the densest concentration of pits is in glaciomarine sediments, but some pits are in the smooth seafloor areas (postglacial mud), again testifying to active impact. A sunken barge (D) is also evident on the image. Parrott et al. (2010) reported on studies at the site that included a repeat multibeam-sonar survey. This detected little change, except a single new iceberg furrow. Comparison with the map in Bulletin 532 (Fig. 46 *in* Shaw et al., 1999b) reveals the value of multibeam sonar in determining the true morphology of the seafloor.

At the north end of the platform that extends along the east side of the Baie Verte Peninsula, bedrock ridges trend to the northeast, separated by low-relief seafloor imprinted by iceberg furrows and pits (Fig. 87). Except in Manful Bight (Fig. 87), backscatter strength is high, suggesting the low-relief seafloor, with depths commonly 60–100 m, is gravelly. The impression gained is that the iceberg furrows (*see* arrow in inset, Fig. 87) and pits are modern.

Inner shelf north of the Baie Verte Peninsula

Detailed maps of the remainder of this region are contained in Shaw et al. (1999b). Offshore from La Scie two depth-dependent zones were identified: an outer zone of iceberg-turbated glaciomarine mud; and an inner zone, less than 90 m depth, with bedrock escarpments up to 25 m high and shallow-water areas of wave-mobilized sediments.

Baie Verte, a 20 km long embayment on the north side of Baie Verte Peninsula, has been intensively surveyed, mainly due to the early discovery of anomalous levels of gold in grab samples (Shaw et al., 1990c). Preliminary assessments of the data were presented in cruise reports (Shaw and Wile, 1990; Shaw et al., 1990a), and comprehensive accounts of the surficial sediments off La Scie and in Baie Verte were published by Shaw (1991, 1992). Southwest of the bedrock sill at the mouth of the bay (depth 192 m) a bedrock trough has a maximum depth of 324 m. In contrast, inner Baie Verte is shallow and gently shelving. Bedrock

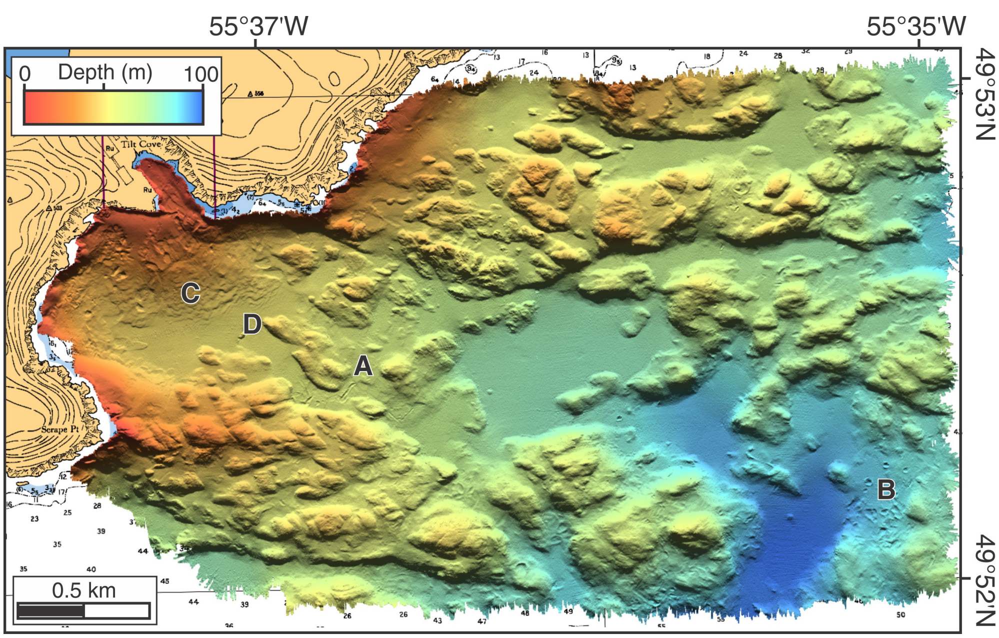


Figure 86. Shaded-relief image of Tilt Cove area. Location shown on Figure 72. Letters are features discussed in the text. Background from Canadian Hydrographic Service (2003c).

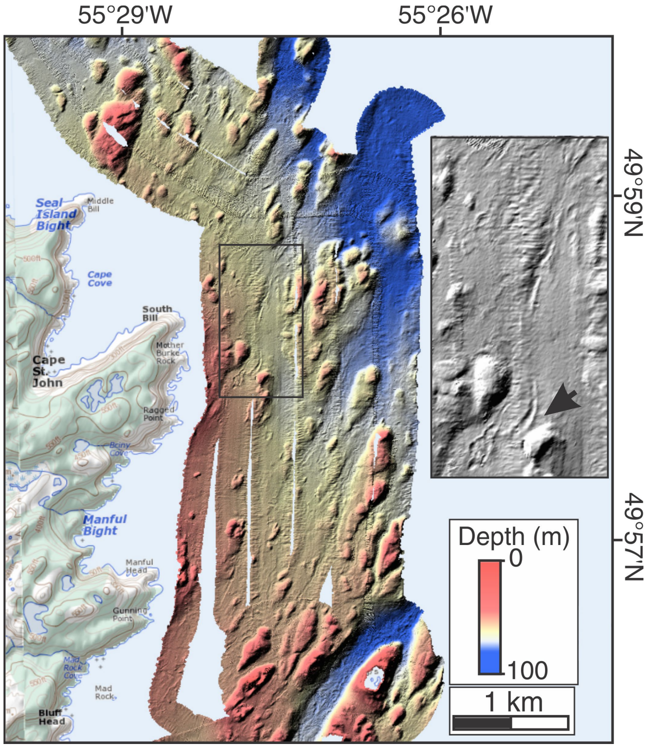


Figure 87. Seafloor morphology off Cape St. John. Location shown on Figure 72.

exposures are common in the outer bay, and less so in the inner bay. Glacial diamicton does not occur in the outer bay, but in the inner bay it overlies bedrock, commonly covered by a veneer of postglacial sediments, either muddy gravel or gravelly mud. Stratified glaciomarine sediments in the deep outer bay attain a thickness of 70 m, and in the inner bay form a thin (5–10 m), conformable drape over either bedrock or glacial diamicton. Gravity cores that intersected this unit show that it is composed of grey, gravelly, sandy clay. An undetermined thickness of postglacial mud occurs in the deep trough of the outer bay. In the inner bay, ponded deposits of mud or sandy mud occur in the deeper areas, grading into sandy mud or muddy sand close to the coast. Wave-energy levels increase from the inner bay to the outer bay, and the sediments on the shallow embayment shelves of the outer coast are sand and gravel.

REGION 11: WHITE BAY TO THE STRAIT OF BELLE ISLE

Setting

This region (Fig. 88) comprises the inner shelf of the east coast of Northern Peninsula and off the west coast of Baie Verte Peninsula. It lies within the Humber tectonic zone. The remarkably linear coastline is aligned with a major extension fault (Fader et al., 1989), separating the basement rocks onshore from Carboniferous rocks offshore. Carboniferous rocks therefore occur offshore everywhere, except north of Canada Bay and around Bell Island, where upper Proterozoic-Cambrian rocks occur (Fader et al., 1989). Like Notre Dame Bay, this region is impacted by both extensive sea ice and icebergs brought south in the Labrador Current. The Geological Survey of Canada has conducted surveys in the southern part of this region (Shaw et al., 1999b). The available multibeam-sonar coverage is, unfortunately, confined to a very narrow coastal fringe (Fig. 89), although several lines extending into deeper water suggest that the geomorphology of the unmapped areas may be very interesting.

Coastlines

The coast is mostly very linear, but is interrupted by a series of relatively small fiord-like inlets that contain deposits of glaciomarine and postglacial mud. Hampden River enters the bay via a fiord-head sand and pebble-cobble delta with extensive subaerial distributary mouth flats.

Inner shelf: overview

The region is divisible into two parts. In the south, White Bay (Fig. 88) extends 82 km in a northeastward direction from its head in Hampden Bay. White Bay has a deep, central basin that reaches a maximum depth of 536 m east of Jackson's Arm, and progressively shallows northward to 400 m east of Little Harbour Deep. In the north, the linear coastline fronts directly on Notre Dame Bay, which was mapped by Dale and Haworth (1979). These authors depicted a narrow zone of bedrock close to the coasts south of Canada Bay, a wide bedrock area to the north of there, and an area of postglacial mud down the middle of White Bay. The area south of 50°30'N was mapped by Shaw et al. (1999b), as part of their 1:250 000 scale map. This map is broadly similar to that of Dale and Haworth (1979) except that a zone of glaciomarine mud is shown between the bedrock and postglacial mud units.

White Bay

White Bay contains sediments of varying thickness in its deep, central area. At the mouth of the bay, 30 m of postglacial mud overlies 5 m of glaciomarine mud and about 25 m of glacial diamict containing internal reflections, suggestive of at least three subunits. Thicknesses increase southward to a maximum of 60 m of postglacial mud and at least 60 m of glaciomarine mud off Jackson's Arm. Due east of Sop's Arm the bedrock valley is narrow and contains only 40 m of postglacial mud, but at least 170 m of the glaciomarine unit. The glaciomarine and postglacial mud units thin greatly on the eastern flanks of the basin. Truncated internal reflections at the surface of the postglacial unit indicate that the sediments have been eroded by as much as 8 m. Figure 90 shows a sleeve gun record along a line transverse to the bay

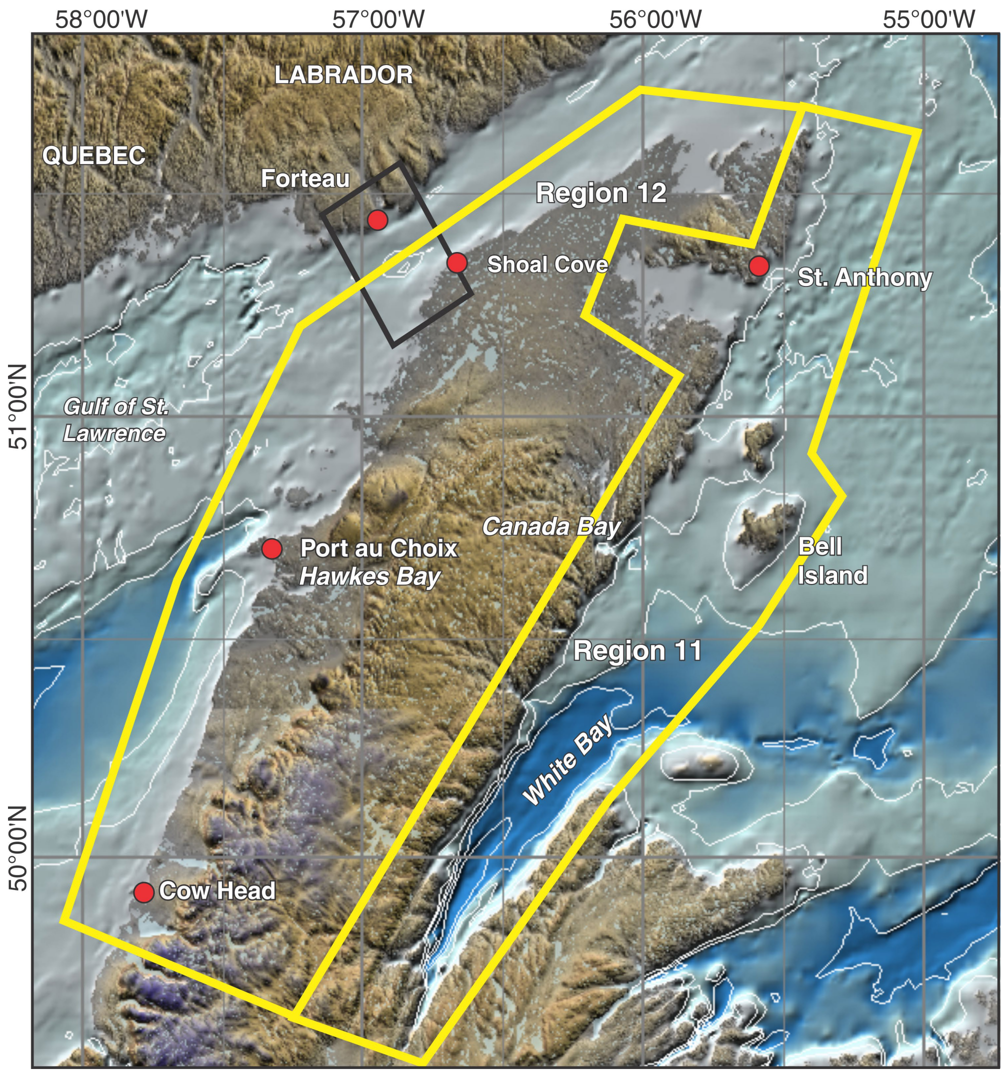


Figure 88. Regions 11 and 12.

(location on Fig. 89); the Huntec DTS record (inset, Fig. 90) shows the acoustic stratigraphy in more detail. The arrow indicates a synclinal structure on the east side of the bay. Shaw et al. (1999b) published five radiocarbon dates on marine pelycypod samples from the postglacial mud unit in core 90013-035 (Fig. 89a) collected in a depth of 367 m, arguing that the youngest glaciomarine sediments at the site had an age of 10.7 ka and the sedimentation rate greatly declined after ca. 4 ka.

Inner shelf off Northern Peninsula

The narrow strip of multibeam-sonar data (Fig. 89a) shows that a zone of bedrock extends 3–7 km out from the coast of Northern Peninsula, and mostly lies above 100 m depth. The steep outer face of the bedrock zone plunges to depths of 300–400 m, where smooth seafloor is encountered. The multibeam-sonar imagery reveals very complex bedrock structure, with a wide range of geomorphic expressions, but mostly with a strong northeast grain (i.e. coast-parallel). Several small basins within the bedrock zone, and also the large bays such as Canada Bay and Great Harbour Deep, contain ponded postglacial mud or sandy mud characterized by a smooth seafloor and low backscatter, overlying glaciomarine mud. In shallow areas (<100 m) numerous small perched basins contain sand (low backscatter). In deeper water, bedrock ridges have a veneer of glaciomarine mud strongly imprinted by iceberg furrows and pits, to a maximum depth of 150 m. In one area a wide, northeast-aligned ridge may be morainal in origin. The great diversity of bedrock terrains cannot be described in detail here, so vignettes from several areas are presented.

Coney Head area

Figure 89b shows the seafloor in the vicinity of Coney Head. Two principal seafloor terrains are present. Bedrock predominates in the shallow areas to the west. The

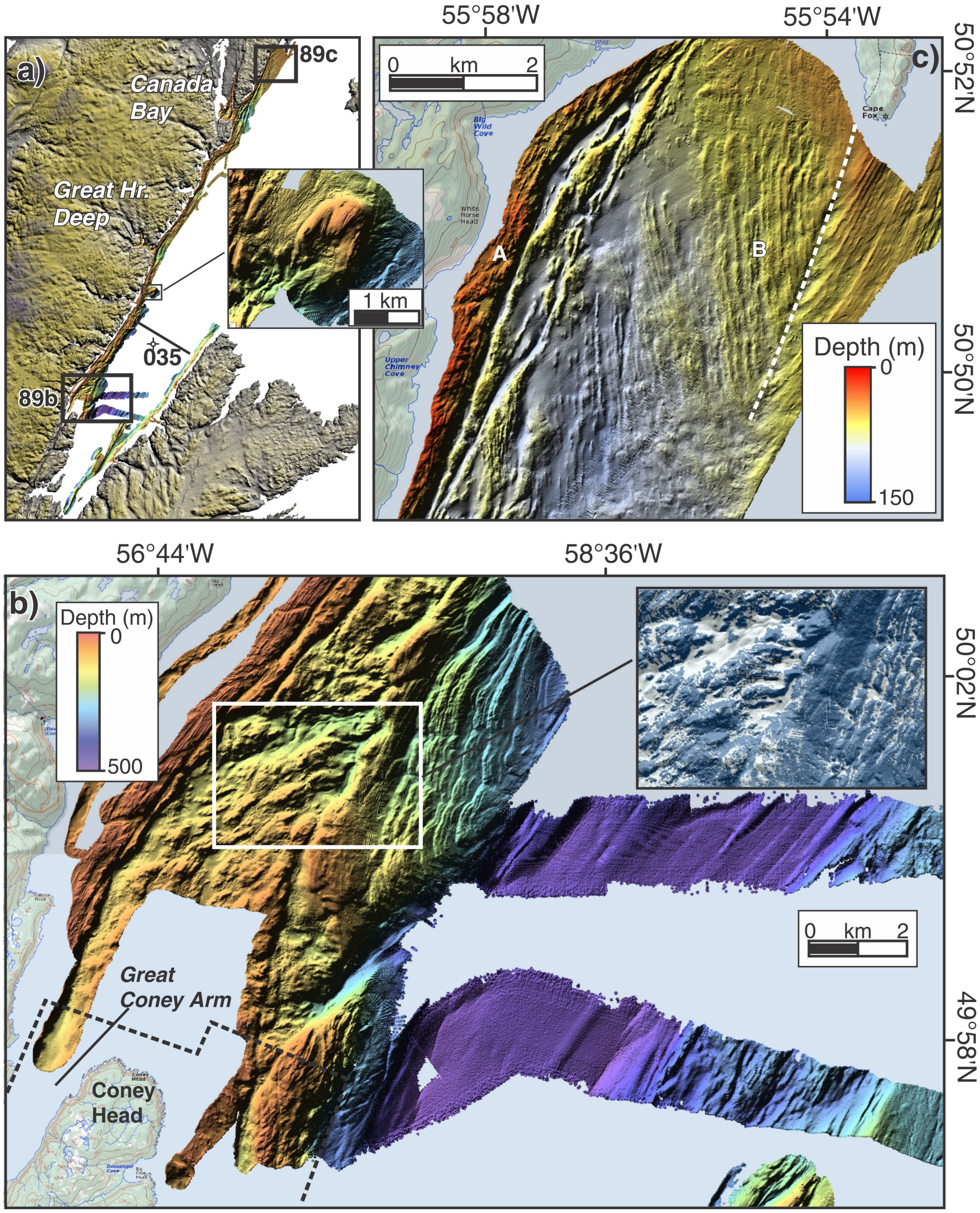


Figure 89. White Bay multibeam-sonar imagery. **a)** The available coverage is indicated, together with the locations of enlarged coverage (b and c), the sleeve-gun seismic-reflection cross-section of Figure 90, and the location of piston core 90013-035 (Shaw et al., 1999b). The inset enlargement on Figure 89a shows a large isolated bedrock knoll offshore from the main bedrock zone. **b)** Coverage off Coney Head, at the south side of the survey area, with an inset image showing backscatter strength distribution in part of the area; the light tone indicates low backscatter. The dashed line indicated the limit of the interpretative map by Shaw et al. (1999b, their Fig. 39). **c)** A series of bedrock ridges trend to the northeast (A). A change in bedrock lithology elsewhere is indicated by the presence of numerous small ridges with relief of about 5 m, organized into an anticlinal or synclinal form (B) the axis of which is shown by dashed white line. The Quaternary cover is greater in the south, where the seafloor in depths of about 100 m is imprinted by iceberg pits.

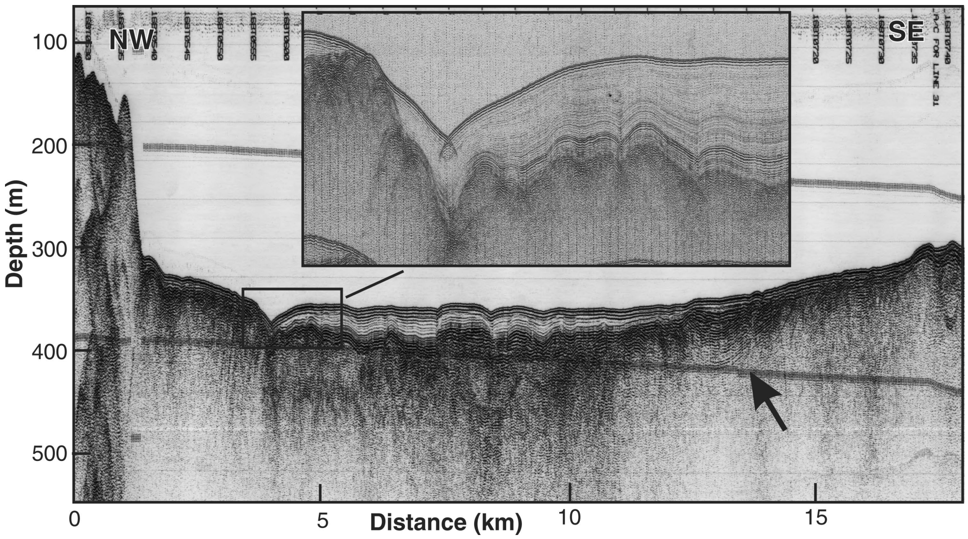


Figure 90. Sleeve-gun seismic-reflection record across White Bay (cruise CCGS *Hudson*, 90013, Shaw and Wile (1990)) with part of a Huntec DTS record showing postglacial mud overlying glaciomarine mud.

backscatter-strength map shows predominant high backscatter strength, with numerous small areas of low backscatter corresponding to areas of smooth seafloor between bedrock highs (*see* insert, Fig. 89b). These areas are likely mud. Areas landward of the 100 m isobath were surveyed in detail by Shaw et al. (1999b) who found that the seabed surrounding Coney Head — outside the multibeam-sonar coverage — comprises a veneer of postglacial muddy gravel less than 5 m thick, small patches of postglacial sand or mud, and bedrock outcrops. Gravel ripples and iceberg pits are present within the muddy gravel. Great Coney Arm contains postglacial sandy mud, less than 10 m thick, that overlies glaciomarine sediment up to 42 m thick. Grab samples and seabed photographs in water depths ranging from 33 m to 123 m show that the seabed consists of sand and gravel. The gravel forms a surficial armour, several clasts thick and encrusted with *Lithothamnion* sp. A comparison with Figure 39 in Shaw et al. (1999b) reveals the limitations of the approach used in portraying the complex nature of the seafloor before the use of multibeam-sonar systems.

East of the bedrock zone, in water depths averaging 470 m, a cover of postglacial mud is characterized by a smooth seafloor with low backscatter. The postglacial mud and underlying glaciomarine sediments overlie streamlined landforms created by fast-flowing glacier ice moving northeastward, out of White Bay. Areas south of the Coney Head area were described by Shaw et al. (1999b).

West side of the Baie Verte Peninsula

The map by Shaw et al. (1999b) shows a zone of bedrock adjacent to the coast. A narrow swath of multibeam imagery from this region shows this interpretation to be largely correct; however, it also shows that between bedrock highs, and down to depths of about 120 m, the seafloor is intensely marked by iceberg pits; and off the northwestern tip of Baie Verte Peninsula, streamlined glacial landforms (megaflutes) in water depths of 170 m to 250 m and oriented at 50°, are indicative of strong glacial ice flow out of White Bay, toward Notre Dame Bay.

Conche Harbour area

This area (Fig. 89c) lies at the north end of the available multibeam-sonar coverage. In the western part of the imagery, a series of bedrock ridges trend to the northeast (A). A change in bedrock lithology elsewhere is indicated by the presence of numerous small ridges with relief of about 5 m, organized into an anticlinal or synclinal form (B). Fader et al. (1989) showed Lower Carboniferous rocks offshore here. Areas of smooth seafloor between the ridges, in depths of 80–110 m, have low backscatter strength and probably comprise muddy sand. The Quaternary cover is greater in the south, where the seafloor in depths of about 100 m is imprinted by iceberg pits.

REGION 12: STRAIT OF BELLE ISLE TO BONNE BAY

Setting

This region (Fig. 88) is in the Humber tectonic zone (Colman-Sadd et al., 1990). Bedrock onshore consists of Cambrian and Ordovician shelf-facies carbonate rocks; similar rocks are found offshore (Fader et al., 1989). The region is characterized by extremely low relief in the coastal regions, except where the Highlands of St. John (504 m) impinge the coast north of Port au Choix. The Long Range Mountains lie far inland in the north-central part of the region, but approach the coast farther south. Offshore, the shelf is shallow with gentle gradients for the most part. The surface circulation is southward in the north of the region, but in the southern part it is northward (Drinkwater and Gilbert, 2004). Only a small portion of the southward-moving Labrador Current enters the gulf via the Strait of Belle Isle (Wu et al., 2012). At its north end the Strait of Belle Isle has a –62 m sill that prevents the larger icebergs travelling south in the Labrador Current from entering the Gulf of St. Lawrence.

This region has an extreme relative sea-level history compared with the rest of the island. Based on data from either side of the strait, and from the east side of Northern Peninsula (Shaw et al., 2002b) relative sea level was about 130 m higher than now about 13 000 years ago (Bigras and Dubois, 1987; Clark and Fitzhugh, 1992; Grant, 1994) and has been falling since. Thus, the sill was considerably deeper in the past, and larger icebergs were presumably able to transit the strait. North of Bonne Bay a series of small fiords isolated from the ocean by falling relative sea levels are located in re-entrants in the Long Range Mountains. The region is archeological significant, being the site of both the Port au Choix archeological site (Bell and Renouf, 2004), and the L'Anse aux Meadows National Historic Site World Heritage Site.

Over the years, a considerable amount of research has been focused in the Strait of Belle Isle regarding connections between Labrador and the Island of Newfoundland. There were early plans were for a utility tunnel to bring Labrador electricity to the Island of Newfoundland, and later plans for a fixed link. More recently, plans were finalized for an updated version of the utility link. This is the only region for which no government or publically accessible multibeam data are available, so descriptions are based primarily on published data.

Coastlines

In the north, beyond the limit of the Long Range Mountains, the carbonate terrain is flat lying, and gradients at the coast are extremely low. Fringing gravel beaches are backed by successive beach ridges that rise in elevation inland, reflecting the monotonic fall of relative sea level over many thousands of years. Sandy coastal systems are well developed at Shallow Bay, where sandy tombolos link to headlands and shoals, and beaches are backed by the most extensive coastal dunes north of St. George's Bay. The dunes, the bulk of which are located in Gros Morne National Park of Canada, contain organic horizons and buried trees indicative of periods of stability. Blowouts in the dunes are oriented to the northeast, indicative of the prevailing winds. Forbes (1984) noted the multiple bar system developed on cobble- and boulder-strewn tidal flats at Hawkes Bay.

Inner shelf

The bedrock geology of the crossing area is summarized in the report by Woodworth-Lynas et al. (1992) who showed that the Palaeozoic sediments (limestone, shale, and sandstone) overlie Precambrian strata (*see* map 90-006 *in* Woodworth-Lynas et al. (1992)). The surficial geology is shown on map 90-006 in the same report. Surficial sediments in the strait consist of thin sand and gravel overlying thin till, and bedrock outcrops. Sand waves and megaripples, dunes, and sand ribbons testify to strong currents in the area, evident on Figure 6 of Wu et al. (2011) which shows very high, extreme currents both near bottom and at the surface.

Perhaps the most interesting aspect of the surficial geology is the presence of ribbed moraines in deeper areas of the strait, apparently formed by glacier ice flowing from the Labrador side. These were not present in depths shallower than about 90 m near the south coast (*see* cross-section of the strait on Figure 15 of Woodworth-Lynas et al. (1992). Fields of De Geer moraines on land nearby were formed by glacier ice based on the Island of Newfoundland (Grant, 1994). Whereas the cause for the absence of moraines between the two types is uncertain (it may relate to the submarine escarpment just east of the ribbed moraines), the presence of both types is likely indicative of incremental retreat of ice to the northeast (Labrador) and southeast (toward Island of Newfoundland), following separation of Island of Newfoundland and Labrador ice.

REGION 13: BONNE BAY TO PORT AU PORT BAY

Setting

Region 13 (Fig. 91) is in the Humber tectonic zone (Colman-Sadd et al., 1990). The onshore rocks, including the ophiolite complexes, belong to the Taconian allochthon, whereas the offshore rocks are Ordovician to Carboniferous sediments. The boundary between the two zones is a major thrust fault. High relief onshore contrasts with low relief offshore in the shallow waters of the Gulf of St. Lawrence, whereas deep water is found in Bonne Bay and Bay of Islands. The surface circulation is dominated by a northward-flowing current (Drinkwater and Gilbert, 2004; Wu et al., 2011). The area is largely free of the modern icebergs that impact regions 7 to 12, and whereas numerous iceberg furrows and pits are found on the seafloor in many areas, they are largely relict. The major fiords of Bonne Bay and Bay of Islands differ in many respects, particularly including degree of anthropogenic impact: Bonne Bay is adjacent to Gros Morne National Park of Canada, and is relatively pristine, whereas Bay of Islands, the site of Corner Brook, Island of Newfoundland's second largest city, has evidence of pollution and submarine landslides triggered by humans.

This region has been extensively mapped with multibeam sonar (Fig. 92). Early mapping by the Geological Survey of Canada scientists suffers from data problems, principally refraction errors. Nevertheless, it resulted in several outputs, notably a poster describing the marine geology of Bonne Bay (Shaw et al., 2002a) and a paper describing the marine geology of Bay of Islands (Shaw et al., 2000a). The Canadian Hydrographic Service subsequently resurveyed parts of these areas, together with additional areas.

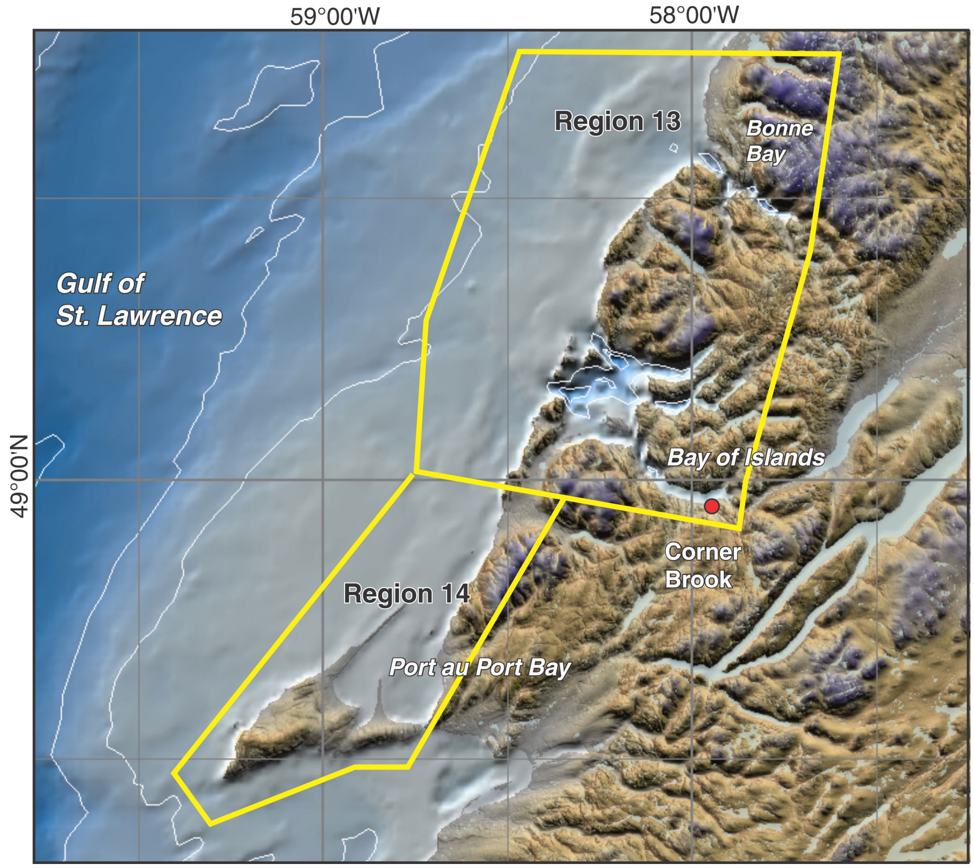


Figure 91. Location of regions 13 and 14. In region 13 the deep fiords at Bonne Bay and Bay of Islands contrast with the shallow bathymetry of the inner shelf. The arcuate moraines off Bonne Bay and Bay of Islands are clearly seen.

The extensive multibeam-sonar coverage described here provides insights into morphology and processes in these two large fiords and reveals the shelf outside the fiords to have a complexity not shown by the previous mapping. The imagery also sheds light on the architecture of the large submarine moraines located offshore from the two fiords, and reveals littoral transport pathways that terminate in submarine fans at fiord mouths.

Ground-truthing is limited to a survey of Bonne Bay (Shaw et al., 1999a) and several surveys of Bay of Islands (Shaw et al., 1995b, 1999a). The Bay of Islands surveys were confined to Humber Arm, so that the other arms of Bay of Islands have not been groundtruthed except on cruise CSS *Baffin* 89008 (Josenhans et al., 1989), when a line was run into the bay at the south entrance and out via the north entrance.

Coastlines

The outer coasts of this region are characterized by bold, rocky coastlines. In a few areas late Quaternary marine sediments are exposed at the coast, forming terraces. The best example is at Green Gardens, where the eroding terrace (Berger et al., 1992) is fronted by a gravel beach. Inside the fiords, relief is also great and unconsolidated coastlines are well developed only in a few areas, notably near York Harbour, where several small forelands are located.

The inner shelf: overview

The surficial geology of the region is shown on several maps of the Gulf of St. Lawrence. On the map by Loring and Nota (1973) the low-relief area offshore from the fiords has an inner zone of sandy gravel and an outer area of 'gravelly pelitic sand'. Josenhans et al. (1990) and Josenhans and Zevenhuizen (1993) mapped the area close to the coast as till, and showed postglacial reworked sand and gravel farther offshore from Bay of Islands. Josenhans' (2007) map showed more than 50 m of an 'upper till' in arcuate ridges at the mouths of Bonne Bay and Bay of Islands, whereas Josenhans and Lehman (1999) showed most of the inner shelf as till, with large areas of reworked sand and gravel.

None of these maps cited above contain information on the two principal fiords. Bonne Bay (Fig. 92) is a fiord with two arms — the relatively shallow (–95 m) South Arm and the relatively deep (–228 m) East Arm. West of the arcuate shoal at the entrance to Bonne Bay, the shallow shelf of the Gulf of St. Lawrence has low relief. Bay of Islands (Fig. 92) is a large fiord with three principal arms: North Arm, Middle Arm, and Humber Arm. These arms converge in a series of deep (–297 m) troughs near the entrance. To the west of the troughs is a series of islands, beyond which the seafloor shallows rapidly to an arcuate morainal ridge. Yet farther beyond this moraine the inner shelf has depths of 60–80 m. Middle Arm bifurcates into two branches: Penguin and Goose arms.

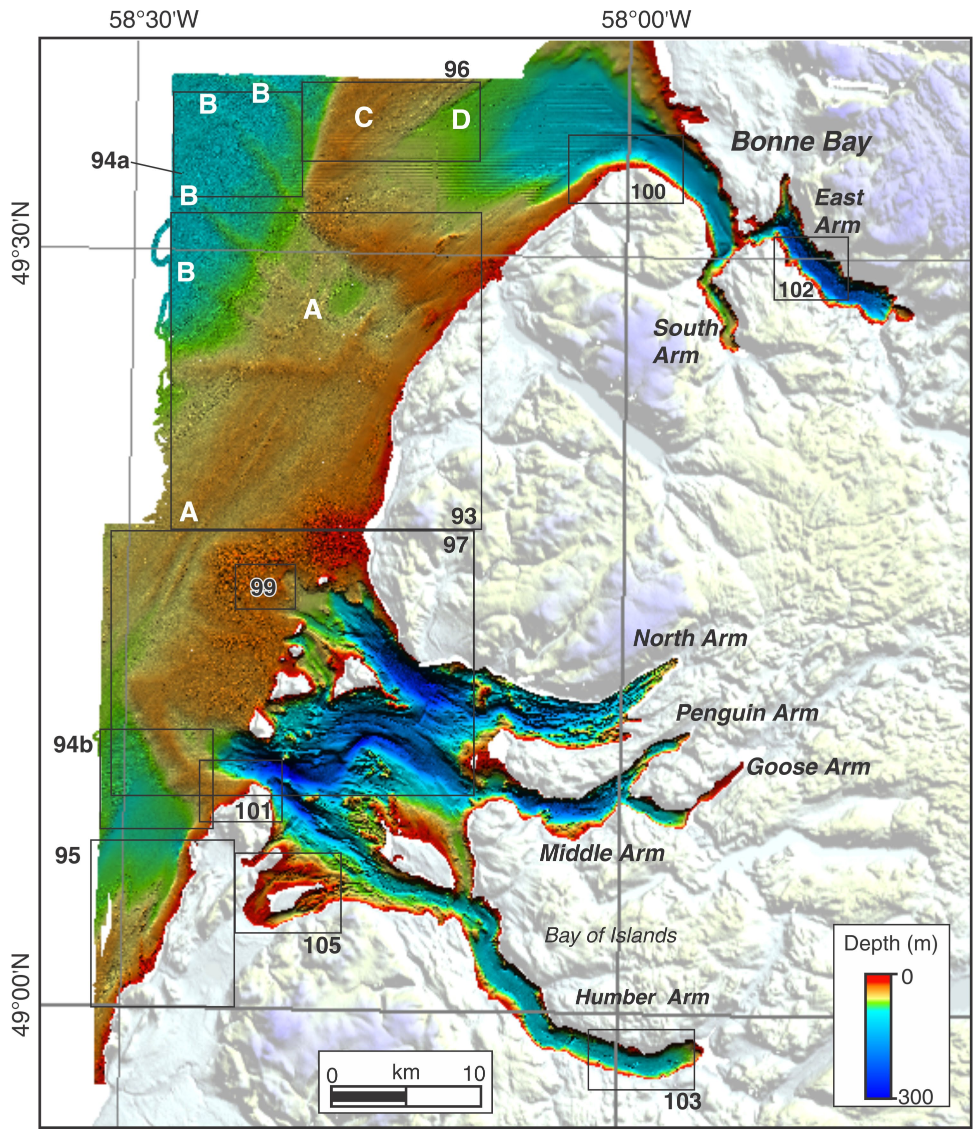


Figure 92. Multibeam coverage in region 13, showing contrasts between the two fiords: Bonne Bay, with a shallow South Arm and deep East Arm, and Bay of Islands, with multiple arms and a chain of islands at its wide entrance. Beyond the two large arcuate moraines at the fiord mouths, the shelf is marked by northeast-trending bedrock ridges with transverse morainal ridges superimposed. *See* text for key to letters. Boxes shows locations of Figures 93, 94, 95, 96, 97, 99, 100, 101, 102, 103, and 105.

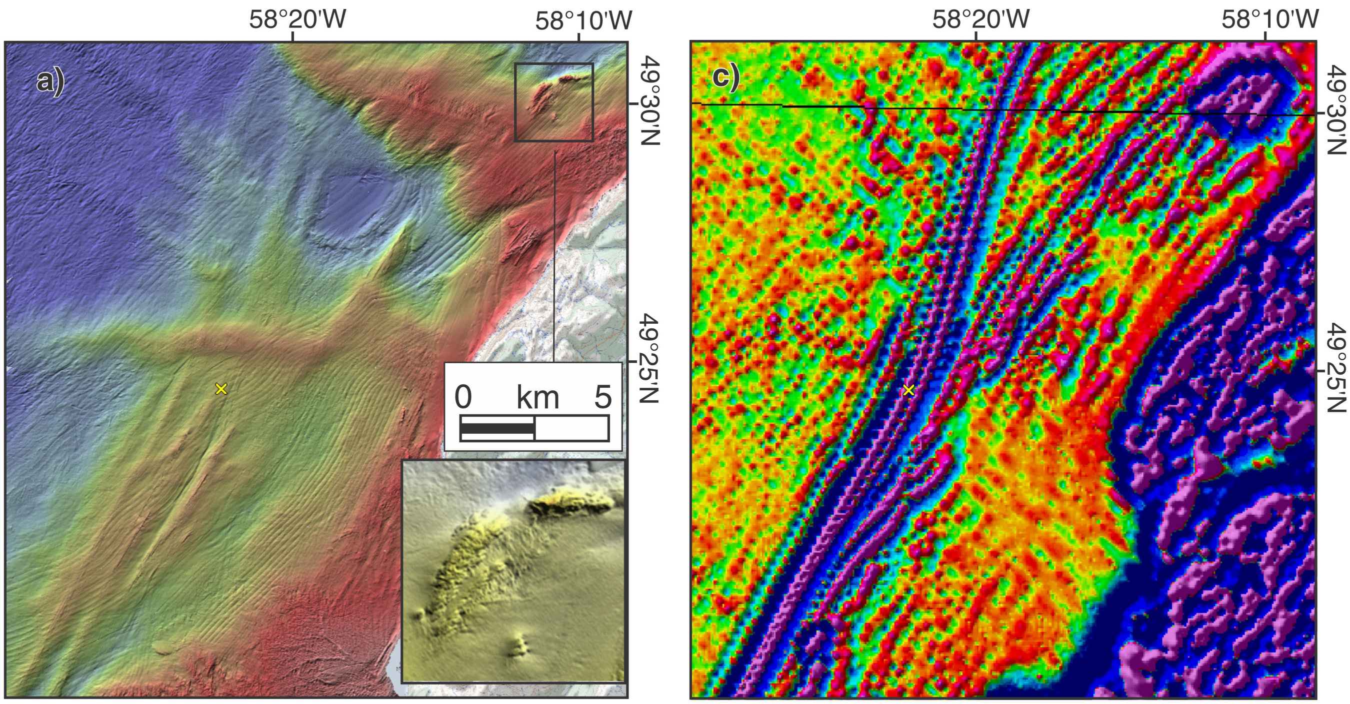
The surficial geology of region 13 is described under the following headings: inner shelf seaward of the submarine moraines; fiord-mouth submarine moraines; submarine fan development at fiord entrances; late-glacial and modern processes inside the fiords; evidence of postglacial relative sea-level change; and anthropogenic effects.

Inner shelf seaward of the submarine moraines

Bedrock ridges and morainal ridges

Ridges with varying and contrasting orientations are visible west of the moraines (Fig. 92), and are classified into two types. A series of 8–10 m high ridges oriented northeast-southwest consist of bedrock with thin Quaternary cover (A). Bedrock outcrops are evident on the ridge crests in places. Ridges with slightly differing orientation occur to the rear of the Bonne Bay moraine. They have a veneer of till imprinted with iceberg furrows. The pattern of northeast-trending ridges (Fig. 93a) corresponds with the narrow, northeast-trending ridges (Fig. 93b) evident on the recently flown aeromagnetic data (Dumont and Jones, 2013). These are a continuation of the Odd-Twins magnetic anomaly mapped farther south, off Port au Port Peninsula (Waldron et al., 2002). The northeast-trending ridges on the VG2 component of the magnetic field (Fig. 93c) suggest that the bedrock ridges on the multibeam-sonar imagery are the source of magnetism. Also seen on the aeromagnetic data (Fig. 93b, c) is an isolated, strong anomaly in the northeast. This corresponds with an arcuate bedrock ridge that stands up to 40 m proud of the surrounding terrain, and that is believed to be the source of magnetism.

A second series of ridges oriented northwest-southeast is superimposed on the bedrock ridges, and probably consists of glacial diamict (B in Fig. 92). These ridges appear to pre-date the Bonne Bay moraine (Fig. 92, C), and are interpreted



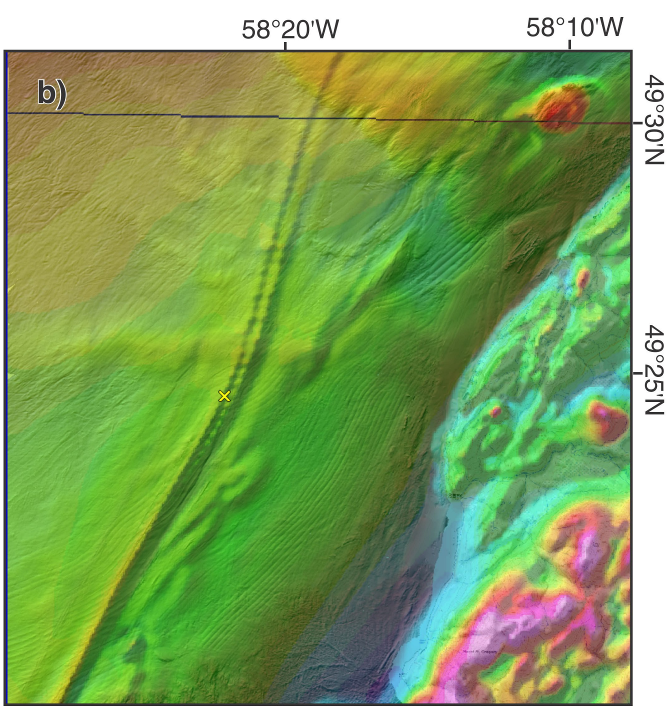


Figure 93. a) Shaded-seafloor relief just south of Bonne Bay, showing northeast-trending bedrock ridges overprinted by northwest-trending morainal ridges; **b)** total magnetic field; **c)** the VG2 component of the magnetic field. Magnetic data are from Dumont and Jones (2013). Locations are shown on Figure 92.

as recessional moraines formed during retreat of ice in the gulf to the northeast (Josenhans and Lehman, 1999), prior to formation of the large arcuate moraines. Other ridges to the rear of Bonne Bay moraine are bedrock (Fig. 92, D).

Iceberg impact

Seaward of the arcuate moraines at the entrances to Bonne Bay and Bay of Islands the seafloor is shallow (50–80 m) and has low relief (10 m). In depths below about 70 m it is intensely imprinted with iceberg furrows in the northwest (Fig. 94a). The furrows average 0.5 m deep, with the deepest at 2.5 m, and have widths varying up to 120 m. They are linear and curvilinear, and typically trend at 045°. These furrows were presumably generated by the northward retreat of grounded ice in the gulf after ca. 14.3 ka (Josenhans and Lehman, 1999). The iceberg furrows southwest of the Bay of Islands moraine (Fig. 94b) average 4 m deep, 100 m wide, lie between 70 m and 80 m depth, and are mostly oriented normal to the moraine. The majority of these furrows terminate against the ridge to the southwest in depths of 80 m. It is likely that they emanated from a calving margin at the moraine. The moraine itself has few iceberg furrows.

Seafloor erosion

At the extreme south of the multibeam coverage (Fig. 95) the Quaternary sediment cover on top of a northeast-trending bedrock ridge has been stripped away, exposing the bedrock and creating steep erosional escarpments up to 10 m high. Some aspects of this area are puzzling, for example the presence of a 1 m high ridge (arrowed) on the lip of the erosional escarpment at its north end. This may comprise coarse material from the eroding area. A survey line from cruise CSS *Baffin* 89008 (Josenhans et al., 1989) crosses this area. The cross-section on Figure 95 shows that up to 10 m of postglacial and glaciomarine mud has been removed. This survey also showed that another very similar eroded zone is present several kilometres to the southwest of the multibeam coverage.

The erosion here has some similarities with erosion on top of ridges in Placentia Bay. In that case (Shaw et al., 2013) it was concluded that currents had come from the north. The circulation in the Gulf of St. Lawrence displays a northeast-flowing current in the area shown in Figure 98 (*see below*) (Drinkwater and Gilbert, 2004). The onset of this current is clearly relatively recent, perhaps mid- to late Holocene, since both glaciomarine and overlying postglacial mud have been eroded, but in the absence of coring information and other data there is no information on the precise timing.

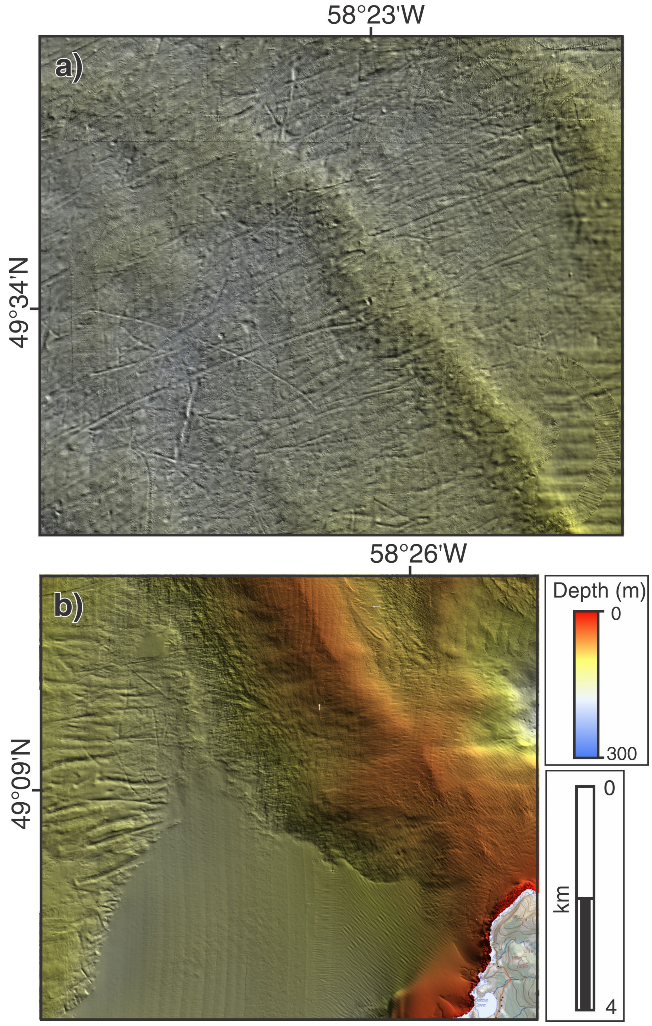


Figure 94. a) Seabed intensely furrowed by icebergs in the northwest of the region, with average depths of 70–80 m. Also shown are several of a series of northwest-oriented ridges; **b)** the furrows southwest of the Bay of Islands moraine in the 70–80 m depth range appear to have emanated from a grounding line located to the northeast, on the moraine. Locations are shown on Figure 92.

Submarine moraines

Bonne Bay submarine moraine

The shaded-relief multibeam-bathymetry image (Fig. 92) shows part of an arcuate shoal that reaches the coast south of Bonne Bay. The digital elevation model (Shaw and Courtney, 2004) shows that the shoal reaches the coast north of the bay, in the vicinity of Green Point. This shoal was interpreted by Grant (1989) as a submarine end moraine. The principal part of the moraine is a broad ridge composed of acoustically incoherent ice-contact sediment that attains a maximum thickness of 30 m (Fig. 96). The ridge extends from a depth of 70 m in the west, and has a steep western face, with a crest at –46 m. A narrower, superimposed ridge is at a depth of 33 m. The east face of the moraine slopes gently to a

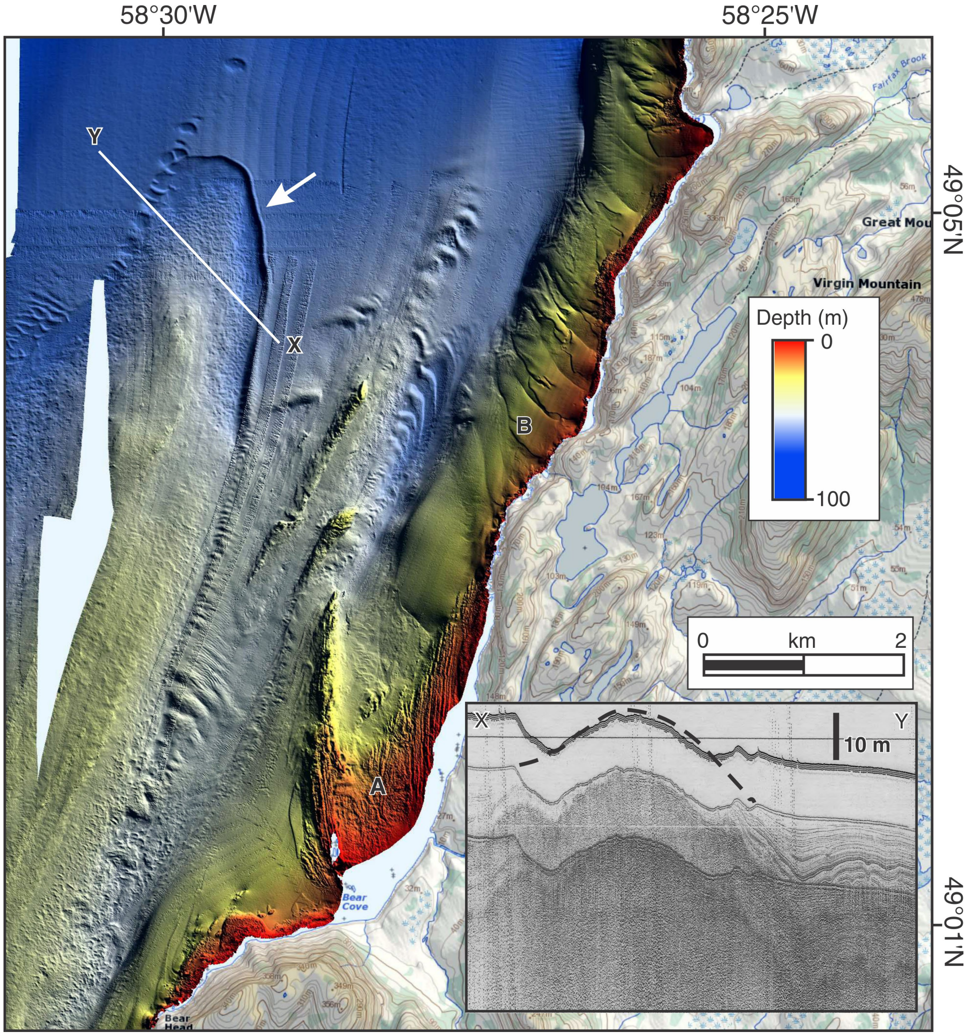


Figure 95. Seafloor erosion south of the entrance to Bay of Islands. The inset show a Huntec DTS (external eel) profile from cruise *Baffin* 89008 (Josenhans et al., 1989). The dashed line on the inset is an approximation of the seafloor before erosion. Also shown are the ridge developed on the rim of the erosional escarpment (arrow), bedrock outcrops near the coast (A), and a prism of sediment, probably sand grading to fine sand with depth, just off from the coast (B). Location is shown on Figure 92.

depth of 46 m. Sidescan-sonar records show that the seabed on the ridge is gravelly, with scattered boulders. It has a smooth appearance on the multibeam imagery, although there are some faint lineations on the crest and steep slope (in the same direction as the iceberg furrows).

The area of unusual texture (A in Fig. 96b) east of the moraine is thin Quaternary sediment over northeast-trending bedrock ridges. The bouldery surface is suggestive of glacial diamict rather than glacial marine mud, i.e. Scotian Shelf Drift Formation rather than Emerald Silt Formation, using the terminology of King and Fader (1986). Grab sample 97060-017 shows that the bottom is covered with bouldery gravel. East of this patterned area the seabed is relatively smooth (B). On the multibeam-backscatter image, the area of fine sediment here is more reflective in the west than in the east. The boundary between the two reflectivity domains corresponds with a change in tone and pattern on the sidescan-sonar records collected during the cruise.

Bay of Islands submarine moraine

The Bay of Islands moraine (Fig. 97) is a complex feature, comprising not only an arcuate ridge of till, but also several trenches formed by glacier ice exiting between the islands. Josenhans et al. (1990), Josenhans and Zevenhuizen (1993), and Josenhans and Lehman (1999) showed that this terminal moraine consists of glacial diamict up to 100 m thick, and that it overrides previously deposited glaciomarine sediments. The simplified cross-section on Figure 98 shows (in succession) the southern part of the moraine, the

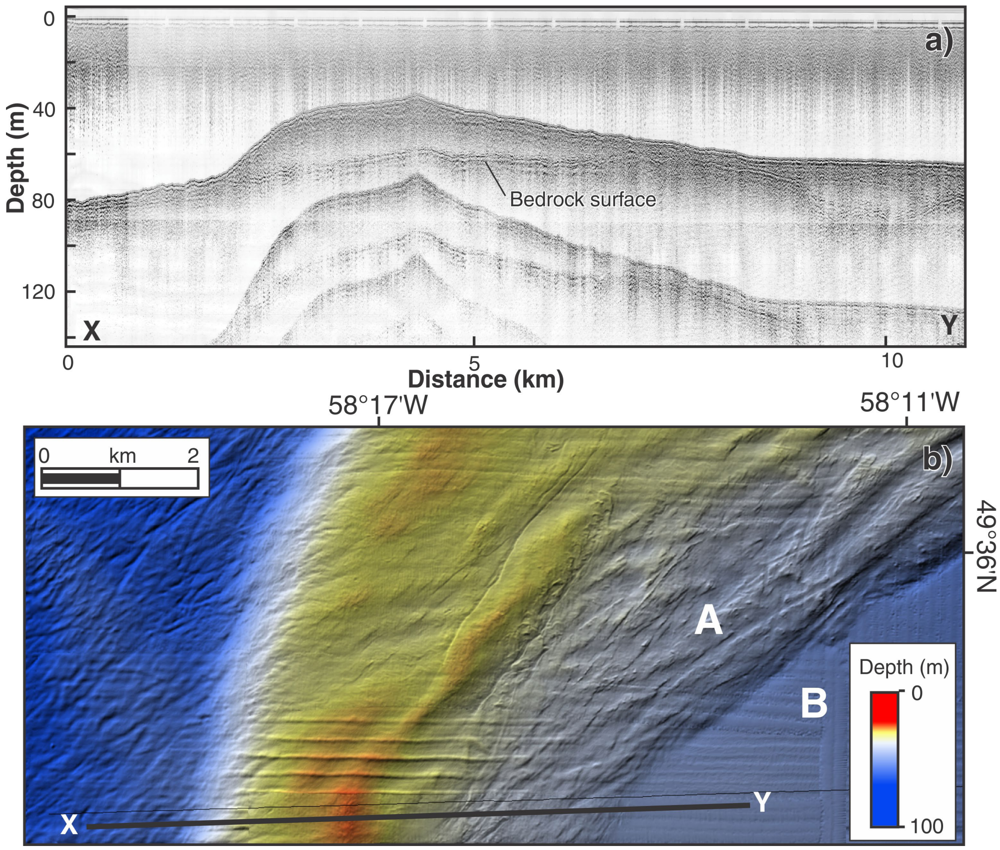


Figure 96. a) Low-frequency seismic-reflection record showing a cross-section across the moraine off Bonne Bay; **b)** shaded-relief image of the moraine, showing location of seismic profile. A and B are features discussed in the text.

trench landward of the moraine, several bedrock ridges, a shallow area with a 20–30 m drape of glaciomarine sediments, the deep trough aligned with North Bay containing thick glaciomarine sediments, the northern part of the moraine, and acoustically stratified bedrock outside the moraine. It is likely that this moraine formed ca. 14 ka when formerly more extensive ice in the Gulf of St. Lawrence retreated to the modern coast.

The Bay of Islands moraine is highly complex in detail. It is lobate, with the lobes extending seaward from gaps between the islands. The surface has an irregular pattern consisting of ridges with a relief of 5 m, most of which are imprinted by random iceberg furrows of varying width (Fig. 99). The furrows tend to be narrower (20 m) and shallower (1 m) than those farther offshore. Near the deep interisland troughs, ramps up to 10 m high are interpreted as grounding-line wedges, formed by the last ice to emanate from the fiord; however, one small ramp-like feature intersected by the survey line of Figure 98 is composed on stratified sediment that thins landward, perhaps reflecting spillover of sediment into the deeper water due to wave action.

Deglaciation of Bonne Bay and Bay of Islands

The arcuate submarine moraines were long recognized, but were assumed to be to mark the limit of late Wisconsinan glaciers in the region (cf. Grant, 1987). Shaw (2003) showed the ice margin that created the Bonne Bay and Bay of Islands moraine, arguing that it formed after retreat of ice that was formerly more extensive (Shaw et al., 2006d). The earliest phase at these moraines would correspond with the 'end lower till ice retreat' phase of Josenhans and Lehman (1999), dated 13.7 ka. Grant (1987) mapped the moraine off Bay of Islands, and showed that the glacier had retreated from the outer coast by 13 700 ± 340 ^{14}C years BP, and was inland from the head of the bay at Corner Brook by 12 700 ± 300 ^{14}C years BP.

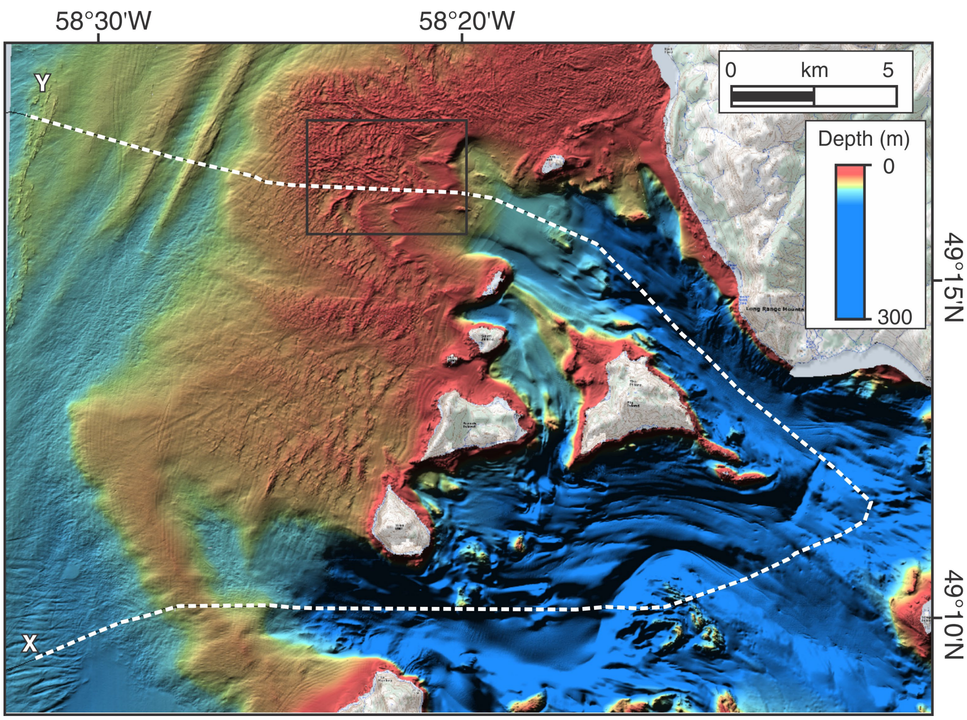


Figure 97. Enlarged view of the submarine moraine at the entrance to Bay of Islands. Also showing bedrock ridges to the west of the moraine (upper left corner). The dashed line shows the location of profile X-Y (Fig. 98). Box indicates location of Figure 99.

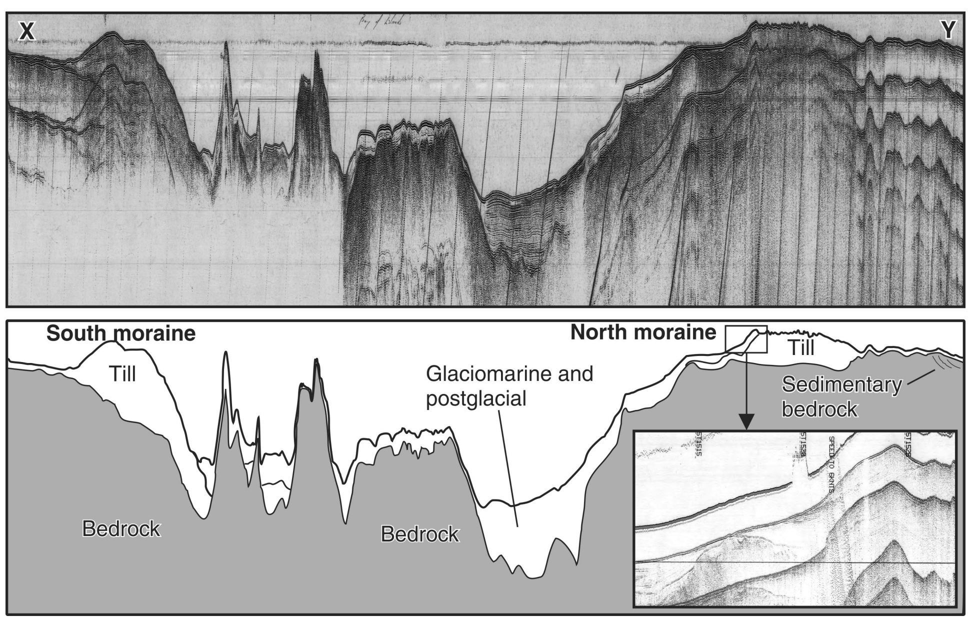


Figure 98. Low-frequency seismic-reflection profile (cruise *Baffin* 89008, Josenhans et al., 1989) that enters Bay of Islands from the west, crosses the deep outer fiord basins, and continues westward into the Gulf of St. Lawrence. Location is shown on Figure 97. Inset is a Huntec DTS record.

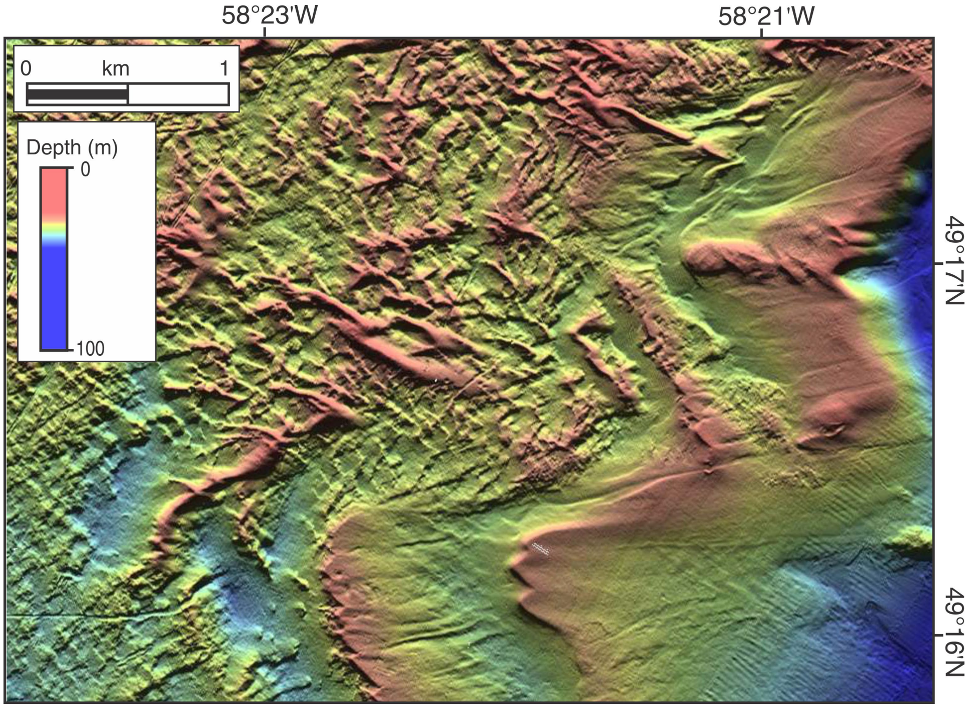


Figure 99. High-resolution shaded-relief image of part of the Bay of Islands moraine. Location on Figure 97.

Littoral and mass-transport processes at fiord entrances

Submarine fans at south entrance to Bonne Bay

The submarine moraine off Bonne Bay reaches the coast at Trout River. Just north of Trout River, in the vicinity of Green Gardens, marine sediments occur as a raised terrace (Berger et al., 1992), erosion of which has formed a gravel beach. Offshore from the beach lies a submarine platform (A in Fig. 100) that is deep and wide in the southwest (–20 m, 1.4 km) and shallow and narrow in the northeast, off Eastern Head (–15 m, 40 m). It has a steep (10°) seaward face and its surface is mostly rippled gravel, with patches of medium sand and scattered bedrock outcrops (B). The rippled gravel is testimony to continuing sediment mobility on the platform. Sparker records (cruise 97060) show a faint progradational structure, with clinoforms extending into the deep water. At the entrance to Bonne Bay, between Western Head and Eastern Head, the platform narrows and shallows, and is flanked by a series of submarine fans (C).

The lobate submarine fans extend about 1 km offshore, to depths of about 130 m. They are gently concave, with slopes of 3–4°. They consist of medium-fine sand, with small amounts of silt, and are incised by channels and levées with a relief of several metres on average. Most of the fans link with 10–15 m deep erosional embayments on the terrace, somewhat akin to the situation off Flat Island spit in St. George's Bay; in many respects they resemble the fans at the distal end of Flat Island spit. East of Eastern Head, the platform is very narrow and steep, and is notched by evenly spaced (~140 m) channels 2–4 m deep that fade out at depths of about 110 m (D). On sidescan-sonar records and on multibeam-backscatter imagery the channels have high reflectivity. The fans are interpreted as having formed as a result of sediment gravity flows generated by the suspension of sand on the littoral platform during storms. An attempt was made to detect change by comparing the 1997 and 2011 surveys; however, the poor quality of the 1997 data due to refraction made this difficult to do.

Submarine fans at north and south entrances, Bay of Islands

A situation similar to that at the south entrance to Bonne Bay exists at the north and south entrances to Bay of Islands — i.e. littoral platforms terminate in submarine fans. At the south side, the moraine (Fig. 101, A) reaches the coast at a headland called The Monkey. Sand (B) derived from reworking of the moraine by waves extends seaward to a depth of 40 m. The sand forms a littoral platform with a break of slope at –40 m. The east end of the platform terminates in a series of submarine fans (C) that extends into the fiord to a depth of 230 m. The largest fan has a channel with berms up to 7 m high and transverse ridges up to 2 m high. As at Bonne Bay, it is surmised that sand from the platform is moved east by wave action and carried down the fans by sediment gravity flows.

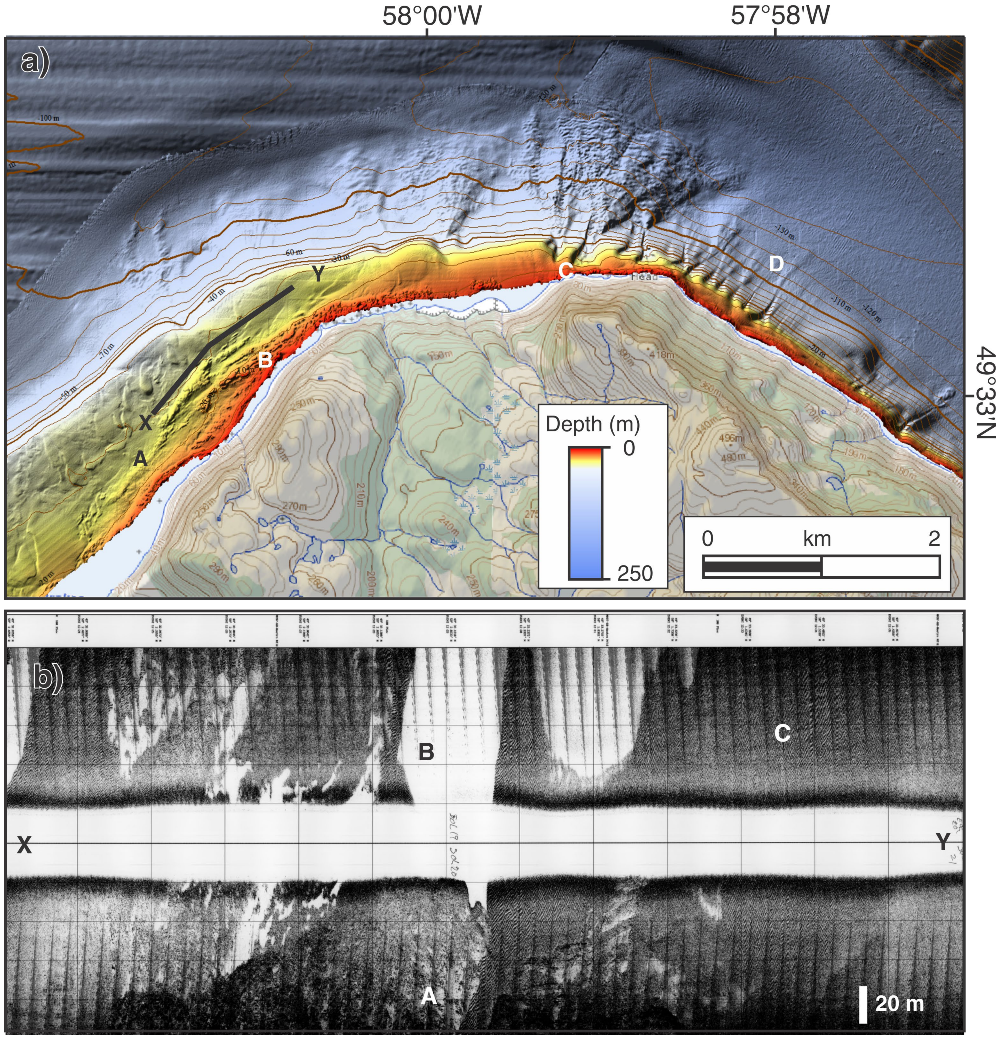


Figure 100. a) Submarine fans at entrance to Bonne Bay. A = sand; B = bedrock; C = amphitheatre at head of fan; D = small fans and channels; **b)** sidescan-sonar record along line X-Y, showing bedrock (A); sand (B), and rippled gravel (C). Location is shown on Figure 92.

Processes inside the fiords

Late Quaternary sedimentation in Humber Arm, Bay of Islands

Humber Arm contains thick glaciomarine sediments overlain by postglacial mud (Shaw et al., 2000a). The glaciomarine sediments are capped by a layer of red mud, probably derived from Deer Lake Basin, connected with the ocean ca. 12.2 ka (Batterson et al., 1993). Piston core MD99-2225 described by Levac (2003) was collected where 16 m of acoustically transparent postglacial mud overlies 6 m of weakly acoustically stratified ice-distal glaciomarine sediment and about 20 m of acoustically stratified glaciomarine mud. The oldest radiocarbon date of 10 980 ± 80 ^{14}C years BP was at 21.7 m in the core; a date at 14.4 m was 9120 ± 90 ^{14}C years BP. Clayey silt in this core from 14.7 m to 16.6 m had grey and reddish-brown banding that Levac (2003) associated with possible slumping from the fiord sidewalls. These dates confirm that high rates of sedimentation prevailed in the fiord in the postglacial period.

Late-glacial mass transport in Humber Arm, Bay of Islands

The steep sidewalls of Humber Arm are notched by numerous V-shaped gullies with relief of about 10 m. The gullies were formed by late-glacial slumping from the fiord sidewalls into the adjacent basin. The slumped material, likely glaciomarine mud, forms a series of stacked, acoustically incoherent lenses overlying acoustically stratified glaciomarine mud and beneath acoustically transparent postglacial mud (Shaw et al., 2000a). The approximate age of these events was estimated at 12–11 ka, a time when relative sea level was rapidly falling from a high of +120 m in the

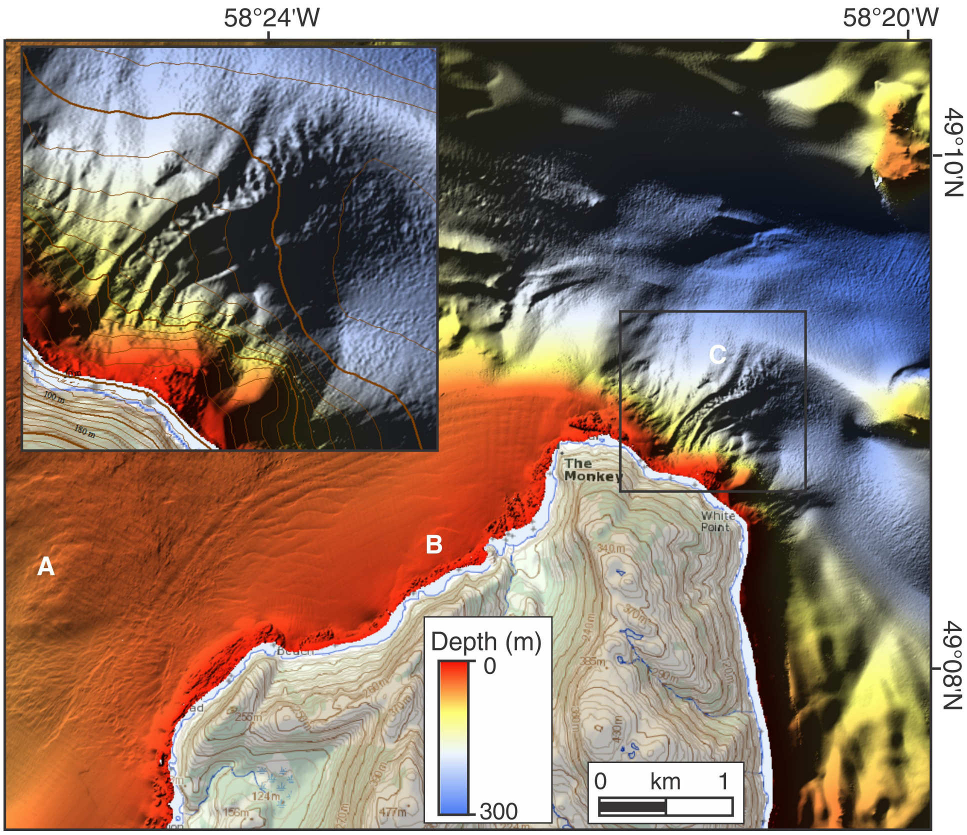


Figure 101. Submarine fans at south entrance to Bay of Islands. Location is shown on Figure 92. Letters indicate features discussed in the text.

area (Grant, 1994; Shaw et al., 2006d). The trigger, it was suggested, was earthquake activity caused by the rebound process.

Active submarine fans in East Arm, Bonne Bay

Bonne Bay (Shaw et al., 2002a) has two arms: the deep (230 m) East Arm and the shallow (130 m) South Arm. East Arm has steep sidewalls and a flat floor. Crossprofiles (cruise 97060) show that the bedrock valley has a V-shape profile, and that total sediment thickness is 100 m, of which the top 15 m is acoustically transparent postglacial mud and the lower 100 m acoustically stratified glaciomarine mud. Groups of overlapping fans with low backscatter are developed on the floor at a depth of about 220 m (Fig. 102). Numbers 1, 2, 5, and 7 slope upward to gullies on the fiord walls, whereas the other fans spread out up to 600 m across the fiord floor with relief of 1 m; one of them (number 3) is depressed in the centre and higher at the rim, much like the fan at Corner Brook. The fans appear to result from sediment transport down gullies on the fiord walls, most of which link up with stream outlets. The process is ongoing, since the fans lie on top of the most recent mud on the fiord floor. The low backscatter may mean either that they have a veneer of recent mud, or that they are composed of fine sediment. The range in morphology suggests that some may have resulted from submarine slides, whereas other relate to sediment transport down gullies seaward of stream outlets.

Active submarine fans, Bay of Islands

Apart from the fans linked with littoral platforms at the entrance to the fiord, at least 13 submarine fans occur in Bay of Islands. Four are located at river outlets, and may be fan deltas of the type described by Prior and Bornhold (1989). Eight fans are located offshore from steep slopes, with no stream outlets nearby, and have a blocky appearance. They may indicate slumping of material into the deep water. The fan at Frenchman's Cove (at the entrance to Humber Arm) appears to have formed offshore from a gravel foreland. The foreland is at the distal end of a shallow (–5 m) submarine platform, and may therefore relate to mass transport due to failures on the steep (23°) slope of the platform margin.

Fluid-escape structure, Humber Arm

A ring-like trench on the seafloor with a diameter of 300 m and depth of 4 m extends downward through the thick Quaternary sediments off Corner Brook (A in Fig. 103; Fig. 104). A zone of acoustic masking occurs 19 m below the centre of the trench, and sampling revealed methane gas in the sediment. It is likely that the feature formed as a result of gas venting at the seafloor, with a source deep down in

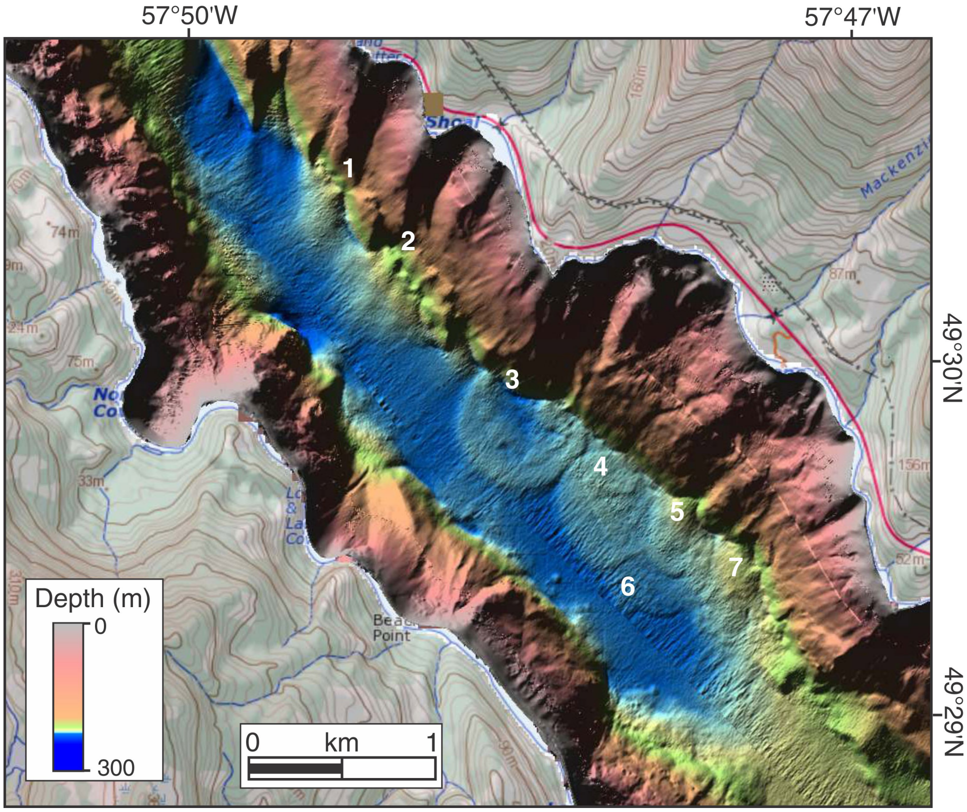


Figure 102. Fans numbered 1 to 7 on the floor of East Bay, Bonne Bay. Location is shown on Figure 92.

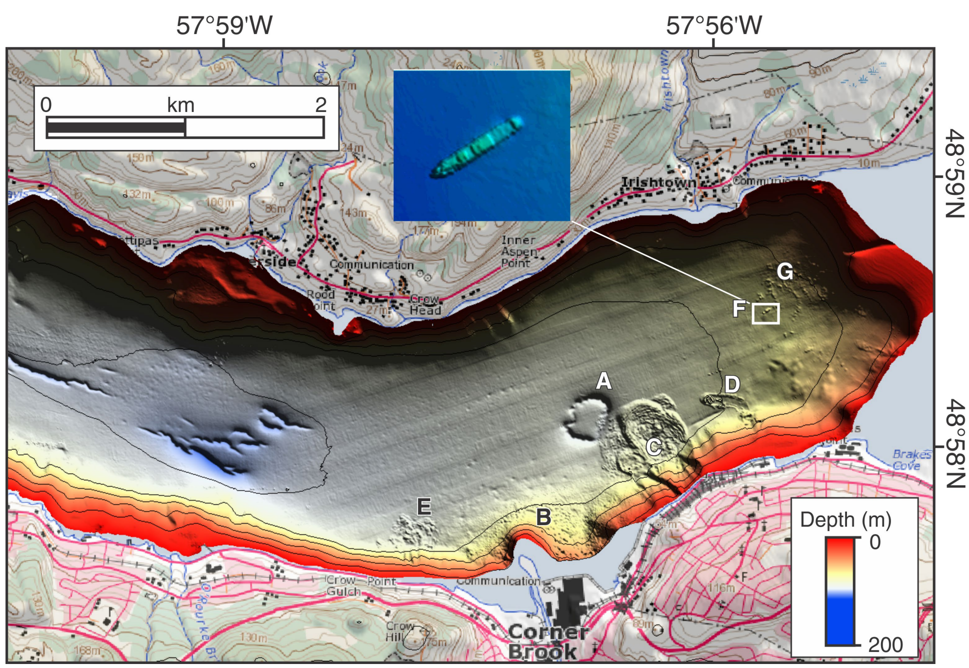


Figure 103. Anthropogenic effects near Corner Brook, Humber Arm. Inset shows multibeam image of the sunken barge (F). Location is shown on Figure 92. Letters indicate features discussed in the text.

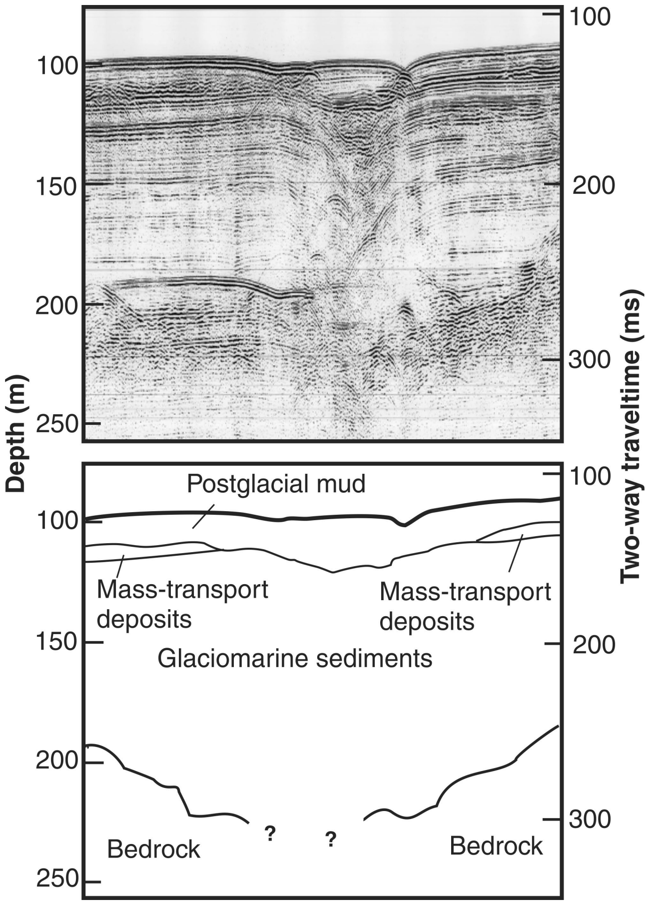


Figure 104. Sparker seismic-reflection profile showing fluid escape, Humber Arm. Adapted *from* Shaw et al. (2000a).

the Quaternary sediments. Late Proterozoic and Cambrian rocks outcrop either side of the bay here, although Early and Middle Ordovician limestone and dolostone occur on land at the head of the bay.

Anthropogenic effects in Humber Arm

Anthropogenic effects are evident in Humber Arm. A large area of the seafloor at Corner Brook is covered by debris from pulp and paper manufacturing (B in Fig. 103), creating a seaward protruding underwater ridge with uneven relief, high backscatter, and masking of acoustic subbottom profiler data. Nearby is a large submarine slide complex (C) produced when extension of the public wharves caused underlying glaciomarine sediment to fail and move into deep water. The complex includes several lobate slide masses, thicker at their rims (up to 4 m) than in their central parts. Together with the V-shaped gullies on the fiord sidewall, the slide complex may be an analogue for the late-glacial submarine slides farther down the fiord. A second slide to the east (D) travelled into deeper water and turned to the left, creating a pair of flanking berms approximately 1 m high. Other smaller slides (E) are linked with road building at the shore. Yet more indications of human impact are a sunken barge (F) at the mouth of Humber River, and numerous dredge spoils in the same vicinity (G).

Evidence of sea-level change

In Bonne Bay, the sill at the entrance to East Bay is trimmed to −15 m, suggesting that this was the depth of the postglacial relative sea-level lowstand. This conflicts with the estimates of Shaw and Forbes (1995 their Fig. 19) that suggest a value between 0 and −5 m. At York Harbour (Fig. 105) barrier beaches and small forelands are nourished by raised deposits of marine sediments in coastal areas. Offshore from York Harbour, submerged terraces and abandoned spits suggest a lowstand of at maximum −15 m. Together these data indicate that the isobase map of Shaw and Forbes (1995) underestimated the curvature of lowstand isobases off southwestern Island of Newfoundland.

REGION 14: PORT AU PORT BAY

Setting

This small region (Fig. 91) lies in the Humber tectonic zone. Rocks exposed on Port au Port Peninsula record the character of the Appalachian deformation front (Waldron and Stockmal, 1991). Offshore from the peninsula, the ‘Odd-twins’ magnetic anomaly extending in a northeasterly direction was found to be caused by magnetite-rich sandstone in the Late Ordovician Long Point Group (Waldron et al., 2002). The region is distinguished by very shallow water in the Port au Port Bay, and gentle gradients in the offshore. The long tapering peninsula of Long Point largely shelters Port au Port Bay from the effect of waves originating in the Gulf of St. Lawrence.

Coastlines

Relief is low, and coastlines are commonly sand or gravel beaches. A major feature is the delta of Fox Island River, south of which is a large barrier beach at Two Guts Pond.

Port au Port Bay

Lithostratigraphy

Port au Port Bay (Fig. 94) was surveyed by the GSC in 1988 (Forbes and Shaw, 1989) and detailed maps are contained in the report by Earth and Ocean Research Ltd. (1989). It contains a suite of Late Quaternary deposits resulting from glaciation, ice wasting, and postglacial fluvial and sediment supply. Total Quaternary sediment thickness exceeds 50 m in East Bay (*see* Fig. 109), reflecting glacial sediment supply and proximity to the dominant fluvial source, Fox Island River. Forbes et al. (1993) identified

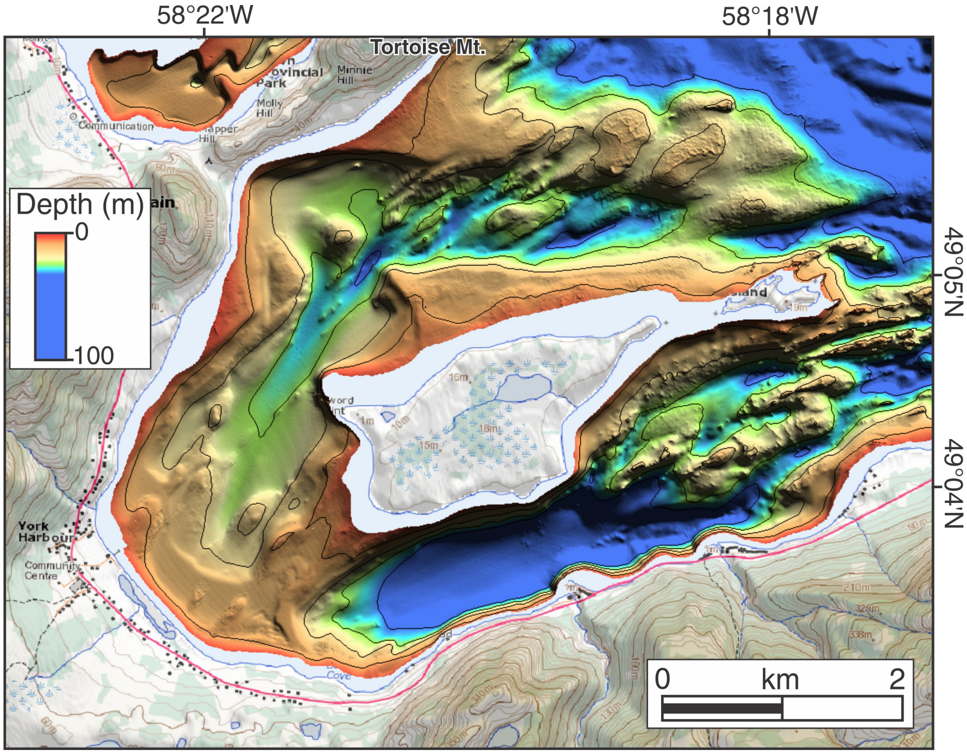


Figure 105. The York Harbour area showing evidence of submerged coastal deposits. Isobaths are at 10 m intervals. Location is shown on Figure 92.

six seismostratigraphic units: 1) till and other acoustically incoherent ice-contact sediments up to 40 m thick; 2) ice-proximal acoustically stratified deposits; 3) a drape of acoustically stratified glaciomarine sandy silt and clay up to 45 m thick; 4) acoustically transparent postglacial marine and/or lacustrine mud, masked by gas where more than 8 m thick; 5) postglacial marine and/or lacustrine sand and gravel, including foreset and topset beds, forming terraces down to –25 m; and 6) mid- to late Holocene transgressive deposits of sand, including a progradational or spillover wedge near the tip of Long Point.

The simplified surface sediment map (Fig. 106; *see also* Fig. 4 *in* Forbes et al., 1993) shows several muddy basins, with sediment grain size coarsening with decreasing water depth. More detailed versions in Earth and Ocean Research Ltd. (1989) show patches of gravelly boulders that represent seafloor glacial diamict and bedrock outcrops off Shoal Point and north of Fox Island River. The age of glaciomarine sedimentation is constrained by a series of radiocarbon dates ranging back to 13 710 ± 230 ^{14}C years BP (conventional age) (Forbes et al., 1993).

Marine placers in Port au Port Bay

Placer potential exists in the vicinity of Fox Island River. Emory-Moore et al. (1988) demonstrated heavy-mineral concentrations in the valley of Fox Island River, with the highest concentration at the beach. Minerals present included chromite and platinum. The work by Forbes et al. (1993) demonstrated the potential for chromite to occur offshore from the river mouth and this was borne out by further sampling (Emory-Moore et al., 1992).

Evidence of relative sea-level change in Port au Port Bay

Various researchers have published on the relative sea-level history of the Port au Port–St. George's Bay area (Brookes, 1969; Brookes et al., 1985; Forbes et al., 1993; Shaw and Forbes, 1995; Bell et al., 2003, 2005), where relative sea level fell from +44 m to a postglacial lowstand of –25 m at ca. 9.5 ^{14}C ka BP before rising once more. The early Holocene relative sea-level lowstand formed the large lowstand delta off Fox Island River (*see* Fig. 8 *in* Forbes et al., 1993) that formed with a former water level of about –25 m. There is, however, the possibility that the relative sea-level lowstand was slightly higher than in adjacent St. George's Bay, and that the delta was graded to the level of a lake in East Bay, as the sill is also at approximately –25 m. West Bay was certainly a lake until its –14 m sill was overtopped ca. 6 ^{14}C ka BP. The paleogeography of the bay ca. 9.5 ka as depicted in Forbes et al. (1993) suggests that former lake sediments should exist in parts of the bay, and coring would provide more evidence of relative sea-level changes.

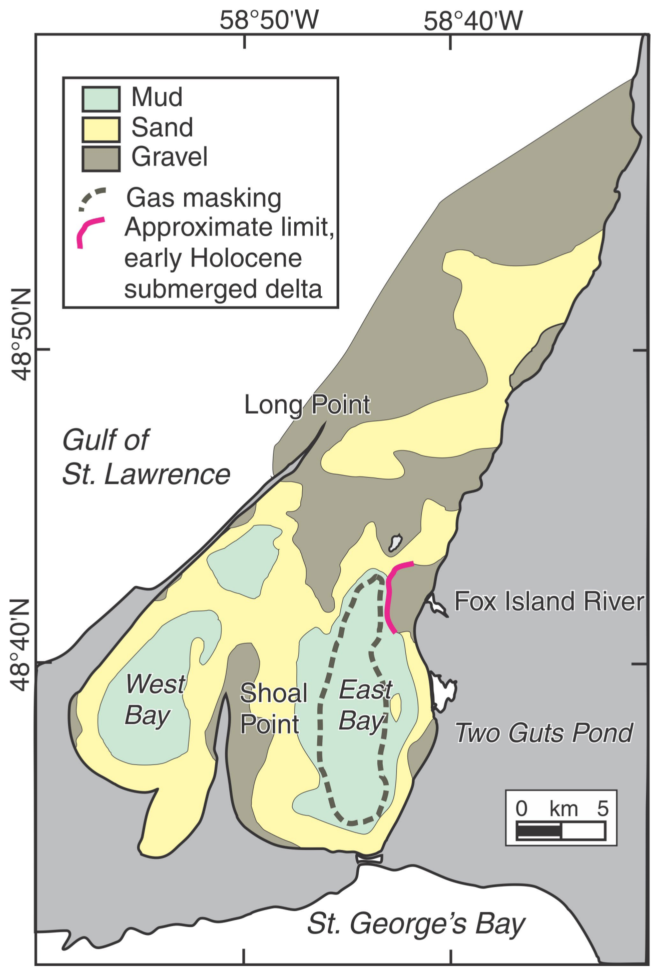


Figure 106. Simplification of surficial sediment distribution in Port au Port Bay. Detailed maps are contained in Earth and Ocean Research Ltd. (1989).

Inner shelf west of Port au Port Peninsula

Loring and Nota (1973) show a zone of gravelly, poorly sorted sand near the coast and 'gravelly pelitic sand' farther offshore. Maps by Josenhans and Zevenhuizen (1993) and Josenhans (2007) show reworked sand and gravel in a littoral zone, and till much farther out in the Gulf of St. Lawrence. Subsequently collected multibeam-sonar imagery (Fig. 107a) reveals a seafloor that is much more complex, and dominated by bedrock structure. At the north end of the coverage (Fig. 107b) bedrock ridges A and B correspond with the Odd Twins positive magnetic anomalies (Waldron et al., 2002), traceable on land nearby and caused by high magnetite content in sandstone of the Late Ordovician Long Point Group. The inset map (Fig. 107c) shows a fault in one of the two ridges. The northeasterly strike of bedrock continues to the southwest, where some of the deeper swales probably contain thicker deposits of postglacial sand and gravel, and the submarine bedrock ridges appear to intersect the coast, at least as far south as Three Rock Point, the southern limit of the Late Silurian Clam Bank Formation on shore (Waldron et al., 2002).

DISCUSSION

Complexity of the seafloor in the coastal regions

Previous seafloor maps, whether they adopt a formational approach to classification (e.g. Fader et al., 1982) or a genetic approach (Shaw et al., 1999b), are created by interpolation between geophysical survey lines that may be many kilometres apart. The point made by Figure 6 is that, in terms of geomorphology and texture, the seafloor appears to be vastly more complex on an image derived from multibeam-sonar mapping that on a map created by the earlier approach. Also, the complexity of the seafloor in coastal regions is in contrast to the lack of complexity farther offshore, particularly on The Grand Banks of Newfoundland, where relief is low, the wave-mobilized sediments are predominantly sand and gravel, geomorphic evidence of relative sea-level lowering (e.g. strand lines, barrier beaches) has been effaced (Shaw, 2005), and glacial landforms have either been destroyed or have subdued relief.

Implications for the glacial history of the Island of Newfoundland and the adjacent shelves

The problem of the Northeast Newfoundland Shelf

This publication is restricted to coastal waters, but it is nevertheless relevant to integrate the findings with the wider context of the glaciation of both the Island of Newfoundland and the continental shelves offshore from the coastal regions, given that the signature of glaciation is very strong at the seafloor in most areas. Shaw et al. (2006d) showed the distribution of late Wisconsinan glacial ice at the last glacial maximum and also a series of vignettes that depict the retreating margins at 20 ka, 18 ka, 16 ka, 14 ka, 13 ka, and 12 ka (i.e. conventional ^{14}C years). In their conceptual model for the deglaciation of Atlantic Canada, a major role was played by ice streams. The ice margin lay close to the edge of the continental shelf at the last glacial maximum. Ice streams occupying shelf troughs comprised: 1) a major ice stream in the Laurentian Channel; 2) secondary streams in the Bay of Fundy–Gulf of Maine, Trinity trough, and Notre Dame Channel; and 3) lesser ice streams elsewhere. Early retreat in the northeast (Notre Dame Bay) and in the southwest (Bay of Fundy–Gulf of Maine) was mainly by calving along embayments. By 18 ka a calving embayment had opened on Emerald Basin (Scotian Shelf). Retreat along channels marooned ice on intervening banks. A major calving episode beginning just before 14 ka removed ice from the Gulf of St. Lawrence, triggering re-advances down secondary troughs. Isolation of an ice cap on the Island of Newfoundland produced radial drainage via fiords and more extensive flow

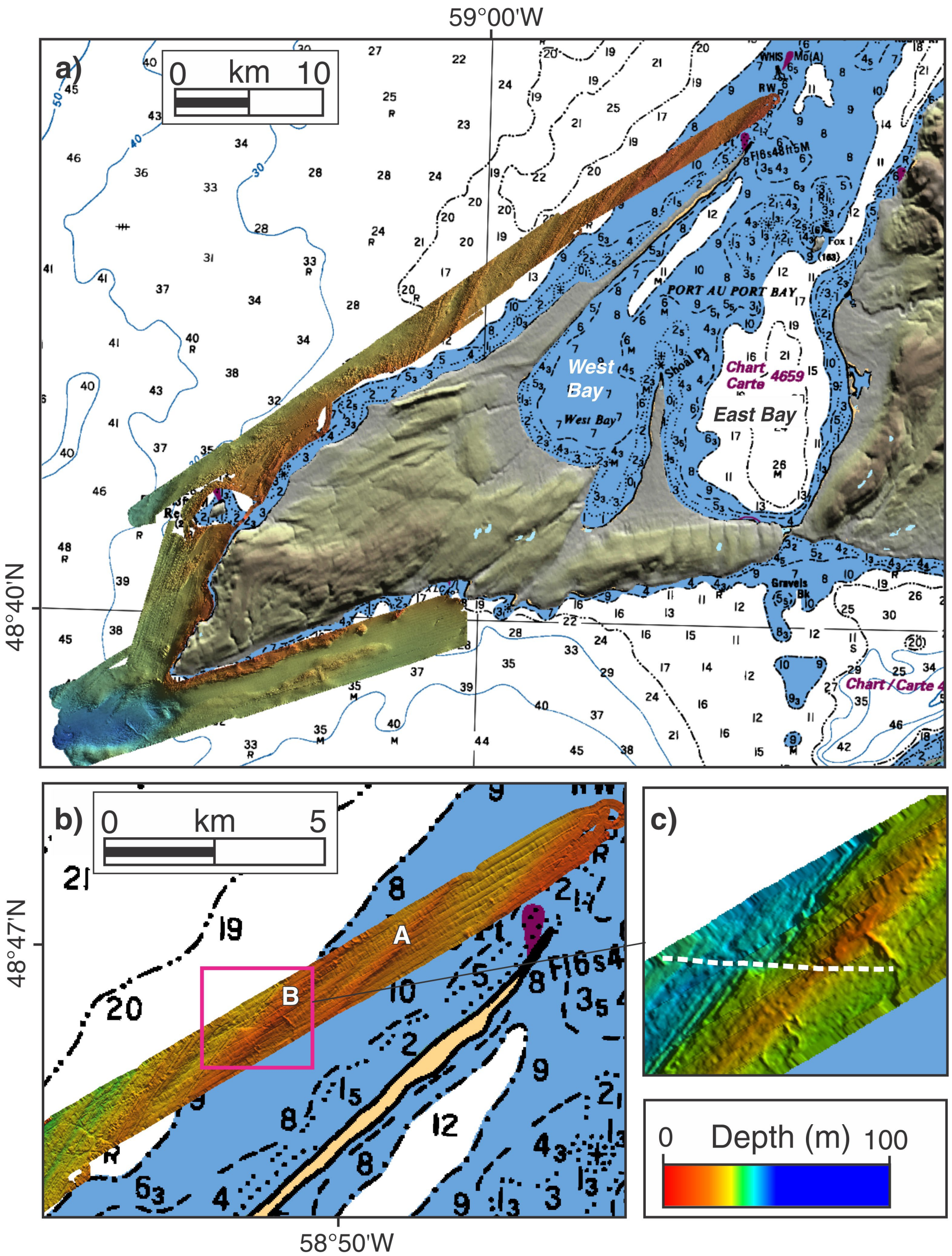


Figure 107. a) Extent of multibeam-sonar coverage on the inner shelf, outside Port au Port Bay; **b)** enlargement of part of the coverage to show northeast-aligned bedrock ridges; **c)** enlargement of part of Figure 107b, showing a fault (trace in dashed white line) with offsets. Background is Canadian Hydrographic Service (2003b).

into the St. George's Bay lowlands on the west coast. By 13 ka most ice was on land, and was ablating by melting rather than calving. Shelf ice caps persisted until 11 ka. This picture is shown in simplified form on Figure 3.

In retrospect, this attempt at synthesis has some weaknesses. For example, E.L. King and G.D.M. Cameron (pers. comm., 2013) have subsequently argued that late Wisconsinan ice on the The Grand Banks of Newfoundland was more restricted in its extent than Shaw et al. (2006d) suggested. Also, Huppertz and Piper (2009) depict the Oxygen Isotope Stage 2 margin around the Island of Newfoundland not as extensive on The Grand Banks of Newfoundland as in earlier stages. On the other hand, more recent work shows that ice margins were located at the mouths of Hawke Saddle, Notre Dame Channel, and Trinity trough not only at the classical last glacial maximum, but earlier (Roger et al., 2013).

A more serious problem was the absence of chronological data pertaining to the Northeast Newfoundland Shelf (Haworth et al., 1976), acknowledged in Shaw et al. (2006d) by their question mark on their last glacial maximum limit in this region. The chronology of retreat was based on dates from a single section of piston core 78023-020 (Dale and Haworth, 1979), from which it was inferred that ice retreat on this shelf appeared to be very early. Bulk samples of glaciomarine mud dated at 25.5 ka and 22 ka and believed to be too old were reassigned ages of 18–19 ka and 17 ka respectively, by Mudie and Guilbault (1982), although Scott et al. (1989) dated foraminifera from the same section of core at 21 800 ^{14}C years BP (conventional ^{14}C years).

It is pertinent, therefore, to assess any new data from the Northeast Newfoundland Shelf that might shed light on glacial history and thus provide context for the new findings in coastal areas. One source of new evidence is single-beam bathymetry data from Olex Ltd. (Fig. 108), collected by numerous vessels and compiled by Olex Ltd. As shown on Figure 1, the Northeast Newfoundland Shelf comprises a series of banks and troughs. With the exception of Hamilton Bank, strictly part of the Labrador Shelf, the banks are relatively deep compared with those off Nova Scotia, for example. Belle Isle Bank averages 270 m and Funk Island Bank averages 300 m. The banks are separated by troughs (Cartwright Saddle (maximum depth ~580 m), Hawke Saddle (575 m), Notre Dame trough (500 m), and Trinity trough (350 m). Analysis of the data from Olex Ltd. shows that much of the irregular terrain in the inner shelf is bedrock with thin Quaternary sediments. By contrast, a large area of smooth seafloor between Notre Dame Channel and Downing Basin, extending to the southwest and also to the northeast, onto Belle Isle Bank, has a cover of Quaternary sediments, typically more than 100 m thick, and up to 150 m thick (Fig. 108; *see also* figures in Shaw et al. (2012)).

The bathymetric imagery from Olex Ltd. contains evidence of glaciation. Megascale glacial lineations on the southeast side of Hawke Saddle (Fig. 108) have a relief of about 15 m and a pattern that is splayed at the east end, suggestive of ice flow out of Hawke Saddle toward the edge of the continental shelf (A, Fig. 109). A series of wedge-shaped bodies at the west end of Hawke Saddle, in depths of 280 m to 540 m, comprise acoustically incoherent sediment,

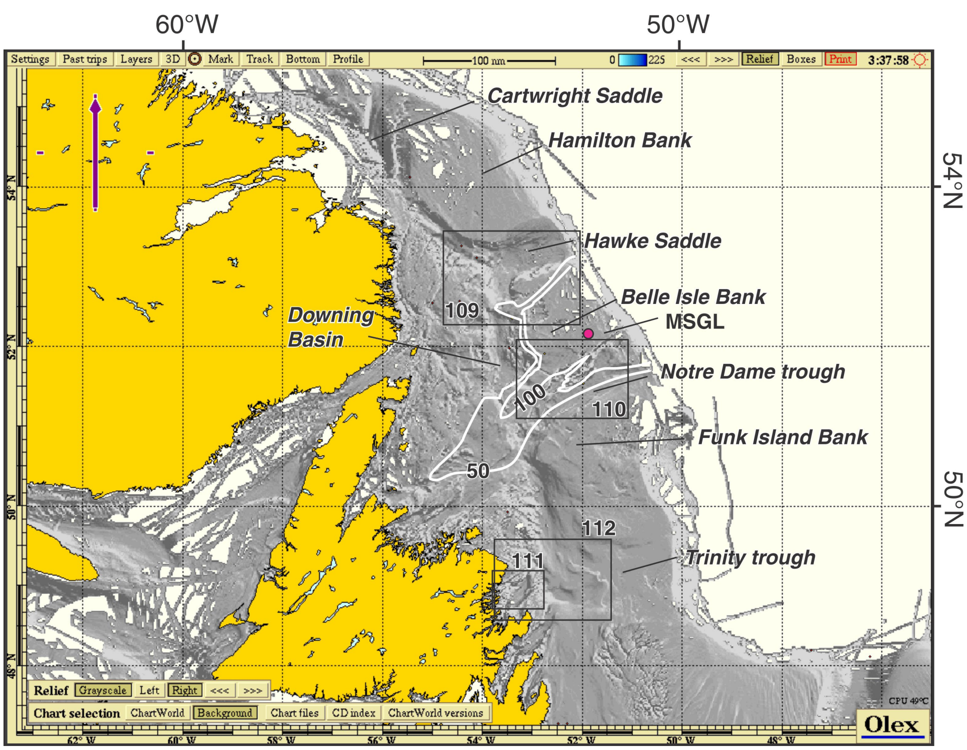


Figure 108. Image of the bathymetry off northeast Island of Newfoundland, exported from the bathymetric imagery system from Olex Ltd. Also showing thickness in metres of Quaternary sediments in the vicinity of Notre Dame trough and Belle Isle Bank. The red dot shows the location of ridges within the Quaternary section on an airgun seismic-reflection record. These ridges are interpreted as megascale glacial lineations, and indicate fast-flowing ice. Also showing locations of Figures 109–112.

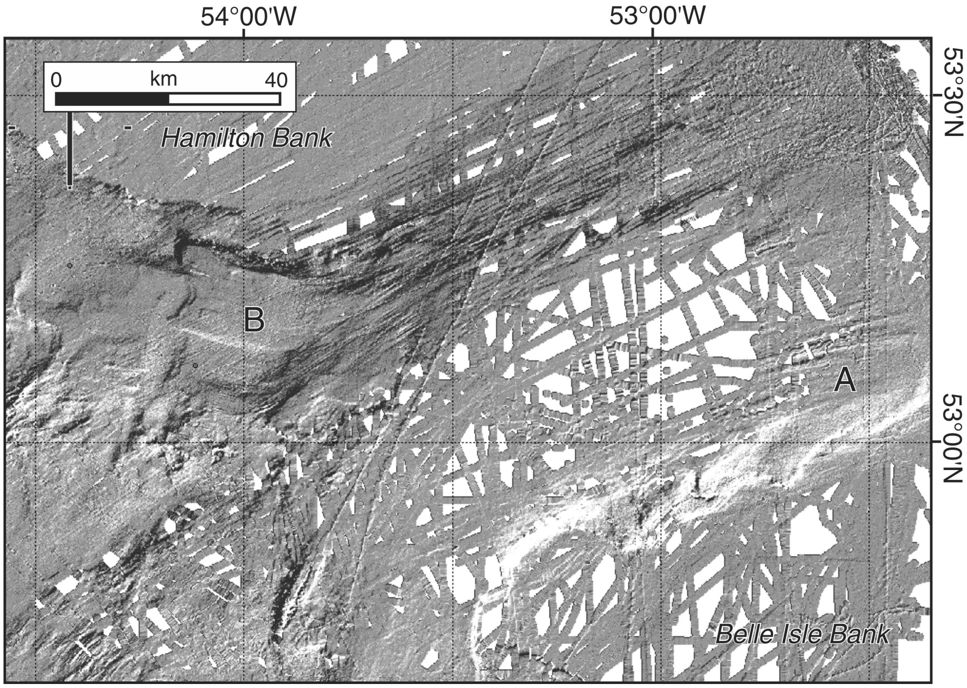


Figure 109. Bathymetric imagery in Hawke Saddle from Olex Ltd., showing megascale glacial lineations on the southeast side (A). Wedge-shaped sediment bodies at the west end of Hawke Saddle (B) are interpreted as grounding-line wedges. Location is shown on Figure 108.

probably glacial diamict (B, Fig. 109). They are interpreted as grounding-line wedges, formed when ice occupying Hawke Saddle retreated to the west.

An area of complex morphology in Notre Dame Channel that includes a 120 m high escarpment (A in Fig. 110), transported sediment bodies (B, C), arcuate berms (D), and mass-transport deposits (E), was interpreted by Shaw et al. (2012) as having been formed by a combination of glaciotectonism and submarine mass transport. It is noteworthy that at the edge of the escarpment, a berm is present, with relief of 50 m. Two curving ridges (F) to the south of this complex morphology have relief up to 40 m. They may have formed at the margin of ice flowing out of Notre Dame Channel.

Ridges on the floor of the trough outside Bonavista Bay (Fig. 111), in waters depths of about 400 m, have relief of about 5 m, and are interpreted as megascale glacial lineations formed by grounded glacial ice flowing to the northeast. They are a continuation of the field of megascale glacial lineations observed on multibeam sonar (Fig. 64). They lie inshore from a bedrock shoal with depths of about 230 m. Faint ridges with relief of about 5 m on the floor of the trough in inner Trinity Bay may also be megascale glacial lineations formed by northeast-flowing ice.

A large ridge previously interpreted as the Trinity Bay submarine moraine (Cameron and King, 2011) is located in Trinity trough (Fig. 112). The bathymetric imagery from Olex Ltd. shows several subdued ridges immediately to the east, perhaps representing earlier grounding-line wedges formed when ice in the trough retreated to the west. Cameron and King (2011) dated the formation of the main moraine to 16.1 ka. Elsewhere on the bathymetric imagery from Olex Ltd., several lobate ridges in Notre Dame Channel, and a transverse ridge in this channel, west of Fogo Island, may represent ice margins formed during ice retreat along Notre Dame Channel.

A possible sequence of events inferred from this evidence, combined with geophysical data from various GSC surveys, is indicated on Figure 113, and is as follows.

In early phases of the Quaternary, ice flowed toward the shelf edge in a broad trough between Hamilton Bank and Funk Island Bank (red lines on Fig. 113). When the thickness of the stacked Quaternary sediments (Shaw et al., 2012) is taken into account, then this palaeotrough was as deep as Hawke Saddle and Notre Dame trough are today. Supporting evidence for fast-flowing ice in this palaeotrough comprises a series of ridges with relief up to 8 m in an airgun record from cruise 75018 about 70 m depth in a 120 m thick Quaternary succession. These are tentatively interpreted as megascale glacial lineations (Clark, 1993).

In the later Quaternary, fast-flowing ice occupied troughs along Notre Dame Channel and Hawke Saddle (Fig. 113), eroding the Quaternary sediments and creating steep sidewalls along the north side of Notre Dame Channel. This is an example of ice-stream switching, as on the Norwegian margin (Dowdeswell et al., 2006). Less dynamic ice may have been present on the intervening banks at this time.

Because of erosion in Notre Dame Channel, blocks of the stacked Quaternary sediment on the steepened flanks failed, and slid into the deeper water along a plane of décollement that was the upper surface of underlying Cenozoic sedimentary rock (Shaw et al., 2012).

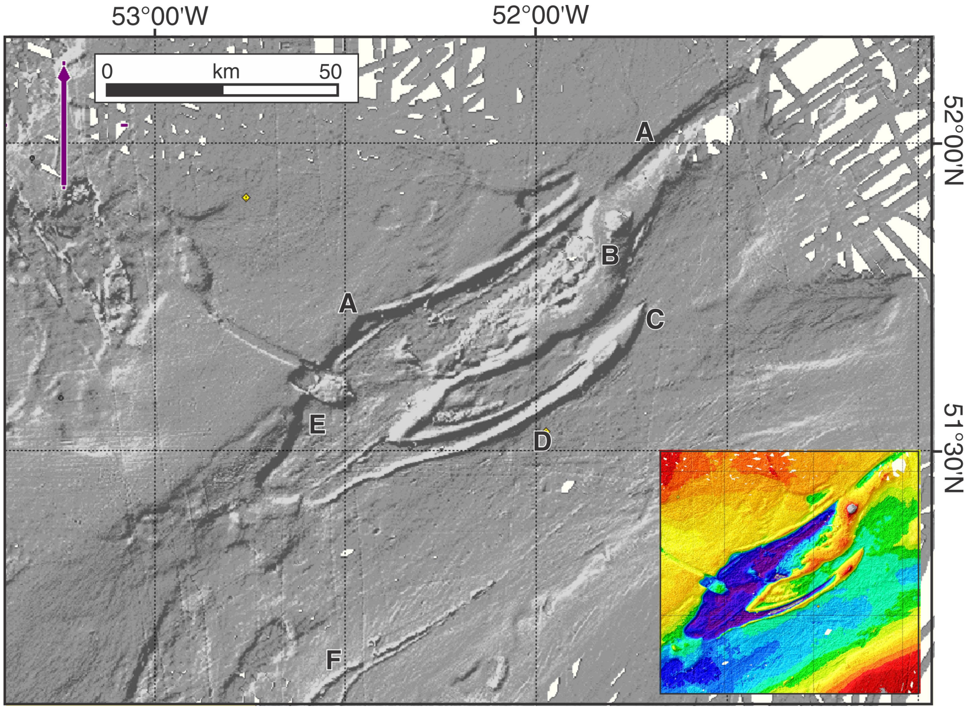


Figure 110. Bathymetric imagery of complex topography in Notre Dame Channel from Olex Ltd., interpreted by Shaw et al. (2011) as having been formed by a combination of glaciotectonic and mass-transport processes. The image at bottom right is a coloured shaded-relief rendering of the data. Location is shown on Figure 108. Letters refer to features discussed in text.

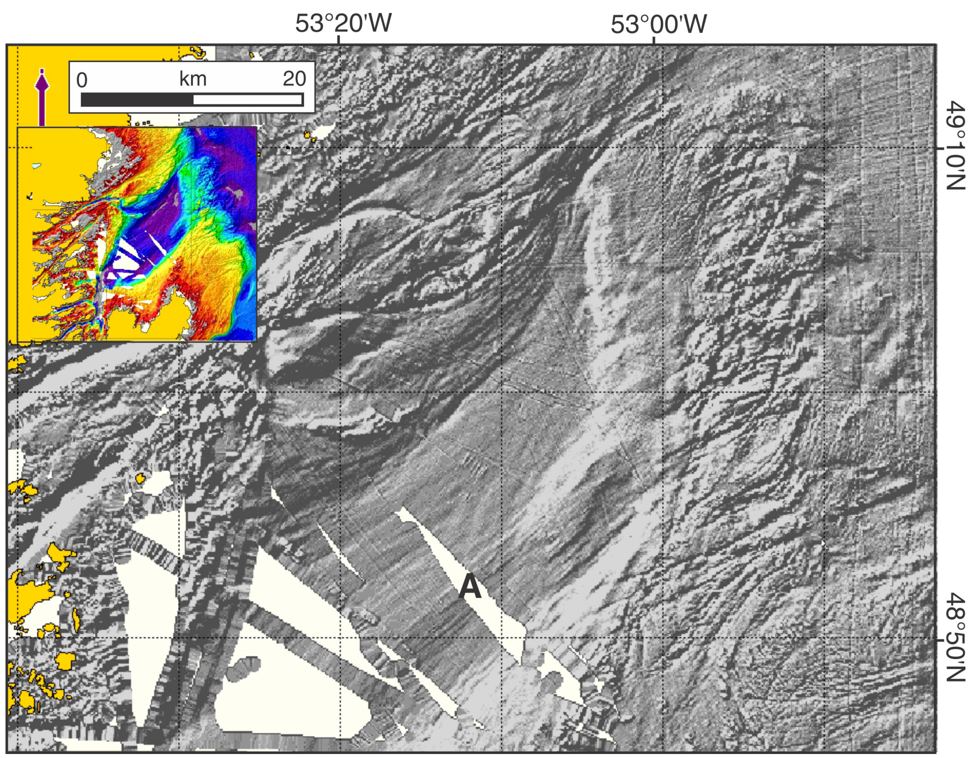


Figure 111. Bathymetric image of ridges (A) on the floor of the trough outside Bonavista Bay, Olex Ltd. interpreted as mega-scale glacial lineations formed by grounded glacial ice flowing to the northeast. The image at upper left is a coloured shaded-relief rendering of the data. Location is shown on Figure 108.

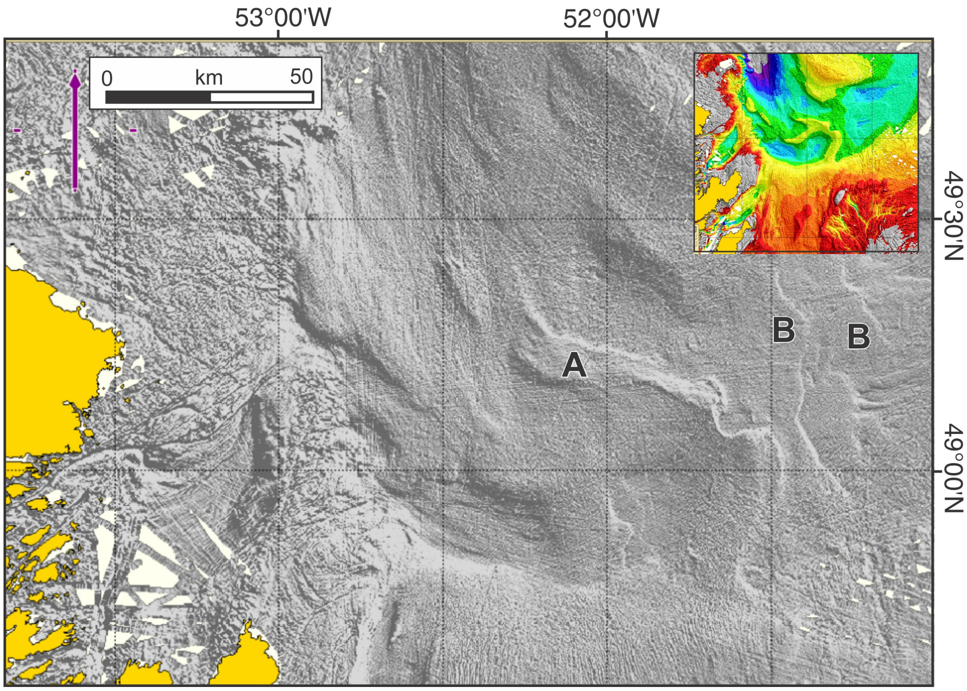


Figure 112. Bathymetric image of the Trinity Bay submarine moraine (A) from Olex Ltd., described by Cameron and King (2011). Several subdued ridges (B) immediately to the east perhaps represent earlier grounding-line wedges formed when ice in the trough retreated to the west. The image at upper right is a coloured shaded-relief rendering of the data. Location is shown on Figure 108.

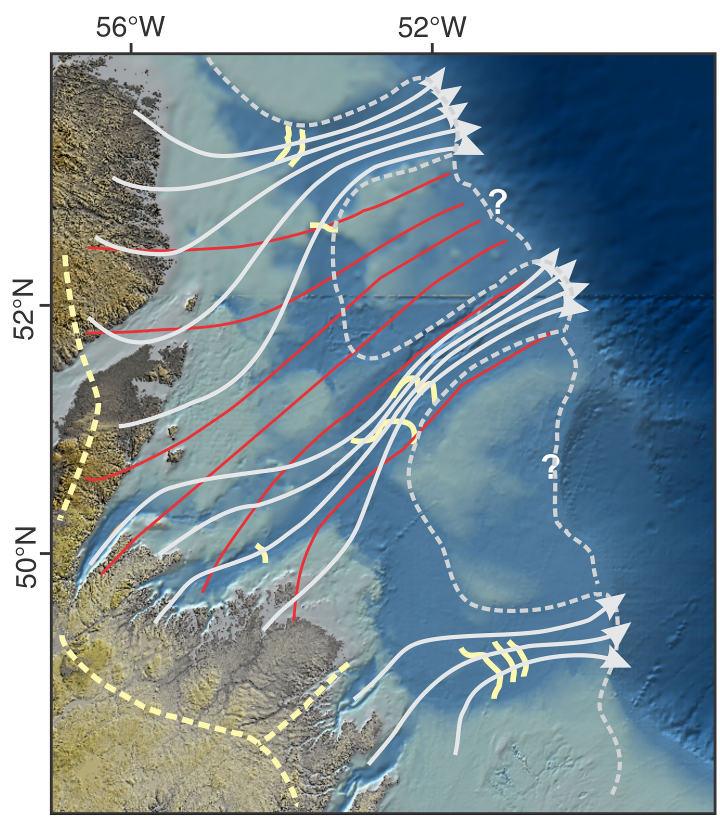


Figure 113. Shaded-relief image of the Northeast Newfoundland Shelf showing an early phase of ice flow (red lines), the most recent phase (white arrowed lines), probable moraines formed during the last ice retreat (yellow lines), the possible extent of ice at last glacial maximum (white dashed line), and the approximate position of glacial divides at last glacial maximum. The white dashed line with question marks indicates an alternative ice margin at last glacial maximum.

A readvance of fast-flowing ice down Notre Dame Channel impacted large-mass transport deposits, generating the glaciotectonized terrain (Fig. 110) as described by Shaw et al. (2012). The ice movement may also have been responsible for the berm along the crest of the failure escarpment and the creation of the pair of ridges (Fig. 110, F).

The retreating ice margin was landward of the site of core 78023-020 by 21.8 ka. The margin stabilized briefly at the west side of Hawke Saddle, at several locations in Notre Dame Channel, and at the Trinity moraine (ca. 16 ka BP). Renewed mass wasting of stacked Quaternary sediments on the flanks of Notre Dame Channel is evidenced by the rectilinear block described by Shaw et al. (2012).

This (somewhat speculative) model would suggest that at last glacial maximum, the margin was not at the shelf edge except at trough mouths (Fig. 113). This depiction has parallels with last glacial maximum margins farther north, off the coast of Labrador, described by Josenhans et al. (1986) and Josenhans and Zevenhuizen (1987).

Flow-parallel landforms and the last glacial maximum margin

The megascale glacial lineations, crag-and-tail features, and other streamlined landforms in inner Notre Dame Bay (region 10) and Bonavista Bay (region 8) support the model described for ice distribution at last glacial maximum suggested by Figure 113. These landforms occur in areas of strong convergence upstream of the fastest flowing ice. This convergence is also well demonstrated in Placentia Bay (*see also* Shaw et al., 2009). In all three areas, the streamlined bedforms offshore appear to be aligned with streamlined features on adjacent land areas, mapped from SRTM imagery (e.g. Liverman et al., 2006).

The idea that the presence of convergent streamlined landforms confirms the conceptual model of Shaw et al. (2006d) leaves an unanswered question, namely age and duration of the ice flow. Evidence from elsewhere suggests that terrestrial ice streams have short life spans. For example, the terrestrial Dubawnt Lake Paleo-ice stream operated only from 9000 a BP to 8200 a BP (Stokes and Clark, 2003). The question is, therefore, whether the inferred flowline patterns indicate flow at last glacial maximum, or whether they formed rapidly at some later stage in deglaciation.

Flow transverse landforms: stabilization of the margin at modern coasts

De Geer-style moraines in Placentia Bay (Shaw et al., 2009), and ribbed moraines in the Strait of Belle Isle (Woodworth-Lynas et al. (1992), indicate the incremental retreat of ice sheet margins just offshore from modern coasts, but as a rule, the evidence in most regions shows a stabilization of retreating ice margins near modern coasts ca. 14 ka (e.g. Shaw et al., 2000b; Shaw, 2003).

The multibeam data available for the present study confirms that the formation of large arcuate fiord-mouth moraines is largely a phenomenon of the west and southwest coasts. The new data for Bay of Islands and Bonne Bay, combined with previously existing subbottom profiling data, show in high resolution the architecture of the large arcuate moraines at the mouths of those fiords. An important distinction between these moraines and those of the southwest coast is that they do not pass laterally into acoustically stratified glaciomarine sediments, probably because they terminate on a very shallow (~80 m) shelf. Since the large fiords of Notre Dame Bay are without arcuate fiord-mouth moraines, it was concluded by Shaw et al. (2000b; 2006d) and Shaw (2003) that the retreating ice sheet did not stabilize at fiord mouths, although it did halt part way down Bay d'Espoir (Shaw et al., 1999a). The recently available multibeam-sonar data confirm this finding, but additionally show that the retreating ice sheet halted once in Halls Bay, several times in Green Bay, and several times in Connaigre Bay. In all instances the halt was at a fiord constriction, and thus the model by Syvitski and Shaw (1995) is applicable.

Chronology

The earliest conventional radiocarbon dates from marine sediments in coastal waters that indicate deglaciation are 14 580 ± 90 ^{14}C years BP (Beta-88458) in region 4, southwest coast (Shaw et al., 2000b), 13 380 ± 90 ^{14}C years BP (TO-2399) in Notre Dame Bay (Shaw et al., 1999a), and 13 710 ± 230 ^{14}C years BP (Beta-88061/ETH-5040) in Port au Port Bay (Forbes et al., 1993). In St. George's Bay (region 1), Bell et al. (2003) reported shell dates from onshore extending back to 14 370 ± 100 ^{14}C years BP (TO-4356). These dates could be interpreted as supporting the conceptual model shown on Figure 7, with ice sheets extending far offshore at last glacial maximum, and retreating to modern coasts ca. 14 ka; however, the dates from the study area say little about the limit of ice at the last glacial maximum, the evidence for which is to be found on the outer edges of the continental shelves.

Revision of relative sea-level histories

Information from recent multibeam sonar surveys show that the broad pattern of relative sea-level change around the Island of Newfoundland as provided by Shaw and Forbes (1995) and Shaw et al. (2002b) remains largely unchanged. New geomorphic data from Holyrood, at the head of Conception Bay, indicate a lowstand depth of approximately −22 m, whereas extrapolation of isobases in Shaw and Forbes (1995) indicates approximately −18 m in that area. At several locations elsewhere, however, greater departures from the earlier estimates are evident.

In Placentia Bay (region 9) a submerged erosional platform revealed by multibeam-sonar surveys (Potter and Shaw, 2009c) off the west side of the Avalon Peninsula reaches depths of −40 m. This prompted Cameron and King (2011) to redraw isobases in southern Island of Newfoundland. Whereas it is to be expected that the lowstand on this platform would be lower than at the Argentia area (−16 m according to Shaw and Forbes (1995)), a −40 m lowstand seems excessive, because during a relative sea-level lowstand in such an exposed area, the process of platform formation — erosion of glacial landforms and infilling of depressions by sediment — may occur some unknown depth below mean sea level.

On the west coast of the island, at Bonne Bay, the lowstand estimate based on new multibeam-sonar data is −15 m, whereas the estimate of Shaw and Forbes (1995) is between 0 and −5 m. At outer Bay of Islands, the suggested lowstand of −15 m is, however, not greatly different from the previous estimate (−12.5 m). Together these data indicate that the isobase map of Shaw and Forbes (1995) underestimates the curvature of lowstand isobases off southwestern Island of Newfoundland: i.e. the isobases may curve more to the north, parallel to the west coast of the island.

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