Late Quaternary Glacial Geology of the Bay of Fundy,

Coastal Southwest New Brunswick

by

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ABSTRACT

The physiographic environment of the southwestern Bay of Fundy nearshore has been shaped by a complex history of glaciation, sea level fluctuations, and modern processes. Interactions between a disintegrating ice sheet and changing sea level produced sequences of glacial and glacimarine geomorphic landforms, providing evidence of former ice margin positions and sea level.

A total of 427 sub-bottom lines representing over 6,500 km of seismic track lines from the research vessels CSL Heron and CCGS Frederick Creed, over the 2007 to 2009 survey seasons, were examined for the Bay of Fundy area between the St. Croix and St. John Rivers. The landforms identified from the data include eskers, deltas, a moraine, slump deposits, glaciofluvial and outwash channels, pockmarks and a flute formed by localized readvance. Stratigraphy for the Bay of Fundy represents bedrock, till, stratified glacimarine, unstratified glacimarine sediment, sand and gravel, and modern Holocene muds. Natural gas obscures sub-bottom profiles for some areas due to acoustic masking. The sequence is divided by an unconformity at depths ranging from - 40 to - 89 m below relative sea level separating a lower sequence of bedrock and glacial/deglacial sediments from modern Holocene marine mud. This surface was associated with the late Pleistocene to early Holocene regression/transgression, which was the result of local isostatic rebound following deglaciation.

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CHAPTER 1

INTRODUCTION

Benn and Evans (1998) refer to the Quaternary period as having experienced as many as 21 glacial and interglacial cycles. The most recent period of glaciation for North America is termed the Wisconsinan glaciation, lasting approximately from 100000 to 10000 yrs ago. The period of time from approximately 10000 yrs ago to the present is a warming event called the Holocene. Much evidence of past glacial-interglacial cycles is eradicated by succeeding glaciations and most surface deposits and morphology are the result of the last glacial episode and recent events during the Holocene.

At the end of the last glacial maximum the Appalachian ice complex covered much of New Brunswick, the Bay of Fundy (BOF), and Nova Scotia, terminating at the edge of the continental shelf. Deglaciation began approximately 20 ka yr BP, beginning a complex interaction of glacial and oceanographic processes (Shaw et al. 2006). The interaction between the disintegrating ice sheet and the changing sea level produced sequences of glacial and glacimarine geomorphic landforms, providing evidence of former ice margin positions and sea level. The geomorphic landforms include (Figure 1.1) eskers, deltas, kettles and kames, moraines, till, erratics and drumlins, which have been documented by Seaman (2006) and others. Since deglaciation, these glacial landforms have been modified by both water and wind. In the BOF the increasing sea level over the last 10000 yrs, in conjunction with changing currents and the return to amplification of high tides over the last 7000 yrs, has modified and submerged glacial landforms (Dashtgard et al. 2007; Desplanque and Mossman 2004).

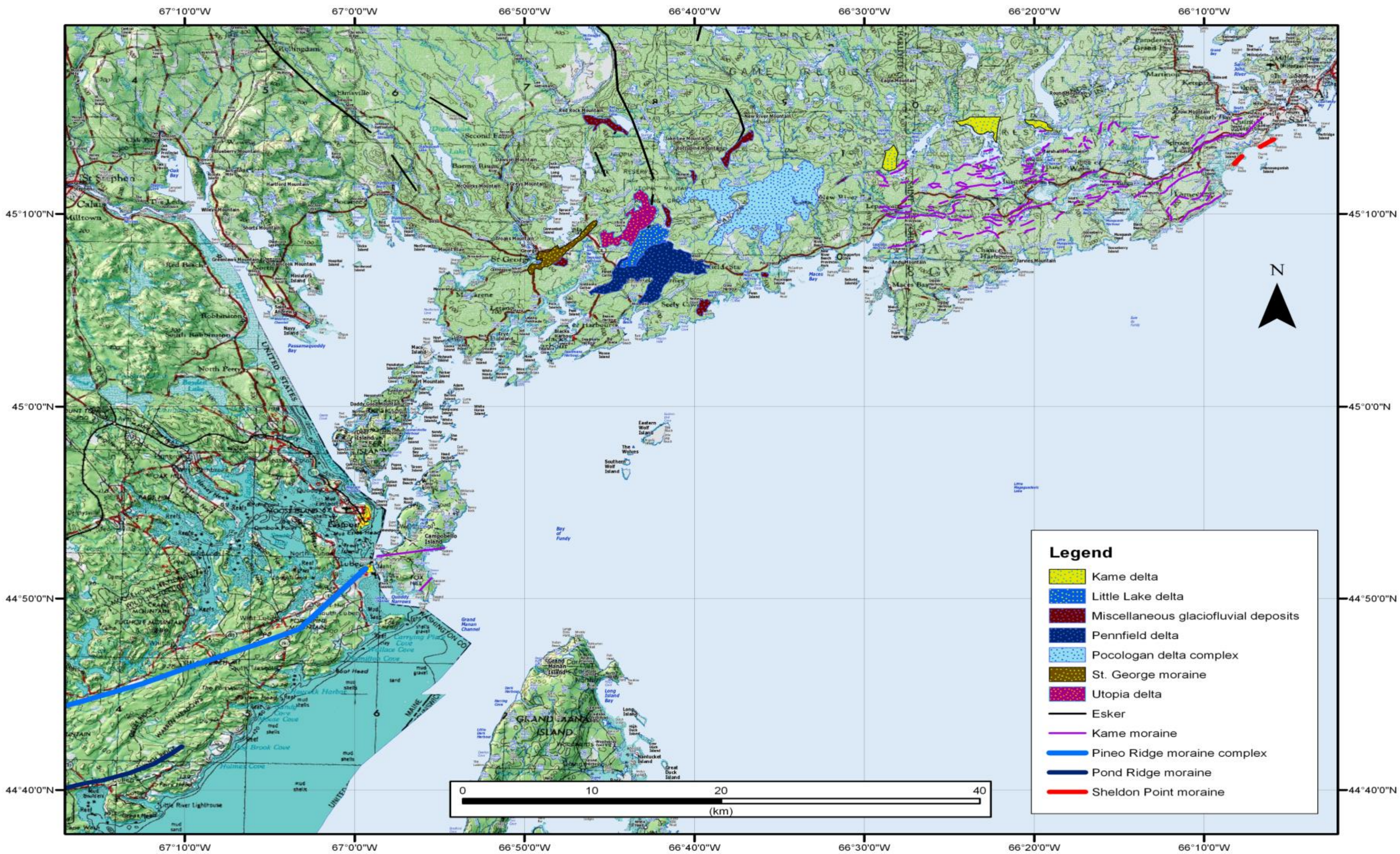

Figure 1.1: Location of study area in Bay of Fundy showing meltwater drainage patterns in southwestern New Brunswick and northeastern Maine (after Seaman 2006 and Kaplan 2007)

Extensive melt water complexes were deposited along coastal areas of New Brunswick during the late Wisconsinan deglaciation (Figure 1.1). The area exhibits geomorphological features formed during the Wisconsinan glaciation and deglaciation. For example, the Pennfield-Pocologan delta complex occurs along the northeastern coast of the BOF in the Maces Bay and Passamaquoddy Bay area, and part of this complex may extend into the Maces and Passamaquoddy bays (Figure 1.1). Extensive deltas and moraines are also found in neighboring eastern Maine. The Pineo Ridge Moraine Complex (PRMC) and the Pine Ridge Moraine (PRM) are prominent features along the northeastern coast of Maine, representing a major stillstand during deglaciation within a confined time frame (Figure 1.1). The PRMC extends into New Brunswick; the Campobello Island moraine segment is a continuation of the PRM found along the coast of Maine (Kaplan 1999).

In 2006 the Geological Survey of Canada Atlantic (GSC Atlantic), in conjunction with the Canadian Hydrographic Service (CHS) and the Ocean Mapping Group (OMG) at the University of New Brunswick, commenced a three-year program to map the BOF. Sub-bottom profiler and multibeam bathymetry and backscatter data were collected simultaneously to provide information on the character and thickness of the subglacial features and overlying sediments on the sea floor. There are many published works on the terrestrial Quaternary geologic record of southwest New Brunswick and its coastline (Rampton et al. 1984, Seaman et al. 1993, Seaman 2004, and others). As well the outer shelf surficial geology of the BOF and Gulf of Maine has been investigated (King and Fader 1986, Bacchus 1993, Barnhardt et al. 1997, Belknap et at. 2002, and others). However, publications on the nearshore submarine geology of the BOF are few.

1.1 Purposes and Scope

The objective of this study is to describe and interpret the nearshore Quaternary geology between Saint John and Grand Manan Island, now submerged along the north shore of the BOF. High-resolution seismic reflections profiles, collected during these cruises, are the primary database discussed here. The seismic profiles (listed in Appendix 1) were interpreted to delineate various seismic facies representing the glacial, glacimarine, and postglacial sediments in the study area, by comparing and contrasting high-resolution seismic reflection profiles using sub-bottom and multibeam bathymetry.

1.2 Location of Sub-bottom

The Bay of Fundy is located between New Brunswick and Nova Scotia, opening into the northeast end of the Gulf of Maine. The study area is located in the BOF along the inner/near shore area of southwestern New Brunswick (Figure 1.2), focused in the area between the St. Croix River and Saint John, New Brunswick, bounded on the north at 45° 27’, on the south at 44°44’, on the east at 65°45’ and to the west at 67°10’.

1.3 Physiography and Bedrock Geology

New Brunswick is located in the Appalachian physiographic region. The Appalachians extend from the northeastern United States and continues through Atlantic Canada to Newfoundland. New Brunswick is the largest and most northern of Canada’s Maritime Provinces, covering an area of 77,440 km^2 (NRCAN 2009). Bostock (1970) has divided New Brunswick into five main physiographic regions (Figure 1.3): the Chaleur Uplands, New Brunswick Lowlands, St. Croix Highlands, Caledonia Highlands

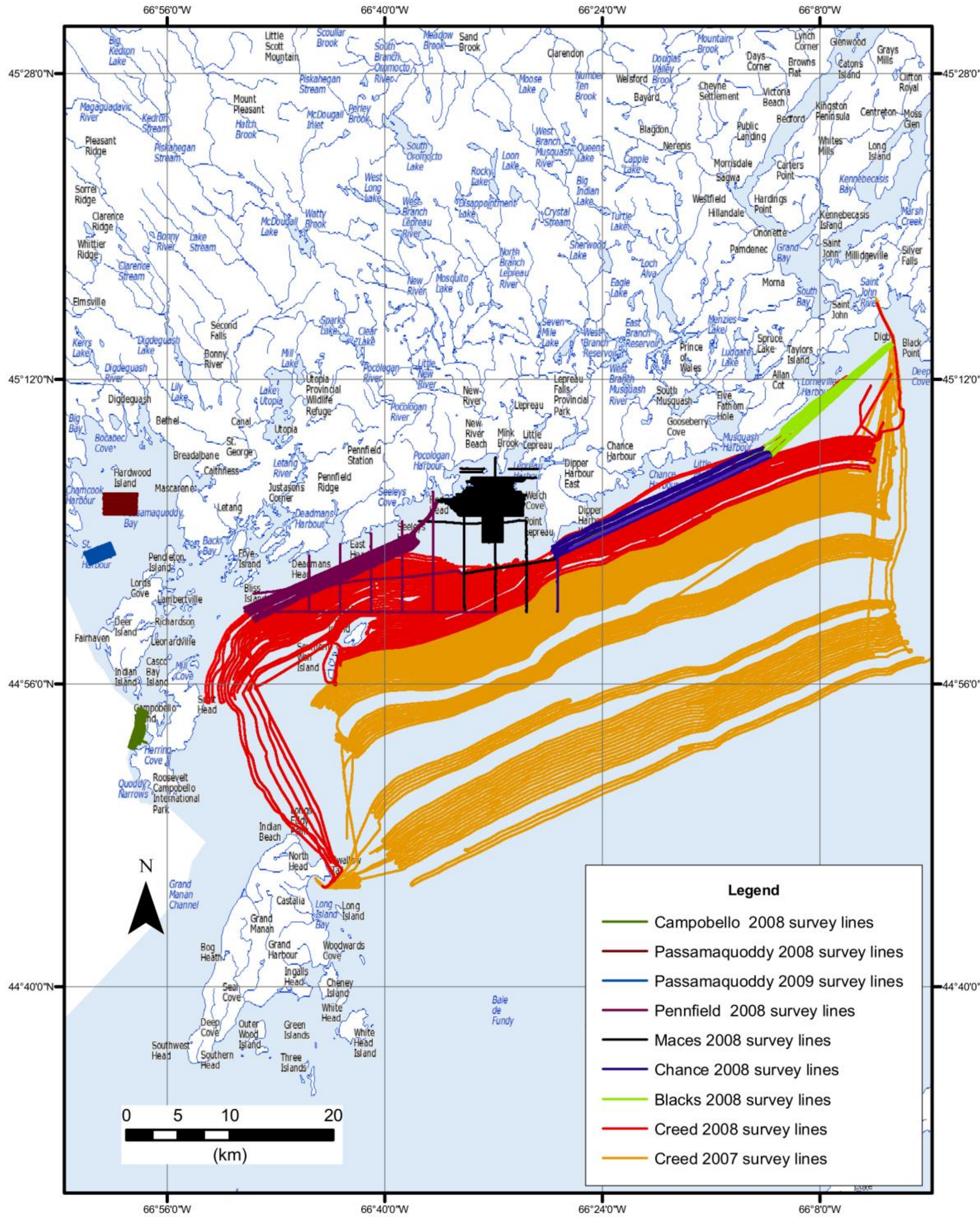

Figure 1.2 Location map of surveys. The surveys in the legend are presented in the order shown in the text.

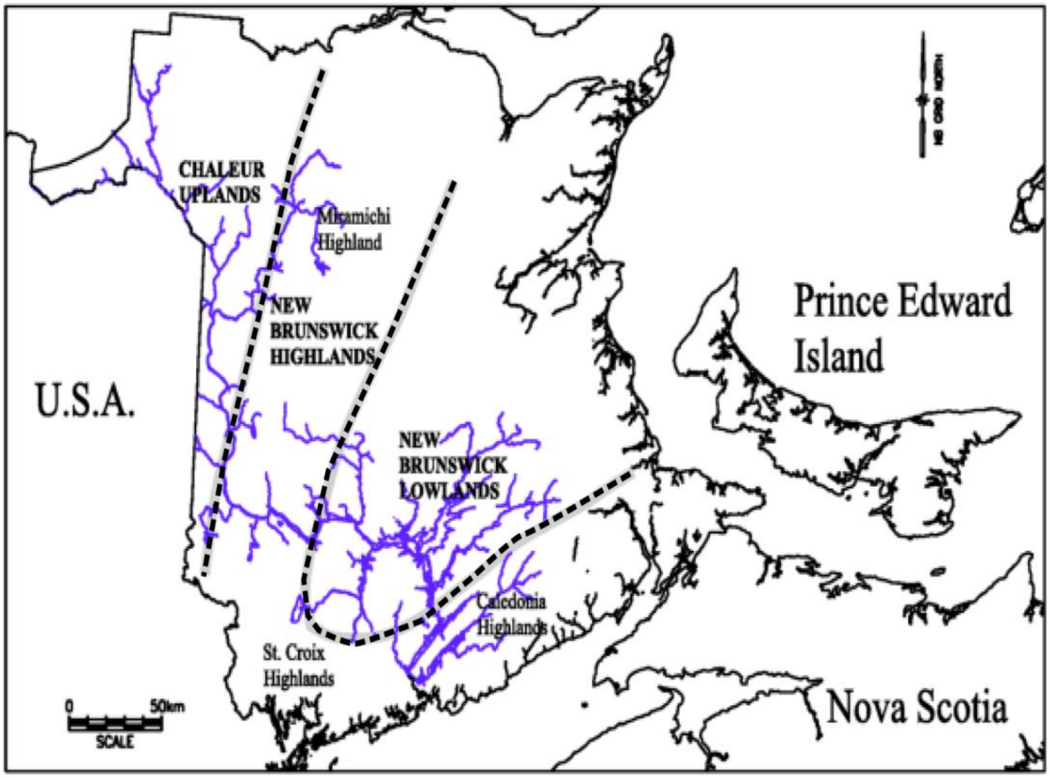

Figure 1.3: Physiographic regions of New Brunswick, delineated by black dashed lines (after Bostock 1970).

and the Miramichi Highlands. The Bay of Fundy lies along the seaward edge of the northern Appalachian orogen in the physiographic region called the Fundy Basin (Figure 1.4, King 1972).

Southern New Brunswick is characterized by a triangular basin of low relief, which is bordered by highlands to the west and south (Figure 1.3). The New Brunswick Lowlands generally lie below 120 m elevation (Gadd 1973). The lowlands are underlain by undeformed gently dipping Carboniferous rocks, primarily Pennsylvanian aged sandstones, siltstones, shales, and conglomerates (Seaman et al. 1993). The St. Croix Highlands occur at the apex of the lowlands' triangular basin and extend along the BOF coastal region to Saint John. The St. Croix Highlands reach an elevation of 300 m and are composed of steeply folded Paleozoic sedimentary and volcanic strata, interjected by granitoid rocks of Middle Devonian and Early Carboniferous age (Thibault et al. 1985). To the east of the St. John River along the Bay of Fundy lie the Caledonia Highlands, with folded Precambrian sedimentary and volcanic rocks reaching 300 to 600 m relief (Seaman et al. 1993).

The Bay of Fundy, a branch of the Gulf of Maine (GOM), is a linear funnel shaped water body trending northeast between New Brunswick and Nova Scotia in eastern Canada (Figure 1.4). The bay bifurcates at the northeastern end into Chignecto Bay and the Minas Basin. The BOF is renowned for having some of the highest vertical tidal ranges in the world; tidal ranges exceed 15 m at the head of the bay (Desplanque and Mossman 2001). This results from the wedge shape and the decreasing depth towards the top of the bay, resulting in near resonance with the Atlantic Ocean tides.

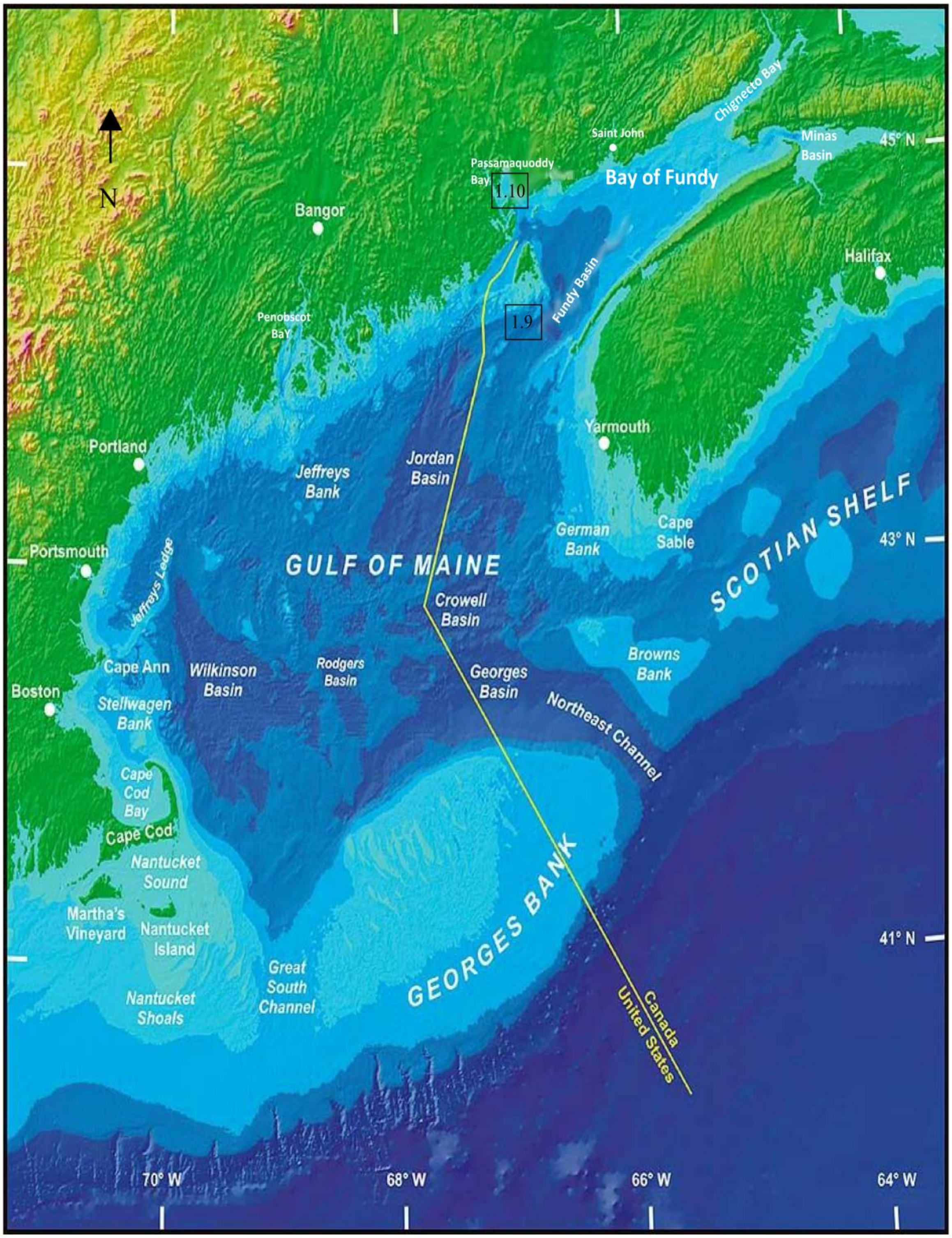

Figure 1.4: Gulf of Maine and Bay of Fundy physiographic regions. The rectangles show the location of subsequent figures in the text (after Cooper, 2009).

The tides move more than 100 km^3 of water in and out of the bay each day. Circulation within the bay results from tidal currents and the counterclockwise flowing gyre that is the dominant circulation force within the neighboring GOM (Desplanque and Mossman 2004). With over 1400 km of coastline, the bay is approximately 145 km in length, varying in width from 100 km at the base, tapering to 48 km wide at the northeastern end and with an average depth of 75 m (Tagg and Uchupi 1966; Swift and Lyall 1968; Fader et al. 1977; Desplanque and Mossman 2001).

The basement rocks of the BOF comprise two different terranes, the Avalon and Meguma terranes (Figure 1.5). The Avalon terrane is located along the New Brunswick side of the BOF and is characterized by low-grade metamorphic, igneous and sedimentary rocks. The Avalonian rocks consist of metamorphosed Mid-Late Proterozoic rocks, Late Proterozoic rifted cratonic volcanic arcs and younger rocks of Cambro-Ordovician, Silurian-Early Devonian and Triassic age (Uchupi and Bolmer 2008). The Meguma terrane is exposed along the southern shore of the BOF on the Nova Scotian side. It is characterized by a thick (>10 km) succession of Cambrian to Ordovician turbiditic meta-sandstones and shales overlain by thinner Silurian to Devonian volcanic rocks and shelf sediment (Wade et al. 1996).

The BOF is a fault bounded half graben, the Fundy Basin, with faults on the northwestern margin. The Fundy Basin was formed along the eastern margin of North America at the time of the breakup of Pangaea during the Mid- and Late-Triassic (Wade et al. 1996). It was initiated during the Appalachian orogeny, 286 – 300 Ma. yrs ago, with sediment infilling of the Fundy Basin beginning in the late Triassic and early Jurassic.

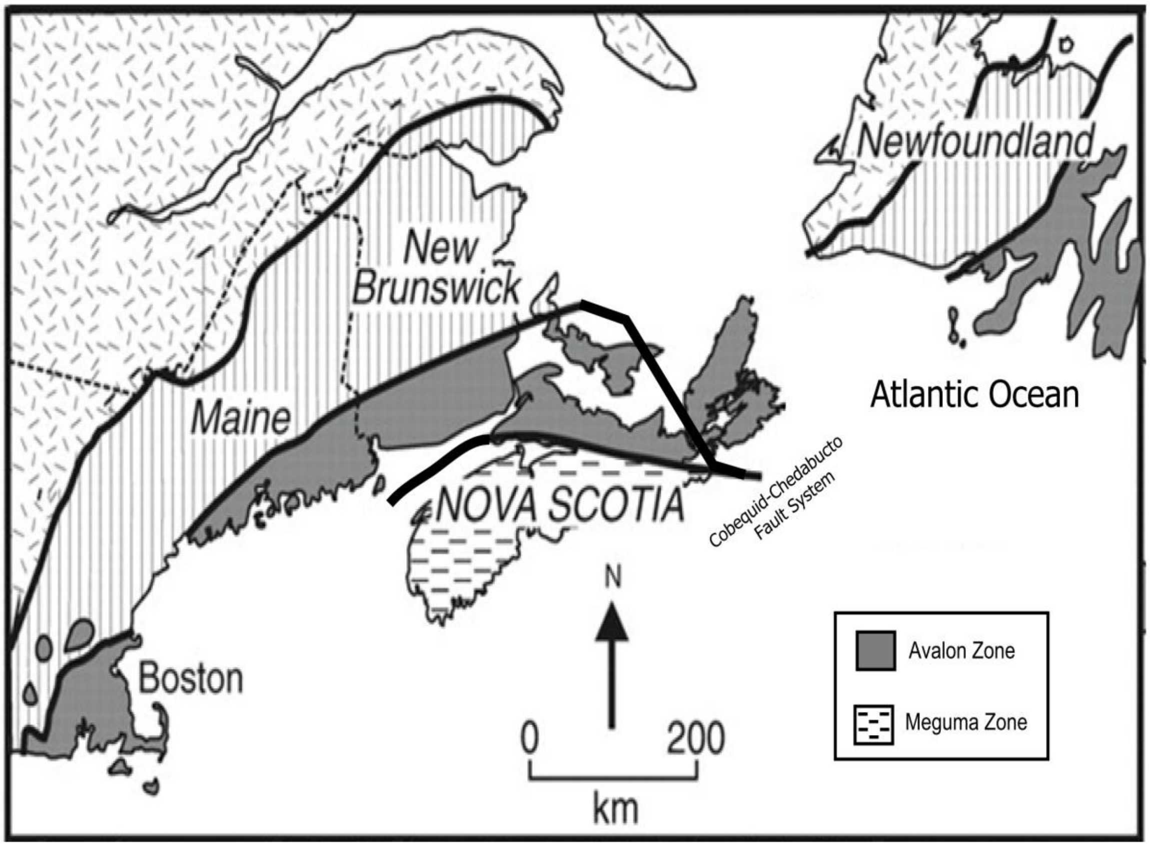

Figure 1.5: Components of the northern Appalachian Orogen, showing the location of the Avalon and Meguma terranes in Atlantic Canada (after Murphy et al. 2004)

Following this period of sediment infilling, the basin went through a period of volcanic eruptions, with basaltic lava laid over the Triassic strata (Desplanque and Mossman 2001). Succeeding the volcanic activity was primarily clastic sedimentation in low lying areas (Desplanque and Mossman 2001).

At the boundary between the Gulf of Maine and the Bay of Fundy is an archipelago of 20 islands (Figure 1.2); the largest of these is Grand Manan Island followed in size by Campobello and Deer Islands (Legget 1979). Grand Manan Island and the surrounding smaller islands lie at the junction between several major tectonic components of the Northern Appalachian orogen (Miller et al. 2007). The archipelago occupies nearly half the entrance to the BOF, which divides into two channels, Grand Manan Channel and Southeastern Channel (Hachey and Bailey 1952)

Several large rivers empty into the BOF (Figure 1.6), which receives the drainage from an area of approximately 64,000 km^2. The major inputs of fresh water include drainage from the Saint John area, the Passamaquoddy area, the Chignecto area, the Minas area, and the Annapolis area (Hachey and Bailey 1952). The St. John River is the largest input of fresh water into the BOF. Beginning on the border of the Province of Quebec and the State of Maine, it is approximately 673 km long and flows southward across New Brunswick, forming a deep trench of varying widths (Thibault et al. 1985; Hachey and Bailey 1952). The preglacial river outlet was at Saints Rest Beach, but the river was rerouted to its present location through the Reversing Falls during the Wisconsinan glaciation (Figure 1.6, Flaherty 1989). The river is an important part of the landscape of southwest New Brunswick, draining an area of more than 38,000 km^2 into the BOF (Hachey and Bailey 1952; Rampton et al. 1984; Thibault et al. 1985).

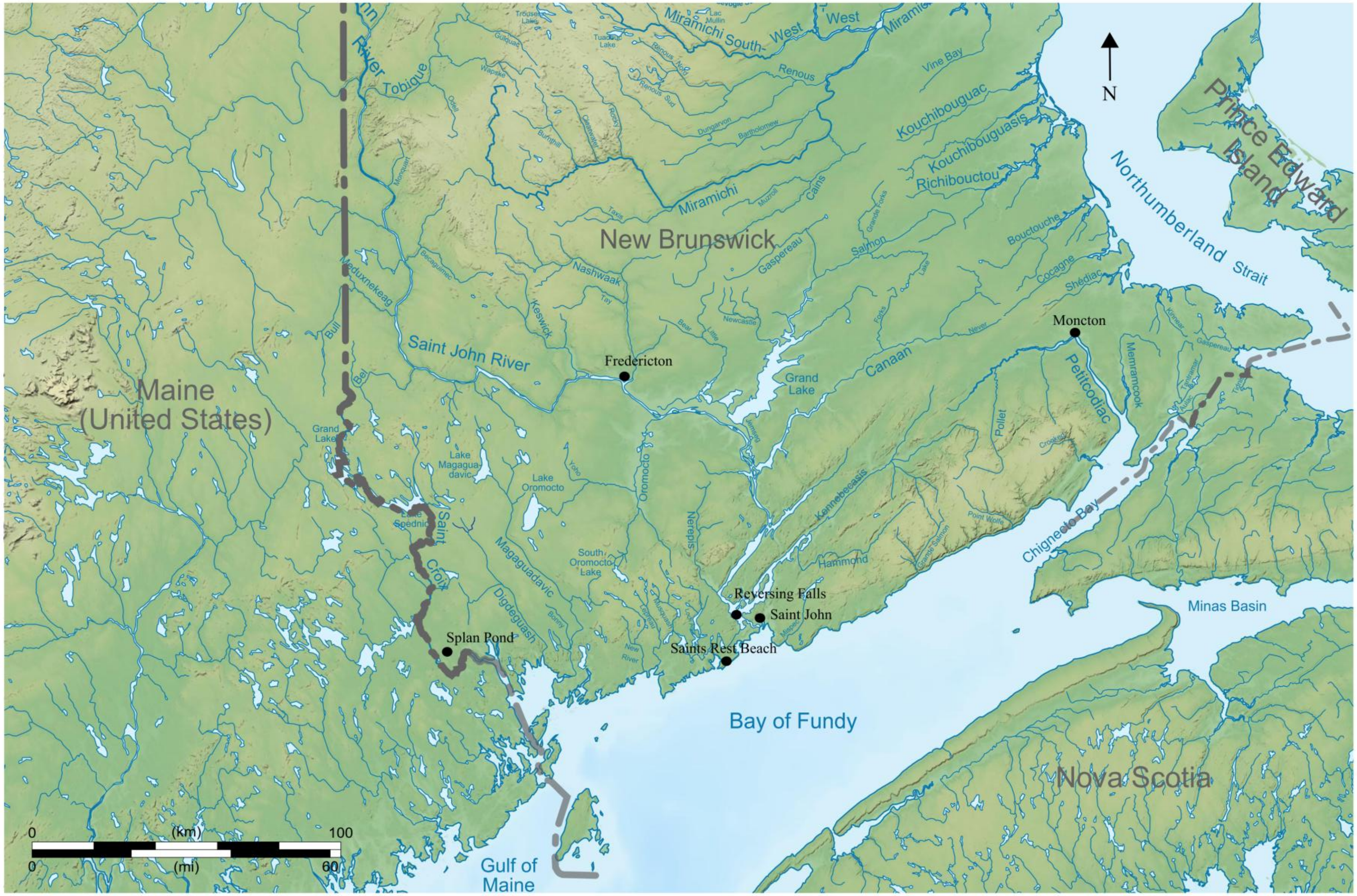

Figure 1.6: Rivers of southern New Brunswick (after Semchur, 2011).

The Passamaquoddy Bay area, located close to the entrance of the BOF on the southwestern shores of New Brunswick receives, fresh water through the St. Croix, Magaguadavic and Digdeguash rivers (Figure 1.6; Hachey and Bailey,1952; Trites and Garrett, 1983).

1.4 Glacial history

During the Pleistocene epoch, several large local glaciers developed in the Maritime region of Canada. The name given to these glaciers is the Appalachian Ice complex, referring to small local glaciers that restricted the advancement of the larger continental Laurentide Ice Sheet into New Brunswick and the Maritime provinces (Stea 2004). New Brunswick was glaciated several times during the Quaternary, with at least two interglacial periods. However, due to the erosive nature of glaciers, the only sediments found, with few exceptions, are of late Wisconsinan age (NBDNR 2006). The Appalachian Ice complex covered southwestern New Brunswick and by the end of the last glacial maximum it covered the BOF (Figure 1.7) and extended to a terminal position at the edge of the continental shelf off the coast of Nova Scotia (Stea et al. 1998).

Rampton et al. (1984) divided late stage Wisconsinan glaciation in the province into six phases. This division was based upon local ice sheets shifting their dominance over the area as the ice masses went through periods of growth and stagnation. The result was a complex pattern of streamlined erosional features such as rat-tails, glacial striae and dispersal trends of till clasts and matrix geochemistry (Broster et al. 2004; Seaman 2006). These ice-flow features (Figure 1.1) suggest that ice advanced in a southerly to southeasterly direction in the southwestern part of the province.

Deglaciation of Atlantic Canada began in conjunction with rising sea levels. As sea

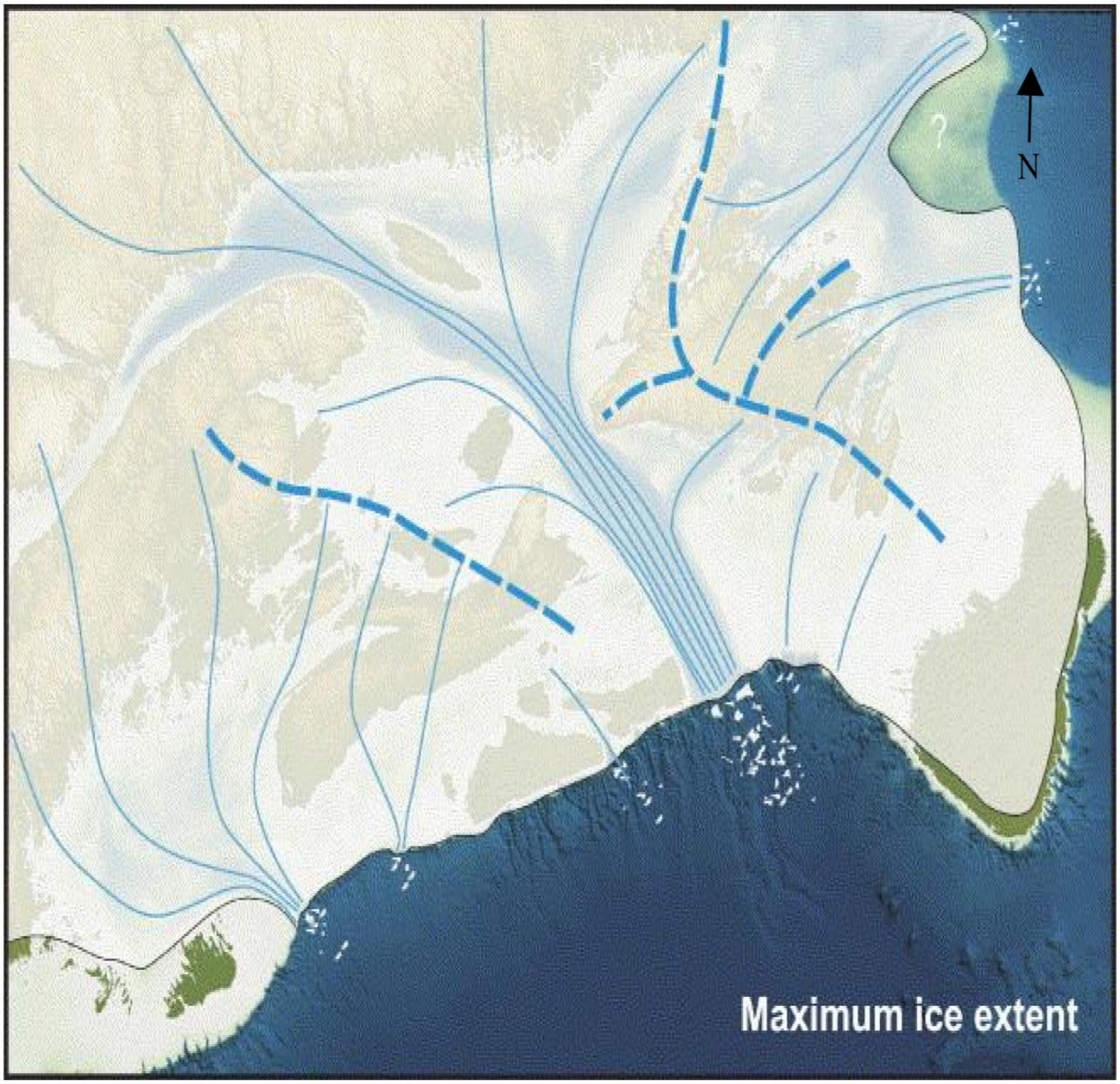

Figure 1.7: Extent of ice coverage at glacial maximum covering New Brunswick, the Bay of Fundy, and Nova Scotia. The thin blue solid lines are generalized ice flow lines; the heavy blue dashed lines are major ice divides (after Shaw et al. 2006).

levels rose, the glacier rapidly broke up and retreated to New Brunswick's present coastline. Shaw et al. (2006) present a conceptual model of ice margins at critical times during deglaciation. At 16800 yrs BP the glacier was rapidly disintegrating and calving in the BOF as ice streams (Figure 1.8). The rapid disintegration led to ice retreat to the BOF coast occurring sometime between 13400 and 16500 yrs BP (Seaman 2006; Dickinson 2008).

The pattern of deglaciation in New Brunswick is based upon mapped glacial landforms, such as eskers, deltas, end moraines, kame moraines and ablation tills (Figure 1.1). These landforms indicate that deglaciation of the southwestern province occurred as a result of stagnation and downwasting.

Eskers are long sinuous ridges composed largely of stratified sand and gravel, deposited by a subglacial or englacial stream flowing within a glacier tunnel formed during glacial stagnation and retreat (Benn and Evans 1998). In New Brunswick a network of eskers 100 km or longer, starting in Maine can be traced through the Woodstock and Millville areas to end in the Pennfield-Utopia delta complex (Figure 1.1). In southwestern New Brunswick, the overall pattern of the eskers is that they are elongated south to southwest, parallel to the last dominant regional ice flow (Stumpf et al. 1997).

During the Late Wisconsinan, the extensive subglacial network of eskers fed into major deltas, found in the Pennfield-Pocologan delta complex along the coast of New Brunswick (Figure 1.1). This proglacial delta system is important in the reconstruction of the glacial history, showing past glacial conditions and changes in sea level (Kiewiet de Jonge 1951; Kaplan 1999).

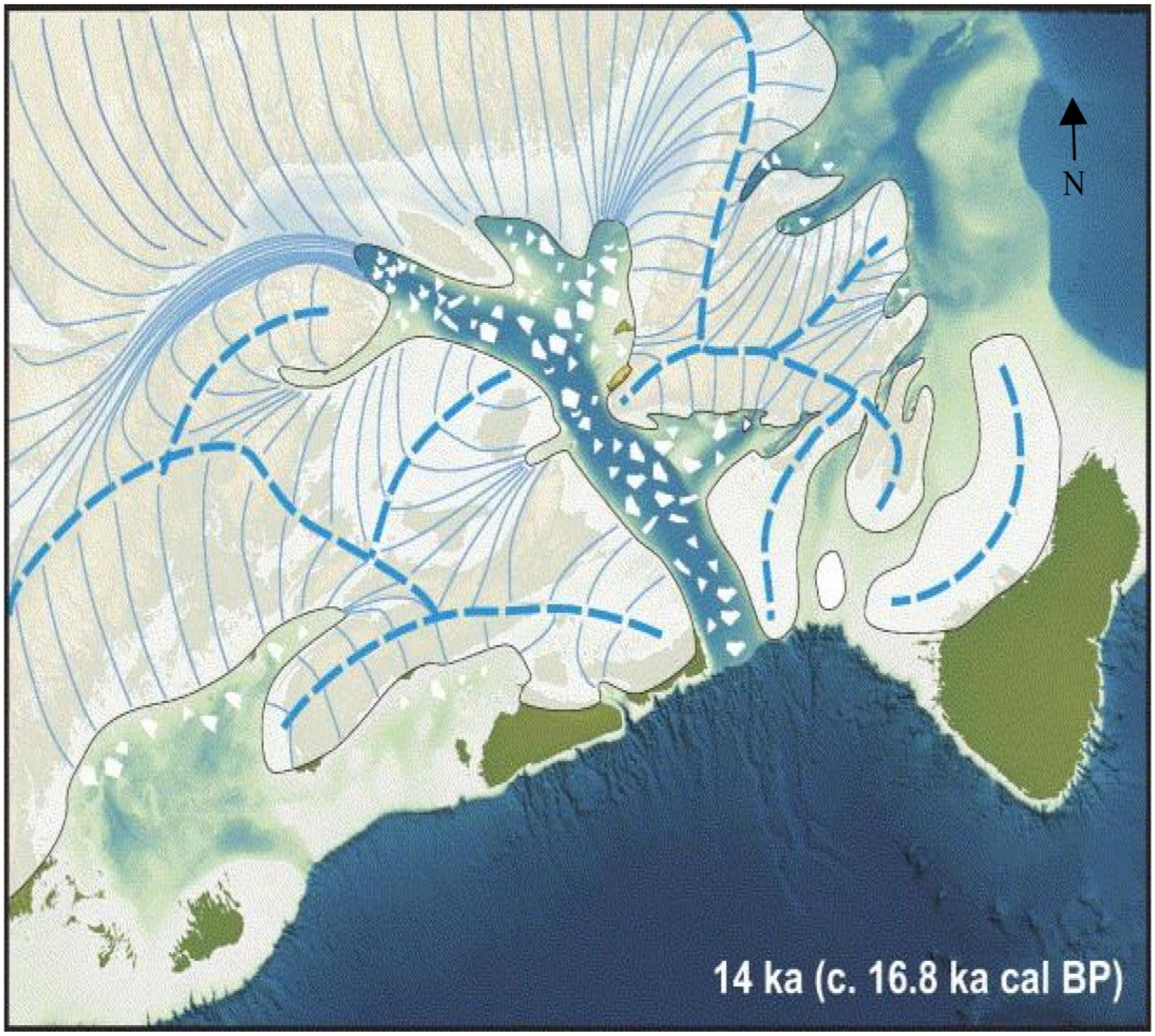

Figure 1.8: Extent of ice at 14 ka yr BP, showing rapid disintegrating and calving glacier in the Bay of Fundy. The thin blue solid lines are generalized ice flow lines; the heavy blue dashed lines are major ice divides (after Shaw et al. 2006).

The Pennfield-Utopia delta complex has three components: the Utopia, Little Lake, and Pennfield deltas (Figure 1.1). These deltaic systems represent a period during deglaciation when the ice margin was relatively stationary, and glacial meltwater was depositing large amounts of glacial outwash into the sea. The large Pocologan delta complex lies within a wilderness reserve, and its sediment history has not been studied in detail (Seaman et al. 1993).

The Utopia delta, situated furthest north of these three deltas (Figure 1.1), has elevations between 65 and 74 m, and is the lowest of the three landforms. A pit exposure at the Utopia delta shows gravelly sand topset and silty sand foreset deposits (Seaman et al. 1993; Seaman 2004). Directly south of the Utopia delta lies Little Lake delta (Figure 1.1), the oldest and highest of the deltas (Seaman et al. 1993). The surface elevation occurs from 85 to 94 m, and the delta is intensely kettled. The internal structure of the Little Lake delta has not been studied. However, the proximity of the Utopia and Pennfield deltas indicates it also is likely of deltaic origin. The largest studied delta, Pennfield, lies between 70 and 85 m asl, and has been studied at an open pit quarry (Seaman et al. 1993). Marine shells discovered at the toe of the delta slope were dated to 13000 yrs BP, indicating the south end was then a marine bench. The southwestern end of the delta has a marine bench cut by wave action at the same height as the Utopia delta (Seaman et al. 1993).

The morainal system in southwestern New Brunswick is comprised of end moraines found along the coast, kame moraines located further inland and a blanket of ablation till located in topographically depressed areas. The largest of these glacial landforms are the linear end moraines and kame moraines that formed transverse to glacial flow (Kaplan 1999). Starting from east to west, this morainal system includes the Sheldon Point moraine in Saint John, the St. George moraine in the Pennfield-Pocologan delta area and the Pineo ridge moraine on Campobello Island and the coast of Maine (Figure 1.1).

The Sheldon Point moraine (Figure 1.1) in Saint John was studied and mapped by Nicks (1988) and was described as a glacimarine end moraine. The maximum height is 45 m asl and it extends to a depth of 90 m below sea level. The depositional record covers a critical period during New Brunswick's deglaciation from 14000 to 10000 yrs BP. Radiocarbon-dated shells from the moraine show the glacier was grounded over a 500 year period from 14000 to 13500 yrs BP (Nicks 1988). Following this short period of grounding, the moraine was modified by two readvances at 12650 and 11200 yrs BP, as determined by dated marine fossils. The formation of the Sheldon Point moraine resulted in the Saint John River changing its course from its previous location at Saints Rest Beach to its present location at the Reversing Falls (Figure 1.6). Nicks (1988) concluded that Sheldon Point moraine is a composite moraine, recording a change from a tidewater glacier to subaerial exposure over a time period of 4000 yrs (Seaman 2006). Situated behind the Sheldon Point moraine at 1.5 and 3 km are two other recessional moraines, the Manawagonish and the South Bay moraines respectively. Both lie within urbanized areas of Saint John and have not been studied in detail (Seaman et al. 1993).

Located along the coast of the BOF between Saint John and St George (Figure 1.1) are major kame moraines (Seaman et al. 1993). Benn and Evans (1998) describe a kame moraine as end moraines deposited in patches from melt water flowing in contact with a moving or decaying glacier. The moraine at St. George is a kame end moraine located to the west of the Pennfield-Utopia delta complex. An exposed pit shows an undeformed bottom layer of laminated marine clays, silts and fine sands. Overlying these deposits are glaciofluvial gravels, which have been sheared and folded by a glacial readvance toward the southeast. The top layer is 1 to 2 m of ablation till with ice-collapsed structures (Seaman et al. 1993). These sediments show initial deposition was within water experiencing currents, followed by ice-contact deposition against a grounded glacier. The grounded glacier went through a period of re-advance, when ice flowed over the moraine, freezing to and contorting the upper layers of sediments in the direction of ice flow. During the final stages of deglaciation the ice melted in place, depositing the top layer of ablation till.

The Campobello Island moraine segment is a continuation of the Pineo Ridge end moraine found along the coast of Maine. The Pineo Ridge moraine is part of a larger complex, which includes a large delta structure (Figure 1.1). Kaplan (2007) concluded that a 100 km long ice margin formed this moraine delta complex. Fossil marine shells place the age of the moraine between 13200 and 12800 yrs BP in Maine (Borns et al. 2004).

Ablation tills form in subglacial environments resulting from deposition of till in place as the glacier melts and downwastes. The ablation tills in southwest New Brunswick are all found inland in topographic depressions and consist of bouldery ablation till (Seaman et al. 1993).

Deglaciation of New Brunswick was suggested by Rampton et al. (1984) to have been completed by 12000 yrs BP. However, there is evidence that residual ice caps reformed and glaciated some areas of New Brunswick during the Younger Dryas

Chronozone (Seaman 2006). Final deglaciation of southwestern New Brunswick may not have been accomplished until sometime after 11000 yrs BP.

Nicks (1988) found that the Sheldon Point moraine Younger Dryas interval shows evidence of glacial readvance with marine shells dated at 11620 yrs BP. These shells are in an overturned, black, stratified, fine sand layer overlain by a thin reddish brown diamicton, which in turn is overlain by gravelly, glacifluvial foreset beds (Seaman 2006). An ice cap present in the Caledonia Highlands may have been the source for this reworking of the Sheldon Point moraine as well as the possible source of an ice dam of the St. John River, creating glacial Lake Acadia. The Caledonia Highlands ice cap was reactivated during the Younger Dryas, obstructing the drainage of the St. John River (Lee 1957; Seaman et al. 1993). The deposition of well developed varves in marine silty sands that blanket much of the New Brunswick lowlands, suggests Lake Acadia occupied a large area of the lower St. John River valley, draining sometime before 11000 yrs BP (Lee 1957; Rampton et al. 1984; Dickinson 2008).

The Younger Dryas Chronozone was globally a time of climatic cooling, occurring near the end of the Wisconsinan glaciation between 10700 and 10100 yrs BP. It is characterized by period of cold climate for 800 to 1000 years in the North Atlantic, reactivating glacial conditions (Borns et al. 2004). A number of sites within New Brunswick record a cool interval during this period, showing evidence of glacial reactivation and advancement (Stea 2004). Evidence for the Younger Dryas climatic cooling is seen with the sediments and palynological history of Splan Pond (Figure 1.6). The sediments include 60 to 80 cm of thixotropic clay that is underlain and overlain by organic sediments (Seaman et al. 1993; Mayle et al. 1993). The palynological evidence shows an environment responding to intense cooling, during which a forest dominated with pine changed to a herb tundra vegetation that was more suited to colder temperatures (Mayle and Cwynar 1995).

1.5 Bay of Fundy Glacial Landforms

Offshore of southwest New Brunswick within the BOF there are several glacial landforms indicating the presence and direction of flow of the Wisconsinan glacier. The bedrock on the floor of Grand Manan basin has streamlined and grooved bedforms recognized in the bathymetry (Shaw et al. 2008). The stoss side of the bedforms is to the northeast and the lee side is to the southwest, indicating southwest flow direction of the glacial ice in this area. These bedforms, found at a depth of approximately 150 metres, have been interpreted as evidence of erosion at the base of an ice sheet (Figure 1.9). Sediments on the southern flank of Grand Manan Basin have elongated mounds with a length to width ratio of 7:1; these are interpreted as megaflutes formed at the base of northeast to southwest flowing ice (Shaw et al. 2008). Relict iceberg scours are found in 110 m deep water to the east and south of Grand Manan Island. On the floor of Passamaquoddy Bay, a drumlin has been mapped by the University of New Brunswick Ocean Mapping Group displaying a northwest to southeast direction (Figure 1.10).

1.6 Post-glacial history

The postglacial history of the study area involves several dynamic factors from isostasy, transgression, regression, tides and currents; sediments deposited by glaciation and deglaciation were modified by these different factors as they changed in influence over time (Belknap et al. 2005). The current stratigraphy of the near shore is a result of interaction of the bedrock, sedimentation rates of glacial and post-glacial sediments, rates

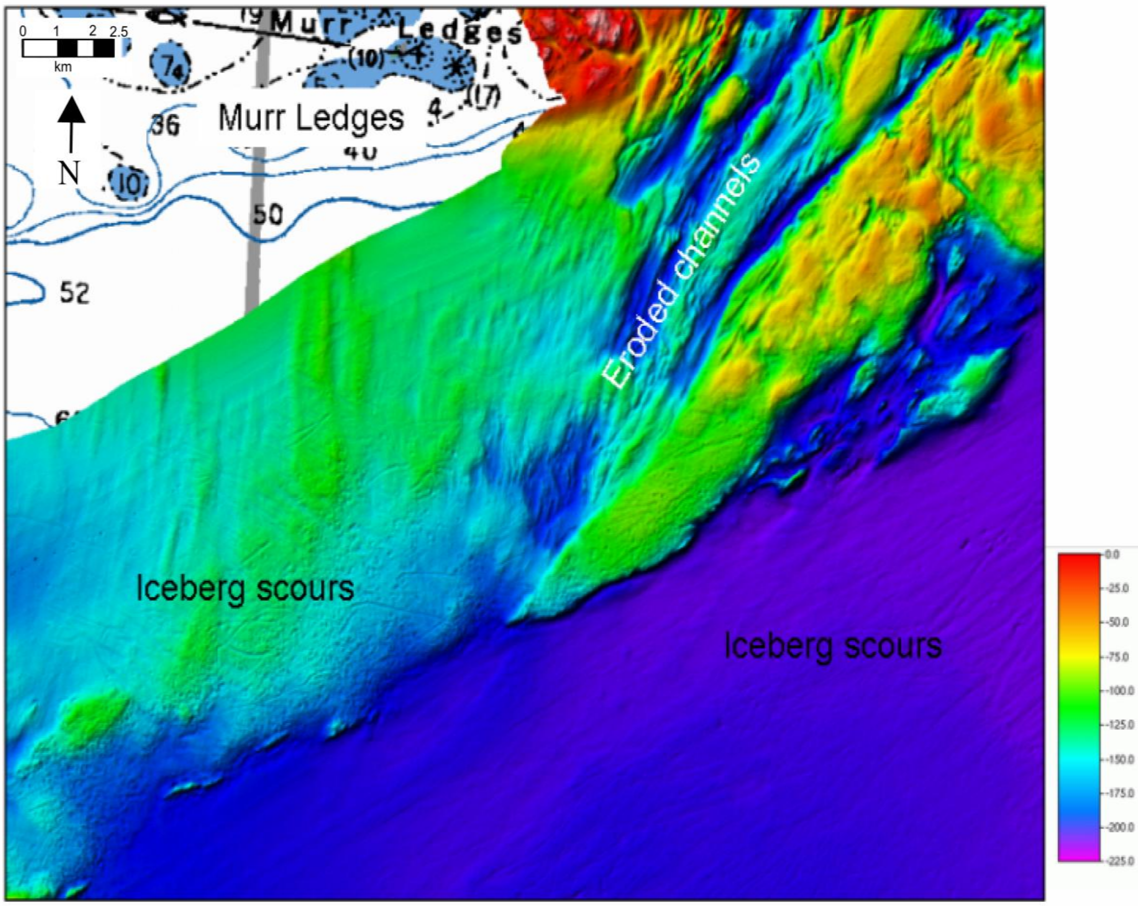

Figure 1.9: Bathymetric map offshore, Grand Manan island in the Bay of Fundy, showing deep trenches eroded into the bedrock, ice scours and pits in the sediment covered areas. See Figure 1.4 for location of area (after Parrott et al. 2008).

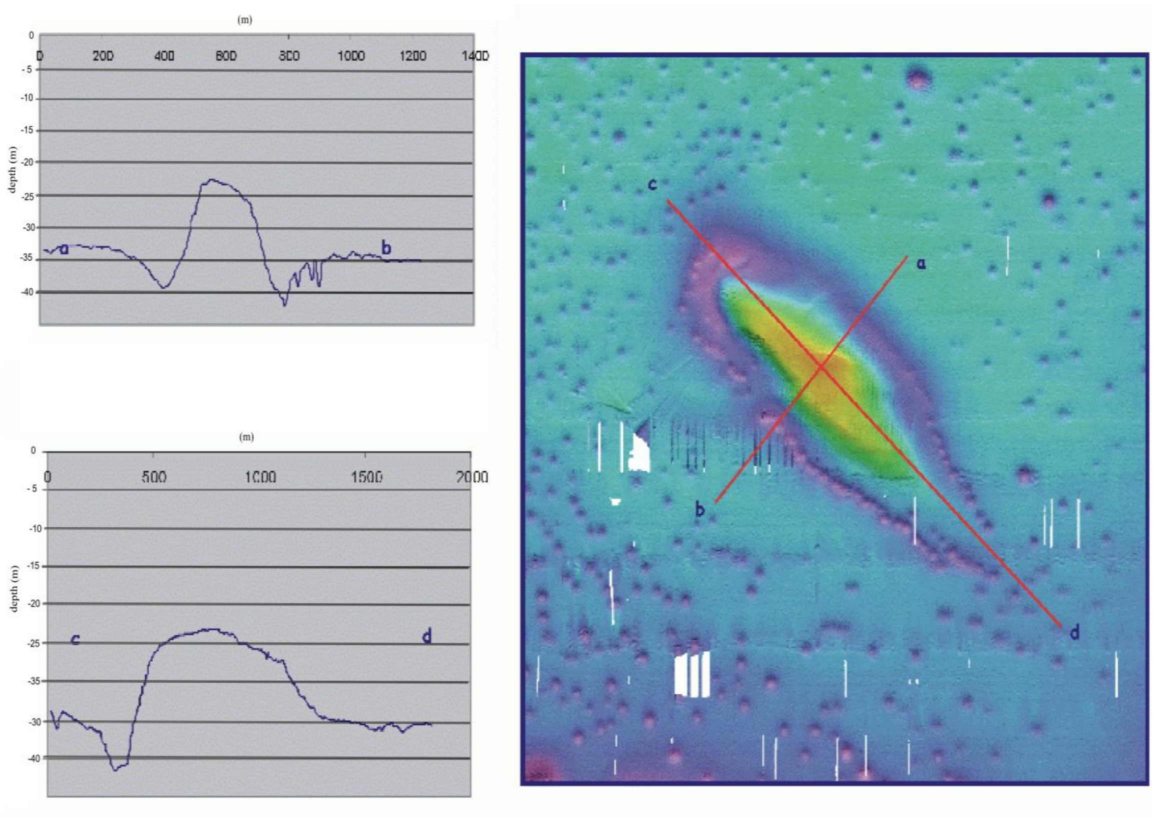

Figure 1.10: Cross section and bathymetry through the Passamaquoddy Bay drumlin. See Figure 1.4 for location of area (after Ocean Mapping Group 2003).

of sea level change, and the direct effects of tidal energy, currents, and mass wasting processes (Belknap et al. 2002).

At the time of the last glacial maximum, sea level worldwide was -120 +/- 20 m lower than at present (Pirazzoli 1991). Globally, sea level began to rise with the melting of the extensive continental glaciers. Within the BOF, sea level fluctuations occurred contemporaneously with deglaciation (Seaman et al. 1993). Just after the time of glacial maximum, when the Appalachian glacier complex overrode the BOF, global sea levels began to rise, contributing to the disintegration of the glacier complex and the pattern of retreat. As the glacier retreated to the coastal areas of the province, the sea inundated the isostatically depressed lands vacated by the disintegrating glacier. Relative sea level along the ebbing ice sheet margin was the result of local isostatic rebound and global eustatic sea level rise (Kaplan 1999). Shipp (1989) summarized the sea level history of the area as that of crustal depression from glacioisostasy, and transgression following deglaciation as sea level rose in concert with the retreating ice margin. The pattern of relative sea level change was controlled by the rate of ice retreat and the level of isostatic rebound following deglaciation (Stea et al. 2001). The isostatic adjustment of land levels and eustatic changes in sea level caused much of southern New Brunswick to be overlain by marine brackish or lacustrine waters during and following deglaciation (Gadd 1973; Rampton et al. 1984; Dickinson 2008). Elevated marine features representing the upper marine limit are marked along the coast by glacimarine deltas and shoreline features. In Atlantic Canada these have been dated within a range of 14 ka to 11.5 ka yrs BP (Wightman and Cooke, 1978). Figure 1.11 shows a digital terrain model of New Brunswick at 80 m asl, with much of southern New Brunswick inundated by water.

In Maine the position of highstand has been recorded from 60 to 132 m above present sea level (Figure 1.12). The marine marks of highstand position in New Brunswick have been recorded from 73 m to as high as 89 m above present sea level (Seaman et al. 1993). The differences can be attributed to the variation in glacial unloading (Bacchus 1993). Following sea level highstand, sea level fell 43 mm/yr until reaching lowstand. In Maine this happened ~12.5 ka yrs BP at approximately 55 – 60 m below present sea level (Belknap et al 2005), which would have made parts of the Maine inner shelf emergent during that time (Bacchus 1993). The resulting lowstand created shelf terraces located in the inner shelf of Maine with a transition to shelf valleys seaward (Shipp 1989). Since low stand, sea level has been gradually rising (Seaman et al. 1993). With this second transgression, the greatest morphological changes occurred along the coast (Uchupi and Bolmer 2008). These changes came in the form of reworking and winnowing of sediments by the action of wave and storm currents (Caron et al. 2004). The amount of erosion is related to pre-existing topography, depth of erosion, wave energy, sediment supply, erosion resistance, tidal range and the rate of relative sea-level change (Belknap and Kraft 1981).

The tides of the BOF are known for having some of the highest vertical tidal ranges in the world (Figure 1.13). The tidal amplitude increases from the mouth of the BOF to the head of Chignecto Bay, the mean tidal range changing from 5 m to over 12 m respectively (Hachey and Bailey1952; Dashtgard et al. 2007). This is the result of the proportions of the BOF basin, the narrowing and shallowing at the head of the bay, and the near resonance of the BOF tides and the Atlantic Ocean (Hachey and Bailey1952).

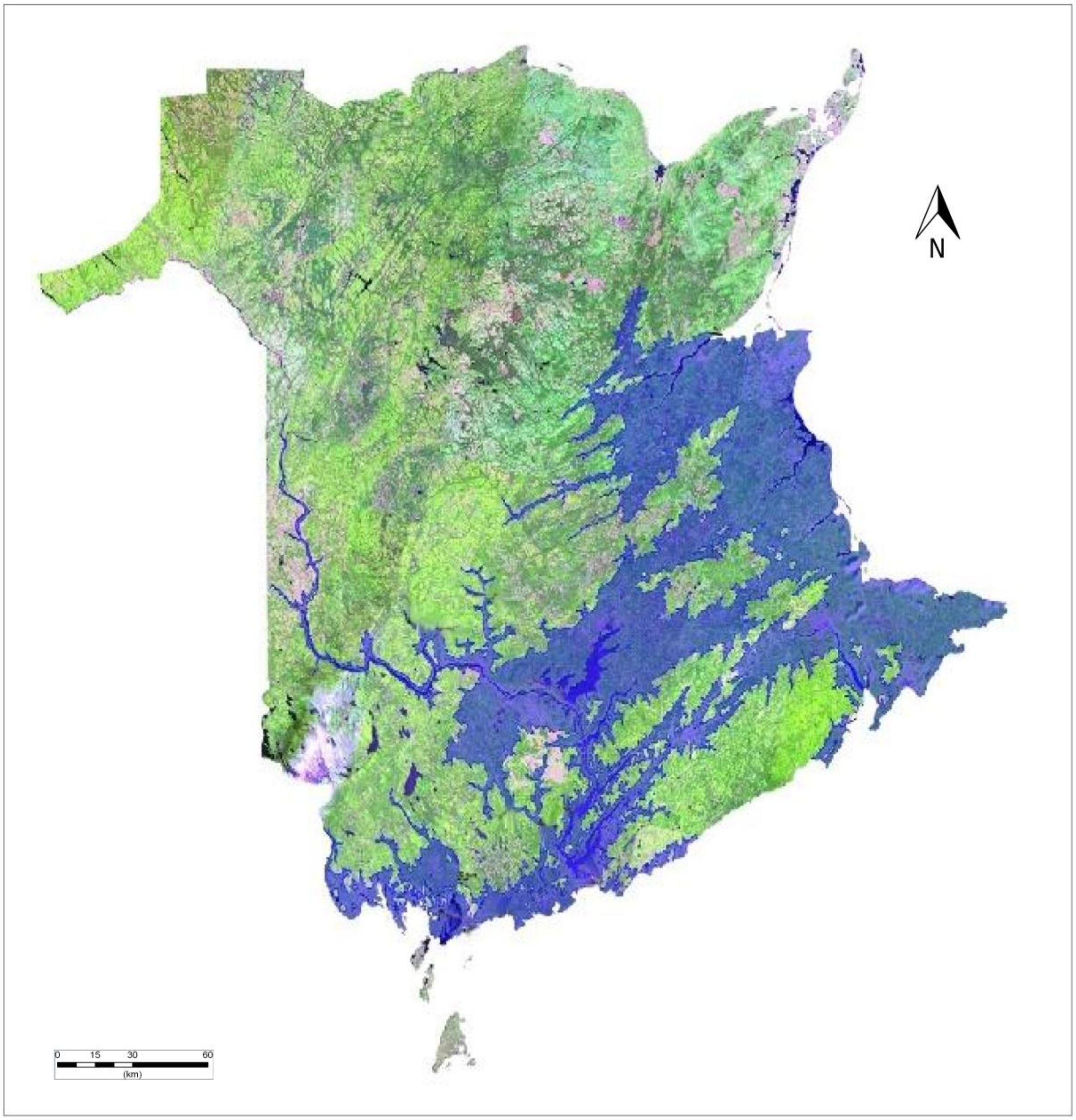

Figure 1.11: Digital terrain model of New Brunswick at 80 m asl, showing southwest New Brunswick inundated by water (after Dickinson, 2008).

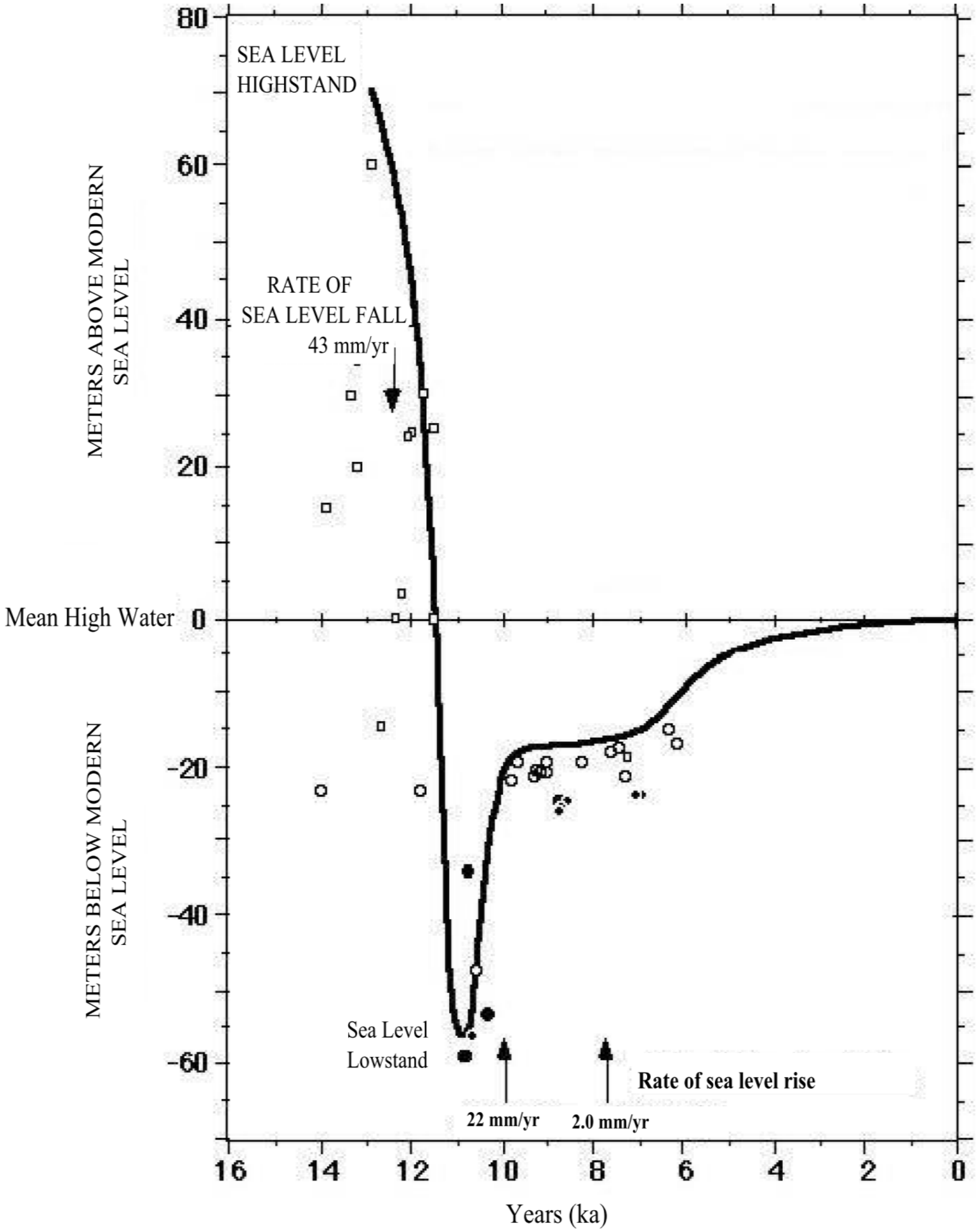

Figure 1.12: Maine's local relative sea level curve for the past 14 ka years B.P. Based on radiocarbon dates of salt marsh peat, plant fragments and marine shells (after Kelley et al. 2005).

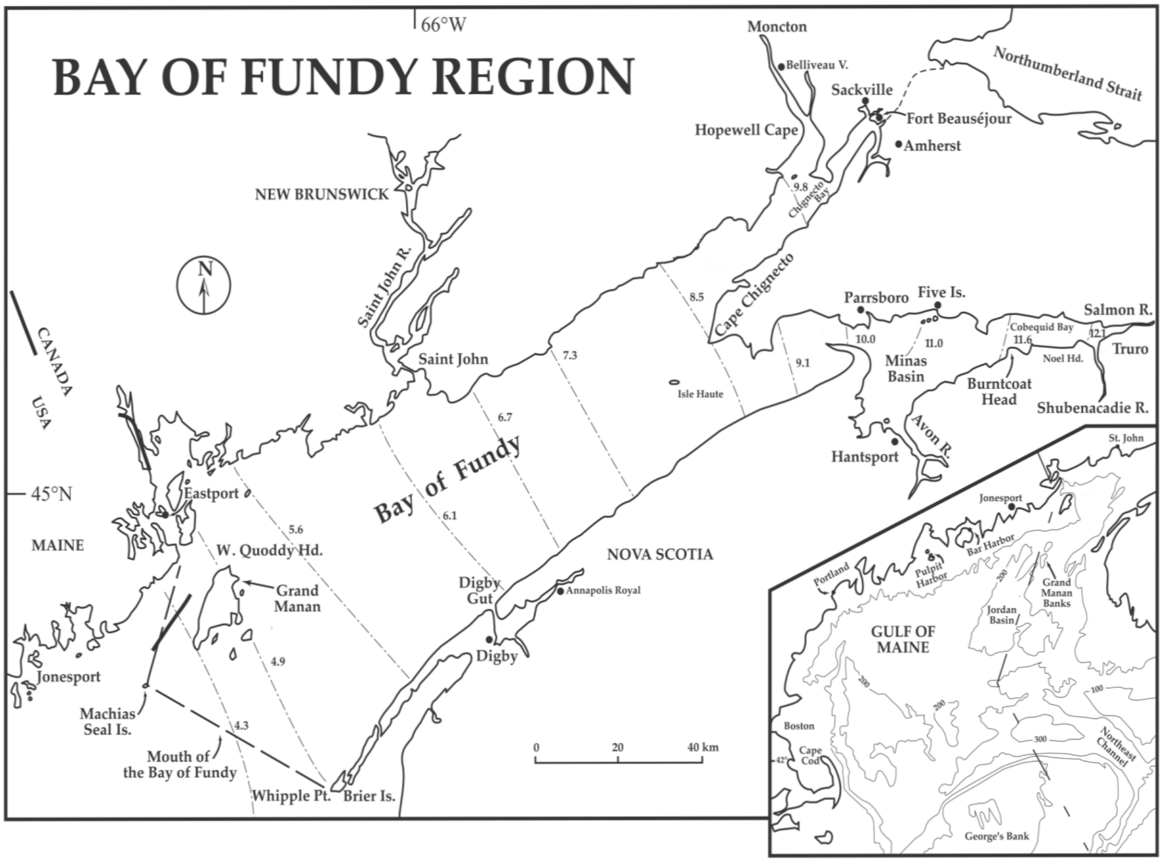

Figure 1.13: Bay of Fundy and Gulf of Maine (inset) mean tidal range, shown by broken lines and represented in metres (after Desplanque and Mossman 2004).

The tidal amplification as it is observed today is a geologically recent event; prior to 7000 yrs ago tides in the BOF were microtidal (Dashtgard et al. 2007). Since deglaciation over the last 14000 yrs, the depth of the bay has changed considerably, increasing the tidal range with depth (Desplanque and Mossman 2004). Scott and Greenberg (1983) suggest water depth over Georges Bank was a controlling factor for tidal amplitude in the BOF. The lowering sea level restricted inflow of water over Georges Bank prior to approximately 6000 yrs ago, resulting in isolation of the Gulf of Maine and Bay of Fundy from Atlantic tidal dynamics (Grant 1970). As sea level progressively rose and submerged Georges Bank, the BOF increasingly came under the influence of tidal forces (Desplanque and Mossman 2004). Mathematical modeling of the Gulf of Maine tidal system indicates that 7000 yrs BP the semidiurnal lunar constituent of the tide, M2, tidal range had reached between 54 - 57 % of the present range; by 4000 yrs BP the tidal range had grown to 78%; and it reached 98% at approximately 1000 yrs BP (Dashtgard et al. 2007; Desplanque and Mossman 2004). The offshore stratigraphy of the BOF is directly related to rising sea level and changes in the tides, resulting in changes in depositional style.

Chapter 2

PREVIOUS WORK IN THE BAY OF FUNDY AND GULF OF MAINE

Important background information to the discussion of the surficial geology of nearshore southwestern New Brunswick comes from geological investigations of local coastal and continental shelf marine and comparable sites from nearby Maine. A description of the local coastal geology of southwestern New Brunswick is presented in section 1.4.

In 1952, Hachey and Bailey compiled the first maps detailing the distribution of sediments in the BOF. The maps were based upon earlier Admiralty charts of the BOF. Swift and Lyall (1968) used continuous seismic reflection surveys to map widespread glacial erosion features and extensive till-like material on the floor of the BOF. The surficial geology of the BOF and Gulf of Maine was most recently investigated in a study by Fader et al. (1977) and King and Fader (1986), who mapped the distribution of sediments in an attempt to understand the processes associated with late Pleistocene deglaciation and the subsequent post-glacial modification by sea level fluctuations. The areas of the Scotian Shelf, eastern Gulf of Maine and BOF were mapped using high resolution seismic surveys, bottom dredge and grab samples, piston cores, bottom photographs, and adjacent land geology. A summary of the regional surficial geology is provided in Table 2.1. Bedrock is overlain by till and Emerald Silt glacimarine sediments, which in turn are overlain by the Sambro Sand formation and a Holocene LaHave Clay deposit (Fader et al. 1977).

Formation	Description	Thickness (m)	Interpretation
LaHave Clay	Medium to coarse-grained, well sorted sand, angular to sub – rounded glacial lag gravels	0 - 70	Post-glacial marine deposition, mud
Sambro Sand	Fine to coarse-grained, sublittoral silty and clayey sand, well sorted grading to angular gravel	0 - 20	Sublittoral sand
Emerald Silt	Fine to coarse-grained clayey and sandy silt	0 – 55	Proglacial silt, either grounded ice shelf , a floating ice shelf or intermittent contact between the ice and seabed
Scotian Shelf Drift	Poorly sorted, cohesive till	0 – 92	Till

Table 2.1: Seismic facies from the eastern Gulf of Maine and Bay of Fundy (after Fader et al. 1977).

The King and Fader (1986) survey located off the Scotian shelf further resolved the Emerald Silt (Table 2.1) into three seismic facies based upon different acoustic appearances. The Emerald Silt facies A and B were interpreted as subglacial in origin. The different morphology of these two is attributed to their respective formations. In facies B dropstones were present, suggesting deposition from settling through the water column from a floating ice shelf. Emerald Silt facies C is interpreted as transitional between facies A and the Scotian Shelf Drift.

Passamaquoddy Bay is located at the southwestern end of the research area, on the border of Canada and the U.S. state of Maine (Figure 1.1). A dense distribution of pockmarks covering the seafloor was first discovered by King and Maclean (1970), who described them as "concave, crater-like depressions that occur in profusion on mud bottoms across the Scotian Shelf". Subsequently, pockmarks have been described worldwide on a variety of geologic settings and in a variety of environments from oceans, seas and also in some lakes (Hovland et al. 2002). The geologic mechanism of formation is still debated; it is believed they are formed by the release of pressurized "fluid" below the seabed typically resulting in circular depressions (Rogers et al. 2006, Gontz, 1999). Pockmarks have a wide range of sizes and shapes, from less than one metre to greater than 700 metres in diameter with depths reaching 45 metres (Gontz 1999). The preservation of pockmarks is widely believed to result from nearbed currents, which inhibit infilling and sediment accumulation within the pockmarks (Josenhans et al. 1978). Although pockmarks are often circular, they have also been found elongated (Hovland et al. 2002). Elongated pockmarks can occur where bottom tidal currents are high with elongation oriented in a preferred direction of the currents (Josenhans et al. 1978; Boe et al. 1988). Another type of pockmark is called 'eyed' and has been described as having an acoustically higher backscatter reflection in its centre. The higher backscatter results from either the presence of either carbonate deposits from biological activity or from erosive processes winnowing away the finer sediments and leaving behind coarser materials (Pecore and Fader 1990; Hovland et al. 2002)

The terrestrial and marine Quaternary geology of Maine has been extensively studied (King and Fader 1986, Bacchus 1993, Barnhardt et al. 1997, Belknap et at. 2002, and others). Modern studies of Maine's geology were initially conducted by Leavitt and Perkins (1935) mapping the surficial geology of Maine. The purpose of that study was the determination of the distribution of gravel for road building. In the 1940's the study was further expanded to include the mapping of fine-grained marine sediments and clays (Trefethen et al. 1947). A massive glacimarine mud unit, Presumpscot Formation, was first described by Bloom (1963). The Presumpscot Formation contains scattered dropstones, is commonly found to be rhythmically bedded, and contains sand and silt laminae in many locations. The unit was mapped along the coast of Maine and into New Hampshire and was estimated to be deposited at the time of deglaciation (Belknap and Shipp 1991). The timing of Presumpscot deposition and thus, marine submergence, has been dated between 13 and 12.7 ka BP by Stuiver and Borns (1975).

Evidence of the retreating glacier is found along the Maine coast with the prominent moraines PRM and PRMC (Figure 1.1, Kaplan 2007; Borns et al. 2004). Marine fossils within the moraines have allowed the time of deposition to be constrained between ~ 13.8 and 13.4 ka BP (Kaplan 2007). The conditions in which the moraines formed are complex and controversial but it has been suggested that they either represent a prolonged stillstand, a readvance or a both a stillstand and a readvance (Kaplan 1999; Kaplan 2007).

The Gulf of Maine (GOM) is located south of the study area (Figure 1.4). Many papers have been published documenting the nearshore and offshore marine geology of the area (Bacchus 1993, Barnhardt et al. 1997, Belknap et at. 2002, and others). These surveys have used seismic reflection profiles and cores to document the offshore marine sediments. One of the earliest nearshore surveys was conducted by Ostericher (1965) in Penobscot Bay (Figure 1.4). Ostericher (1965) obtained a radiocarbon date of 7390 $\pm$ 500 yr BP at the top of glacimarine sediment, establishing a sea level of -18 m. Similar glacimarine units were also found by Schnitker (1974), Folger et al. (1975) and others along the nearshore of Maine. Schnitker's (1974) study of a deltaic like feature at the mouth of the Kennebec River showed a - 65 m lowstand of sea level. Through many studies Belknap et al. (1987), and Shipp et al. (1999) developed a graph (Figure 1.12) depicting the changes in local relative sea level since the deglaciation of Maine. The Maine sea level curve shows the sea in contact with the ice front, flooding coastal Maine from 14000 to 12500 yrs BP (Shipp 1989). From 15000 to 11000 yrs BP relative sea level change was dominated by isostasy, followed by eustasy after deglaciation (Belknap et al. 2002). Along coastal Maine, a submerged shoreline occurs at 50 to 65 m depth that was produced when sea level was at its lowest between 10000 and 8000 B.P (Belknap and Shipp 1991).

Pratt and Schlee (1969) completed a comprehensive survey of the offshore surficial sediments across the GOM. They mapped till, striae, the orientation of boulder trains and drumlins. Their overall interpretation was that glaciers flowed in a southeasterly direction and grounded ice occupied the GOM during the Pleistocene.

In a series of studies, Shipp (1989) and Belknap and Shipp (1991) described seismic facies of Maine's marine inner shelf. Eight seismic facies were described based upon their morphology and intensity of reflections at the bounding surfaces. Table 2.2 summarizes the seismic facies characteristics, frequency and approximate age.

The contact between the glacimarine sediments and the modern marine Holocene sediments in the nearshore GOM has been found to be the position of an unconformity between the Pleistocene and Holocene sediments. The unconformity separating the glacimarine mud from the modern mud resulted from the lowest post-glacial sea level stand near the former ice sheet edge following transgression (Shipp 1989; Belknap and Shipp 1991; Fader 2005). Seabed features, which indicate a lower sea level followed by submergence during transgression, include terraces, erosional surfaces and unconformities, winnowed sediments, absence of fine grained sediment, erosion of glacial till and glacimarine sediment, and muted topography and relatively greater exposures of bedrock (Fader 2005). Investigations in Maine by Shipp (1989), Belknap and Shipp (1991) Belknap et al.(1987), and others have shown that the Pleistocene/Holocene unconformity shows up as a submerged shoreline along the Maine coast at – 60 m in the inner shelf. Further offshore, between – 60 to – 90 m, the contact transitions to a paraconformity, and uninterrupted sedimentation in – 90 m and greater depths. Shipp et al. (1991) suggested that these events occurred at 13000 to 15000 yrs B.P. for the initial submergence; following submergence the coast emerged from 13000 to 11000 yrs B.P. with relative sea level falling to 60 m below present.

Formation	Description	Thickness (m)	Interpretation
LaHave Clay	Medium to coarse-grained, well sorted sand, angular to sub – rounded glacial lag gravels	0 - 70	Post-glacial marine deposition, mud
Sambro Sand	Fine to coarse-grained, sublittoral silty and clayey sand, well sorted grading to angular gravel	0 - 20	Sublittoral sand
Emerald Silt	Fine to coarse-grained clayey and sandy silt	0 – 55	Proglacial silt, either grounded ice shelf , a floating ice shelf or intermittent contact between the ice and seabed
Scotian Shelf Drift	Poorly sorted, cohesive till	0 – 92	Till

Table 2.1: Seismic facies from the eastern Gulf of Maine and Bay of Fundy (after Fader et al. 1977).

Finally the second transgression and submergence began 10800 yrs BP, as isostatic rebound slowed and eustatic sea level rise predominated.

Belknap and Shipp (1991) found that glacimarine sediments dominated the inner shelf and coastal lowlands of Maine between 14000 and 11000 yrs B.P. The subdivided glacimarine sediments fall into three major seismic facies based upon interpretation of their reflection profiles. Bacchus (1993) summarized the seismic facies of Maine's inner shelf, the Gulf of Maine and the Scotian Shelf, relating the differently named seismic facies with each other (Table 2.3). Unit 1 is bedrock; it is always the lowest facies present and is characterized by steeply dipping reflectors. It is interpreted to represent Paleozoic crystalline rocks of the Avalon Terrane and Meguma terranes (Bacchus 1993). Unit 2 is till; it shows a variety of morphologies but the upper surface is often sharp and very irregular. The basal glacial unit (Table 2.3), till (T), is characterized by dense chaotic returns with no internal stratification, displaying a variety of morphologies throughout the Gulf of Maine. Overlying T is massive glacimarine (MGM) or proximal glacimarine (PGM), found to be highly variable in thickness, and deposited as draping blanket over till or bedrock and with many point source reflectors interpreted as boulders. This unit was deposited rapidly in a sub ice-shelf environment and is interpreted as massive glacimarine mud or diamicton (Belknap and Shipp 1991). Within the MGM unit are found ice-rafted debris (IRD), deposited from a floating ice shelf. The second unit (Table 2.3) is the draped glacimarine (DGM) or transitional glacimarine (TGM); it has similar characteristics to the Emerald Silt Facies A with a distinctive draping geometry, occasional point source reflectors, and is usually well stratified. This unit is interpreted

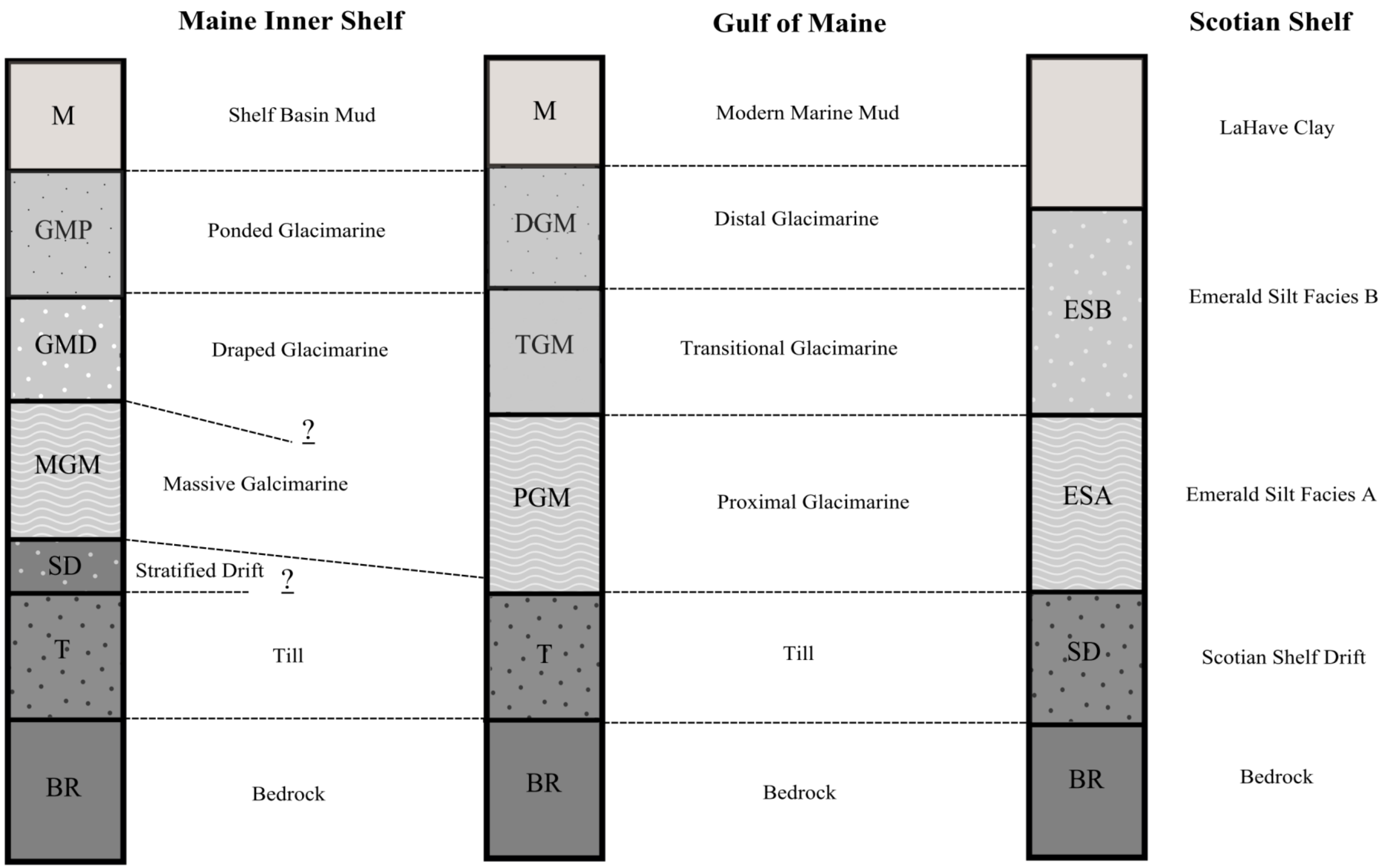

Table 2.3: Correlation of seismic facies from the inner shelf of Maine to eastern, Gulf of Maine and the Scotian Shelf (after from Bacchus 1993).

as a glacimarine interbedded mud, silt and very fine sand; and as having been deposited in a proglacial environment with rapid sedimentation from suspension. The third major seismic glacial facies (Table 2.3) is ponded glacimarine (GMP) or distal glacimarine (DGM). It is acoustically transparent, ponded, with rarely present point source reflectors, and less well stratified than the lower units. It is interpreted to represent ice-distal glacimarine conditions when the glacier was grounded and iceberg influence was reduced (Bacchus 1993). This unit is equivalent to King and Fader's (1986) LaHave clay. A less common unit (Table 2.2) found in Maine is called a thin gravel layer, TGL, which consists of stacked sets of high-intensity reflections that occur within the glacimarine facies (Belknap and Shipp 1991). The TGL unit is represented by well stratified, laterally discontinuous reflections, terminating abruptly in a horizontal direction and often associated with a natural gas facies (Belknap and Shipp 1991, Barnhardt et al. 1997). Belknap and Shipp (1991) describe TGL as ice proximal sediments from gravity flows, found in glacimarine or deltaic sediments and rare in occurring.

The inner shelf stratigraphy of coastal Maine has been categorized into three typical environments: (1) barriers in open embayments, (2) large rivers with lowstand paleodeltas, and (3) estuaries (Belknap et al. 2002). The stratigraphy of each environment resulted from the interaction and influence of the topography, glacial framework, sediment type, isostasy and eustasy and tidal, wave, and mass wasting processes (Belknap et al. 2005). The barriers in open embayments are dominated by refracted waves and storms. Local sediments are the dominant source in this environment, usually from nearby rivers. The second type of environment is large rivers with lowstand paleodeltas. The processes dominating the environment during transgression involve reworking of deposits with rising sea level forming migrating barrier systems. As well, more sediment was added to the environment in the Late Holocene by increasing river input as the sea level decreased (Belknap et al. 2002). The third type of environment found in coastal Maine is estuaries.

Knebel and Scanlon (1985) documented pockmarks in Penobscot Bay, Maine (Figure 1.4). Succeeding geophysical marine surveys by Kelley et al. (1994) have mapped over 2000 pockmarks in the glacial fluvial marine sediments overlying bedrock. The underlying bedrock of the area is metamorphic and igneous, and not known to be hydrocarbon bearing, suggesting a biogenic source from more recent decomposition (Maine Geological Survey 2007).

Chapter 3

METHODOLOGY

3.1 Data collection

The geophysical data were acquired during cruises conducted by the GSC Atlantic, CHS and OMG from 2007 to 2009 with the research vessels CSL Heron and CCGS Frederick Creed. High-resolution single-channel seismic reflection profiles were gathered along approximately 6,500 km of track lines (Figure 1.2). Appendix 1 lists the starting and end coordinates of all the seismic profiles used for this study. In order to measure accurately georeferenced bathymetry the CSL Heron and the CCGS Creed used the "Position Orientation System for Marine Vessels" (POS/MV). This system is a differentially corrected and inertially aided dual GPS positioning system. These systems provide precise position, heading, heave, and attitude data, enabling accurate estimates of seafloor topography from a moving platform. All tides, referenced to mean sea level, were applied using the Department of Fisheries and Ocean Canada WebTide (DFO 2010) prediction model.

The acoustic mapping instruments used in this study include a 3.5 kiloHertz (kHz) Knudsen 320 Marine Echosounder, Kongsberg EM1000, Kongsberg EM1002, Kongsberg EM3002 multibeam and Kongsberg EM710 bathymetric survey. Errors in refraction were corrected using sound velocity profile data collected at the time of the surveys. The EM1000, EM1002, EM3000, EM3002, and EM710 systems recorded depth for multibeam bathymetry and backscatter; the depth values collected were collated and gridded for depth data. These systems also record a mean backscatter value, which was collated and gridded in a similar manner to the depth data.

3.2 Processing

Processing of the multibeam, backscatter and sub-bottom data was accomplished using the OMG and the GSC software tools. The multibeam data were processed using the OMG SwathEd Software Package, allowing for corrections for time delays, manual bathymetric data cleaning, and generating backscatter and multibeam digital terrain models. All bathymetric images in this study use 15 m resolution bathymetry to generate the maps. A higher resolution, 1 m, was available but due to limitations of the ESRI ArcMap (Environmental Systems Research Institute) software used to generate the maps, the lower resolution was used.

Processing of the sub-bottom data was done using the SegyJp2Viewer GSC software developed by Bob Courtney (Geological Survey of Canada - Atlantic). SegyJp2Viewer is a Windows XP program that is used to view and interpret single-channel SGYJP2 files. The seafloor was picked as a horizon and used as the datum, so that all the seismic horizons picked were referenced to the seafloor as the horizon datum. The files are exported as shape files, so that they can be added to digital map sheets compiled in ESRI ArcMap software for digital interpretation and mapping of related features.

For description and interpretation of seismic units, the thickness of each of the units identified on the reflection profiles was calculated assuming an acoustic velocity in water and surficial sediments of 1500 m/s. The velocity increases in glacimarine muds and sands and gravels with speeds of 1500-1800 m/s and 1600-2700 m/s respectively and thus the method underestimates the thickness of these sediments by 6.7% and 16.7 – 23.3% respectively (Davies et al. 1992; Belknap et al. 2005).

Chapter 4

RESULTS

4.1 Study Areas

This investigation of the nearshore stratigraphic framework of the Bay of Fundy in the area between the St. Croix River and Saint John relies on seismic reflection profiles, with multibeam echo sounding providing supporting data. Appendix 1 lists all the seismic lines used in this study. There are six primary study areas (1) Campobello, (2) Passamaquoddy, (3) Pennfield, (4) Maces Bay, (5) Chance and (6) Blacks; naming is based upon the areas found near shore (Figure 1.1). Also, there are two secondary data sets from the CCGS Frederick Creed cruises in 2007 and 2008 features (Figure 1.1). Both these data sets are located offshore of the primary study areas. The data set from Creed 2007 is of poor quality due to a malfunction in the Knudsen instrumentation at the time of data collection. The Knudsen echosounder misfired shots, resulting in poor image quality. Figure 4.1 shows a side by side image of original data and OMG post processing fixed data. Post-processing did provide a useful image, but with a decrease in detail and resolution compared to the other seismic lines used in this study. The 2007 data set is included in this study for its expansion of the locations of natural gas in the Bay of Fundy. All other seismic units other than bedrock are difficult to differentiate.

4.2 Stratigraphy

The stratigraphic framework of the following areas was established through the interpretation of the seismic sub-bottom lines (Appendix 1). The majority of the survey areas were covered by many tightly spaced sub-bottom lines (Figure 1.1), allowing for a continuous interpretation of seismic units within the area.

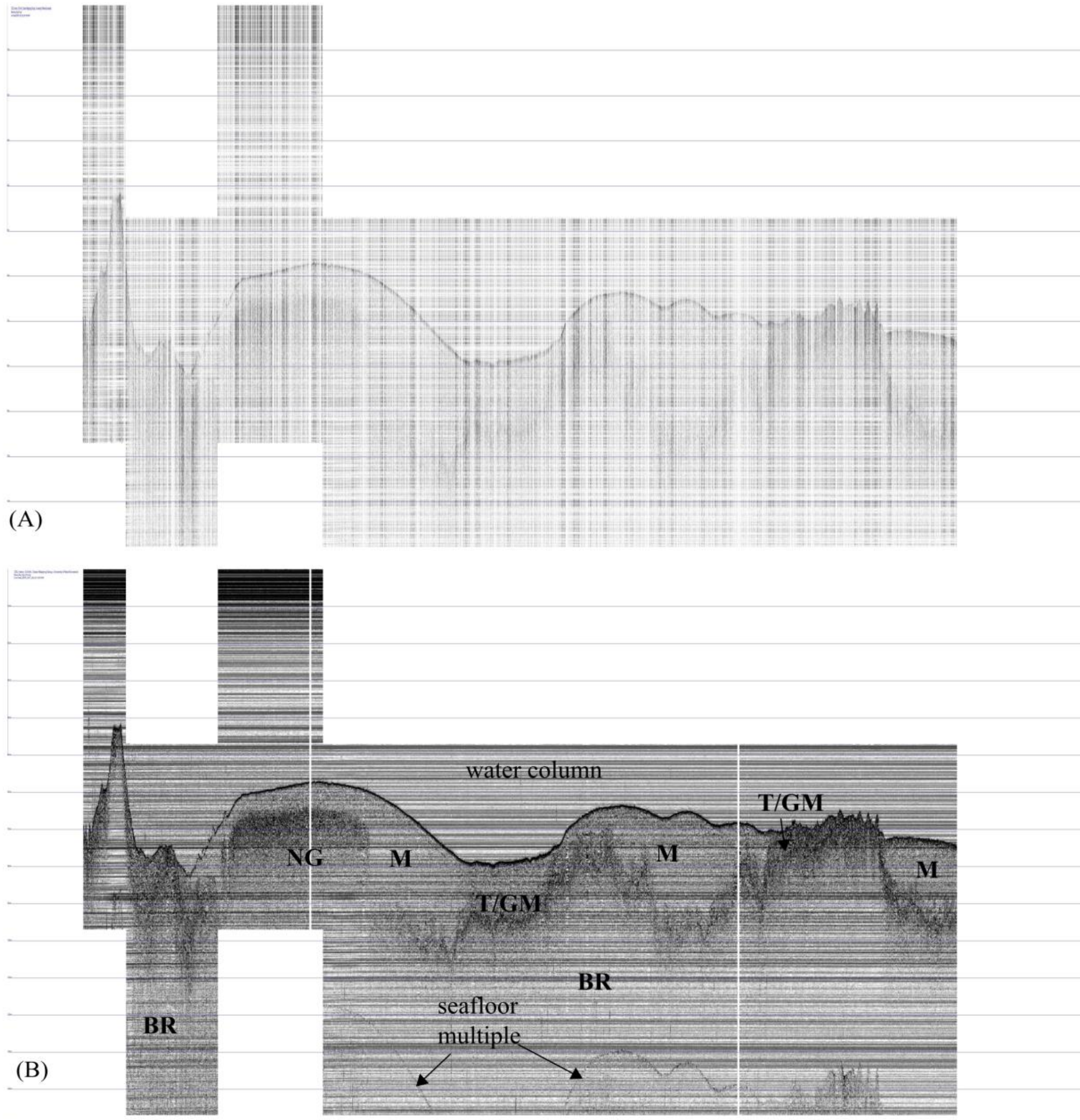

Figure 4.1: (A) Image of a Creed 2007 seismic line with misfiring of the Knudsen echosounder resulting in gaps in the data image. (B) Same seismic line post processing, resulting in an improved image. BR = bedrock, T = till, GM = glacimarine sediment, M = Holocene mud, and NG = natural gas. A seafloor multiple is seismic energy which has been reflected more than once, resulting in a ghosting image. In this case a multiple of the seafloor.

4.2.1 Campobello Survey Area

Campobello Island is located in southwestern New Brunswick at the entrance to Passamaquoddy Bay, close to the border of Canada and the U.S. state of Maine (Figure 1.1). It is the middle island in a chain of three islands, which also includes Deer Island and Grand Manan Island. Campobello Island is 45 km^2 in area, 15 km long and 4.5 km at in width at the widest point.

Bedrock is exposed in the northern half of the island (Figure 4.2); and is composed of Devonian and Silurian rocks consisting of greywacke, slate, siltstone, sandstone, conglomerate and limestone (McLeod 1979; McLeod et al. 1994). Campobello Island's surficial geology includes a kame moraine in the middle, which is likely a continuation of the moraine in adjacent Maine, the Pineo Ridge moraine (Kaplan 1999). To the south of the morainal segments, on the southern tip of Campobello Island is a thin blanket of sand, silt, gravel and clay (NBDNR 2006).

The location of the survey is in Friars Bay, located on the western side of the island (Figure 4.3). The survey includes 12 survey lines (Appendix 1); 11 run parallel to each other in a northeast-southwest direction and one runs perpendicular through these lines towards their southern end. The total length of the lines was 39 km, covering an area of 3.7 km^2. The western limit of the survey was constrained by the international border. The data were collected in the 2008 Heron cruise and also included 15 m resolution bathymetry and backscatter data.

The bathymetry (Figure 4.4) of the Campobello Island survey area shows a gradual increase in depth with increasing distance offshore, from less than – 5 m to – 30 m. At the northwestern end of the survey area, in between Campobello and Deer Islands,

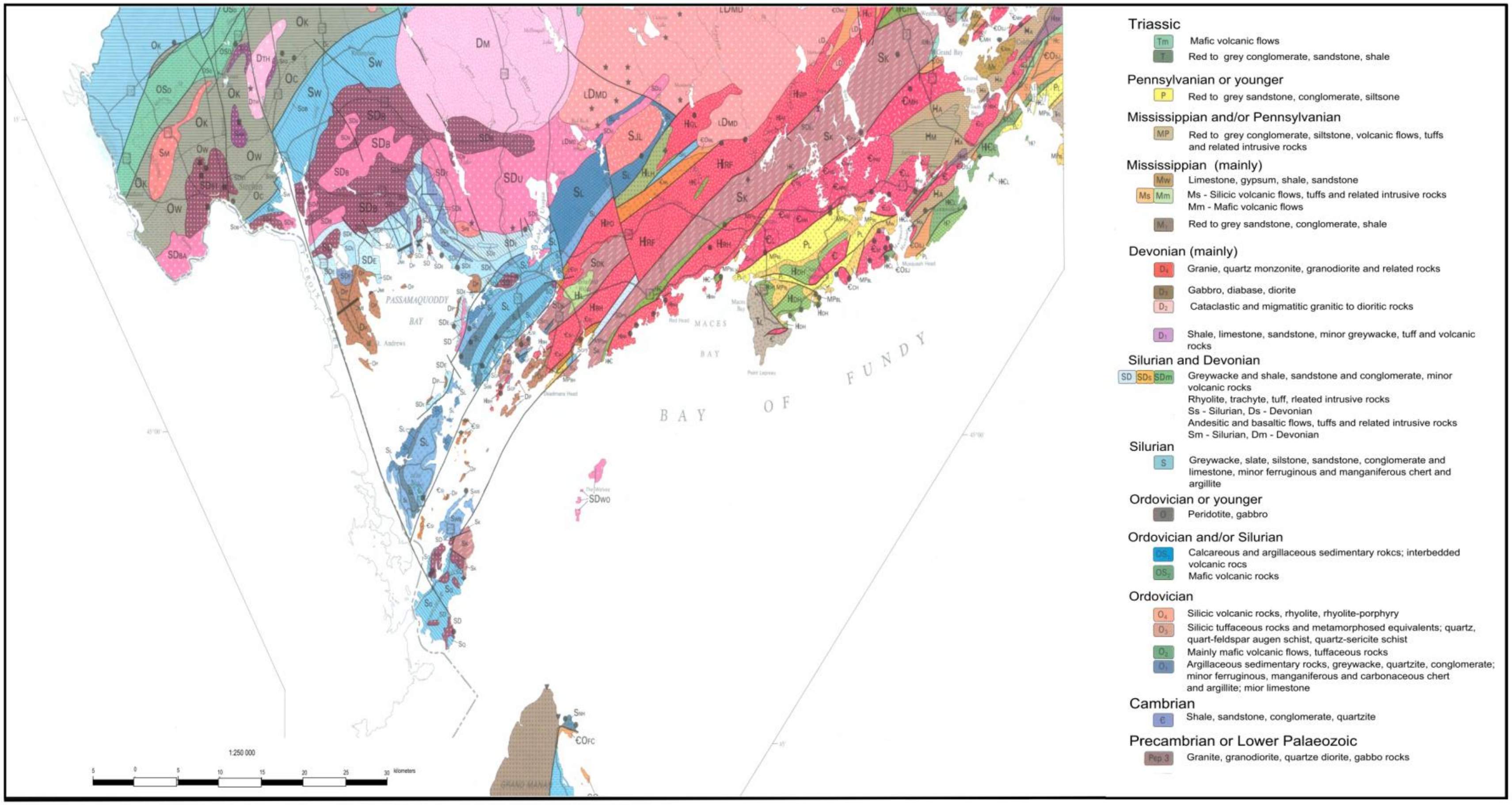

Figure 4.2: Bedrock of southwestern New Brunswick (after McLeod et al. 1994) .

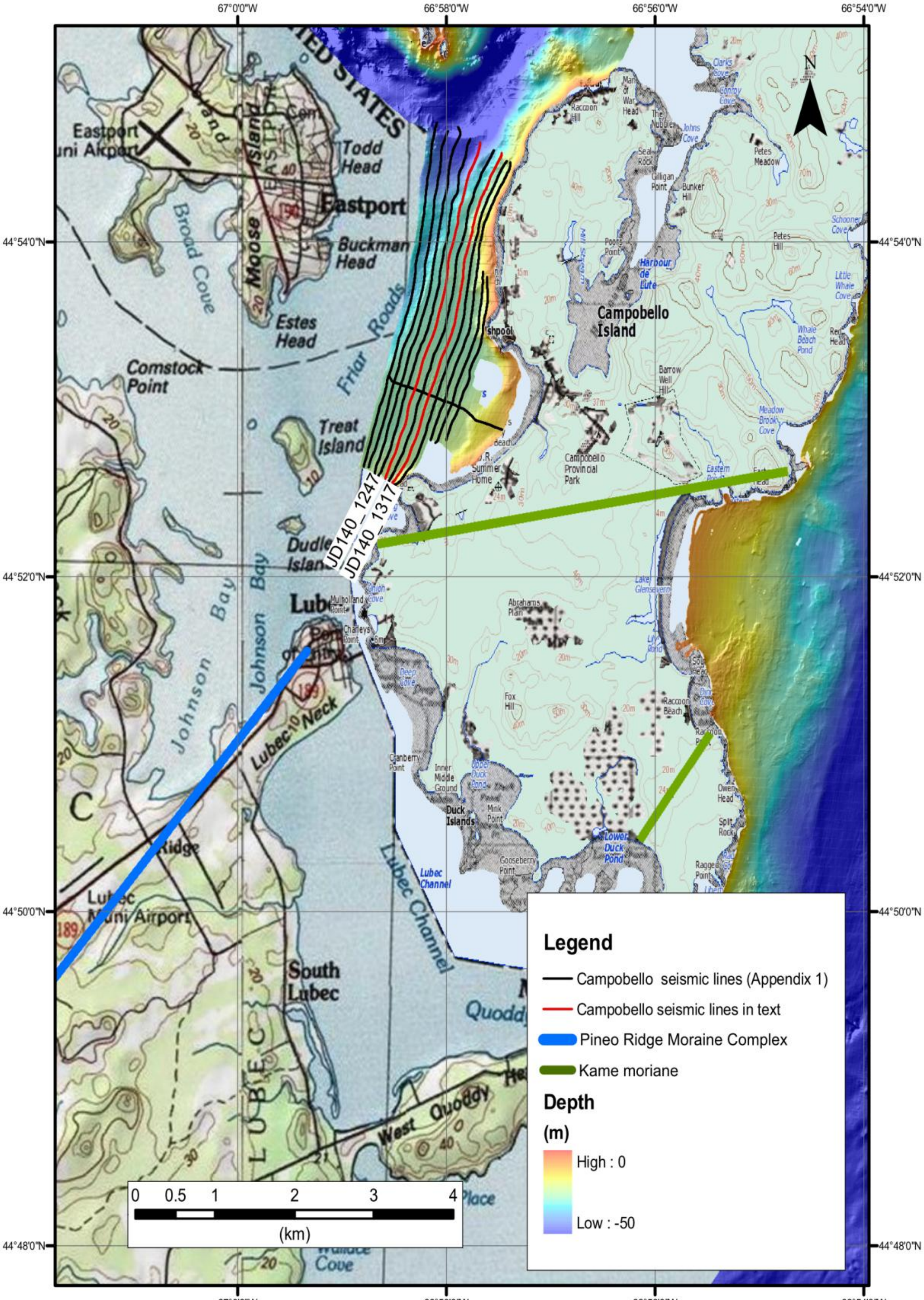

Figure 4.3: Bathymetric map showing Campobello survey area, seismic lines, and glacial fluvial deposits.

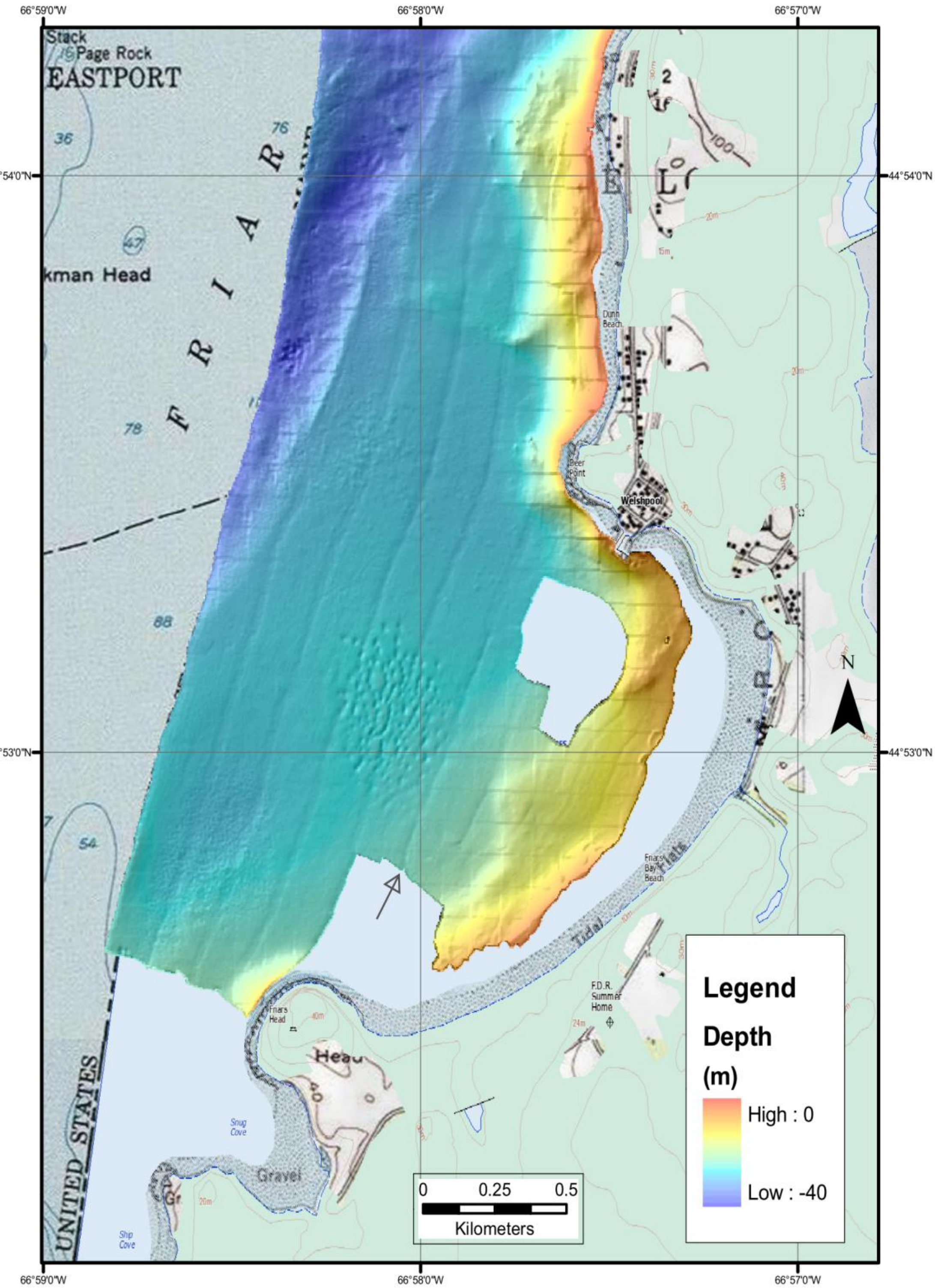

Figure 4.4: Bathymetry of Campobello survey in Friars Bay showing a cluster of pockmarks in the centre of the bay. The two unmapped areas in the bay are fish farms. Note active erosion in northwest due to modern tidal currents. Arrow points to a line parallel to shore; this and other parallel lines are artefacts due to data collection and are not landform features.

the depth steeply drops to – 100 m. The bay is home to two fish farms, located at the northern and southern ends. The only features on the bathymetry map of Friars Bay are a cluster of small pockmarks located in the centre of the bay and anthropogenic scars created by trawling. The pockmarks are located in 40 m water depth, generally oblong, and vary from approximately 5 m to 20 m in length and 5 m to 12 m in width. Some of the pockmarks have merged into linked chains with lengths up to 78 m; however the individual pockmarks are still visible.

The Campobello Island survey area backscatter shows regions of both high and low strength (Figure 4.5). Backscatter strength is directly related to the surface morphology and subsurface sediment type (Lurton 2002). Generally, high backscatter intensity or high reflectivity, a strong acoustic return, is associated with rock or coarse grained sediment and shows as white and light gray tones. Low backscatter or low reflectivity, a weak acoustic return, characterizes finer sediments and shows as black and dark gray tones (Lurton 2002; Oliveira and Hughes Clarke 2007). Nearest to the shoreline the backscatter has the weakest acoustic return, suggesting a seabed of fine sediments such as mud or clay. The area where the pockmarks are located shows a moderate backscatter, medium grey level, suggesting a seafloor of coarser grained sediments. The rest of the bay shows lower backscatter which can indicate a seabed composed of softer sediments such as sand, silt and mud.

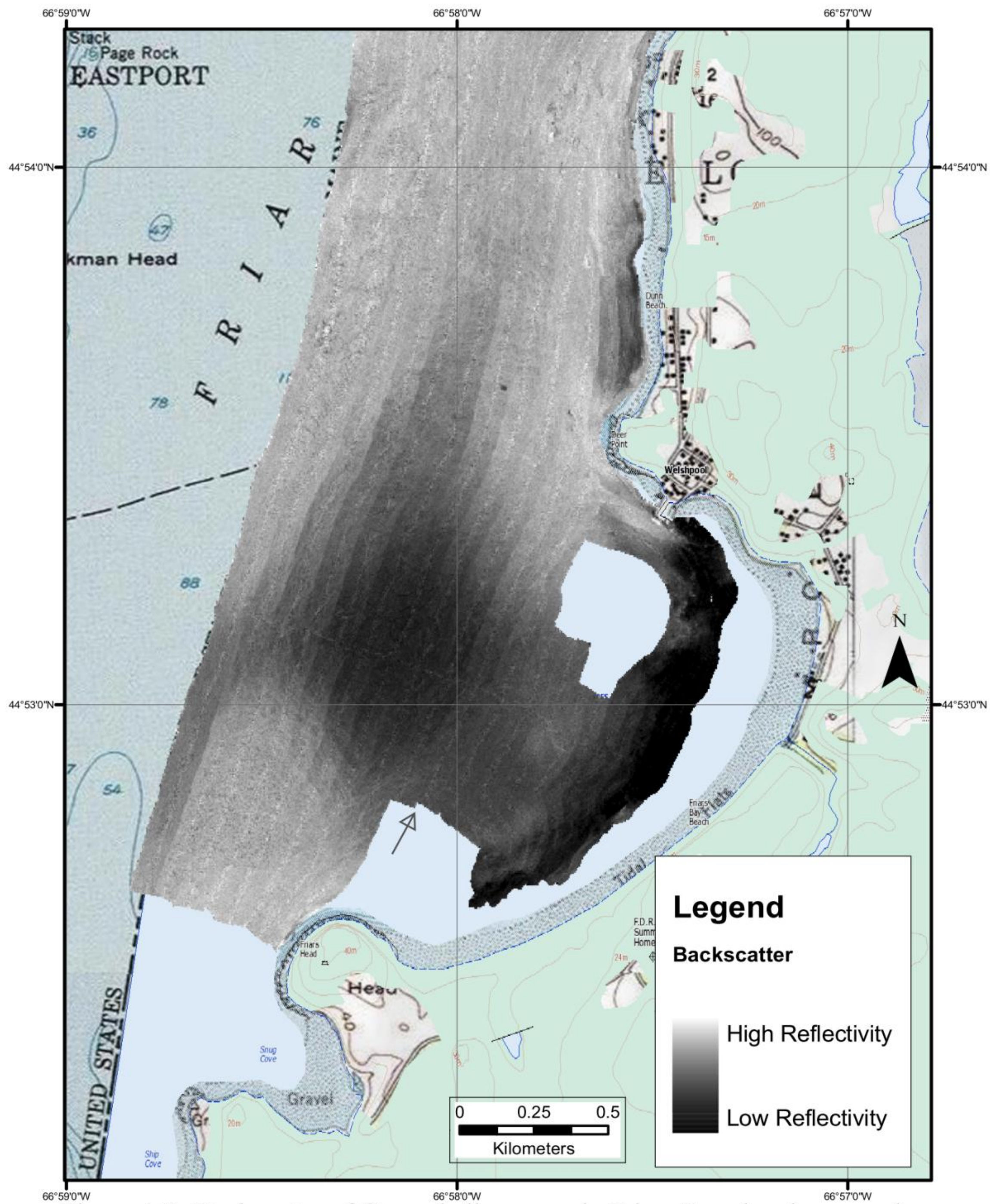

Figure 4.5: Backscatter of Campobello survey in Friars Bay showing a weak acoustic return over most of the bay. The two unmapped areas in the bay are fish farms. Arrow points to a line parallel to shore, this and other parallel lines are artefacts due to data collection and are not landform features.

4.2.2 Campobello Survey Stratigraphy

Three units are recognized from the seismic reflection profiles. Unit 1 is the lowest unit observed, and is interpreted as bedrock (BR). It is characterized by a strong even high intensity return from a highly irregular surface and generally displays a peak and valley morphology. It is present in all of the survey lines; however the bedrock often dips below the maximum achieved penetration of the 3.5 kHz echosounder (Figure 4.6).

Unit 2 overlies the bedrock. The unit is conformably draped over the bedrock unit and displays strong stratification (Figure 4.6). It was found in all of the sub-bottom lines and is interpreted as glacimarine (GM). The bounding surface is irregular in intensity, showing both high and low returns. The stratification varies in thickness (Figure 4.7). The sub-bottom line nearest to shore shows a pattern of internal reflectors that are concave up. The next two lines show two concave up sections separated by a bedrock peak (Figure 4.7). The two separated sections then merge together in the subsequent sub-bottom lines further offshore (Figure 4.7). At the south end of the track lines, the bedrock highs are overlain by concave up internal reflectors (Figures 4.6 and 4.7). Within this unit are found occasional ice-rafted debris (Figure 4.7). In several of the seismic lines pockmarks are seen in profile, showing v-shaped features of 1 to 1.5 metres deep on the seabed surface (Figure 4.6). Internal reflectors (stratification) are thought to represent alternations in the sand and silt content of the sediments, perhaps resulting from strong seasonal effects on sedimentation (Belknap et al. 1987; Belknap and Shipp 1991; and Bacchus 1993).

Unit 3, the uppermost unit, is acoustically transparent and is observed in the north end of the track lines (Figures 4.6 and 4.7). This unit is interpreted as representing

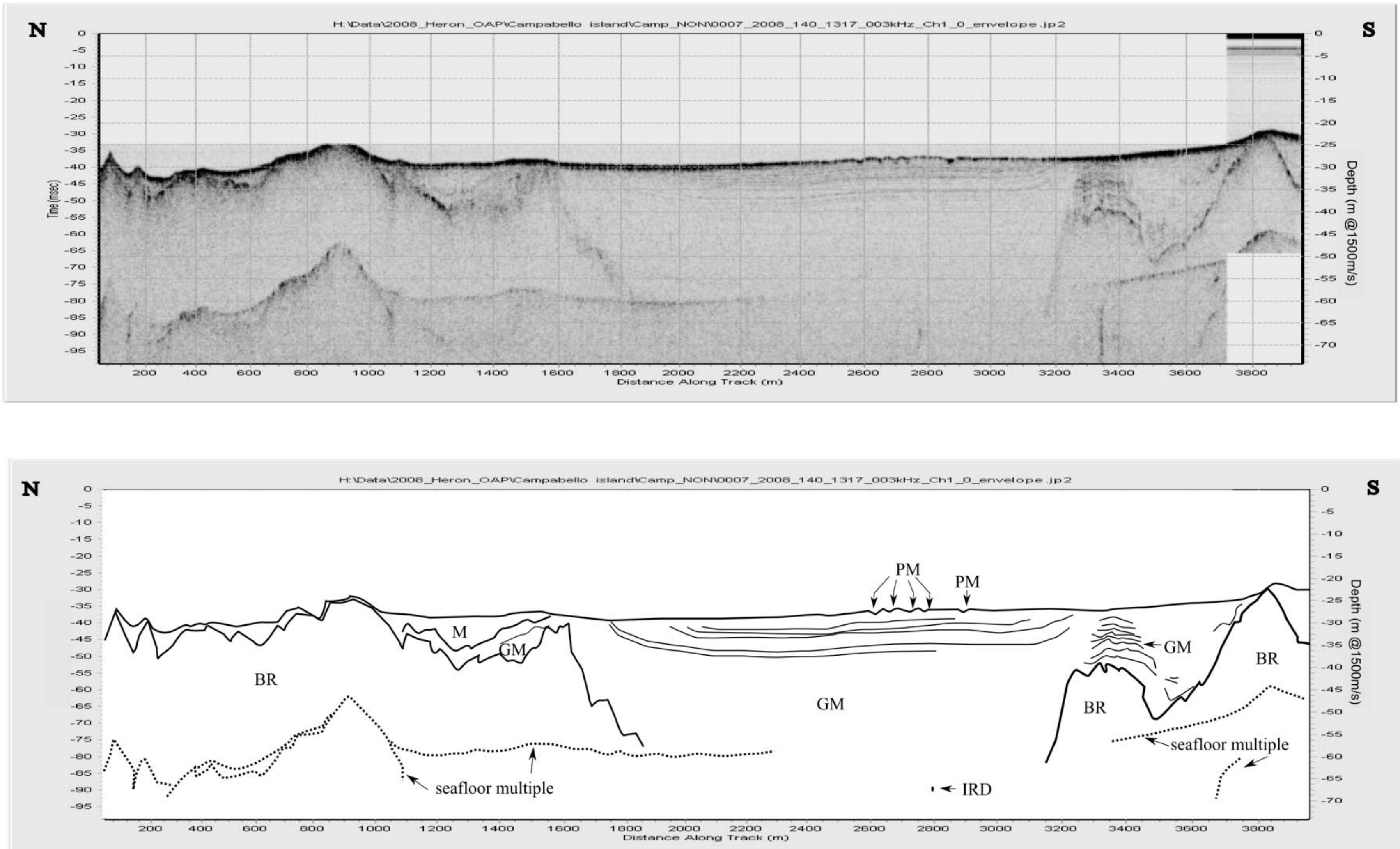

Figure 4.6: Seismic track line JD140_1317 for the Campobello survey area. (A) Original record with (B) the interpreted line drawing. Black arrows point to pockmarks. BR = bedrock, GM = glacimarine sediment, IRD = ice rafted debris and M = Holocene mud . Note this section shows a channel structure with stratification representing alternations in the sand and silt content.

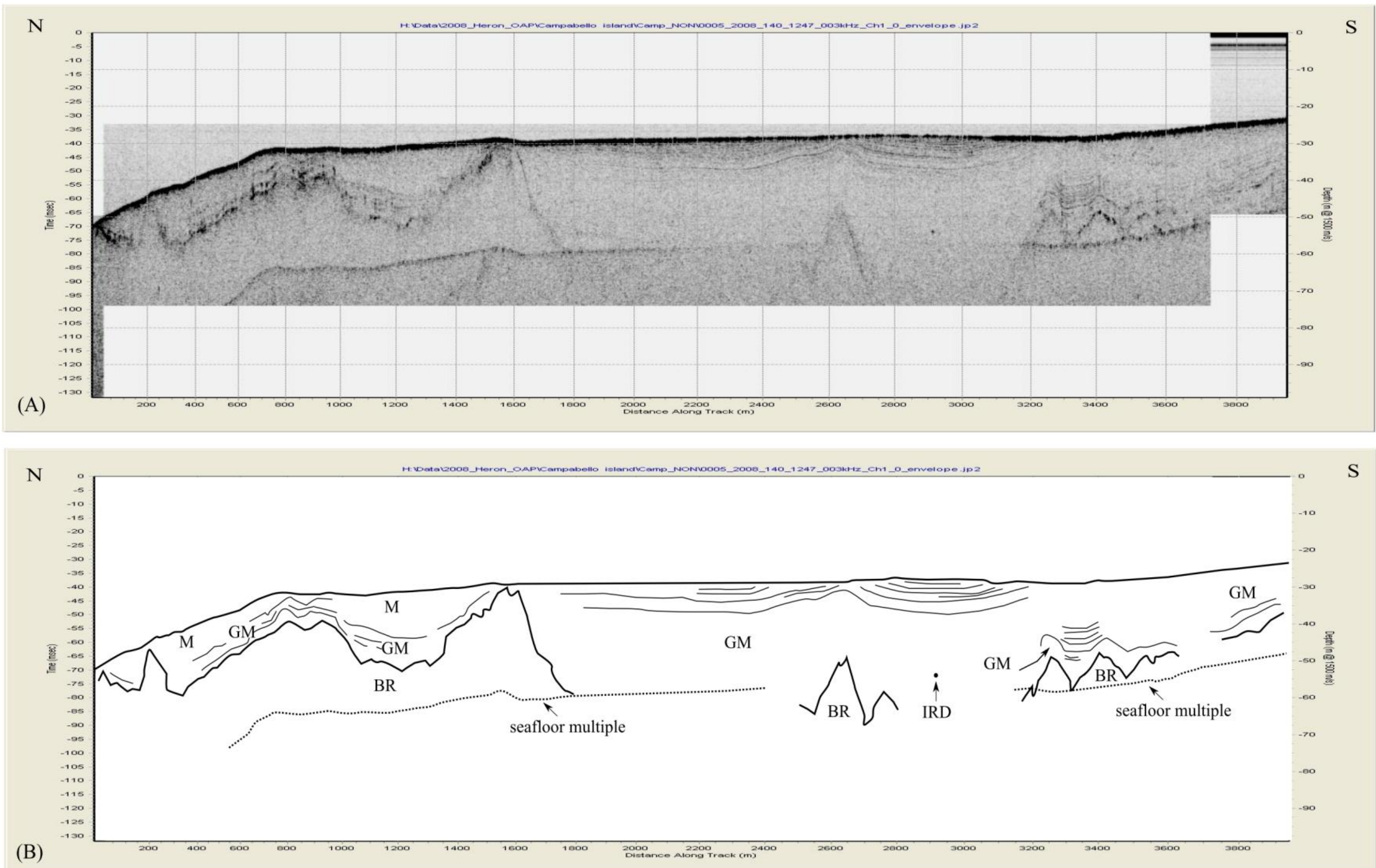

Figure 4.7: Seismic track line JD140_1247 for the Campobello survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, IRD = ice rafted debris and M = Holocene mud. Note this section shows a channel structure with stratification representing alternations in the sand and silt content.

postglacial conditions of mud deposition during the Holocene and is given the designation M.

4.2.3 Passamaquoddy Bay Survey Area

Passamaquoddy Bay is located on the northwest shore of the BOF close to the New Brunswick and Maine border (Figure 1.4). It is a large estuary, with the mouth of the bay restricted by a chain of islands: the two largest are Deer Island and Campobello Island. The main channels leading into the BOF are the Western Passage and Letite Passages. The largest rivers in this area that drain into the bay (Figure 1.6) are the St. Croix, Magaguadavic and Digdeguash rivers (Hachey and Bailey 1952).

The onshore bedrock geology (Figure 4.2) consists of weathered outcrops of Devonian and Silurian volcanic and sedimentary bedrock, with exposures along the coastal shoreline and the islands (Cumming 1968, NBDNR 2006). The region is seismically active, with historically documented earthquakes going back to 1817 (Burke 2004). There are two major strike slip faults in the region: the Lubec and Oak Bay Faults. Plots of the earthquake epicenters in the region form a lineament with a northwest to north trend paralleling the Oak Bay fault, suggesting that the seismic activity may be related to movement on this fault (Rast et al. 1979; Gates 1984).

The surficial sediments (Figure 4.2) are made up of a discontinuous veneer (less than 0.5 m thick) of mainly stony till over bedrock along the shoreline (NBDNR 2006). The Pineo Ridge Moraine complex found in Maine continues across the southern end of Campobello Island (Figure 1.1, Kaplan 1999). Approximately 12 km to the north of Passamaquoddy Bay is the St. George kame end moraine (Figure 1.1).

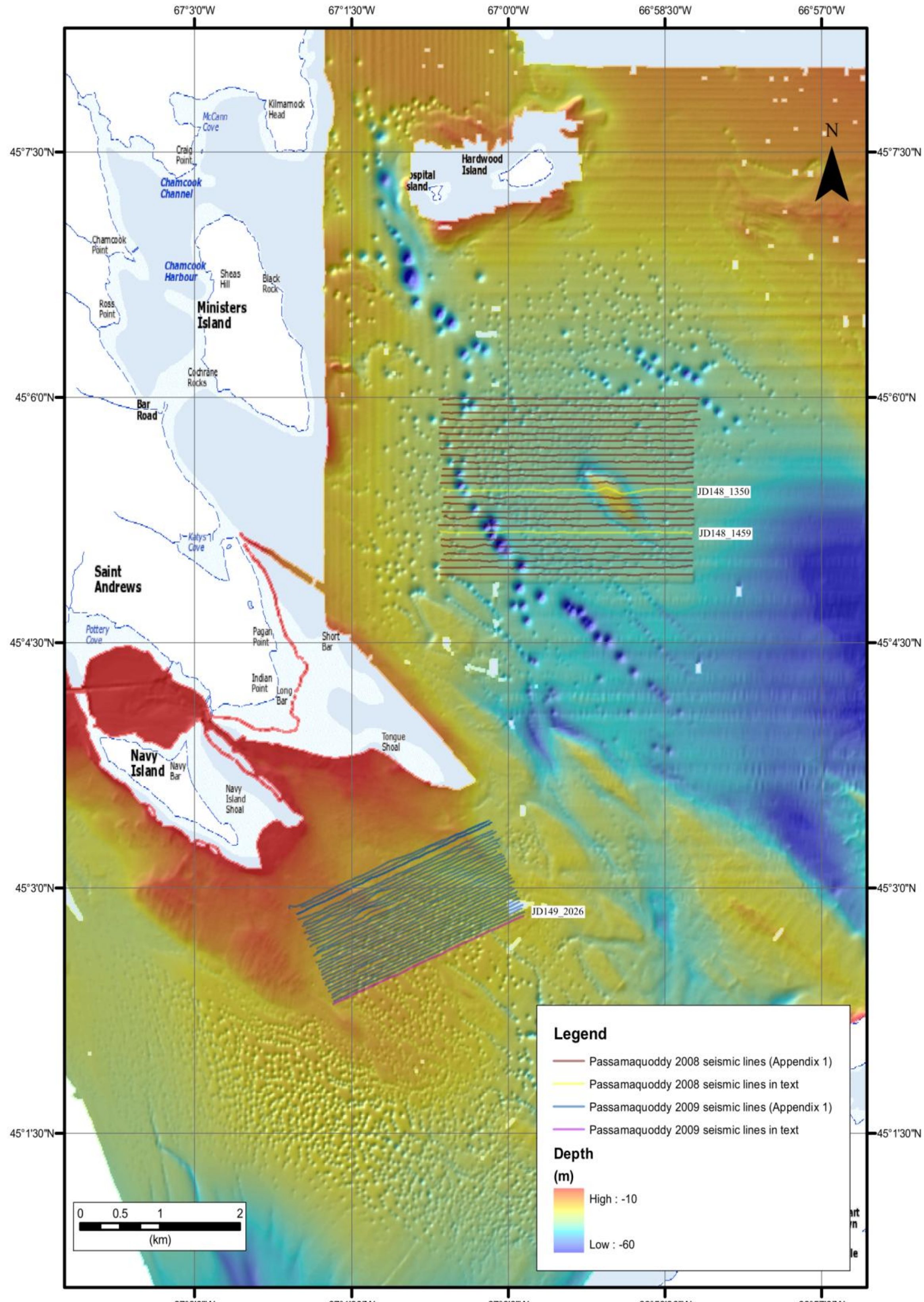

Figure 4.8: Bathymetric map showing Passamaquoddy Bay survey area, and seismic lines.

The floor of Passamaquoddy Bay is densely covered with pockmarks covering an area of 87 km^2; these pockmarks are concave and confined to the Holocene clay layer (Figure 4.9). The pockmarks in this area are calculated to number up to 11000, varying in diameter from 1 m to 300m and reaching depths up to 29 m from the seafloor (Pecore and Fader 1990; Wildish et al. 2008). The pockmarks appear as single features or as groups of longer linked chains. The highest density of pockmarks occurs in two large areas: 1) the central to northern part of the bay, and 2) between Navy and Deer islands, where they trend in linear chains in a predominantly northwest-southeast direction (Figure 4.9). The overall pattern of the pockmarks shows a northwest-southeast linear trend in the high density areas. Also, the linear pockmarks form as an outline around structural highs on the floor of the bay. For example, Figure 1.10 shows a drumlin outlined by a chain of pockmarks.

The offshore study area data are made up of two surveys conducted in 2008 and 2009 (Figure 4.8). The data consist of over 170 km of seismic lines, totaling 61 sub-bottom lines oriented in an east-west and northeast-southwest direction collected from the CSL Heron 2008 and 2009 cruises respectively (Appendix 1). The data also include 15 m resolution bathymetry and backscatter data for the 2008 cruise (Appendix 1).

The bathymetry of the area shows that water depth ranges from less than 2 m near shore to over 90 m. The water depth in the 2008 and 2009 survey area varies between 30 and 40 m, and 20 and 30 m (Figure 4.8). There are several rock drumlin-like features on the floor of the bay, all of which trend northwest-southeast (Figure 4.9). Figure 1.10 shows a cross-section of the drumlin along both the short and long axes.

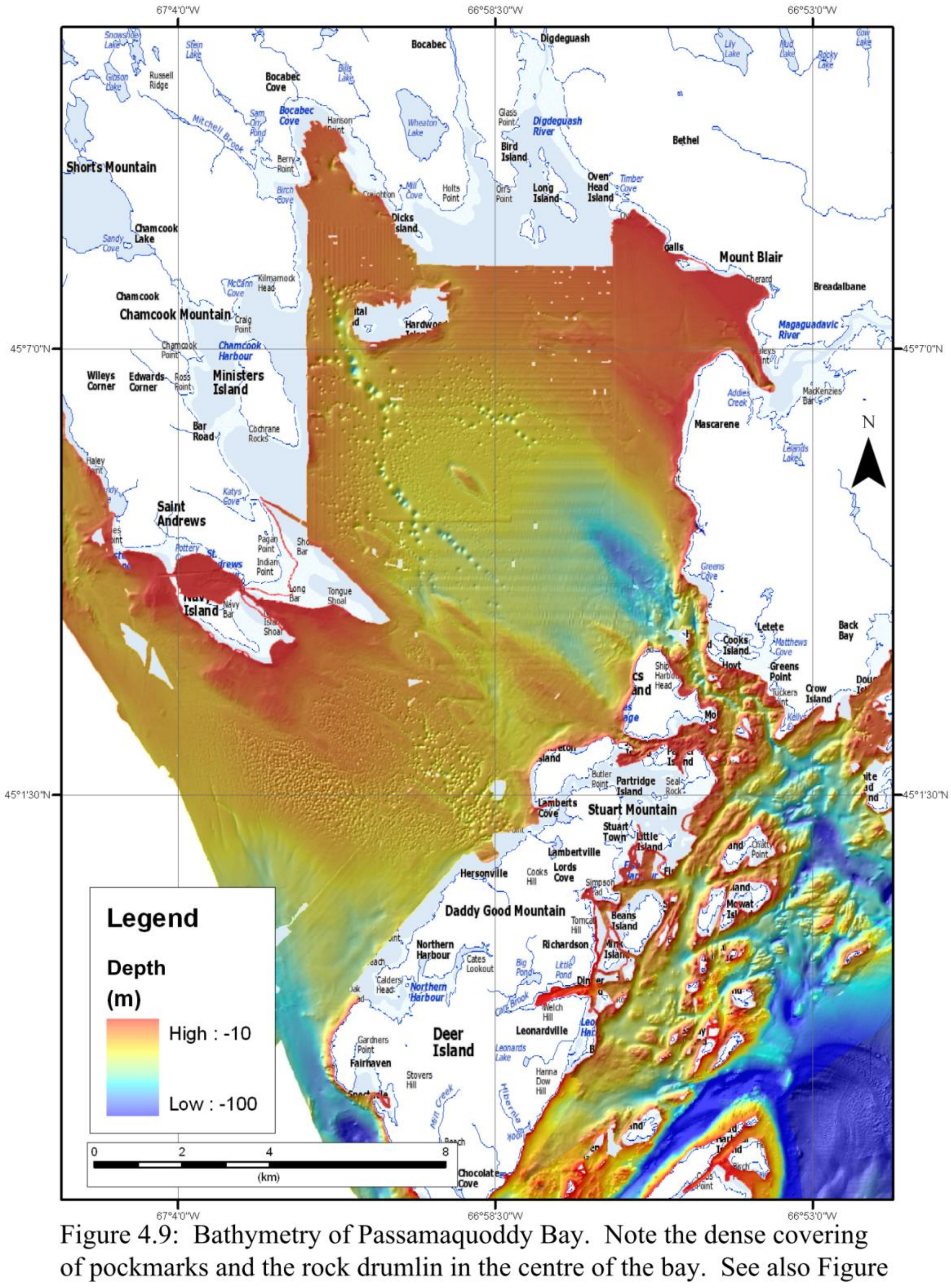

Figure 4.9: Bathymetry of Passamaquoddy Bay. Note the dense covering of pockmarks and the rock drumlin in the centre of the bay. See also Figure 1.10.

The Passamaquoddy survey areas backscatter map (Figure 4.10) shows regions of both high and low backscatter. A strong acoustic return (high reflectivity) shows as white and light gray tones, a weak acoustic return (low reflectivity) shows as black and dark gray tones. The majority of the backscatter in Passamaquoddy Bay is predominantly black and dark gray, low reflectivity, indicating a seabed composed of softer sediments such as sands and mud. However, there are also areas of strong acoustic return. The 2008 survey shows the drumlin with a high reflectivity indicating it is composed of bedrock or till (Figure 4.10). Also, many nearshore areas and channels have high reflectivity, likely submerged bedrock highs or areas where high current activity which has scoured away sediments to expose the bedrock. The heavily pockmarked area between the Navy and Deer Islands shows high points of reflectivity in the centres of the pockmarks; these are described as eyed pockmarks.

4.2.4 Passamaquoddy Bay Survey Stratigraphy

Unit 1 is the lowest unit observed from seismic reflection profiles, and is interpreted as bedrock (BR). For both the 2008 and the 2009 survey data it is characterized by a strong, high intensity return on a highly irregular surface. Figures 4.11 and 4.12 are examples from each survey years. The Passamaquoddy 2009 survey shows the BR unit to be closer to the seafloor surface compared to the 2008 survey where it has thicker sediments overlying it. The 2008 profile shows two rock drumlins within the survey area, one that emerges above the seafloor and one that is buried by the overlying sediments (Figure 4.13).

Unit 2 is observed overlying bedrock and is interpreted as till (T, Figure 4.11). The till surface exhibits a highly irregular morphology, and the upper bounding surfaces

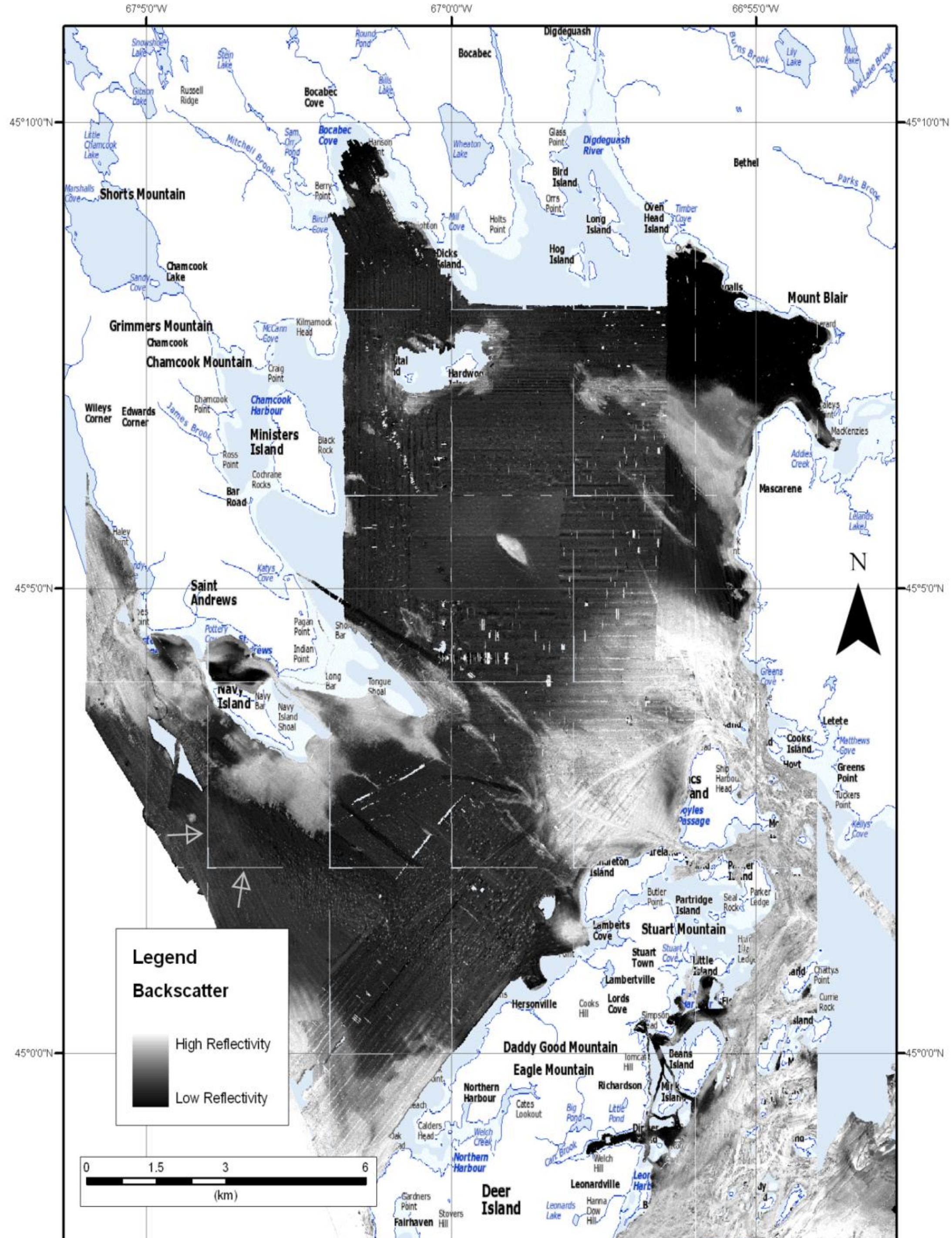

Figure 4.10: Backscatter of Passamaquoddy Bay showing a weak acoustic return over most of the survey areas. Note the strong acoustic return of the drumlin. Grey arrows point to artefacts due to data collection issues which are not landform features.

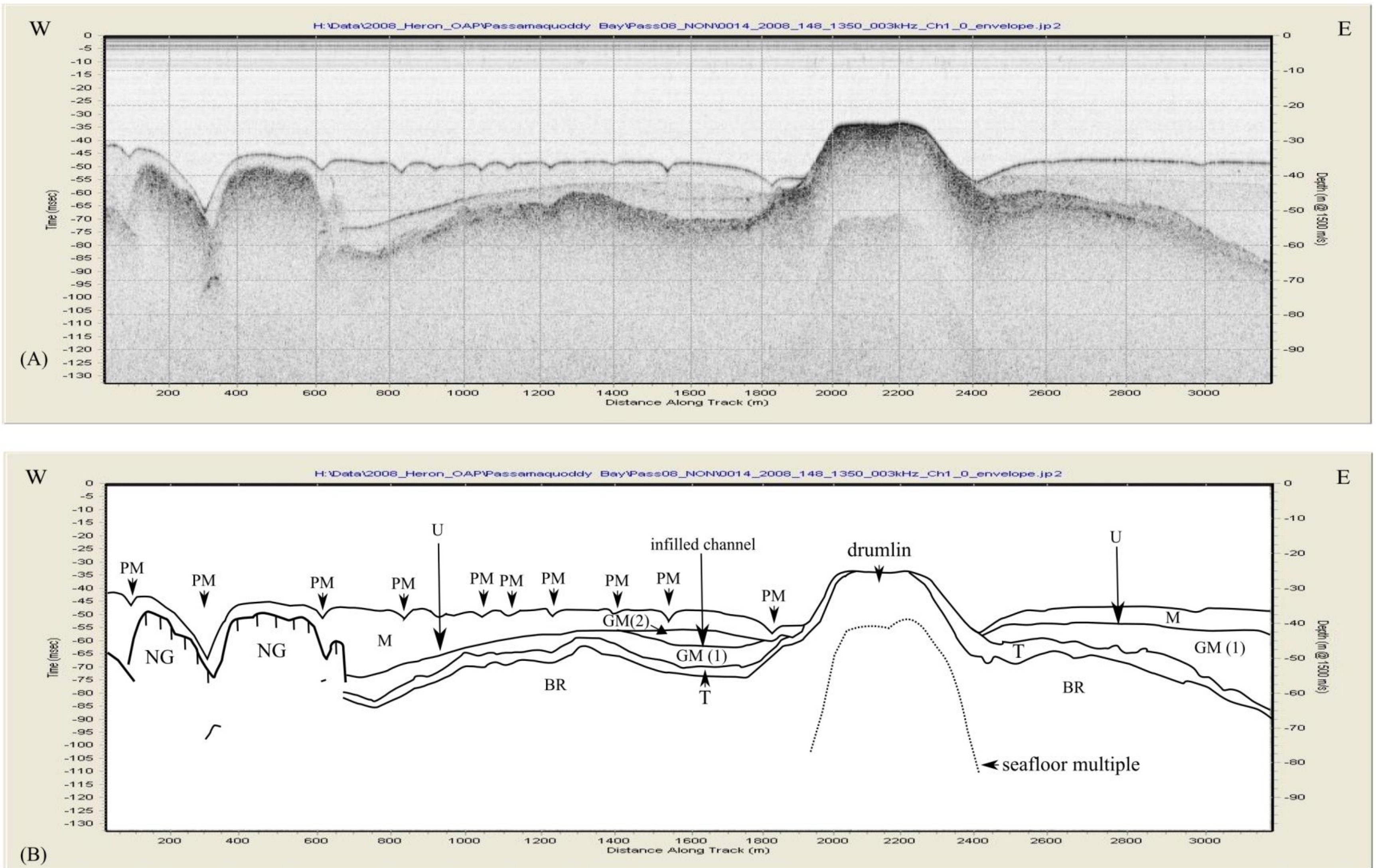

Figure 4.11: Seismic track line JD148_1350 for the Passamaquoddy 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity, M = Holocene mud and PM = pockmarks. Note infilled channel and the rock drumlin (see Figure 1.10).

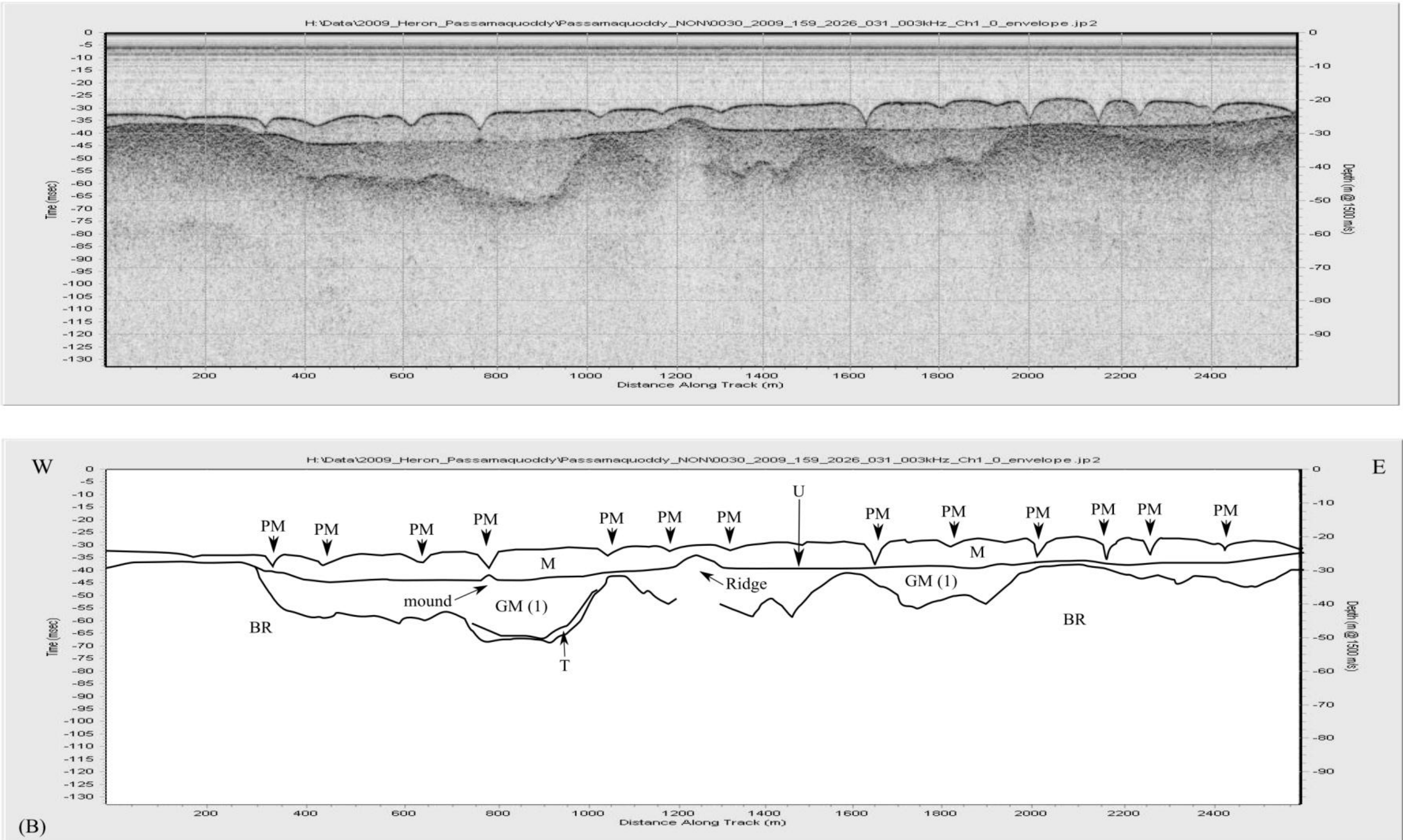

Figure 4.12: Seismic track line JD159_2026 for the Passamaquoddy 2009 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, PM = pockmark, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the mound directly beneath a pockmark at 600 m and the ridge at 1200 m of the line masking the underlying bedrock.

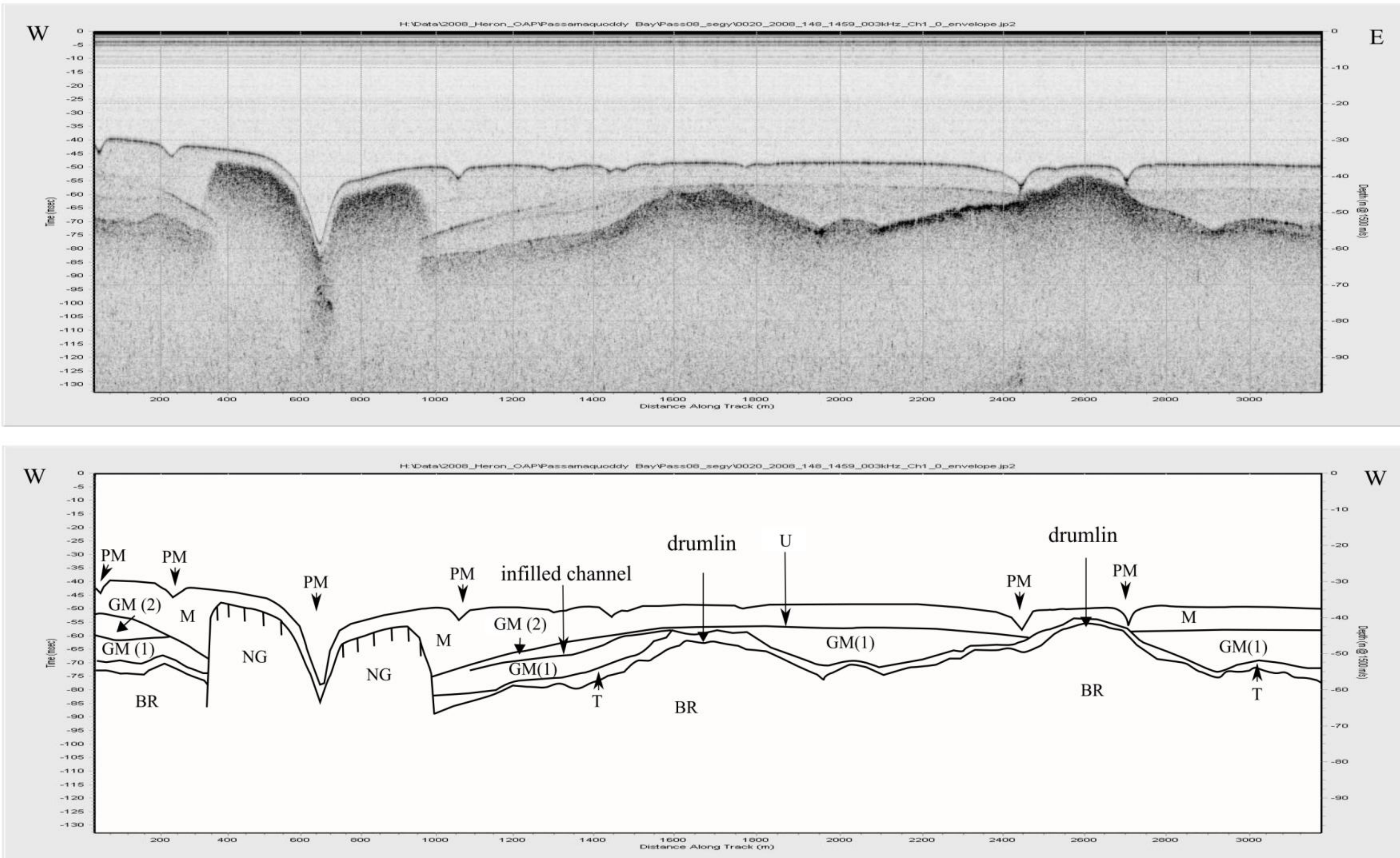

Figure 4.13: Seismic track line JD148_1459 for the Passamaquoddy 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity, M = Holocene mud and PM = pockmarks. Note infilled channel and rock drumlins.

have an intense return (Figure 4.11). In the 2008 survey area the till drapes conformably over the underlying BR unit (Figure 4.12). The rock drumlin located in the middle of the survey has the till draping on its flanks (Figure 4.12). In the 2009 survey area the till unit is present in lenses and does not appear to drape over the BR unit. Due to the similar intense return of bedrock and till, it may not be possible to distinguish till in some reflections.

Unit 3 overlies bedrock and till. It was found in all of the sub-bottom lines in both the 2008 and 2009 surveys; this unit is interpreted as glacimarine (GM-1). The GM-1 unit in the 2008 survey area is draped conformably over the lower units, except at structural highs where it does not cover the bedrock. The upper bounding surface has a moderate return. In the 2008 survey area an unconformity is recognized within 14 consecutive lines (U, Figure 4.11); it appears to be a channel located at 47 m below the sea level, and measures 400 m across and 2.6 m in depth (Figure 4.11). The GM-1 unit in the 2009 survey is either ponded or filling in the BR lows and thinly covering the BR highs (Figure 4.12).

Unit 4 overlies GM-1 only in the 2008 survey. It has a very weak surface return; as a result it is often difficult to see as a continuous surface (Figures 4.11 and 4.13). The lower bounding surface is more intense than the upper bounding surface. Where present it is ponded over GM-1. This unit is interpreted as glacimarine 2 (GM-2).

A ridge is found only in the 2009 survey area, overlying the GM-1 unit (Figure 4.12). The ridge is 585 m long, varying between 155 and 170 m wide, and between 5 to 6 m high. The structure is found in 12 consecutive lines and may extend further offshore into areas outside the survey. The upper bounding surface has an intense return, while the lower bounding surface is not seen. The ridge prevents penetration of the 3.5 echosounder to the bedrock unit directly below it; in Figure 4.12 the bedrock surface is not seen directly below it but can be seen to the east and west. This lack of penetration is due to the composition of the materials, likely of coarser materials such as sand and gravel. The ridge is oriented in a northwest-southeast direction and is surrounded on both flanks by pockmarks, but pockmarks are not present over the ridge itself. The linear nature of the feature and the line of pockmarks along its flanks suggest that the feature likely extends to a total length of 2000 m. Figure 4.14 shows the extent of the ridge mapped and likely extension due to the linear pattern of the pockmarks surrounding the ridge. The location of the ridge overlying the glacimarine unit suggests it may have been deposited during a localized readvance of the glacier in the area and it is recognized as a flute. Benn and Evans (1998) define glacial flutes as elongated streamlined ridges of sediment aligned parallel to former glacier flow.

Figures 4.11 and 4.13 show a feature that occurs as a draping darkened shadow with a very sharp surface return, and characterized by zones of incoherent reflections accompanied by lack of acoustic penetration; it is likely not a lithologic unit. Similar to such occurrences reported elsewhere (Fader 1997), this unit is interpreted as natural gas (NG). Shallow interstitial gas within marine sediments restricts seismic reflection imaging below the gas unit; this effect is called acoustic masking (Fader 1997). The gas attenuates the acoustic energy, blocking off penetration. Figure 4.14 shows the extent of the natural gas in the survey areas. It was only found in the 2008 survey, around the outer margins of the survey area.

Unit 5 is acoustically transparent, and lies flat except where pockmarks are

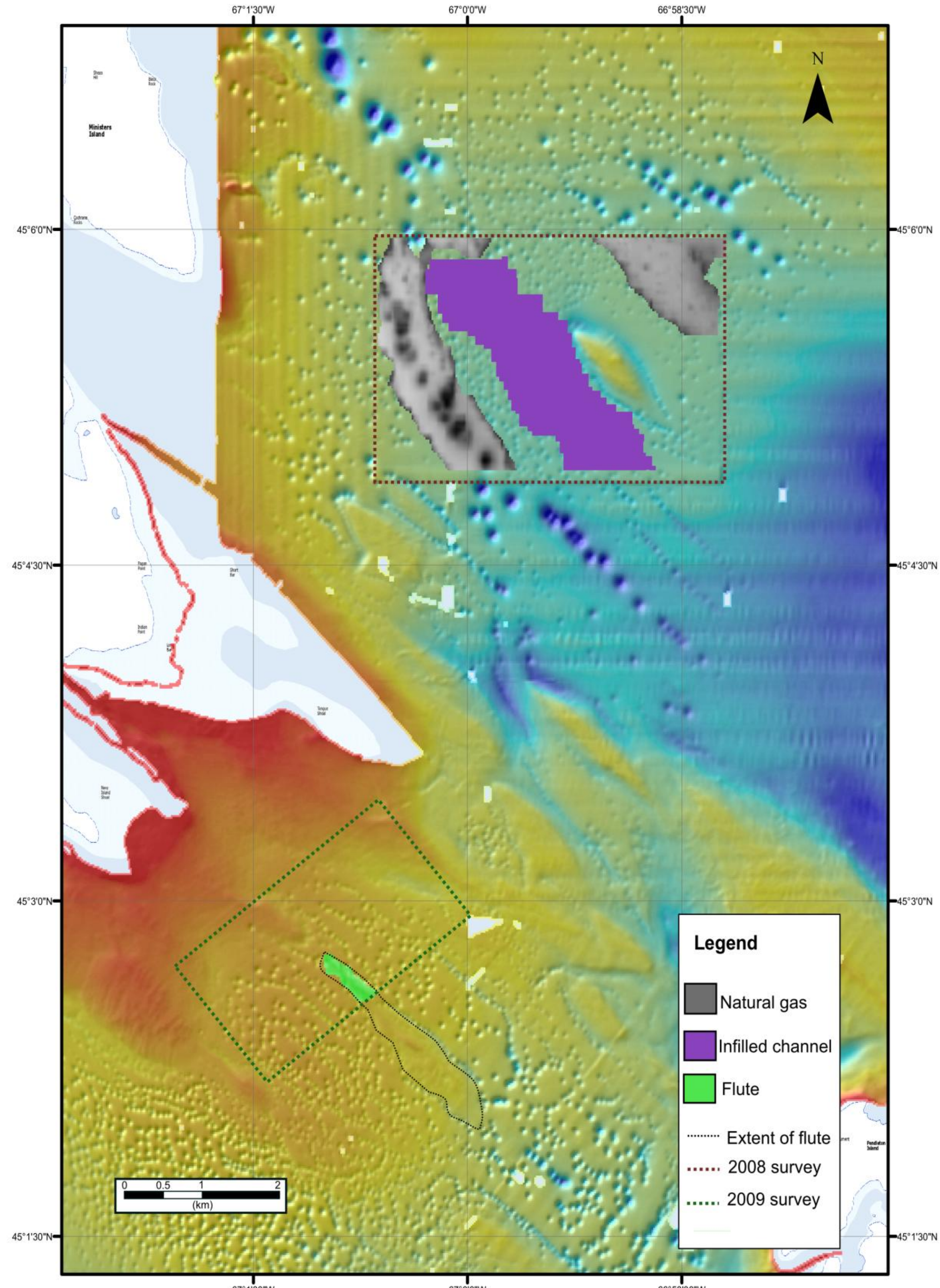

Figure 4.14: Location of natural gas, infilled channel, flute and likely extension of flute in Passamaquoddy Bay.

present. This unit is interpreted to represent postglacial conditions of mud deposition during the Holocene and is given the designation M. This unit is equivalent to King and Fader's LaHave clay(1986). The pockmarks are all developed in this unit and do not extend to any deeper unit. It is always the upper most seismic unit, and is found in all lines in both the 2008 and 2009 surveys (Figures 4.11 and 4.13).   Some of the pockmarks in the 2009 survey extend all the way to the top of the top of the underlying GM unit (Figure 4.11). Also, in some of the pockmarks are seen a cone shaped mound directly below the lowest point of the pockmark lying on top of the GM unit (Figure 4.11). These mounds may have formed during the formation of the pockmark, suggesting that some of the coarser debris fell back into the newly formed depression.

4.2.5 Pennfield Survey Area.

The Pennfield area is located to the east of the Passamaquoddy area and extends from Blacks Harbour to Red Head (Figure 4.15). The shoreline of the survey area is indented with several inlets, including Blacks Harbour, Deadmans Harbour, Seelys Cove and numerous smaller inlets. The largest fresh water input is from the Letang River emptying into Blacks Harbour located at the western end of the survey area (Figure 4.15). At the western end of the survey lie Green Island and Bliss Island and at the southern end lies a small archipelago called The Wolves.

The onshore geology (Figure 4.2) shows a mixture of Precambrian or Lower Paleozoic and Mississippian rocks.   The Precambrian or Lower Paleozoic group consists of granite, granodiorite, quartz diorite, gabbro, sedimentary, and volcanic rocks (NBDNR 2006).

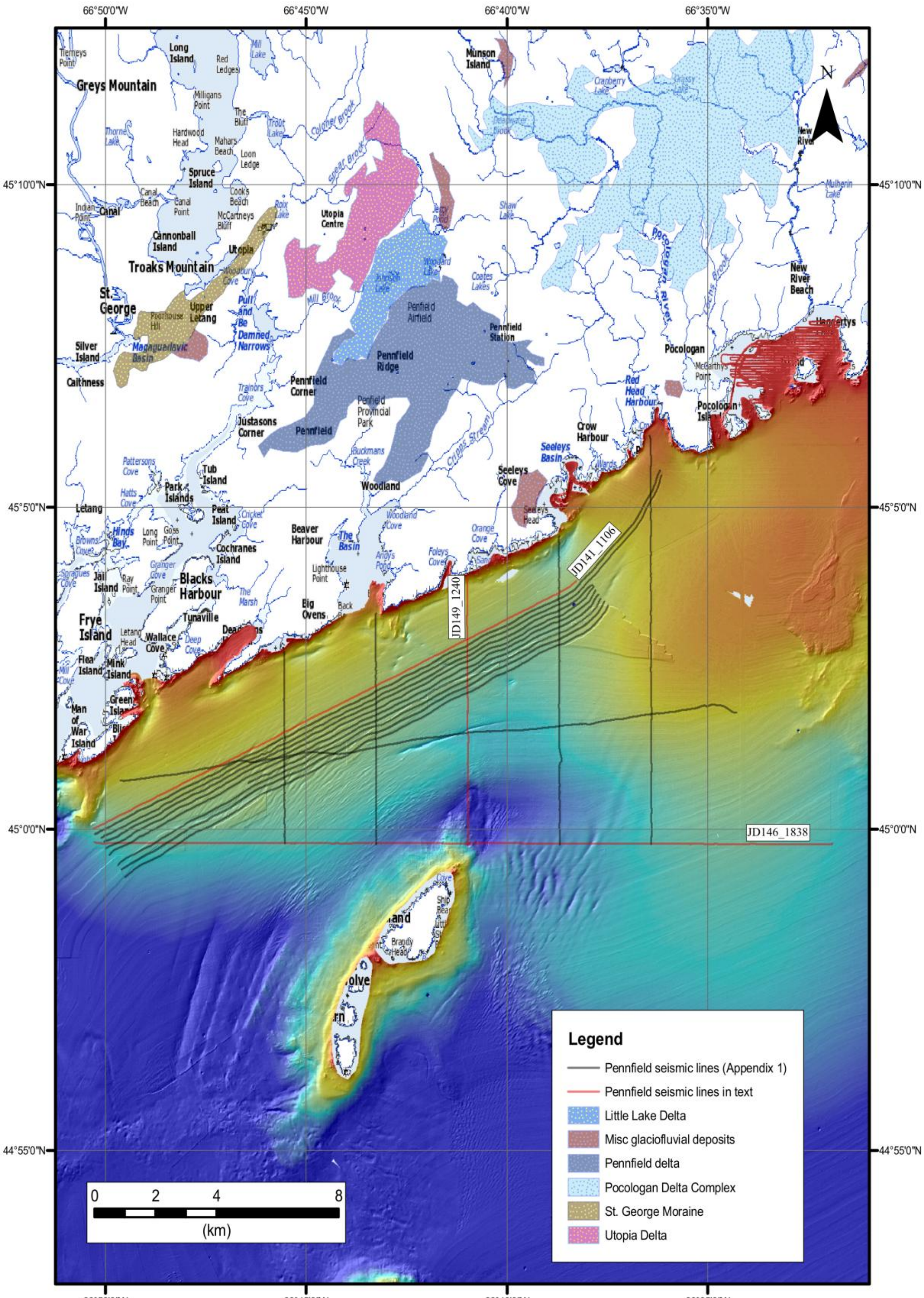

Figure 4.15: Bathymetric map showing the Pennfield survey area, seismic lines, and onshore glacial fluvial deposits. To the south of the survey lines are the Wolves Islands.

The surficial Quaternary geology of the area is dominated by the Pennfield-Pocologan delta complex located approximately 7 km to the north of the offshore study area, and the St. George moraine (Figure 1.1), discussed in the section 1.4 on Glacial geology.

The Pennfield survey totals 273 km of sub-bottom lines covering an area of approximately 50 km^2 (Figure 4.15). The survey includes 11 survey lines running parallel to the coast in a southwest-northeast direction (Appendix 1); 5 lines that intersect these lines in a north-south direction and 2 lines, run in an east-west direction, intersecting the majority of the SW-NE lines at the southwestern end. The data were collected in 2008 by the CSL Heron; the dataset includes 15 m resolution bathymetry and backscatter data (Appendix 1).

The bathymetry of the Pennfield survey displays a gradual increase in depth with increasing distance offshore (Figure 4.16). Located between Eastern Wolf Island and Foleys Cove is an elongated indentation trending southwest northeast, with the dimensions of 3.4 km length and 0.1 km the widest point. Several of the sub-bottom seismic lines go through this feature (Figure 4.15).

The Pennfield area backscatter (Figure 4.17) shows regions of both high and low reflectivity. The majority of the backscatter in the survey area is low, indicating a seabed composed of softer sediments such as sands and mud. There are areas of strong acoustic return surrounding The Wolves archipelago, indicating a seabed of rock outcrops.

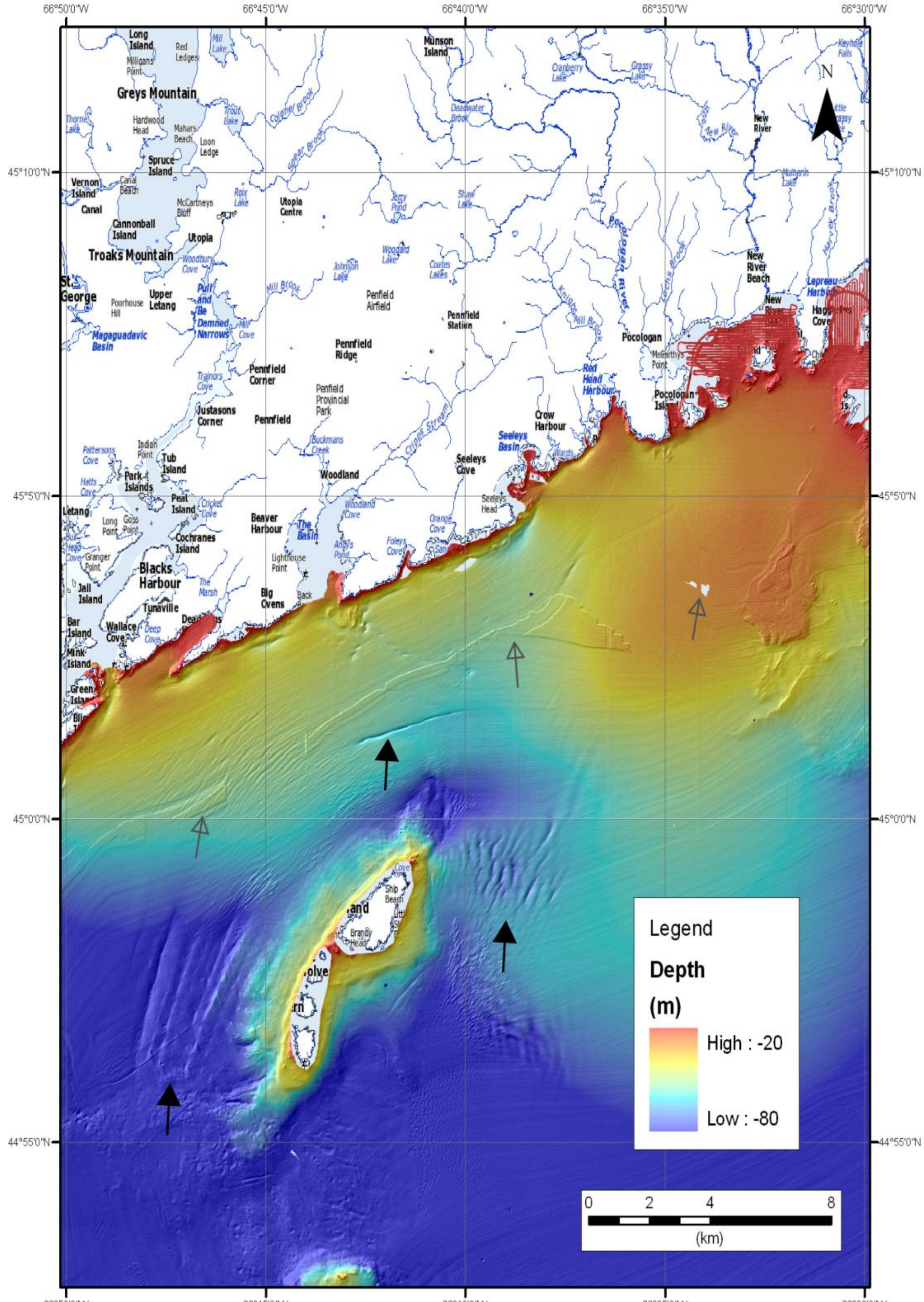

Figure 4.16: Bathymetry of the Pennfield survey area. Solid arrows point to elongated indentations surrounding the Wolves Islands. Grey arrows point to artefacts due to data collection issues which are not landform features.

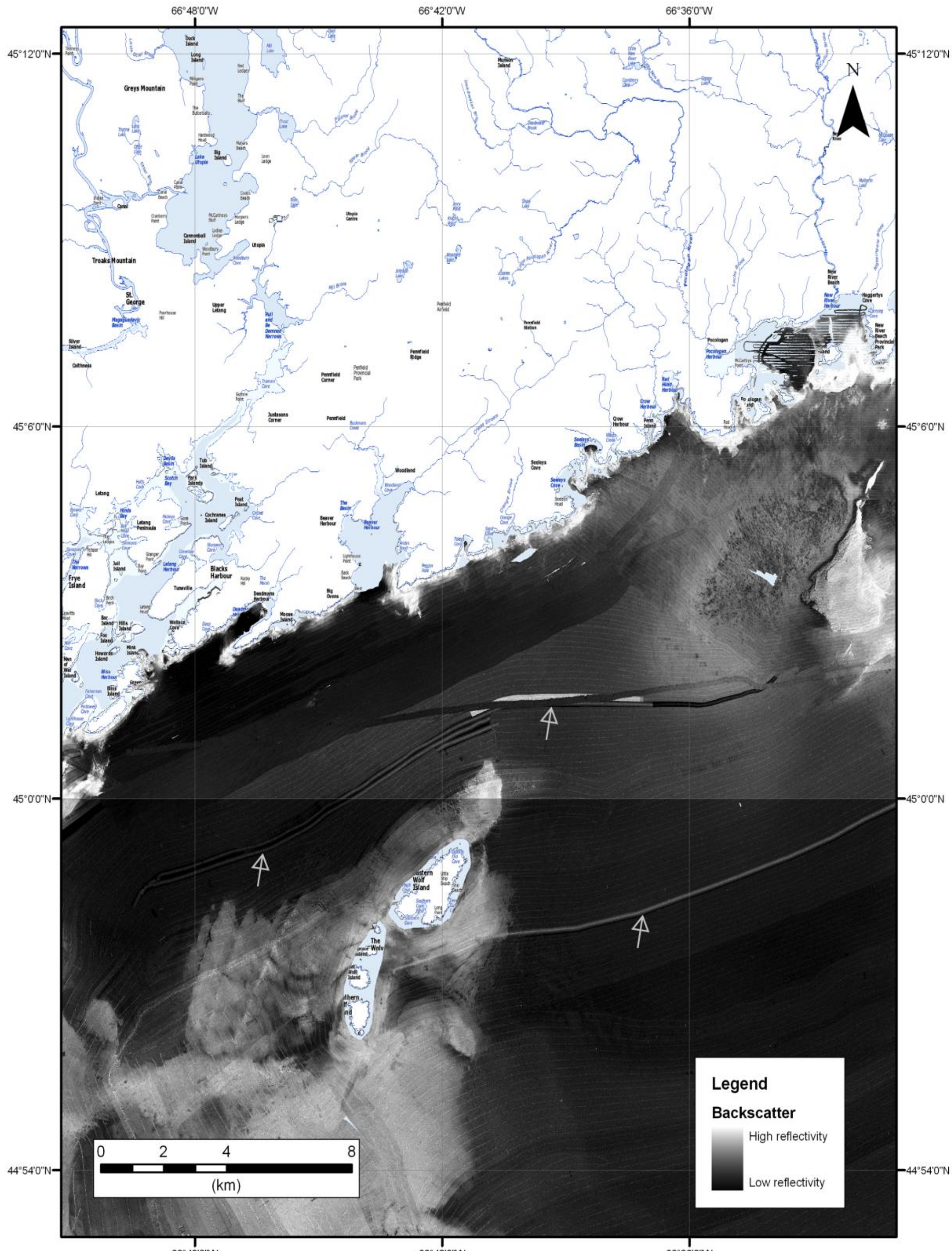

Figure 4.17: Backscatter of the Pennfield survey area showing low reflectivity over much of the survey area, indicating a seabed composed of softer sediments. Surrounding the chain of the Wolves Islands is an area of high reflectivity likely due to nearshore bedrock highs. Grey arrows point to artefacts due to data collection issues which are not landform features.

4.2.6 Pennfield Survey Stratigraphy

Unit 1 is the lowest unit observed (Figure 4.18), and is interpreted as bedrock. It is characterized by a strong, high intensity return on a highly irregular surface and is generally steeply dipping. The bedrock unit often can be traced from near shore outcrop highs (Figure 19) where it dips steeply offshore and then rises and then dips below maximum penetration achieved by the 3.5 kHz echosounder. In all of the SW-NE sub-bottom lines only a very small area of bedrock is observed in peaks where it appears to dip steeply beyond the depth of the seismic window (Figures 4.19 and 4.20).

Unit 2 is observed overlying bedrock in some areas and is interpreted as till (T, Figure 4.20). The till surface exhibits a highly irregular morphology, and both the upper and lower bounding surfaces have an intense return. Where observed the till drapes conformably over the underlying BR unit. Due to the similar intense return of bedrock and till, it is not possible to distinguish till in some reflections.

Unit 3 overlies till and is characterized by a moderate intensity return with stratification, displaying peak and valley morphology that is draped over the underlying bedrock (Figure 4.18). This unit is interpreted as glacimarine and is seen in all the survey lines (GM-1). The internal reflectors within this unit show variable intensities and are conformable with both the upper and lower bounding surfaces. The lower bounding surface has a more intense return than the upper bounding surface. Part of this unit is not seen as it dips below the maximum achieved penetration of the 3.5 kHz echosounder.

Unit 4 overlies GM-1; it is characterized by a weak intensity return, displays a ponded morphology, and is transparent with no visible internal reflectors. This unit is interpreted as glacimarine and is seen in all the survey lines (GM-2). The lower surface

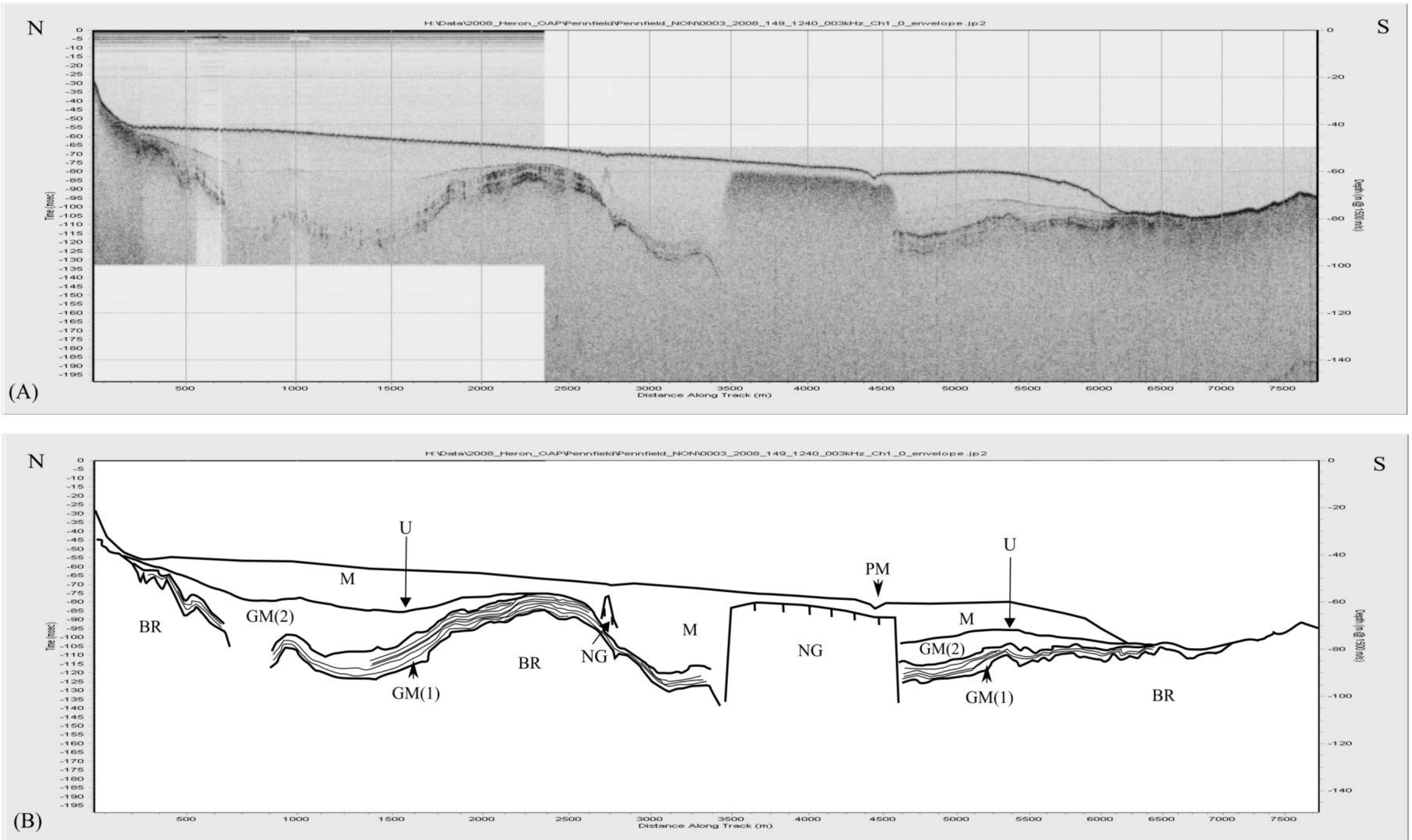

Figure 4.18: Seismic track line JD149_1240 for the Pennfield survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity M = Holocene mud and PM = pockmark. Note stratification in GM(1) representing alternations in the sand and silt content.

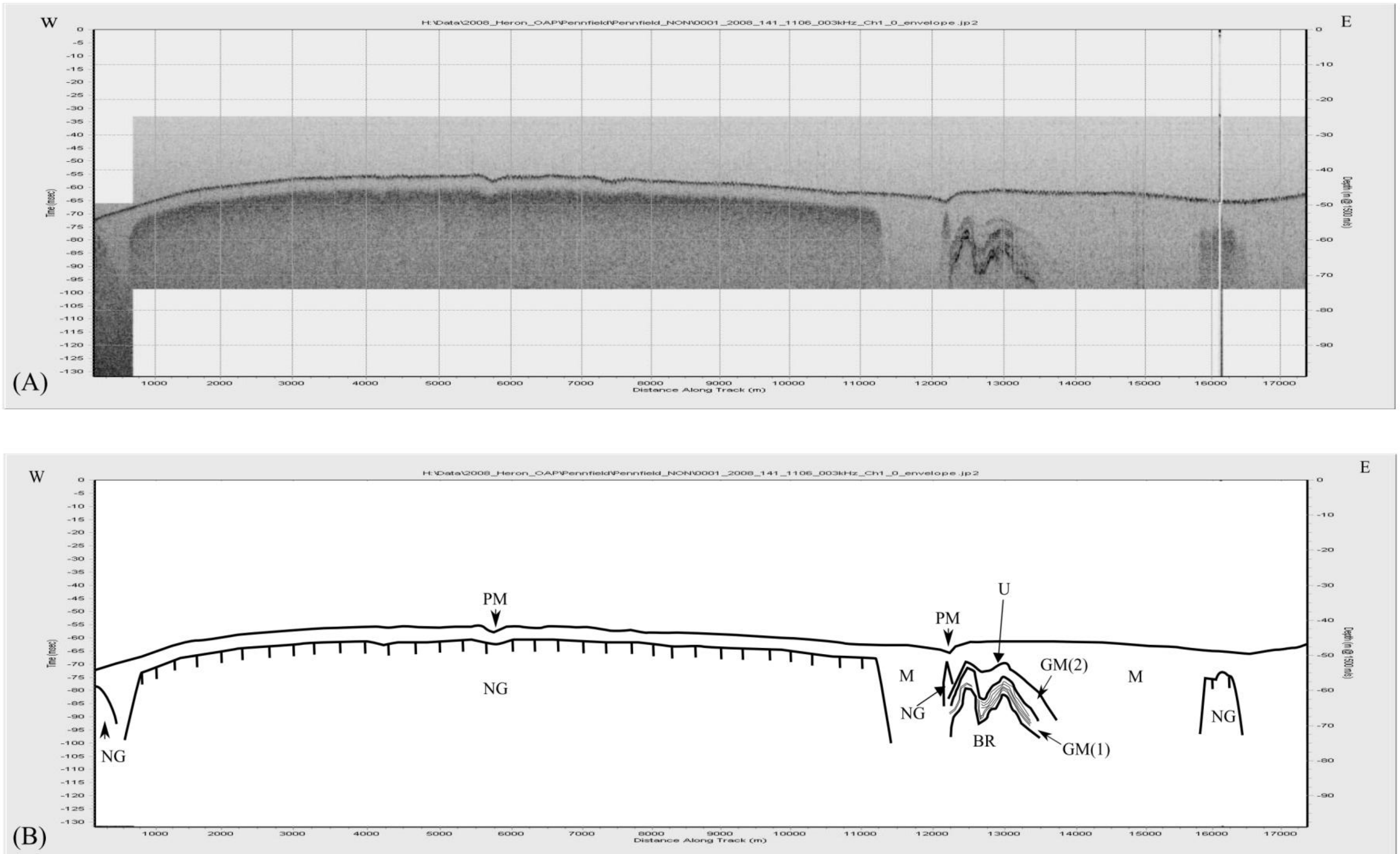

Figure 4.19: Seismic track line JD141_1106 for the Pennfield survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note stratification in GM(1) representing alternations in the sand and silt content.

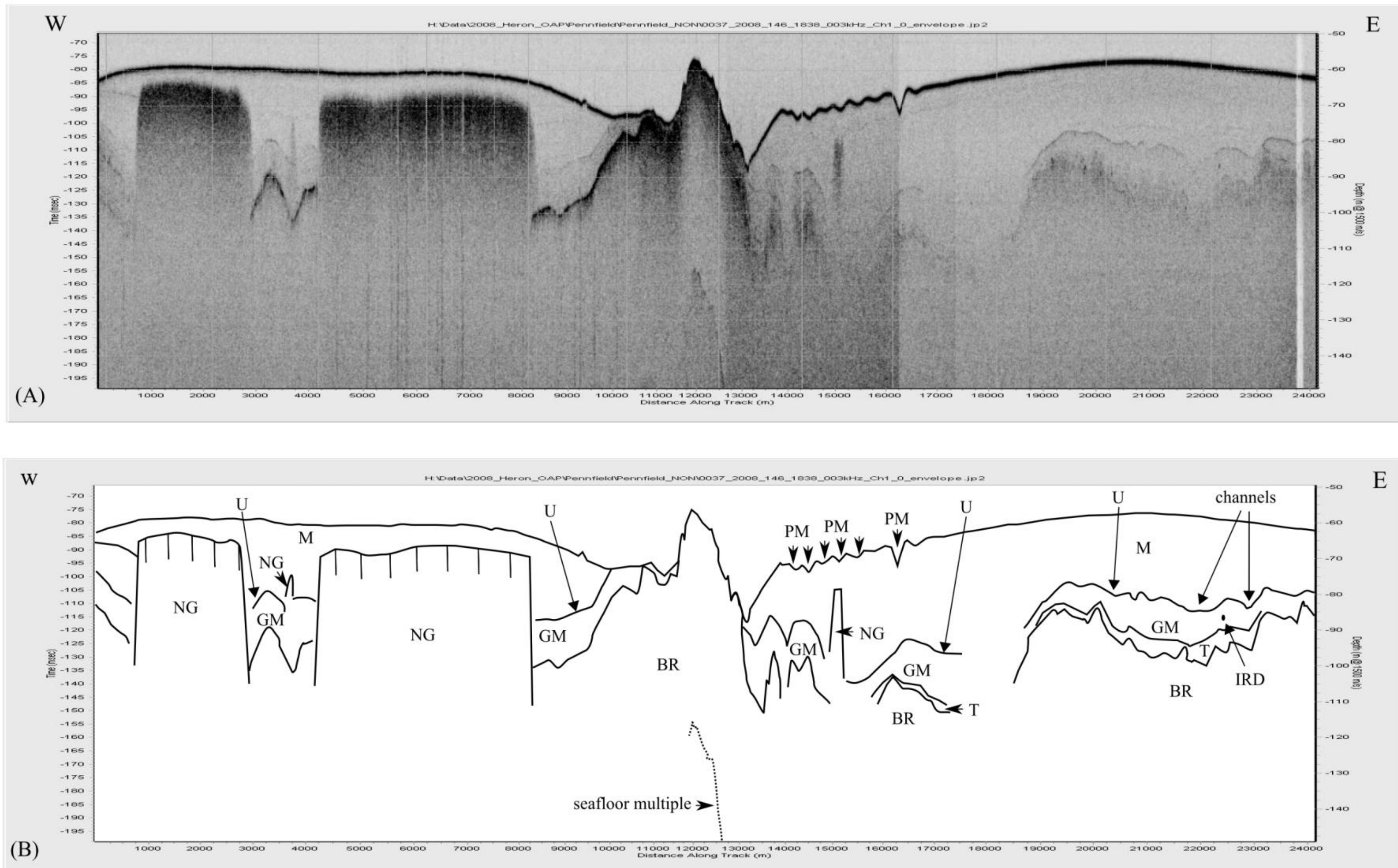

Figure 4.20: Seismic track line JD146_1838 for the Pennfield survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, IRD = ice rafted debris, NG = natural gas, U = Pleistocene/Holocene unconformity M = Holocene mud and PM = pockmarks. Note channels infilled with Holocene mud.

has a more intense return than the upper bounding surface; in some areas it is difficult to distinguish the upper bounding surface (Figure 4.18). An unconformity (U) is recognized at depths of -59 m to -89 m (Figures 4.18 and 4.20), and represents aerial and fluvial erosion.

A feature that occurs as a draping darkened shadow with a very sharp surface return, and characterized by zones of incoherent reflections accompanied by lack of acoustic penetration, is likely not a lithologic unit. It is interpreted as natural gas (NG), and when present obscures all other seismic units below, called acoustic wipeout (Davies et al. 1992), as seen in Figure 4.20. In the Pennfield survey it is present in large areas of the survey lines (Figure 4.19). Figure 4.21 shows the extent of the natural gas mapped in the Pennfield survey area.

Unit 5, the uppermost unit, is characterized as an acoustically transparent and ponded unit (Figure 4.20). It is interpreted as Holocene mud (M) and represents postglacial conditions. The lower bounding surface is marked by an erosional surface and is found at highs of -65 m and lows of – 95 m. Parts of the unit dip below the maximum achieved penetration of the 3.5 kHz echosounder. The elongated indentation seen in Figure 4.16 is seen in two of the sub-bottom lines, which intersect it (Figure 4.18). This feature is interpreted as an elongated pockmark. The natural gas unit below the pockmark shows an indentation directly below it, suggesting it may have resulted from an expulsion of gas. The elongated pockmarks have not been infilled by sediments either due to their recent formation or because they are preserved by the high current activity in this area.

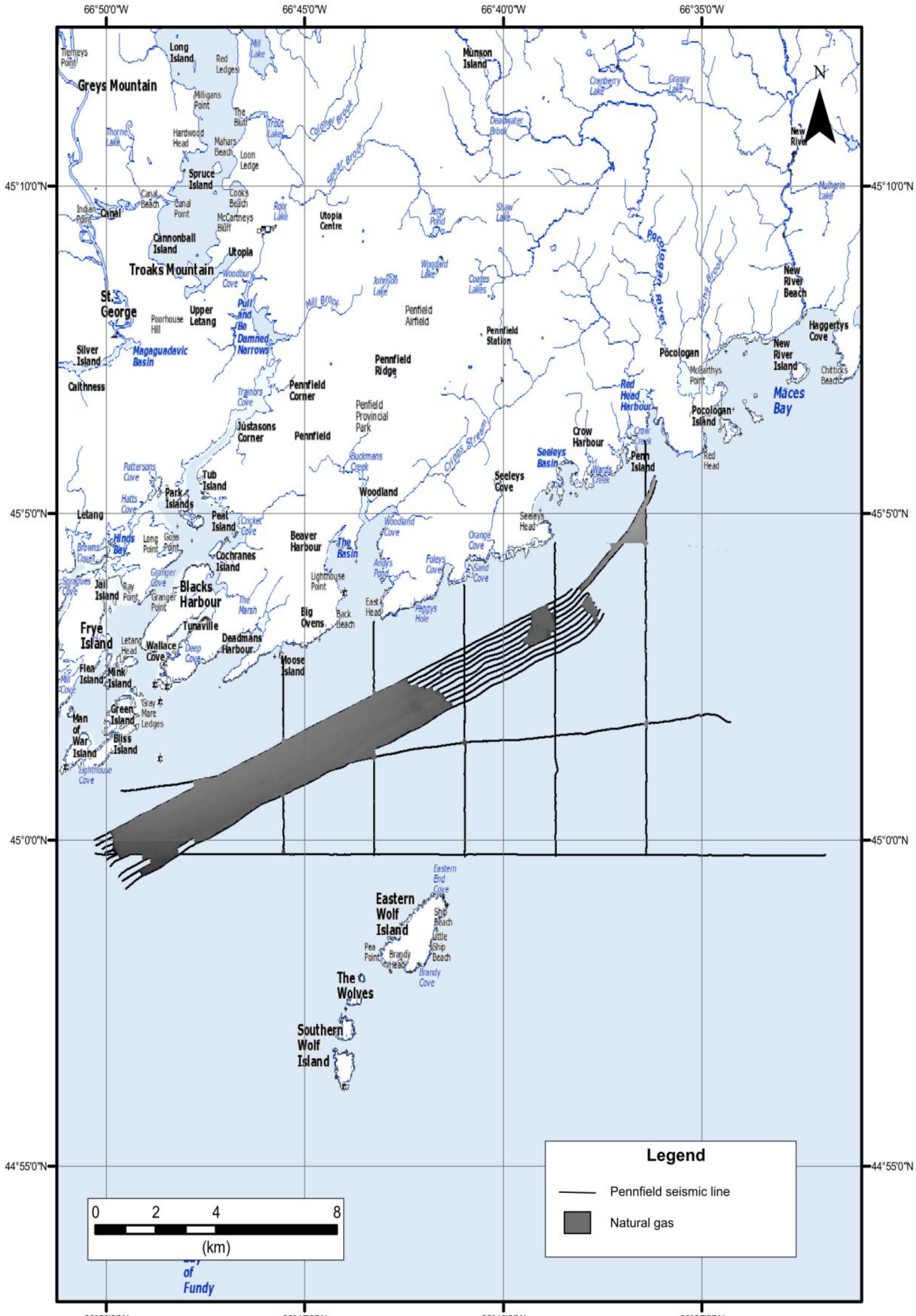

Figure 4.21: Location of natural gas in the Pennfield survey area.

4.2.7 Maces Bay Survey

Maces Bay is a large triangular-shaped coastal embayment on the northwestern coast of the BOF (Figure 4.22). The inshore area of the bay is bisected by New River Beach Provincial Park, which separates Carrying Cove and Haggarty's Cove. There are three islands in the bay, from east to west respectively they are the Pocologan, New River and Salked Islands. There are several small rivers that drain into the bay, the larger of which include Pocologan River, New River, and Lepreau River. Point Lepreau, a peninsula, juts out approximately 3 km at the eastern end of the bay, with the western end being Red Head (Figure 1.1).

The onshore bedrock geology of Maces Bay area (Figure 4.2) consists of Pennsylvanian and Triassic red to grey conglomerate, siltstones, and can include silicic to mafic intrusive rocks. The surrounding surficial Quaternary deposits consist of sand, silt and some gravel and clay along the New River beach side, and a veneer of stony till along the Pt. Lepreau side (NBDNR 2006).

Surrounding the bay onshore is an extensive melt-water complex, discussed in section 1.4 on the glacial history of the area, which was deposited during deglaciation. Directly to the north of Maces Bay lies the little studied Pocologan Delta complex. To the west are the Pennfield, Little Lake, and Utopia deltas, and to the east a series of kame moraines and the Sheldon Point moraine in Saint John (Figures 4.22 and 1.1).

The offshore study area data consist of 371 km of seismic data, totaling 66 sub-bottom profile lines (Appendix 1) oriented east-west and north-south, and collected from the CSL Heron 2008 and CSL Heron 2009 cruises (Figure 4.22). Of the total of 66 sub-bottom lines, 16 are in a north -south direction. The majority of the sub-bottom lines run

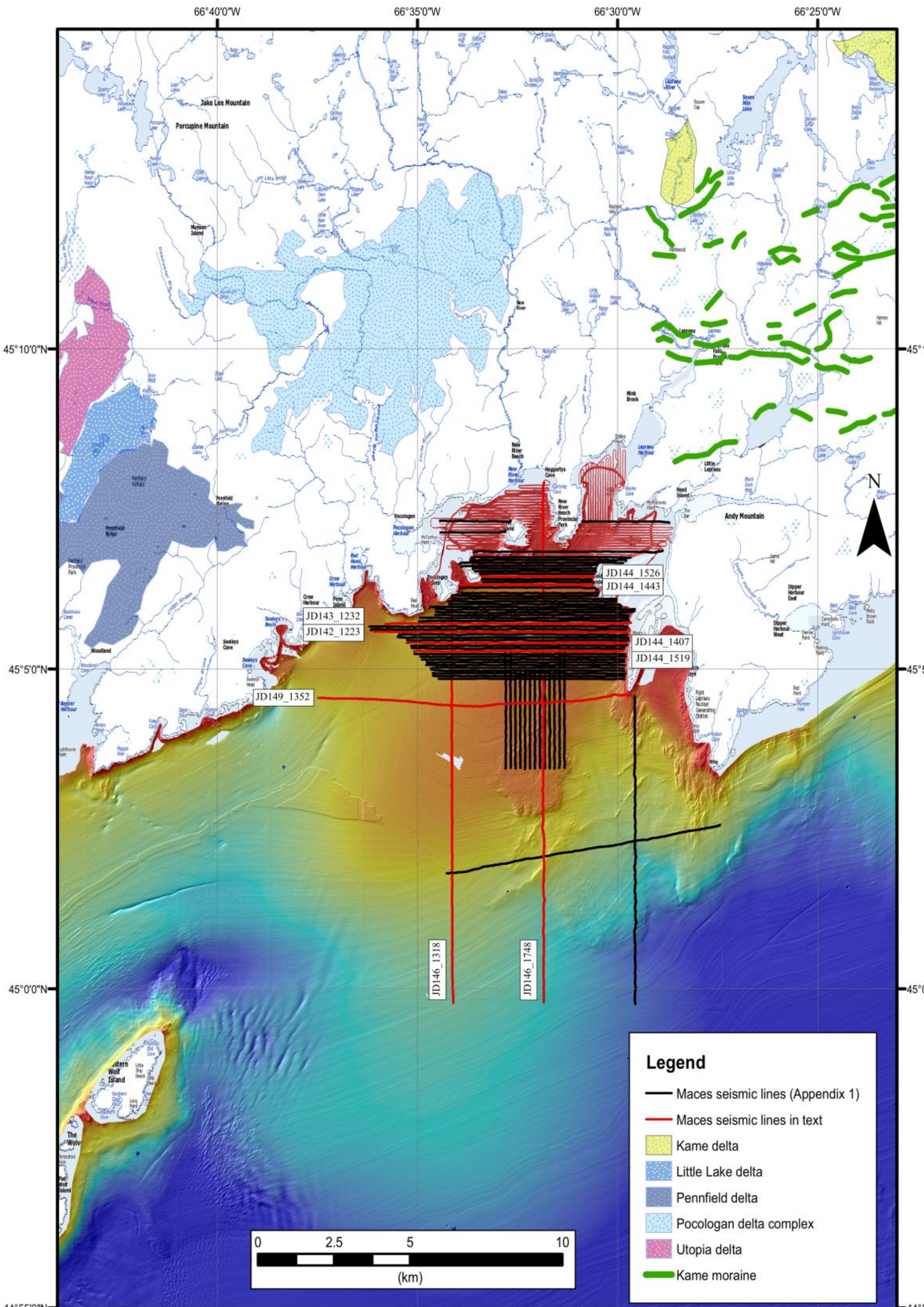

Figure 4.22: Bathymetric map showing the Maces Bay survey area, seismic lines, and onshore glacial fluvial deposits.

in an east-west direction transverse to the direction of glacier advance and retreat, and are located in the northern half of the bay. There are limited sub-bottom data on the southern portion of the bay. The data also include 15 m resolution bathymetry and backscatter data collected by the CSL Heron in the 2008 cruise (Appendix 1).

The bathymetry of Maces Bay displays a gradual uniform increase in depth with increasing distance offshore, from less than - 5m to approximately - 30 m (Figure 4.23). A pronounced feature on the bathymetry data is an arcuate shaped formation (Figure 4.24). Its total area is 5 km^2, with a maximum north-south dimension of 3.2 km and east-west of 2.3 km. At its northern end, is an elongated cardioid feature with both lobes oriented in a southwest northeast direction. There is also a long sinuous ridge like feature, 1.3 km in length, trending in a southwest direction following the western margin of the cardioid shape, which is interpreted as an esker. The larger, broader features may be kames and the wide lobate features are interpreted as a delta. To the northeast of this feature is an elongated feature trending in a southwest northeast direction. At the southern end it widens, and it tapers to the north, its dimensions are 1.1 km long, with width varying between 60 m at the narrowest and 190 m at the widest.

Maces Bay backscatter shows regions of both high and low backscatter (Figure 4.25). The arcuate shape in Maces Bay shows a high reflectivity, which can indicate a seabed either of coarse sand and gravel, or rock outcrops. The elongated feature to the northeast has a very high reflectivity, which indicates it is composed of sand and gravel, or rock outcrops. To the west of the arcuate feature there is an area of low backscatter with mottled high backscatter. Mottled backscatter is caused by a hummocky surface with variable topography and sediment distribution (USGS 2007). The rest of the bay

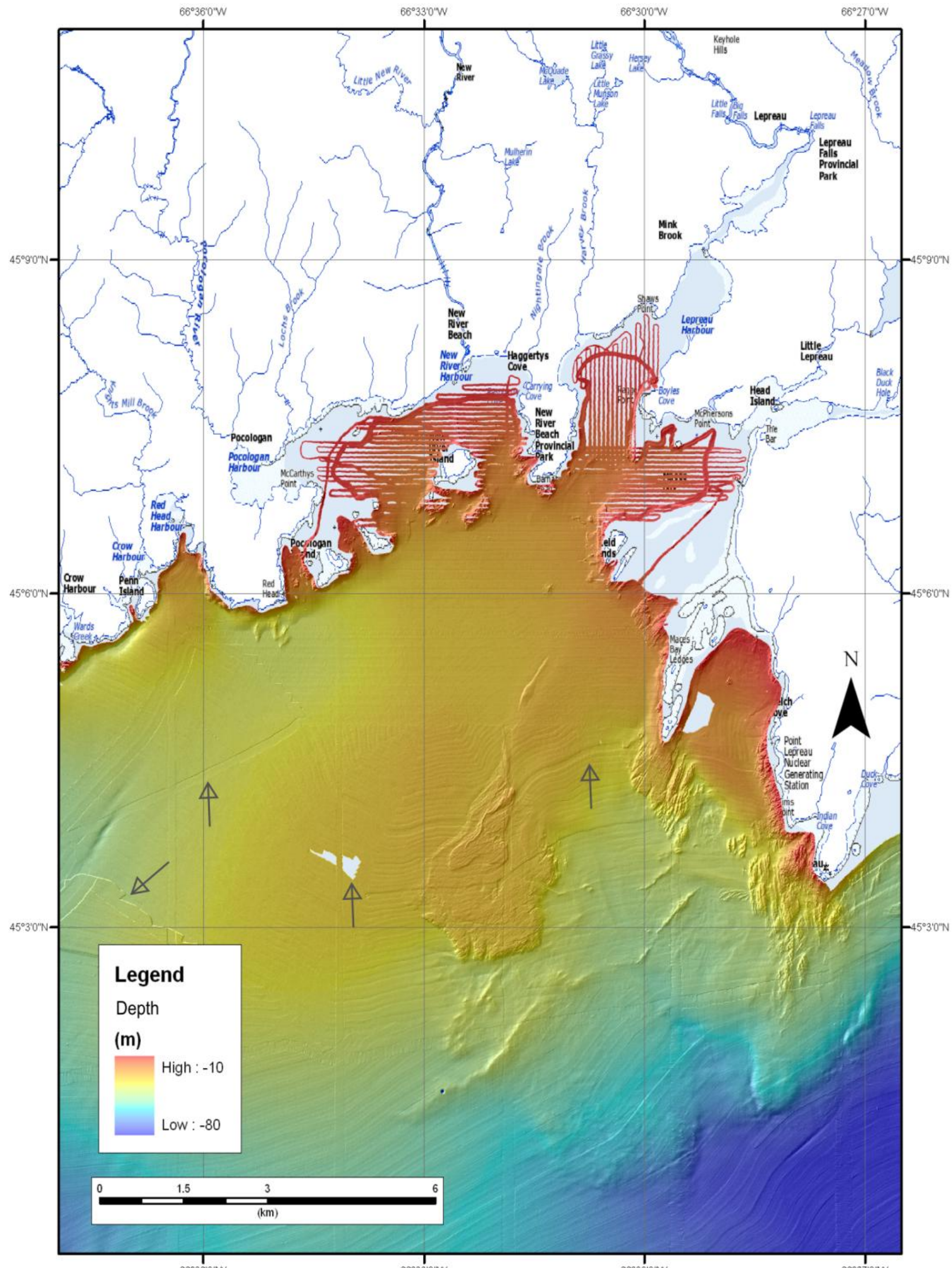

Figure 4.23: Bathymetry of Maces Bay. Note the cardioid shape at the southern end of the bay, with a sinuous ridge at the northwestern end on top; to the south of these features is a northeast trending ridge. Arrows point to artefacts due to data collection which are not landform features.

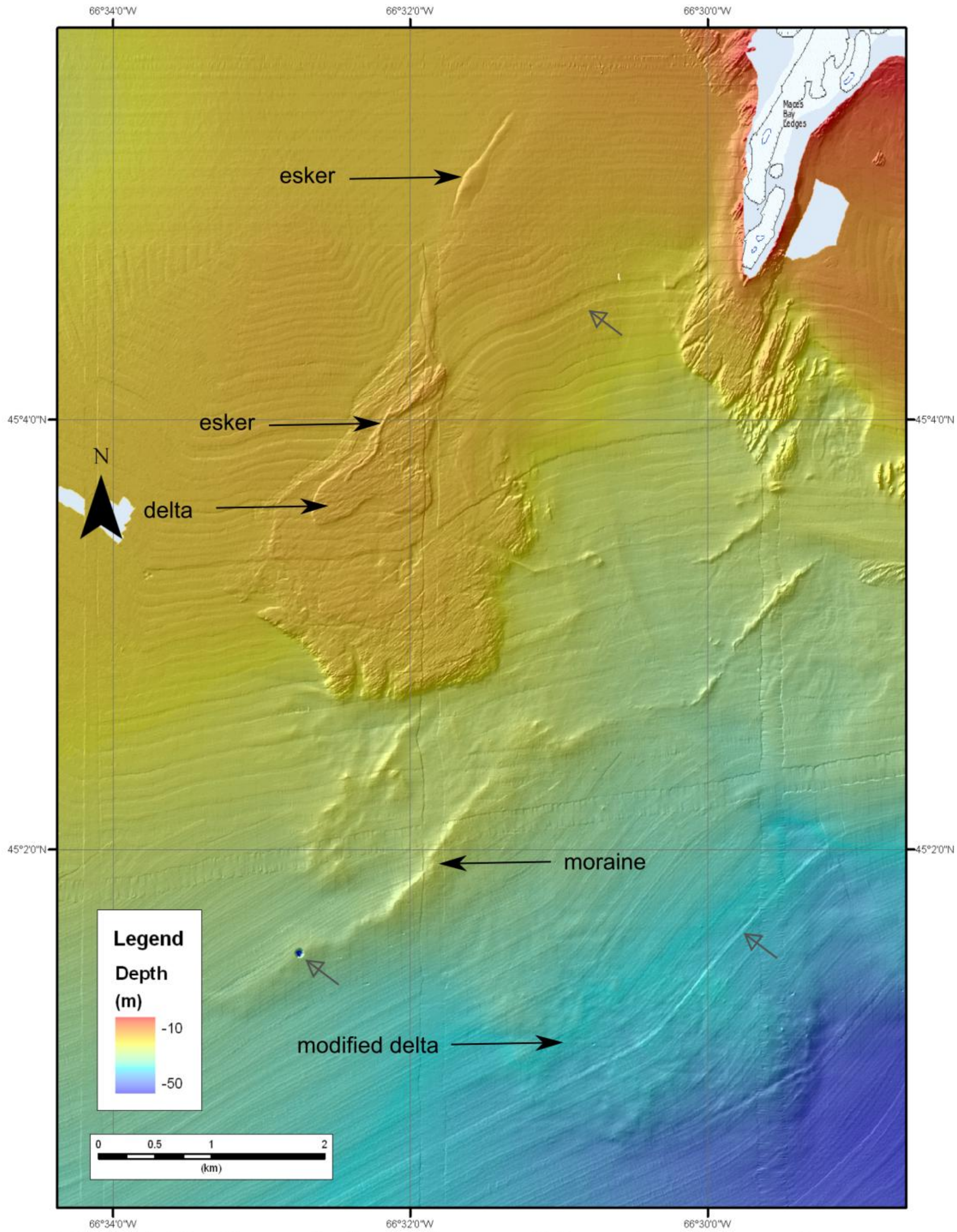

Figure 4.24: Closeup bathymetric map of Maces Bay, arrows point to an esker, deltas and a moraine. Grey arrows point to artefacts due to data collection issues which are not landform features.

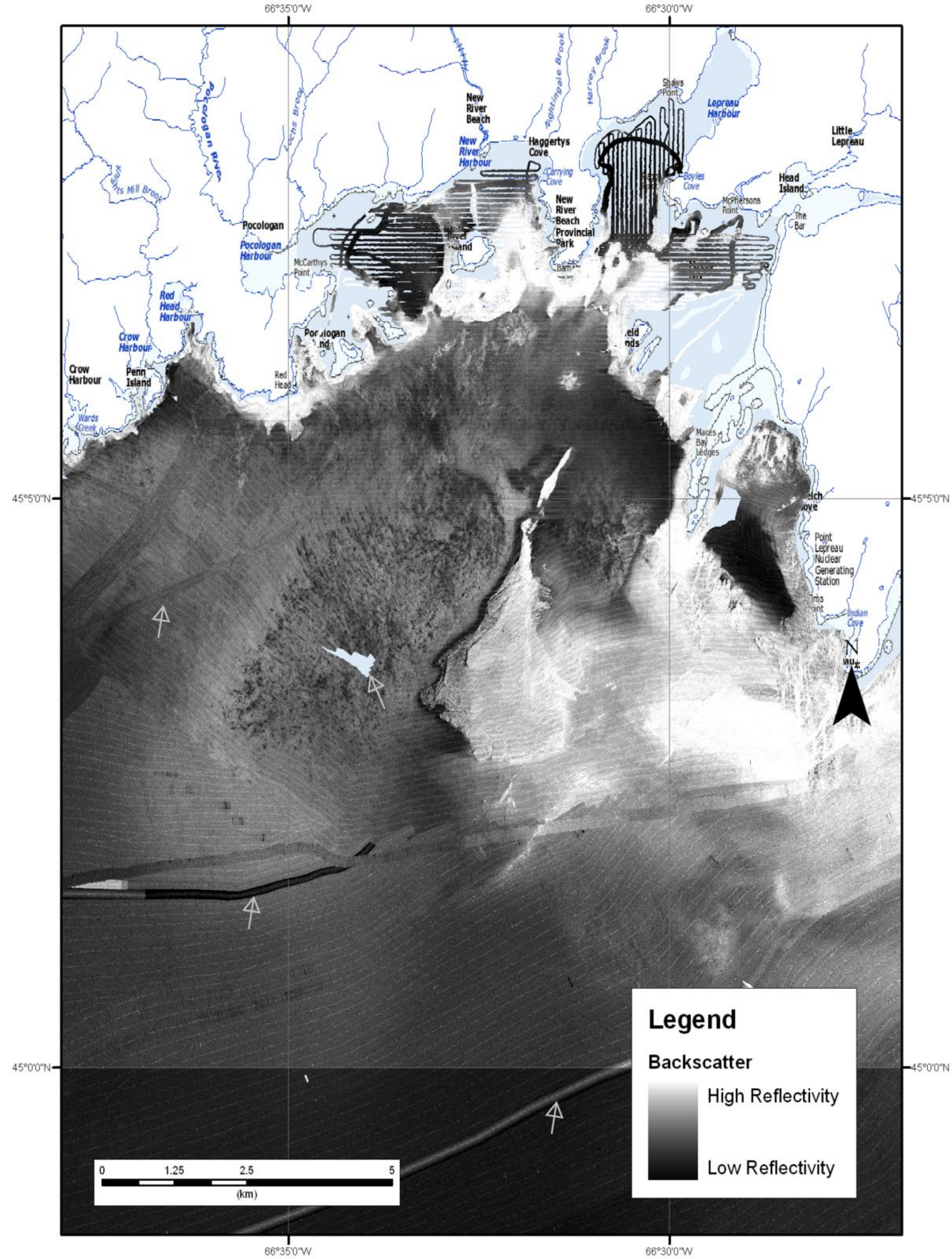

Figure 4.25: Backscatter of the Maces Bay survey area showing a strong acoustic return over the delta complex and moraine to the south, directly to the west of the complex is a mottled return suggesting a hummocky surface, and to the east of the complex is a high acoustic return, likely nearshore bedrock highs. Grey arrows point to artefacts due to data collection issues which are not landform features.

shows low reflectivity which indicates a seabed composed of softer sediments such as sands and mud.

4.2.8 Maces Bay Survey Stratigraphy

Unit 1 is the lowest unit observed, and is interpreted as bedrock (BR). It is characterized by a strong, uniformly high intensity return on a highly irregular surface and generally shows steeply dipping peaks and valleys. The bedrock unit often can be traced from near shore outcrop highs and is found in all of the lines; it often dips below maximum achieved penetration of the 3.5 kHz echosounder (Figure 4.26). In the centre of the bay it appears to surface to the seafloor (Figure 4.25).

Overlying bedrock in limited areas is unit 2, interpreted as till (T). Where observed it is generally thin with the surface exhibiting a highly irregular morphology, conformable with the underlying BR (Figure 4.27). The lower bounding surface has a more intense return; however the upper bounding surface is also intense but shows more variability in the intensity of return and with some high reflective points, likely large boulders. Till may be present in more of the area (Figure 4.26), but it is difficult to distinguish between the bedrock and till as the till is thin, conformably lies on top of the bedrock, and both have an intense return.

Unit 3 overlies BR and T; it is the most massive of all the units in Maces Bay sub-bottom lines. This unit is interpreted as glacimarine (GM) and is seen in all of the survey lines. The lower bounding surface has a more intense return than the upper bounding surface which shows more variability, from an intense return to less intense. The unit contains occasional high point reflectors interpreted as ice-rafted debris (Figure 4.28). Within 31 lines an unconformity (U) is recognized at depths between 35 m and 40 m

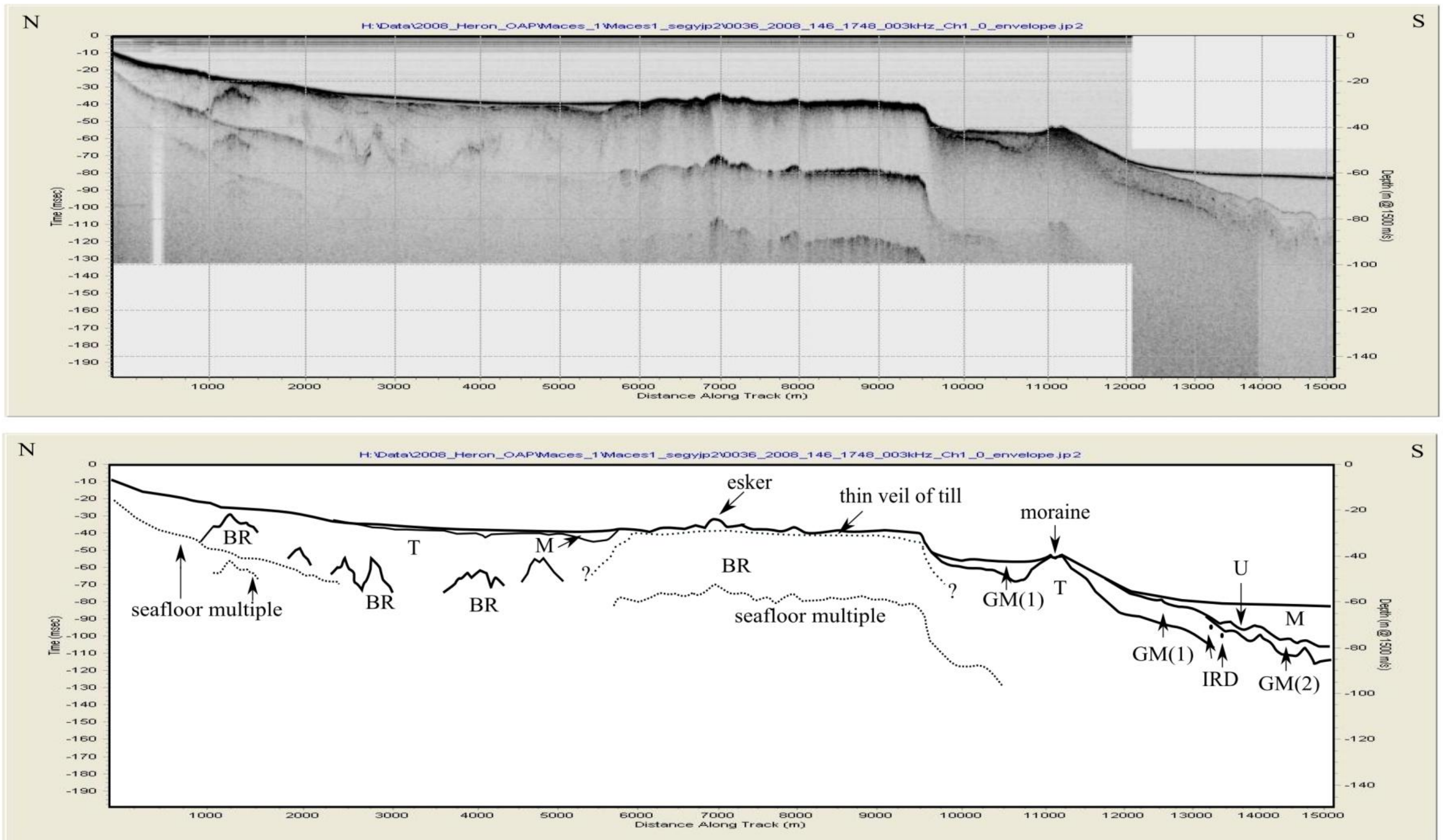

Figure 4.26: Seismic track line JD146_1748 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity, and M = Holocene mud. Note the esker ridge at 7000 m and the moraine at 11000 m.

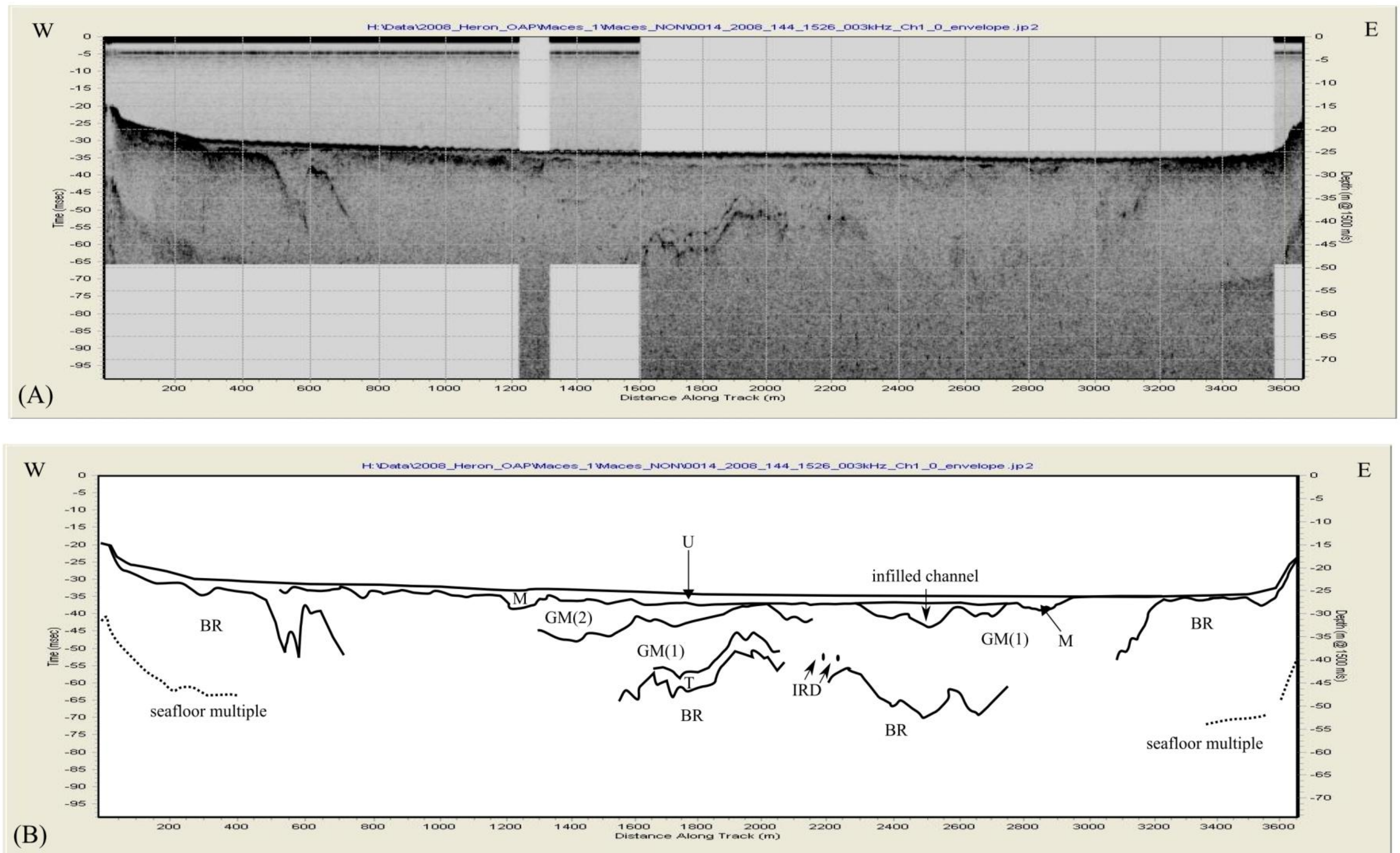

Figure 4.27: Seismic track line JD144_1526 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the infilled channel between 2400 and 2600 m.

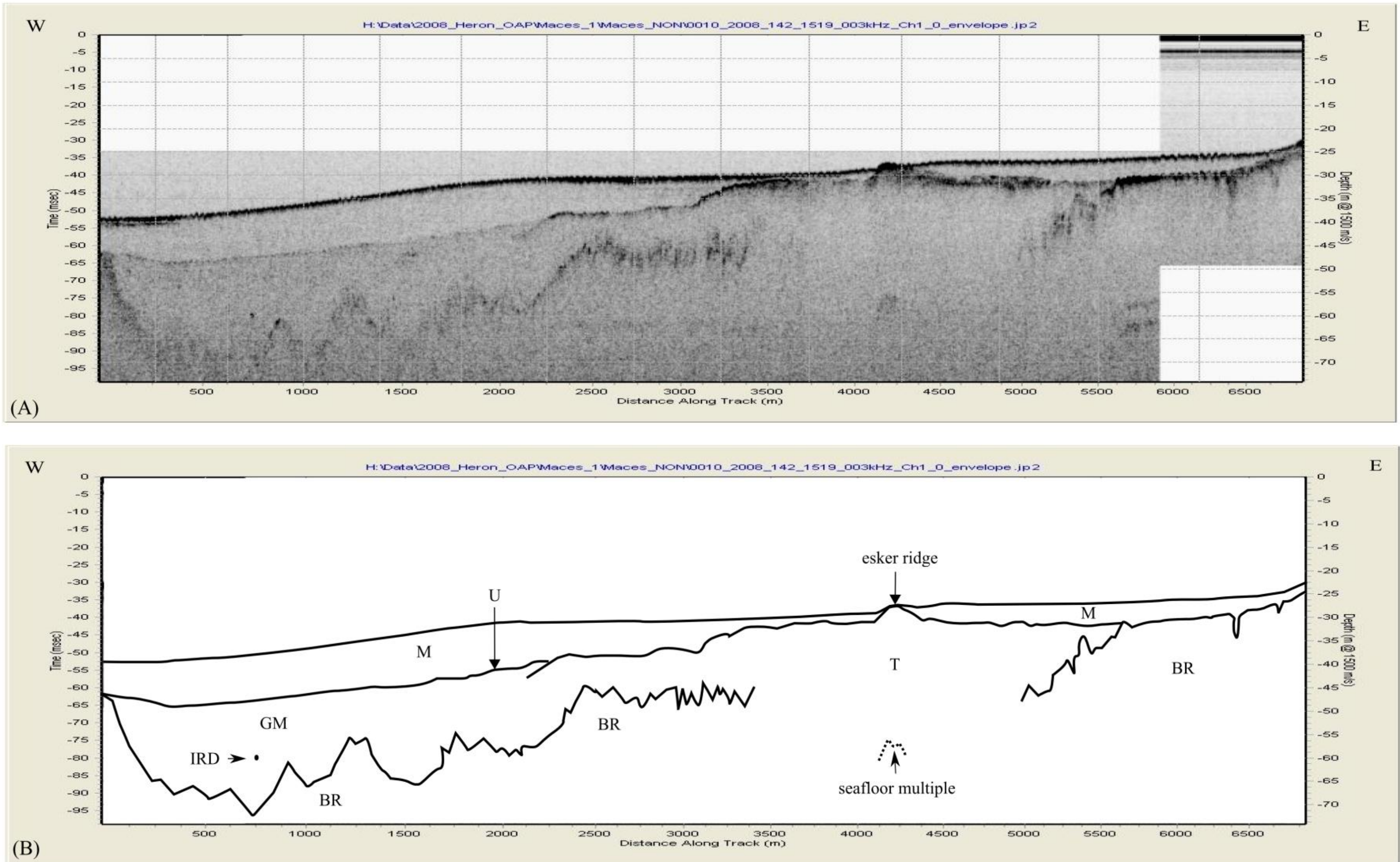

Figure 4.28: Seismic track line JD142_1519 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, T = till, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the esker ridge at 4250 m.

below the sea level that represents fluvial erosion (Figure 4.29). The total length of the fluvial system measures 2.7 km and it runs in a north south direction (Figure 4.30). Where the fluvial deposits are present the upper bounding surface is less intense than the bounding surface without the fluvial deposit. The fluvial channels change in morphology from nearshore to further offshore. The nearest to shore fluvial channels are wide and shallow, showing internal reflections representing stratification within the fluvial system (Figure 4.31). These channels vary from 200 m to a maximum of 600 m in width and from 8 m to 15 metres in depth; they can be seen for just over 850 m running north south. Figure 4.32 shows a fence diagram of 5 sequential sub-bottom lines displaying stratification.

Moving south i.e., further offshore, the fluvial system becomes double channels, representing a braided area (Figure 4.29). The braided channels vary from 400 m to a maximum of 600 m in width, 4 m in depth and are likely infilled with sand and occasional gravel. The braided channels measure just over 500 m in length, in a north south direction. Following the braided channels are distinct V-shaped channels, this shape is observed in 2 of the seismic lines (Figure 4.33). These channels show in-fill. The final channel observed is a wide and shallow channel (Figure 4.34), which can be seen for 1600 m in length. The channels furthest south are widest, starting at 150 m in width and narrowing to 600 m for the furthest to the south; the depth varies from 6 m to 1.5 m. These channels are infilled with Holocene mud, as opposed to the cut and fill deposition of the other channels. Figure 4.30 is a map of the extent of the glaciofluvial deposits mapped in Maces Bay.

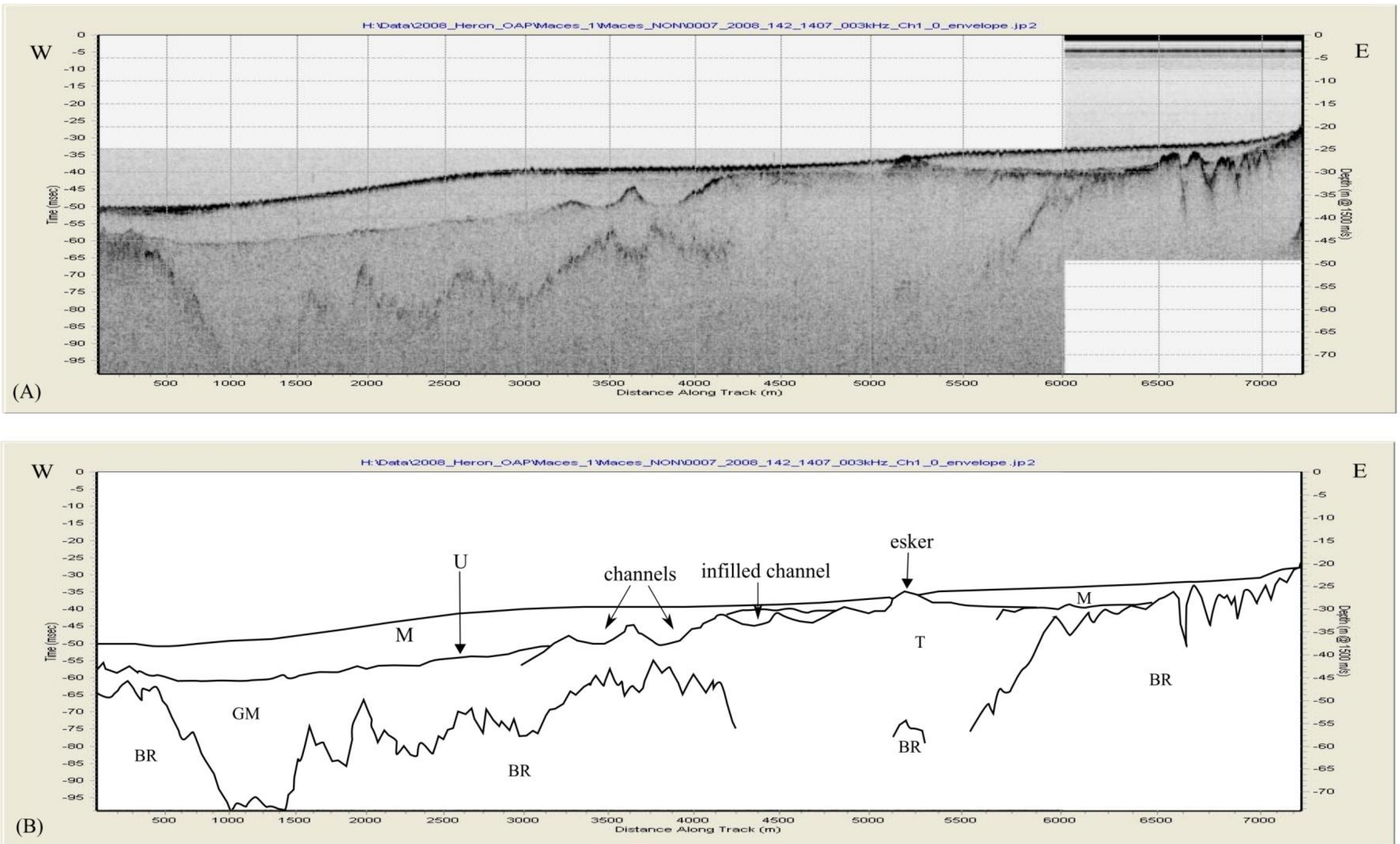

Figure 4.29: Seismic track line JD142_1407 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the two channels between 3200 and 400 m, the infilled channel between 4250 and 4900 m, and the esker rigde at 5250 m.

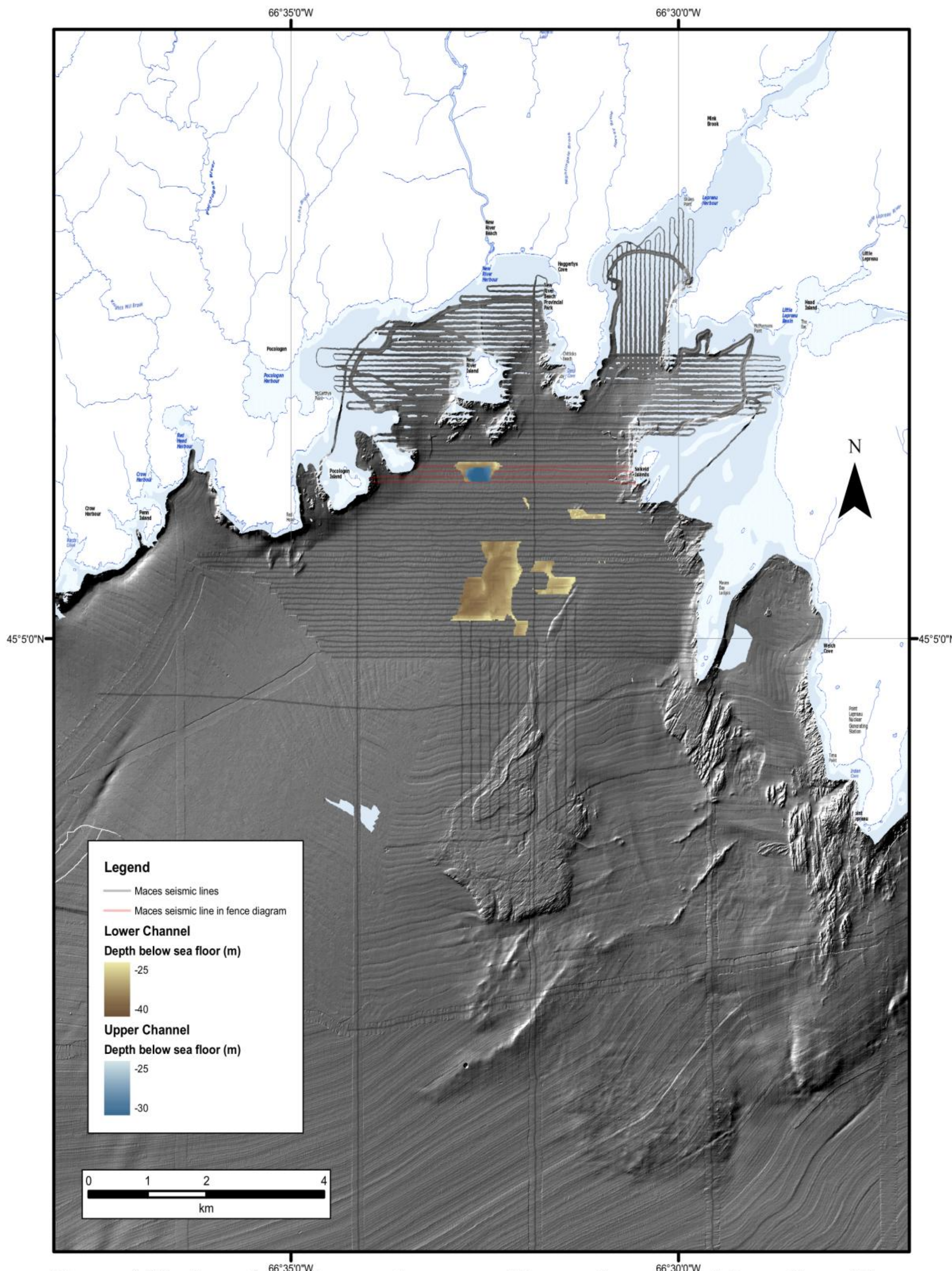

Figure 4.30: Location of mapped upper and lower channels in Maces Bay. The red seismic lines refer to fence diagram Figure 4.32.

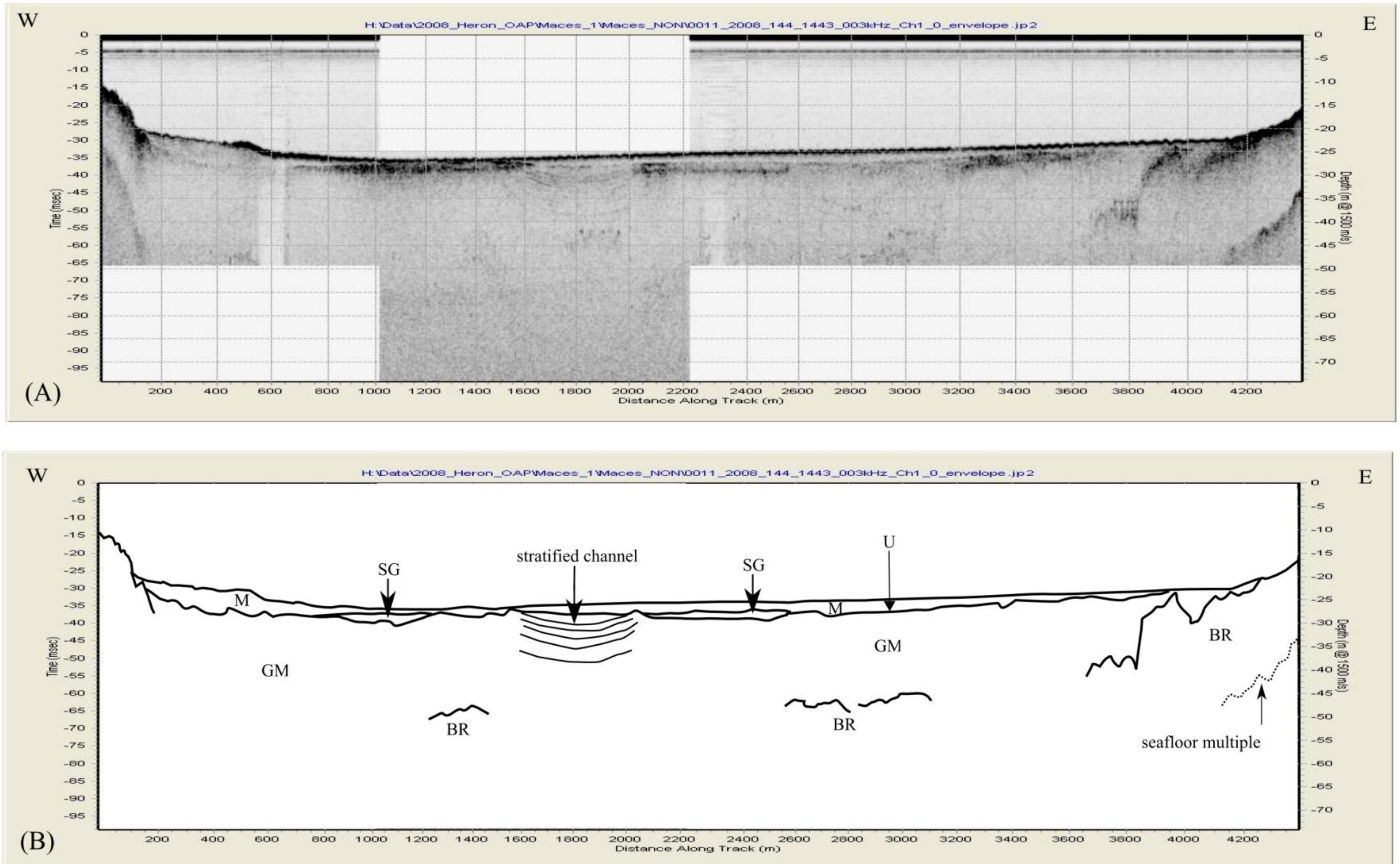

Figure 4.31: Seismic track line JD144_1443 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, SG = sand and gravel U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the 200 m wide stratified channel between 1600 m and 2000 m.

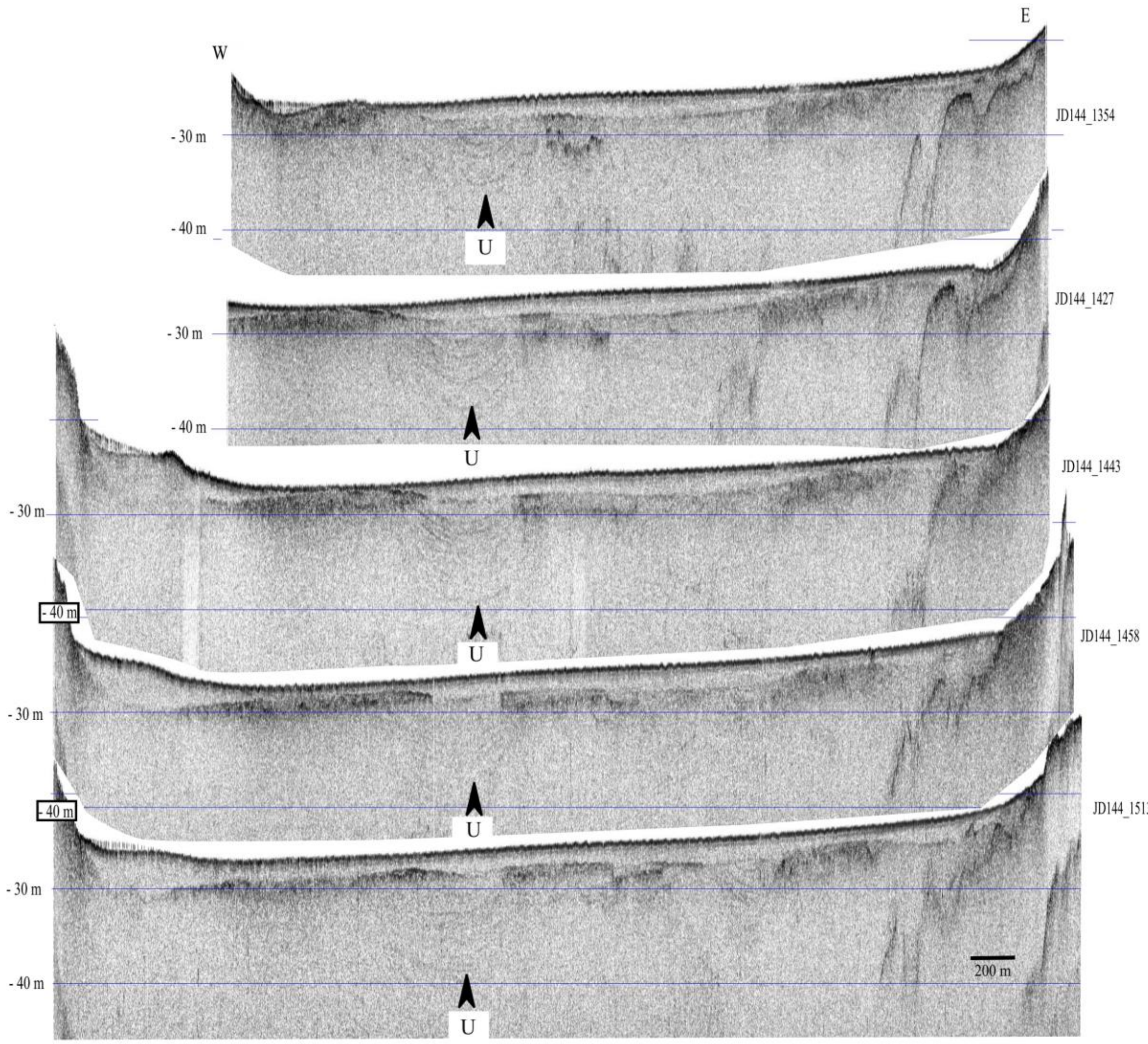

Figure 4.32: Seismic profiles presented as a fence diagram showing channel, lines go from nearshore to offshore. Arrows point at U = Pleistocene/Holocene unconformity.

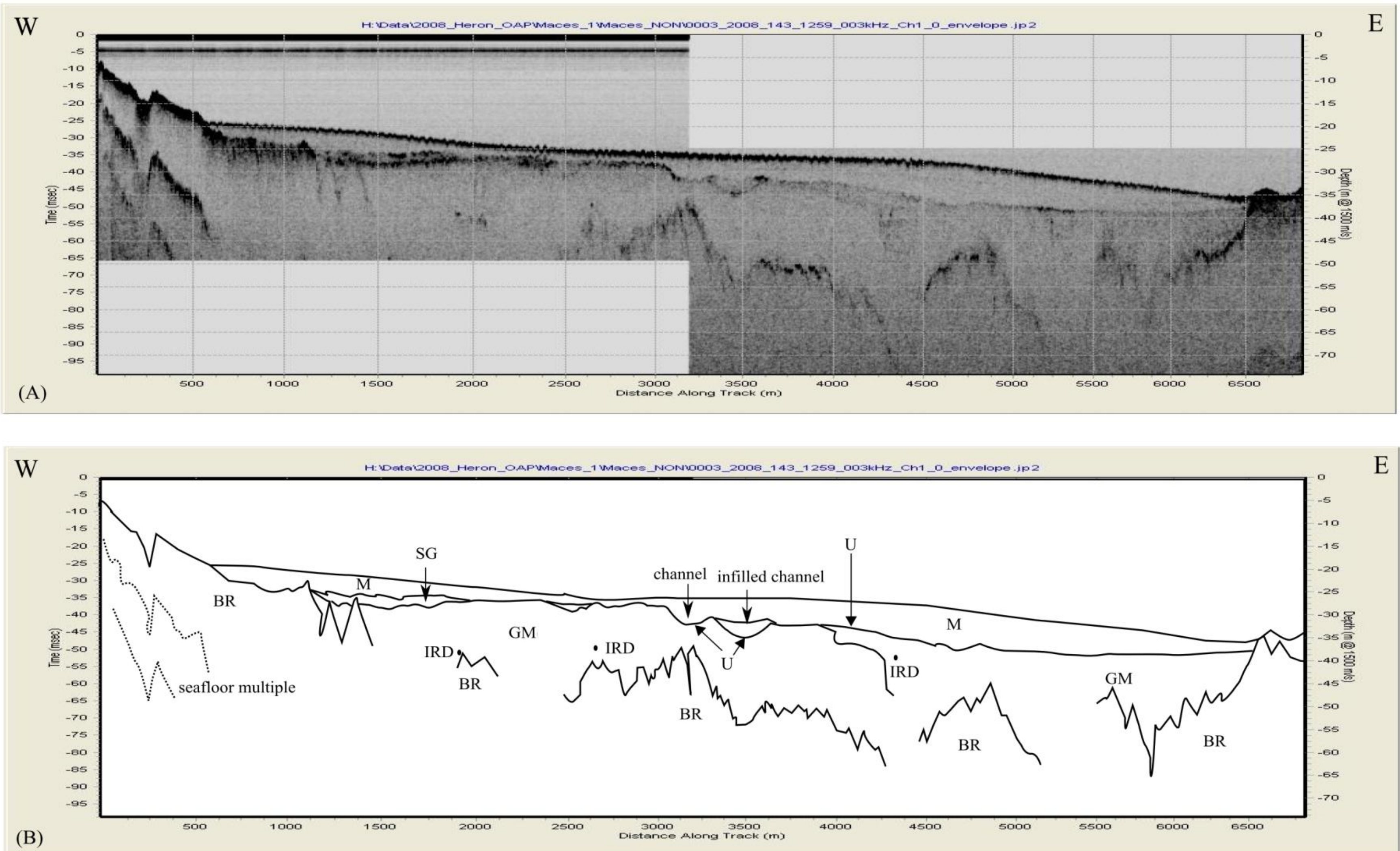

Figure 4.33: Seismic track line JD143_1259 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the two different channels between 3000 and 3500 m.

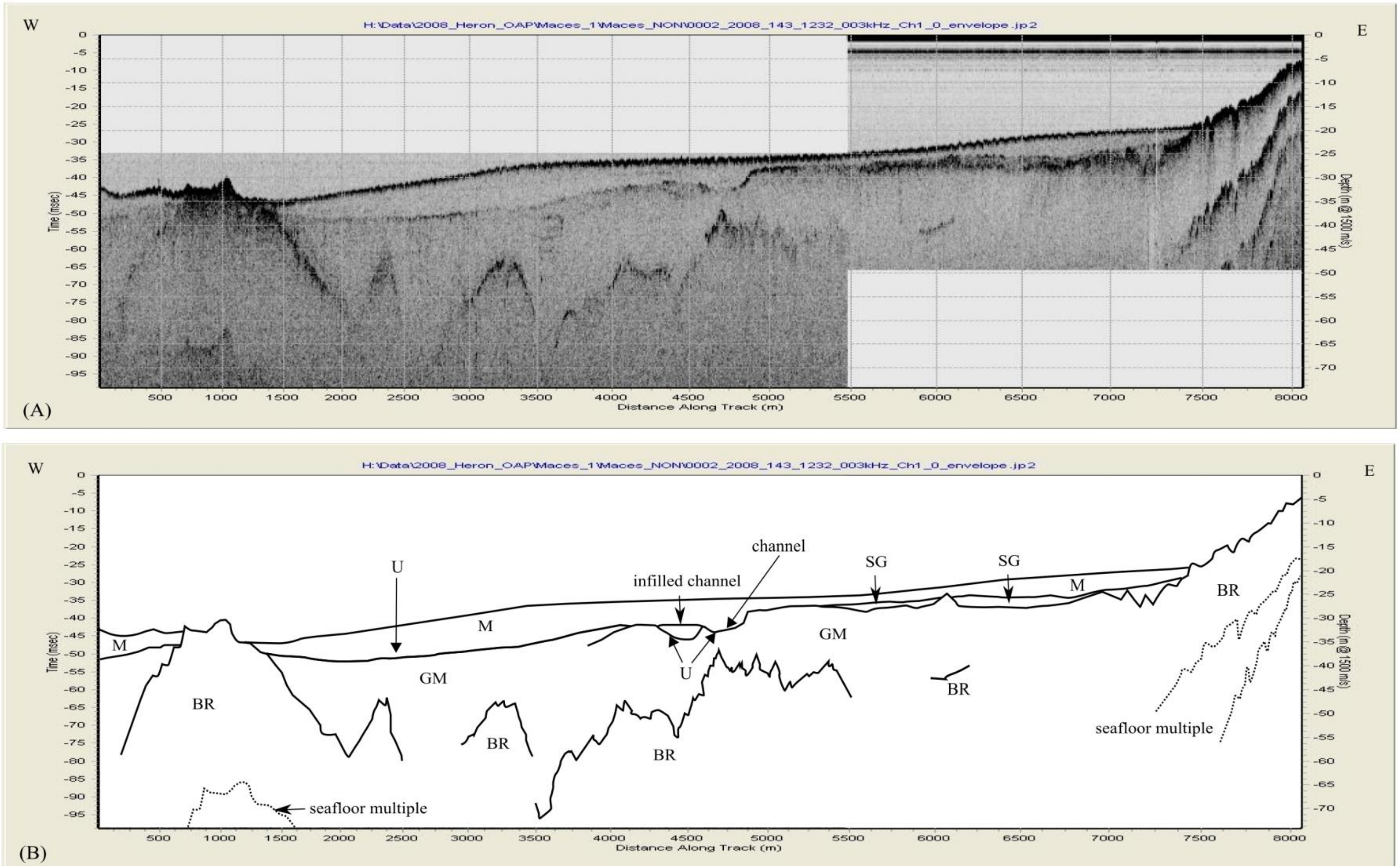

Figure 4.34: Seismic track line JD143_1232 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, SG = sand and gravel, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the two channels located between 4250 and 5000 m.

Natural (NG) was detected in three lines (Figure 4.35) located in the outer western flanks of the bay.

Unit 4 has a strong surface return and is found as lenses overlying GM in some areas (Figure 4.31). This unit is interpreted as sand and gravel (SG).

Unit 5 is an acoustically transparent, flat-lying unit that is always present as the upper most seismic unit and is found in all lines. It increases in thickness further offshore (Figure 4.36*)*. This unit is interpreted to represent a postglacial Holocene mud, and given the designation M. This unit is interpreted as and equivalent to King and Fader (1986) LaHave Clay.

4.2.9 Chance Survey Area

The Chance survey area (Figure 4.37) is located from Split Rock near the entrance of Musquash Harbour to Point Lepreau. The shoreline of the survey area is indented with several inlets, including Musquash Harbour, Chance Harbour, Dipper Harbour and many smaller coves (Figure 4.37). The largest fresh water input is from the Musquash River (Figure 1.6).

The onshore geology of this area consists of a mix of Precambrian or Lower Paleozoic and Triassic bedrock (Figure 4.2). The Precambrian or Lower Paleozoic group consists of granite, granodiorite, quartz diorite, gabbro, volcanic, limestone, quartzite, argillite, dolomite, conglomerate and gneissic rocks. Local exposures of Triassic red to grey conglomerate, sandstone, and shale outcrop at Point Lepreau (NBDNR 2006).

The surficial geology of the area is dominated by weathered, glacially scoured and polished bedrock that has area blanketed by mainly stony till (Rampton 1984).

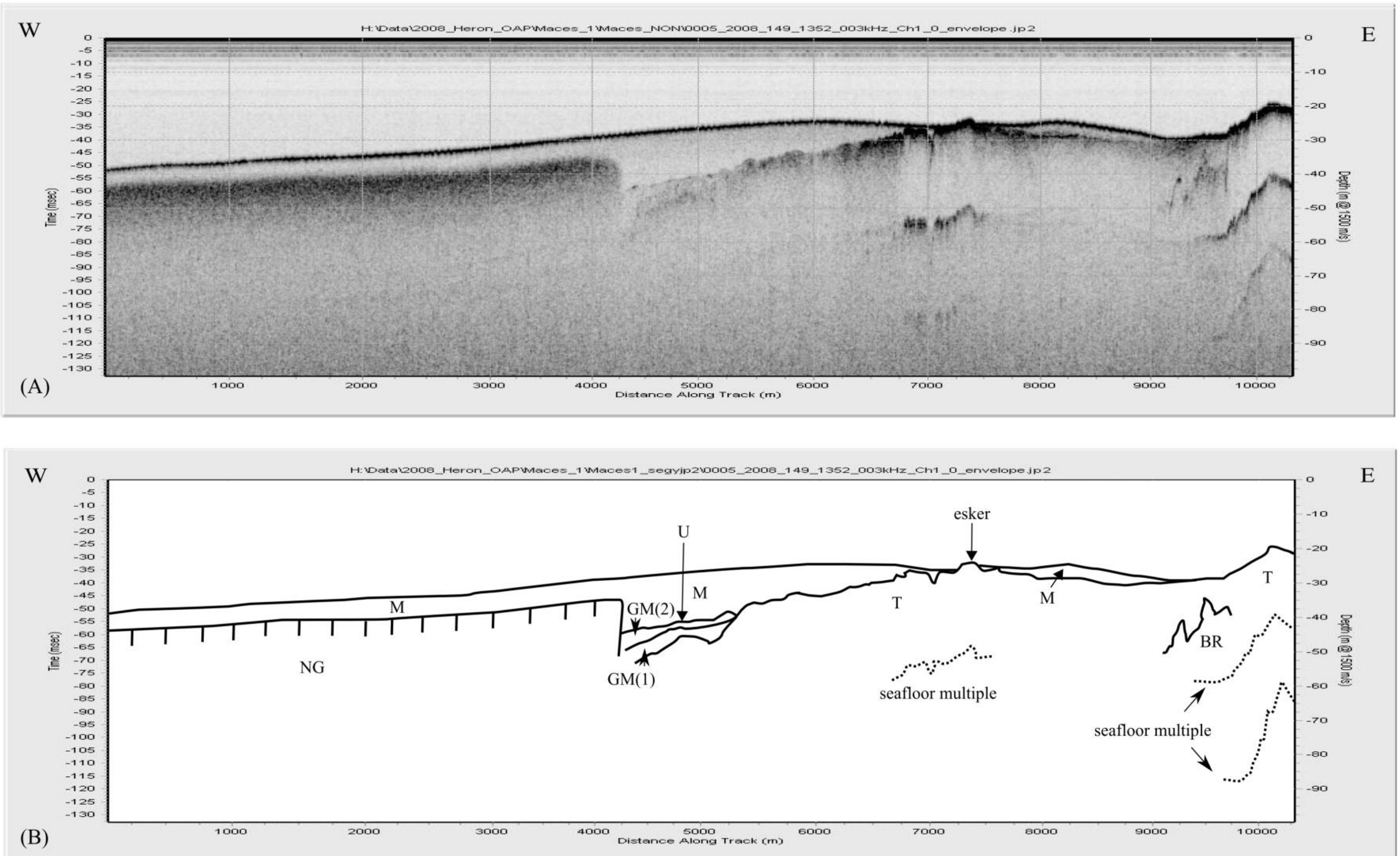

Figure 4.35: Seismic track line JD149_1352 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the esker ridge at 7500 m, and to the east a reflector interpreted as bedrock trending downwards.

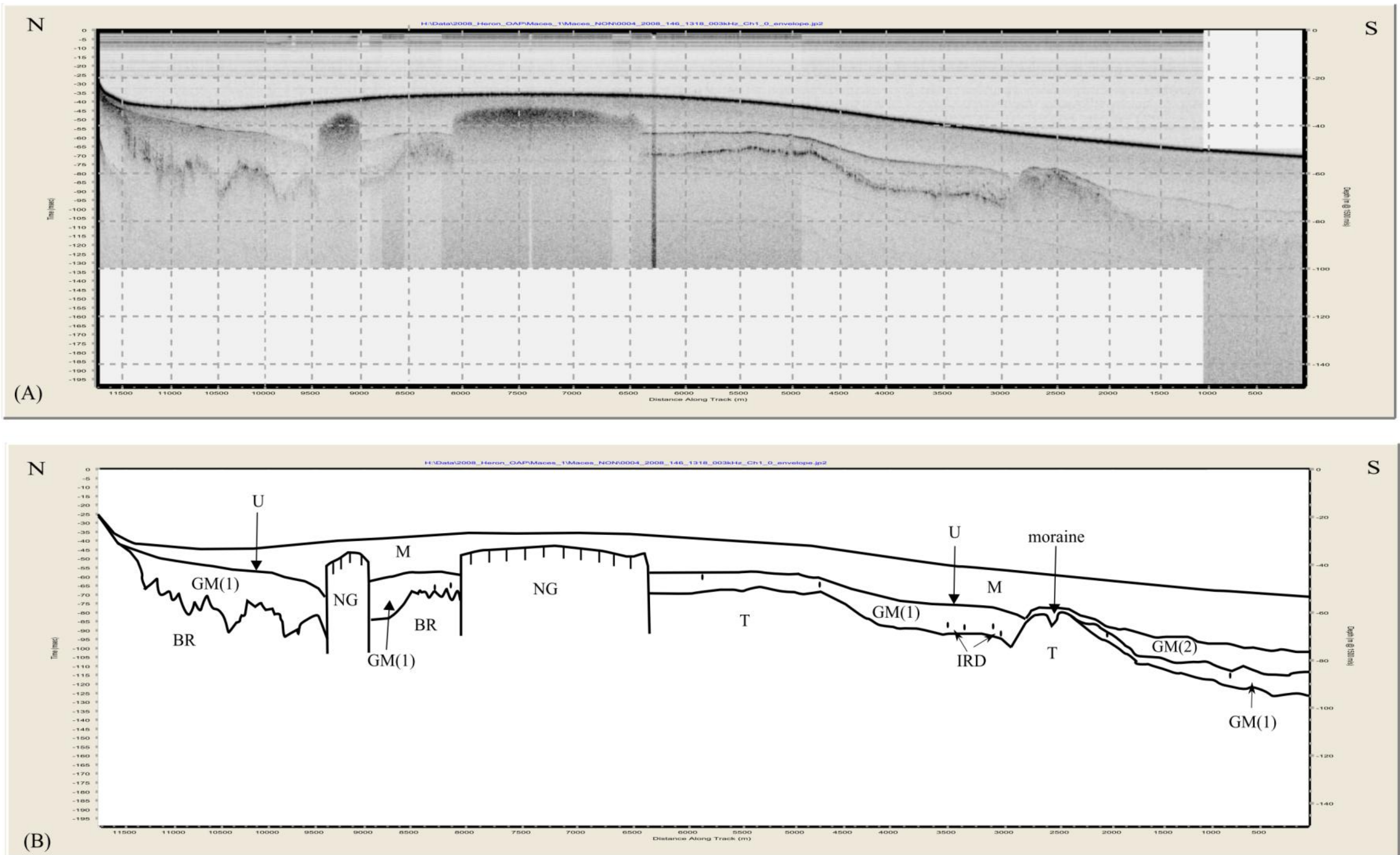

Figure 4.36: Seismic track line JD146_1318 for the Maces survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, IRD = ice rafted debris, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note moraine at 3500 m.

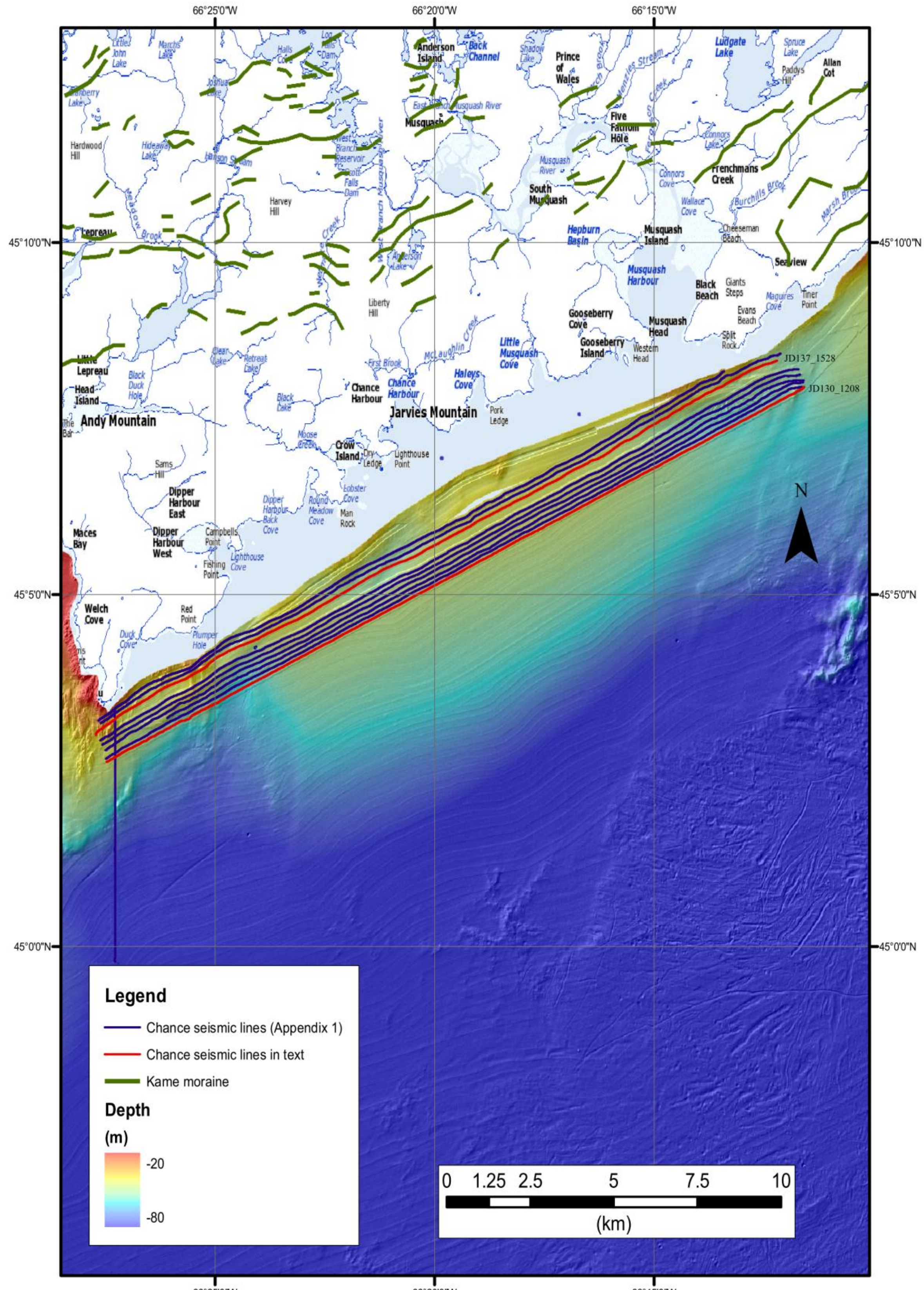

Figure 4.37: Bathymetric map showing Chance survey area, seismic lines, and onshore glaciofluvial deposits.

The onshore surficial geology has a series of kame moraines (Figure 1.1), as discussed in chapter 1.4.

The survey offshore of the Chance study area covers an area of 31 km^2, including 10 survey lines (Appendix 1) totaling 208 km of sub-bottom (Figure 4.37). Nine of the survey lines run parallel to the coastline in a northeast to southwest direction, and the north-south lines intersects these at the western end off Point Lepreau. The data were collected in the 2008 by the CSL Heron and includes 15m resolution bathymetry and backscatter data.

The bathymetry of the Chance survey (Figure 4.38) shows an increase in depth from nearshore to the offshore, with a bedrock high at the western end off Point Lepreau. To the south of the survey area in 100 m depths of water, iceberg scours can be recognized.

The backscatter of the Chance area shows regions of both high and low backscatter strength (Figure 4.39). Offshore data from Pt. Lepreau and Split Rock demonstrate a high backscatter return, indicating a seabed composed of either coarse sand, gravel, or rock outcrops. For the rest of the survey area, backscatter is low reflectivity, indicating a seabed composed of softer sediments such as sands and muds.

4.2.10 Chance Survey Stratigraphy

Unit 1, the lowest unit observed, is characterized by a strong, high intensity return on an irregular surface, and generally displays a peak and valley morphology in the the survey lines (Figure 4.40). Unit 1 is interpreted as bedrock (BR). At the western end of the survey lines, near Point Lepreau, bedrock outcrops at the sea floor and dips steeply eastward, where it eventually disappears below the maximum achieved penetration of the

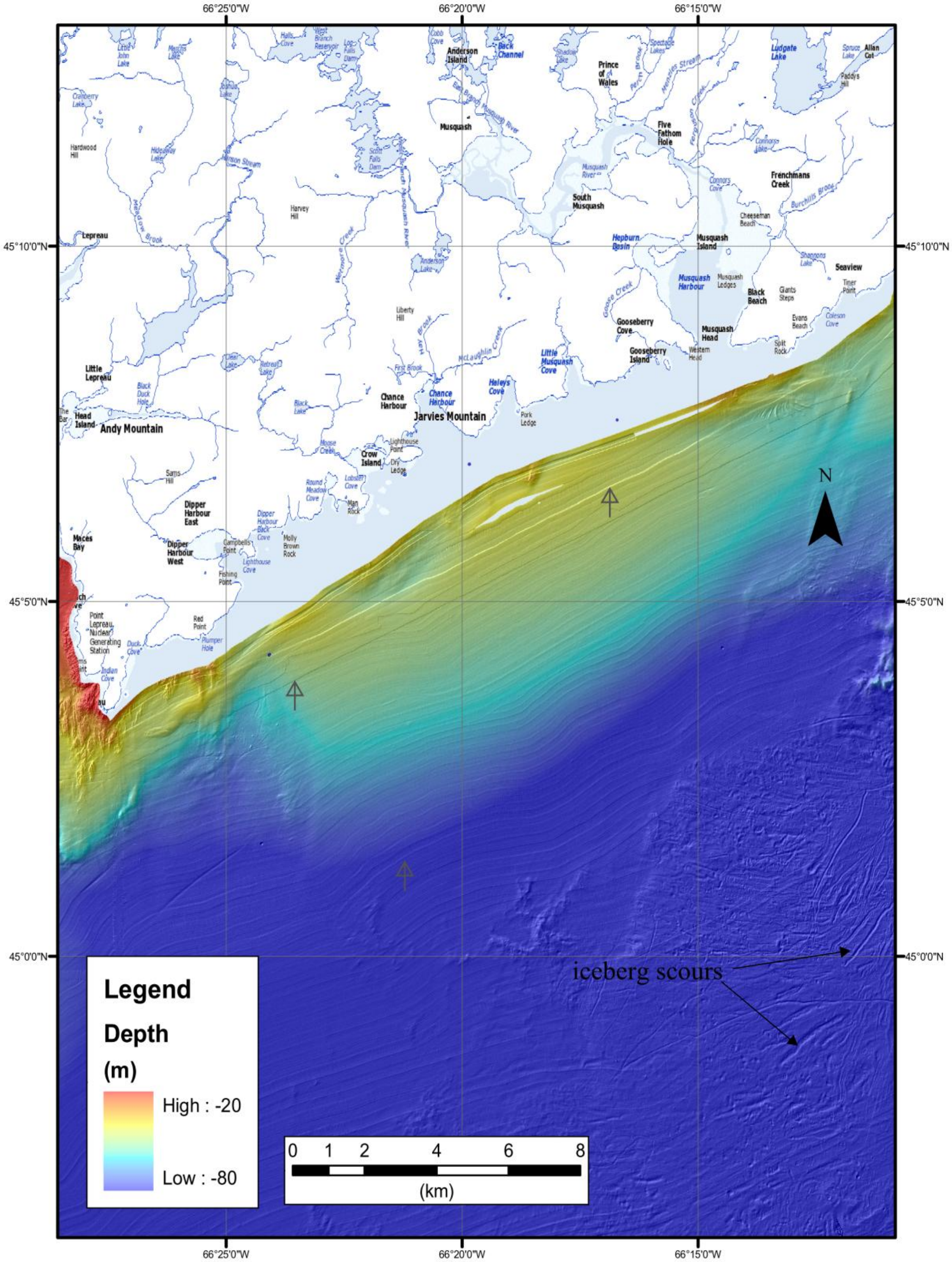

Figure 4.38: Chance survey area bathymetry. Note the iceberg scours. Grey arrows point to a parallel line to shore, this and other parallel lines are artefacts due to data collection and are not landform features.

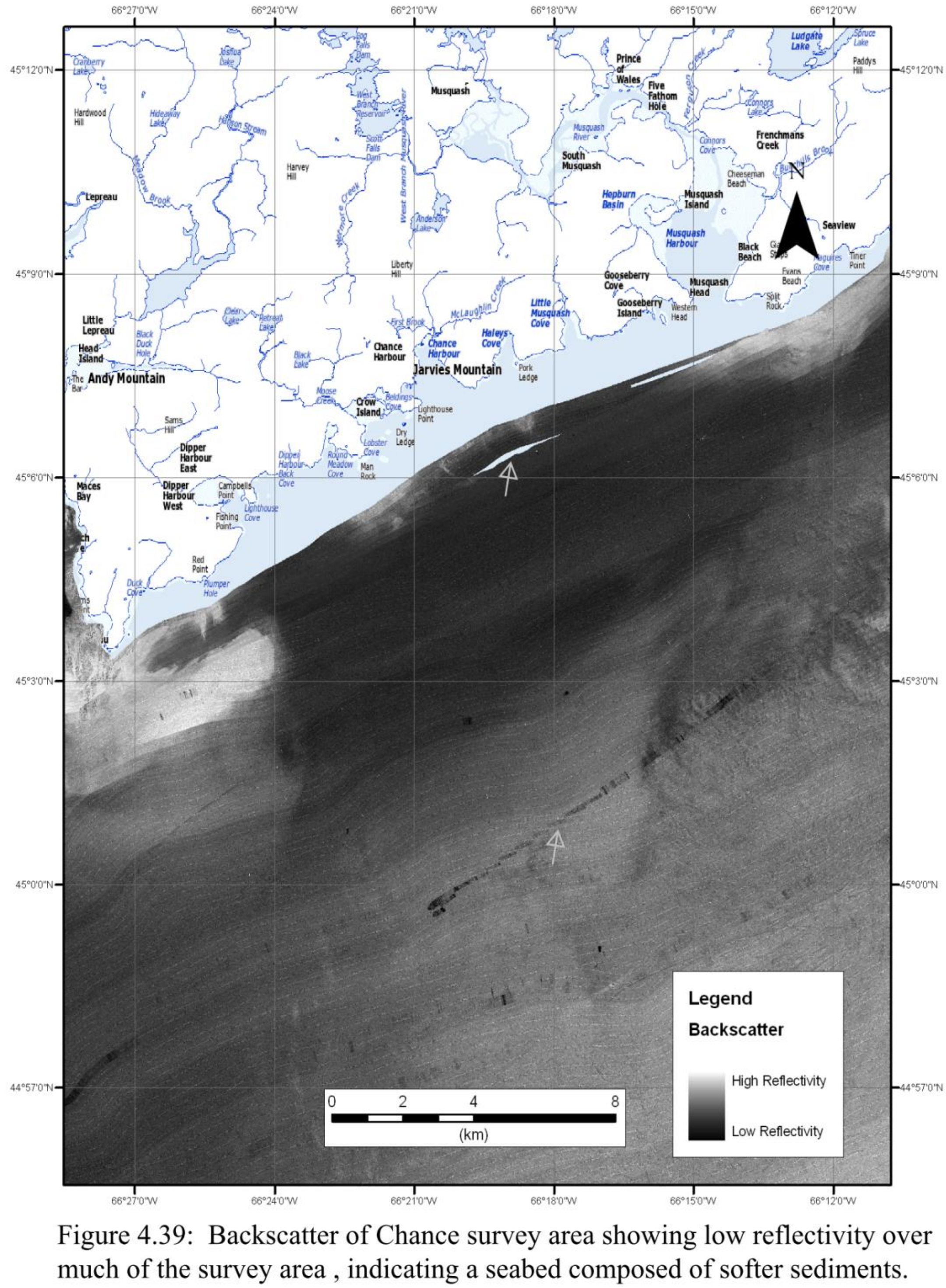

Figure 4.39: Backscatter of Chance survey area showing low reflectivity over much of the survey area , indicating a seabed composed of softer sediments. High reflectivity is observed at the western end off shore Point Lepreau likely nearshore bedrock. Grey arrows point to artefacts due to data collection issues and which not landform features.

3.5 kHz echosounder. Bedrock reflection reappears in the centre and eastern end of the survey at the mouth of Musquash Harbour. The BR unit is found in all the lines and is stratigraphically the lowest unit present.

Unit 2 overlies BR and is characterized by a moderate intensity return with stratification, displaying peak and valley morphology that is draped over the underlying bedrock (Figure 4.40). This unit is interpreted as glacimarine and is seen in all the survey lines (GM-1). In some of the lines, internal reflectors are observed (Figure 4.41); where seen they are conformable with both the upper and lower bounding surfaces. The lower bounding surface has a more intense return than the upper bounding surface. Parts of this unit dip below the depth of the survey. Within this unit are found occasional ice-rafted debris (IRD).

Unit 3 overlies GM-1 and is characterized by a moderate intensity return, displays a ponded morphology, and is transparent with no visible internal reflectors (Figures 4.40 and 4.41). This unit is interpreted as glacimarine and is seen in all the survey lines (GM-2). The upper bounding surface has a more intense return then the lower bounding surface and in some areas it is difficult to resolve the difference between the two units. The upper bounding surface is marked by an erosional surface and is found at lows of -55 m and highs of -50 m.

Unit 4, the uppermost unit, is characterized as an acoustically transparent and ponded unit. This unit is interpreted to represent Holocene postglacial deposition following glaciation, Holocene mud (M), it is equivalent to the LaHave Clay of King and Fader (1986). Parts of the unit dip below the depth of the survey window, as do the lower units. As this unit goes from nearshore to offshore it begins as two large lobes

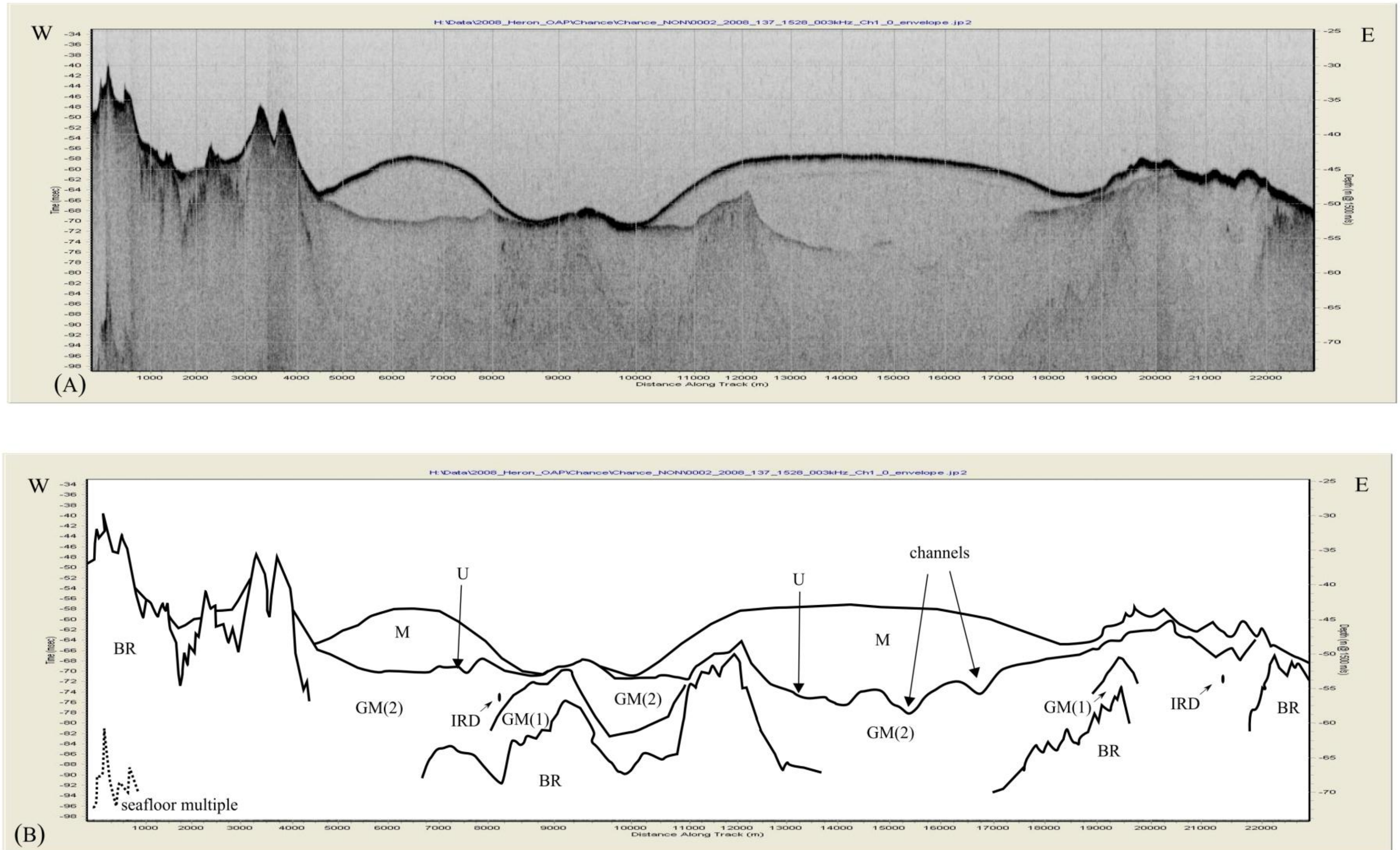

Figure 4.40: Seismic track line JD137_1528 for the Chance survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the two large lobes of Holocene sediment representing Holocene fans and channels infilled with Holocene mud.

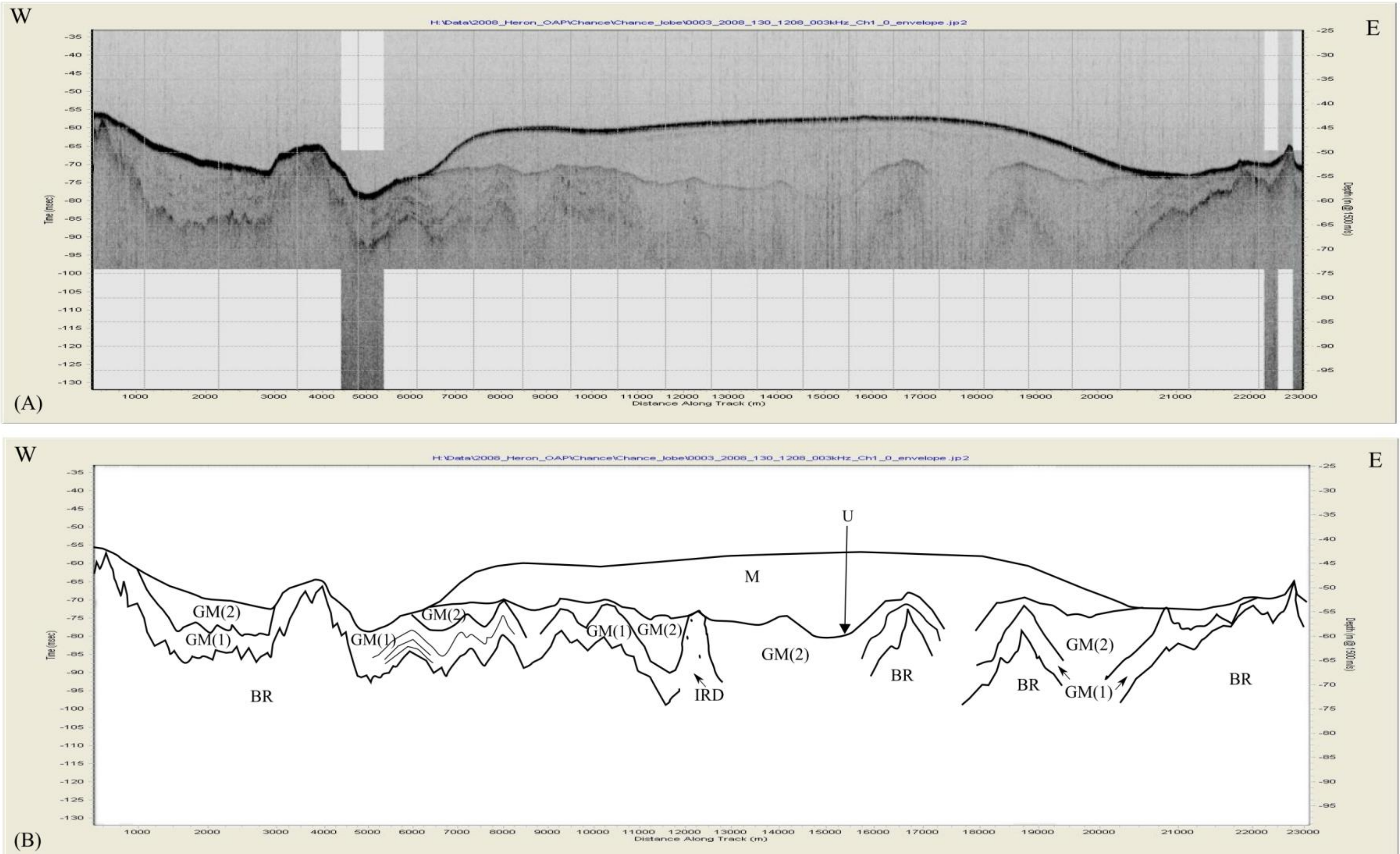

Figure 4.41: Seismic track line JD130_1208 for the Chance survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the single large lobe of Holocene sediment representing Holocene fans.

(Figure 4.40) which coalesce into one lobe in the sub-bottom lines further offshore (Figure 4.41).

4.2.11 Blacks Survey Area

The Blacks survey area is located between the Saint John Harbour and Musquash Harbours (Figure 1.2). The shoreline is bisected by Lorneville Harbour. There are two small island located in the study area, Partridge and Manawagonish Islands. Partridge Island is connected to the land by a manmade breakwater named Negro Point Breakwater. The river input is from the Saint John River, discussed previously in Chapter 1.3, Physiography and Bedrock geology.

The onshore bedrock geology consists of Cambrian-Ordovician conglomerates dominated by sandstones, siltstones and shales (Figure 4.2; Tanoli and Pickerill 1988). The surficial terrestrial geology of the area is dominated by bedrock outcrops, with blankets of morainal materials, mostly till, varying in thickness, and large ridges of ice-contact stratified drift (Seaman et al. 1993). Kame moraines oriented in a southeast and northwest direction are located between Saint John and Musquash Harbour. The Sheldon Point moraine (Figures 1.1 and 4.42) onshore is described as a glacimarine end moraine (Nicks, 1988) and discussed in more detail in Chapter 1.4.

The offshore study area covers an area of 15 km^2 (Figure 4.42). The closest seismic line is less than 400 metres from shore and the furthest line is 1.5 km offshore. The data are made up of 12 survey lines parallel to the shoreline oriented northeast to southwest, totaling 155 km (Appendix 1). The data were collected in 2008 by the CSL Heron; the data set includes 15m resolution bathymetry, and backscatter data.

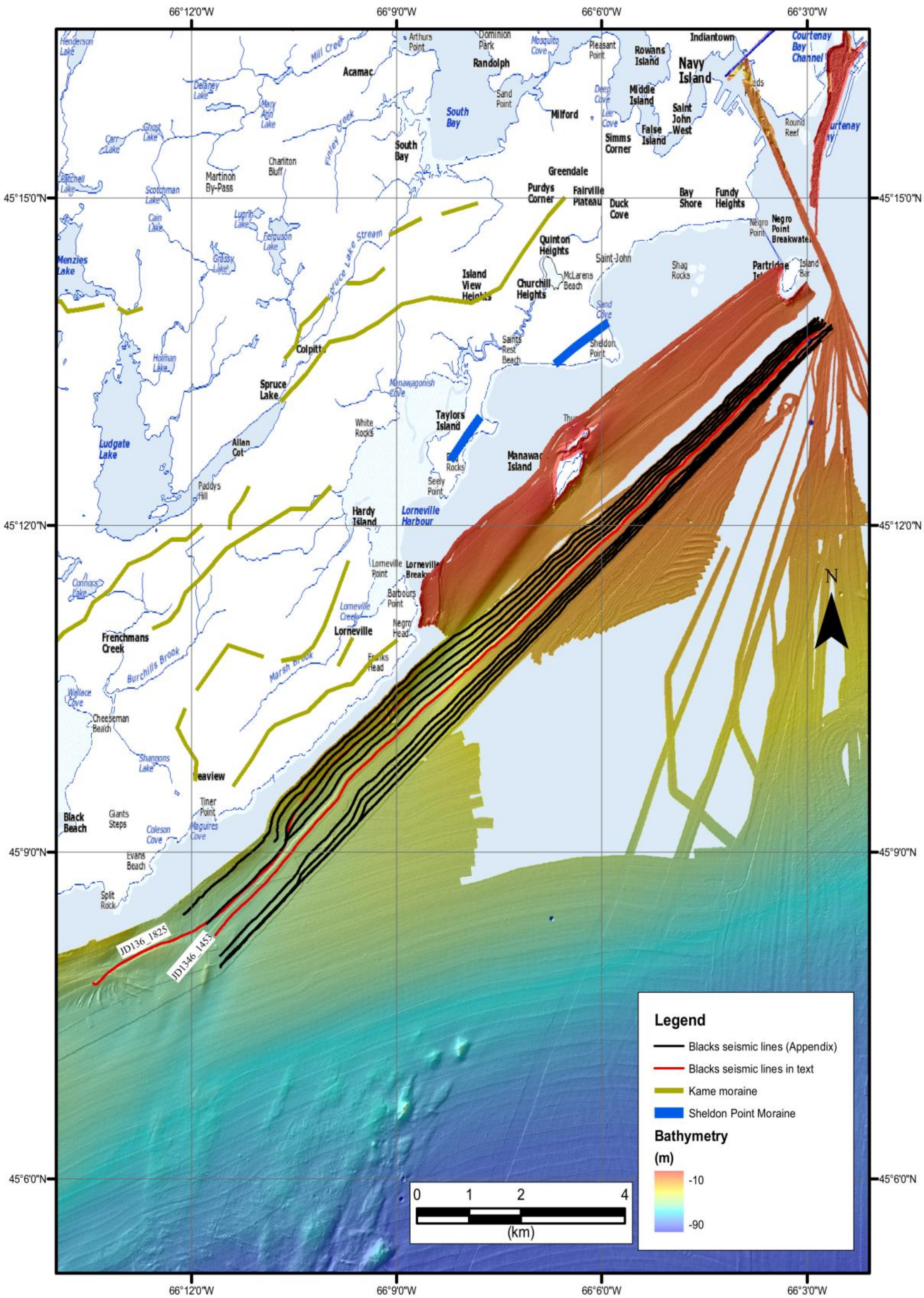

Figure 4.42: Bathymetric map showing Blacks survey area, seismic lines, and onshore glaciofluvial deposits.

The bathymetry (Figure 4.43) of the Blacks area gradually increases with increasing distance offshore from -15m depth, near the Saint John Harbour to a maximum depth of -55 m. The bathymetry is quite uniform, although one distinguishing landform is found offshore of Split Rock. This landform (Figure 4.44) has nine approximately 500 m long, tendril-like features and several smaller tendril-like features extending outward. To the east of this landform is a drumlin trending northeast-southwest, with dimensions of 442 m in length and 90 m in width (Figure 4.44). Over the sub-bottom area the backscatter (Figure 4.45) gradually grades from low nearest to Saint John harbour, to medium and high off the shore of Split Rock near Musquash Harbour. This backscatter indicates a soft muddy basement within the Saint John harbour to a seabed of coarse sand, gravel, or rock outcrops along the near shore between Lorneville and Musquash Harbour. The backscatter of the tendril like features off Split Rock shows a medium acoustic return, likely indicating sand.

4.2.12 Blacks Survey Stratigraphy

Unit 1 is the lowest unit observed in all the lines (Figure 4.46), and is interpreted as bedrock (BR). It is characterized by a strong, high intensity return on a highly irregular surface and generally displays a peak and valley morphology in the survey line.

Unit 2 overlies BR. This unit is interpreted as glacimarine (GM). It was found in all of the sub-bottom and varies in thickness from a thin mantle covering over the BR to a massive unit (Figures. 4.46 and 4.47). Thicker units cannot be followed to the bedrock as the Knudsen instrumentation settings for depth were not set large enough at the time of the survey. Within this unit are found occasional ice-rafted debris (IRD). There are stacked sets of high-intensity reflections beside the natural gas unit, only found in this

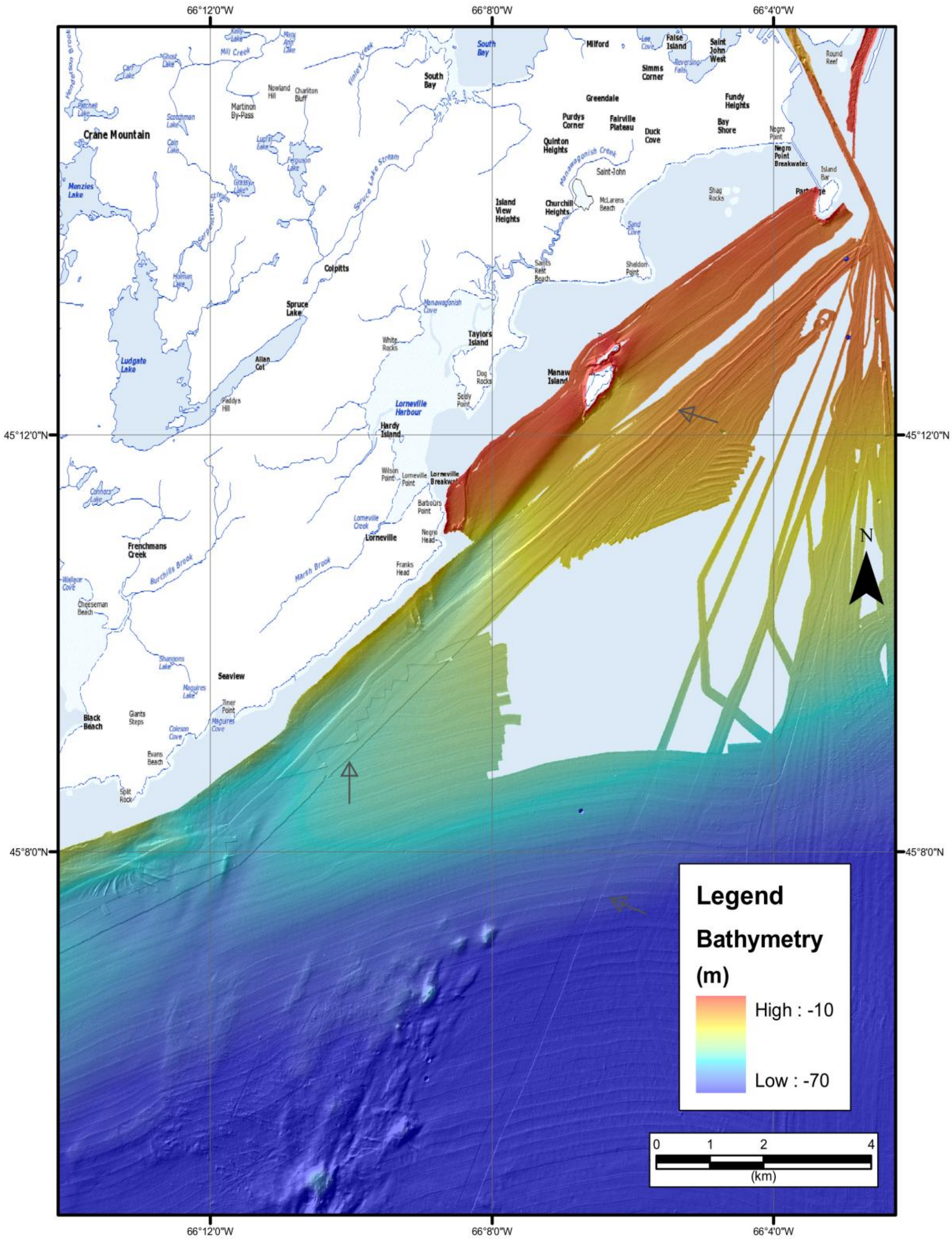

Figure 4.43: Blacks survey area bathymetry. Grey arrow points to a line parallel to shore, this and other parallel lines are artefacts due to data collection and are not landform features.

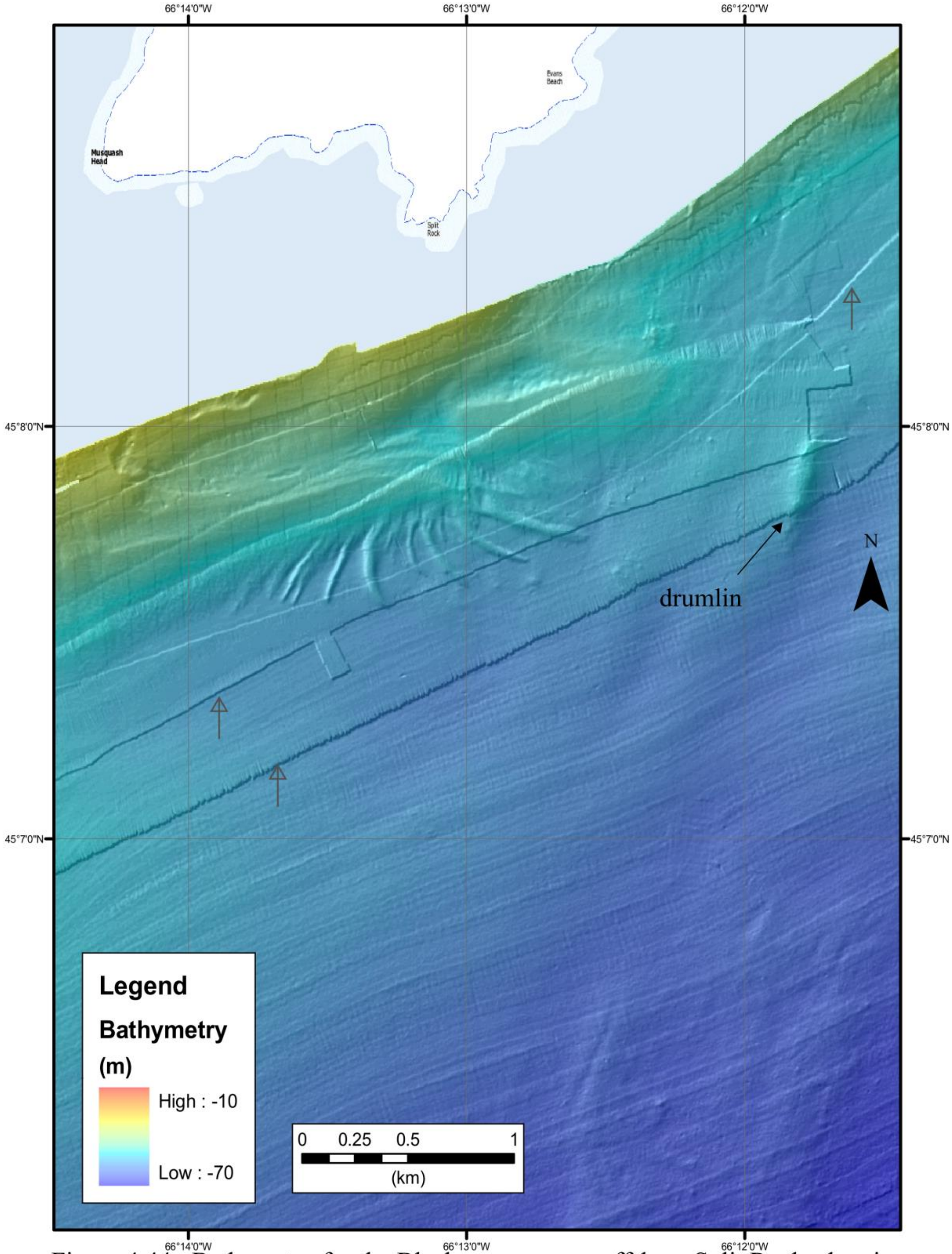

Figure 4.44: Bathymetry for the Blacks survey area offshore Split Rock, showing a drumlin. Grey arrow points to a line parallel to shore, this and other parallel lines are artefacts due to data collection and are not landform features.

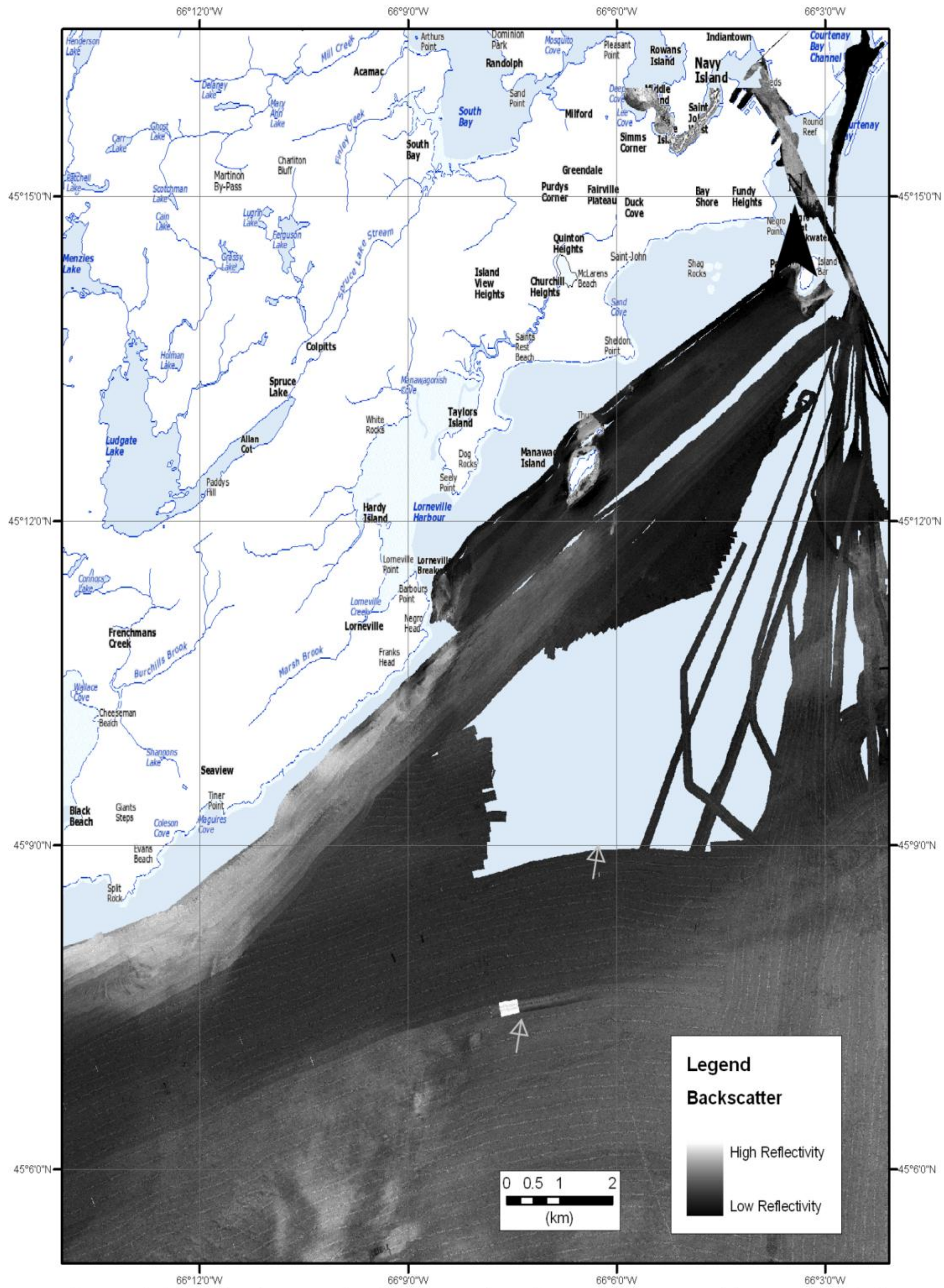

Figure 4.45: Blacks area backscatter showing moderate and low reflectivity over much of the survey area, indicating a seabed composed of softer and coarse grained sediments. Grey arrows point to artefacts due to data collection issues which are not landform features.

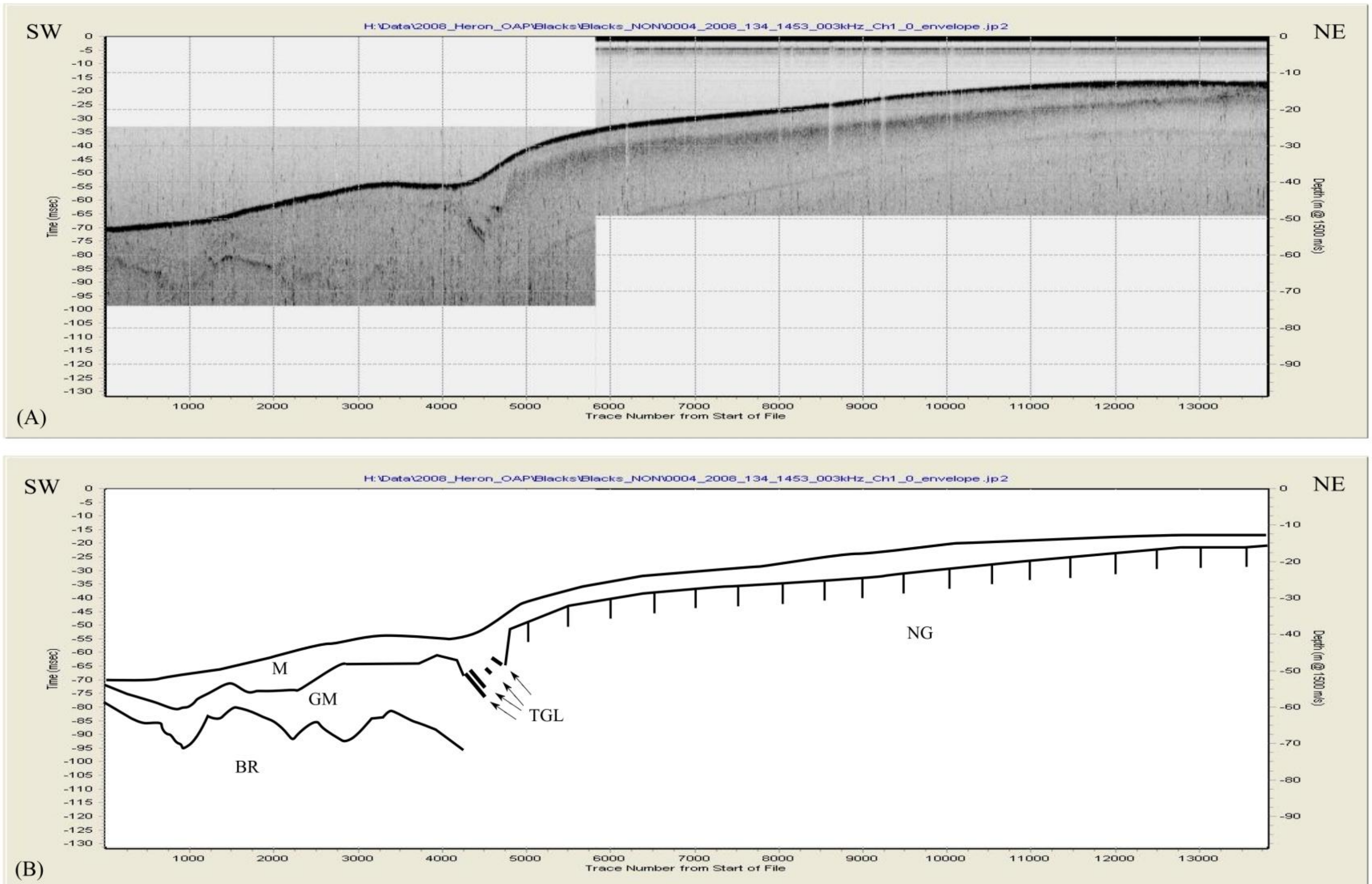

Figure 4.46: Seismic track line JD134_1453 for the Blacks survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, M = Holocene mud, NG = natural gas, and TGL = thin gravel layer.

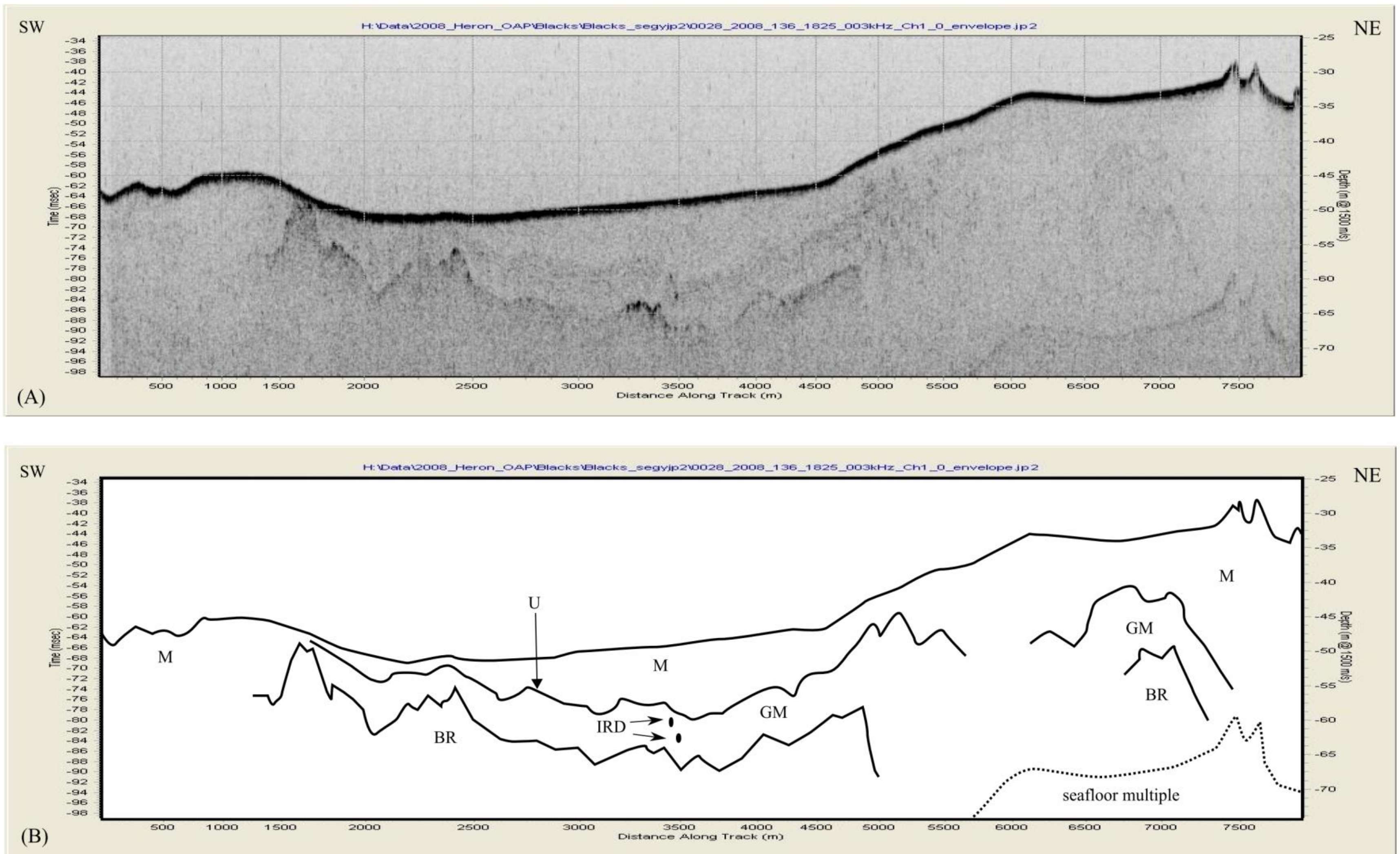

Figure 4.47: Seismic track line JD136_1825 for the Blacks survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity and M = Holocene mud.

this survey (Figures 4.46). The feature can be observed in the nearest to shore line sub-bottom line and the following 7 sub-bottom lines. It is found overlying the glacimarine unit, varying in depths from 45 m to 62 m; the length varies from 100 m to 400 m. All the units are approximately 1 m in thickness and the number of layers seen within each sub-bottom line varies from 1 to 5 m, lying horizontally and angled towards the northeast. This unit is interpreted as a thin gravel layer (TGL). Figure 4.48 shows the extent of theTGL mapped in the Blacks survey area. The TGL feature is located 3 km offshore of the Sheldon Point moraine. At the time of the moraine formation, large amounts of sediments and water were being seasonally discharged in the area; slides, slumps and sediment gravity flows transported sediments down the foreslope of the morainal bank into the proximal basin (Powell 1981). Similar features have been recognized in Maine by Barnhardt et al. (1997), and identified as thin gravel layer facies (Table 2.2).

A draping darkened shadow with a very sharp surface return is identified on the northeast side of track line JD134_1453 (Figure 4.46). This feature is interpreted as natural gas (NG) and is found in all the lines in the Blacks area that run from Saint John harbour to Negro Head point (Figure 4.48).

Unit 3 is acoustically transparent, and lies flat except where the thin gravel layer and the NG units meet; at this point it has a depression. In this area the unit dips seaward. It is always the upper most seismic unit and found in all lines (Figures 4.46 and 4.47). This unit is interpreted to represent postglacial sediments of Holocene mud and is given the designation M. This unit is interpreted as equivalent to King and Fader's (1986) LaHave clay.

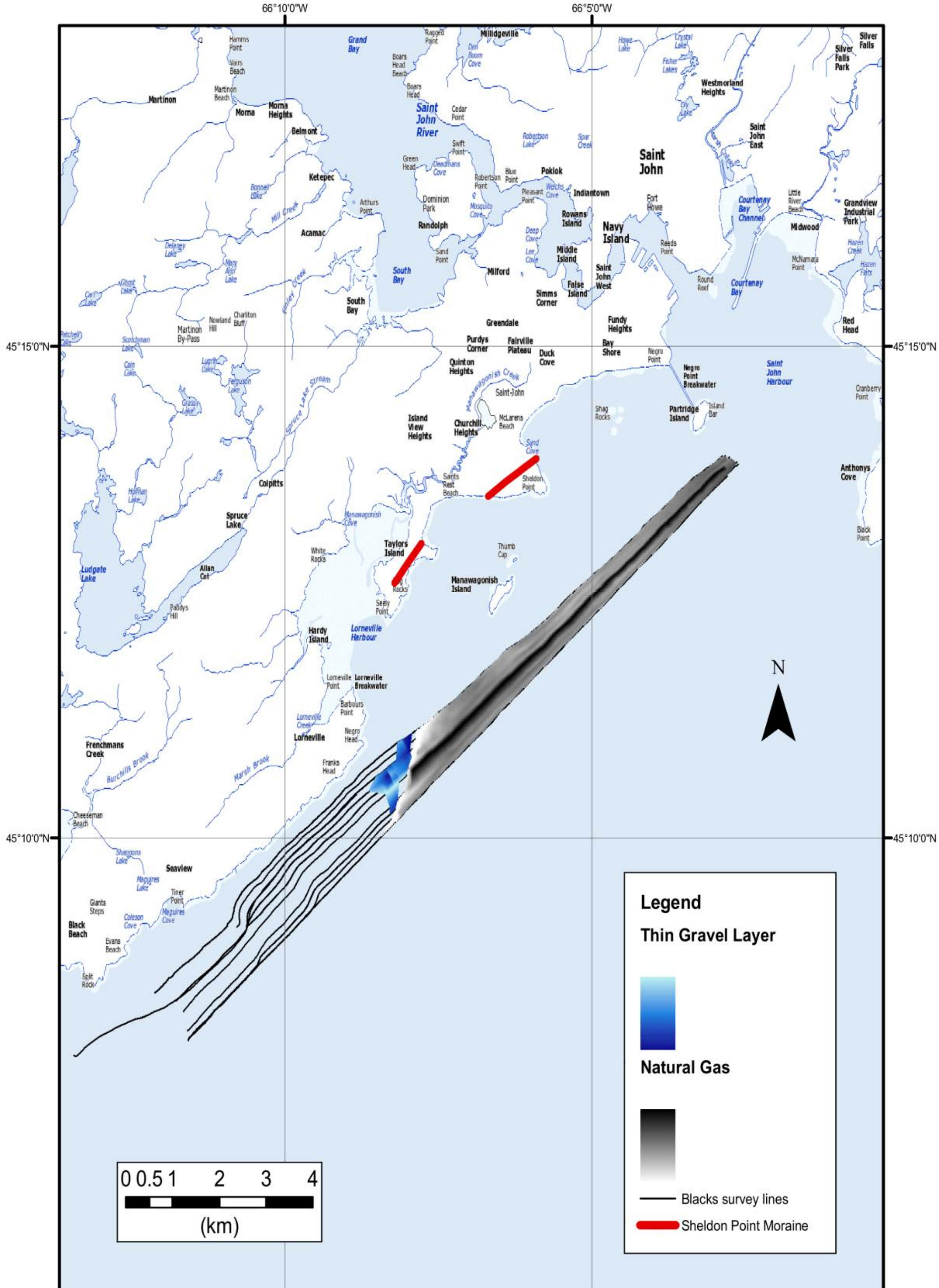

Figure 4.48: Location of thin gravel layer and natural gas for the Blacks survey area. The thin gravel layer likely represents a series of debris flows formed during the Sheldon Point Moraine formation.

4.2.13 Creed 2008 Survey Area

The Creed 2008 survey area runs offshore from Campobello Island and the northern tip of Grand Manan Island to offshore Lorneville Harbour near Saint John (Figure 4.49). The lines extend further offshore from the Pennfield, Maces, Chance, and Blacks surveys (Appendix 1). There are a total of 90 sub-bottom lines, which cover an area of approximately 780 km^2 and total over 1890 km of seismic lines. Many of the Creed 2008 sub-bottom lines include ship turns; these sections of the sub-bottom lines were not included in the study as they are distorted and unusable. The survey was conducted on board the CCGS Frederick G. Creed during the 2008 cruise. The majority of the lines run parallel to the shoreline in a southwest-northeast direction, only 6 turn and continue on in a southeast-northwest direction going from offshore Campobello Island to Grand Manan Island. The data also include 15 m resolution bathymetric and backscatter data collected by the CCGS Frederick Creed in 2008.

The bathymetry of the Creed 2008 (Figure 4.50) survey is a continuation from the Pennfield, Maces, Chance, and Blacks surveys. The Pennfield bathymetry further offshore shows a gradual increase in depth with increasing distance from the shoreline. To the south of the Maces Bay lobate feature is a sinuous ridge running in a southwest to northeast trend that is 3.2 km in length. The Creed 2008 data offshore Maces Bay and Chance shows large lobes of sediment extending from nearshore to offshore (Figure 4.16). At the eastern end of Eastern Wolf Island are a series of elongated indentations trending southwest to northeast, varying in length from less than 100 m to 800 m.

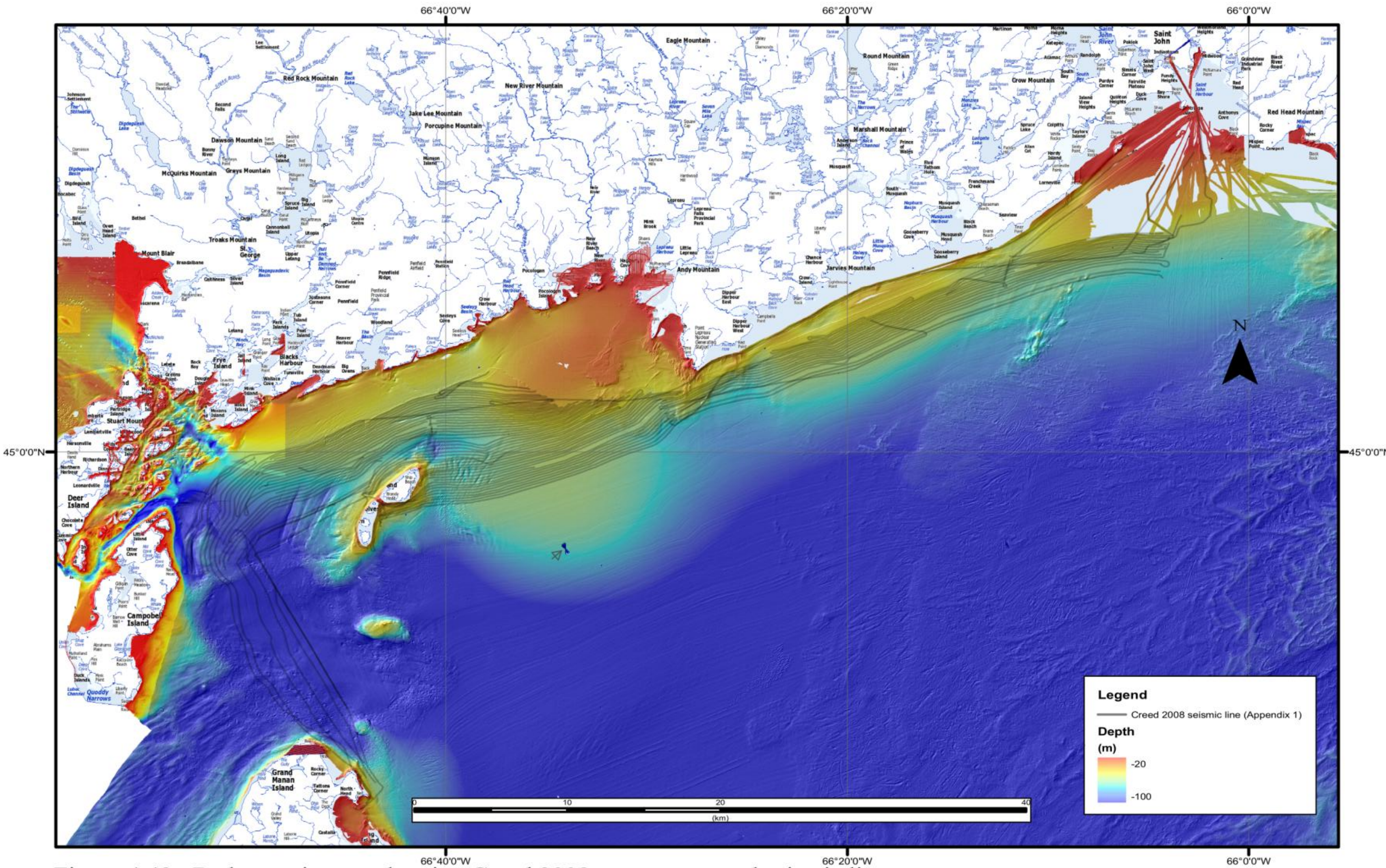

Figure 4.49: Bathymetric map showing Creed 2008 survey area and seismic lines.

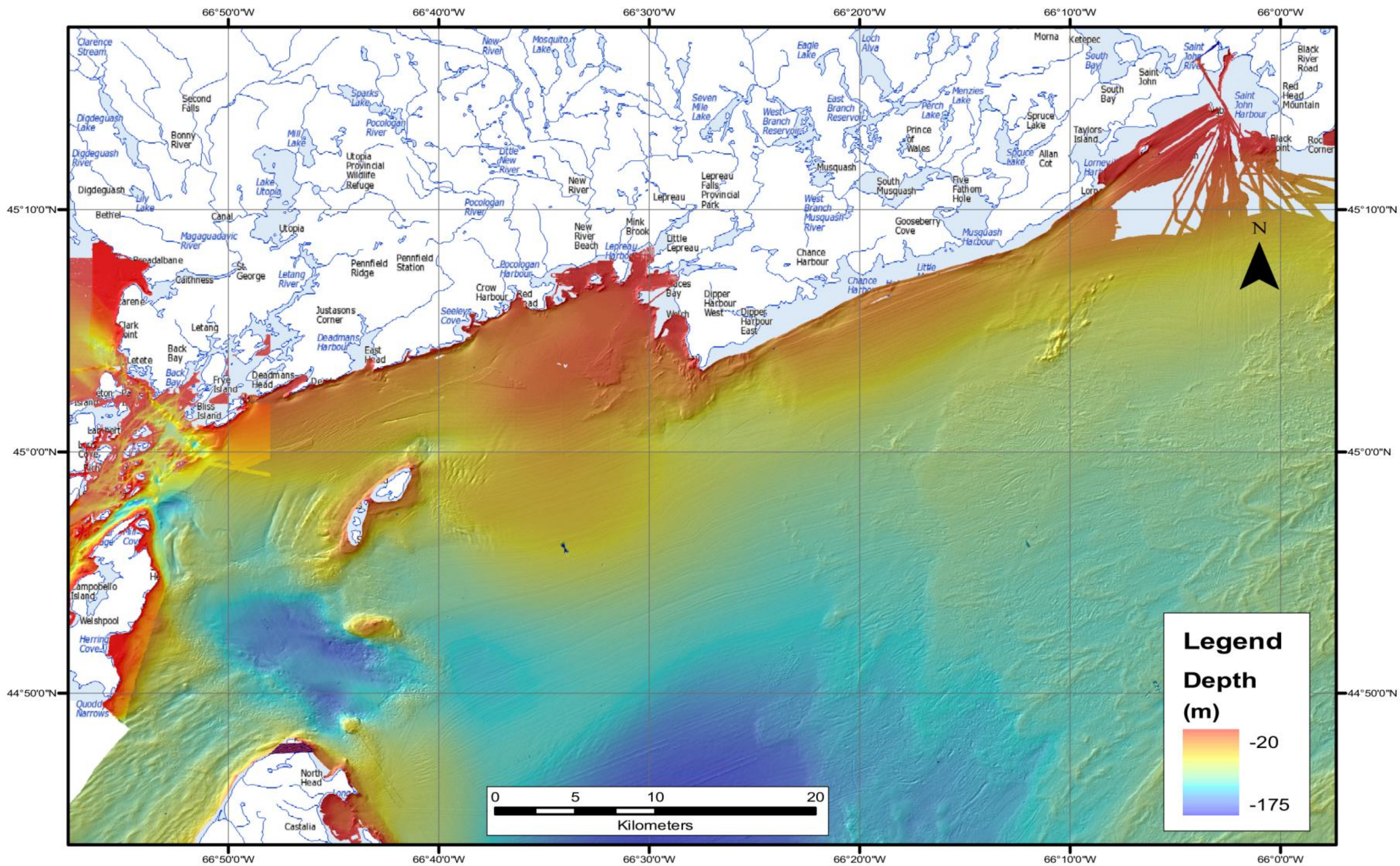

Figure 4.50: Creed 2008 survey area bathymetry.

On the south western side of the Wolves Islands is a series of elongated pockmarks, ranging in length from 3.5 to 4.5 km (Figure 4.16). Between the Wolves and Grand Manan islands is a basin reaching 150 m at its deepest; surrounding the basin area the seafloor is pitted and gouged by iceberg scours (Figure 1.9). Scours and pits have been documented in the Bay of Fundy at depths ranging from 225 to 90 metres (Parrott et al. 2008). On the western side of the Southern Wolf Island is a series of longer elongated indentations (Figure 4.50) trending southwest to northeast. They vary in length from 2500 m to 4500 m and are of unknown origin.

4.2.14 Creed 2008 Survey Stratigraphy

The stratigraphy of the Creed 2008 data has been divided up according to the nearer shore surveys presented in the earlier sections, which includes Campobello, Pennfield, Maces, Chance and Blacks combined. The Creed 2008 data (Appendix 1) is not as good quality as the Heron data, the poorer quality image sub-bottom images show less resolution.

The Creed 2008 seismic lines that are near Campobello Island are comprised of 6 lines that go in a northwest to southeast direction, from the northern tip of Grand Manan island to the northwestern side of Campobello Island (Figure 4.51). Unit 1 is the lowest observed unit and is interpreted as bedrock (BR, Figure 4.52). It is characterized by a strong, even, high intensity return on a highly irregular surface and generally displays a peak and valley morphology. It can be followed from nearshore bedrock highs near the northern tip of Grand Manan Island (Figure 4.52). Unit 2 drapes over the bedrock, displaying a pattern of internal reflectors, and is interpreted as glacimarine (GM, Figure 4.52) similar to Friars Bay in the Heron 2008 Campobello survey (Figure 4.6).

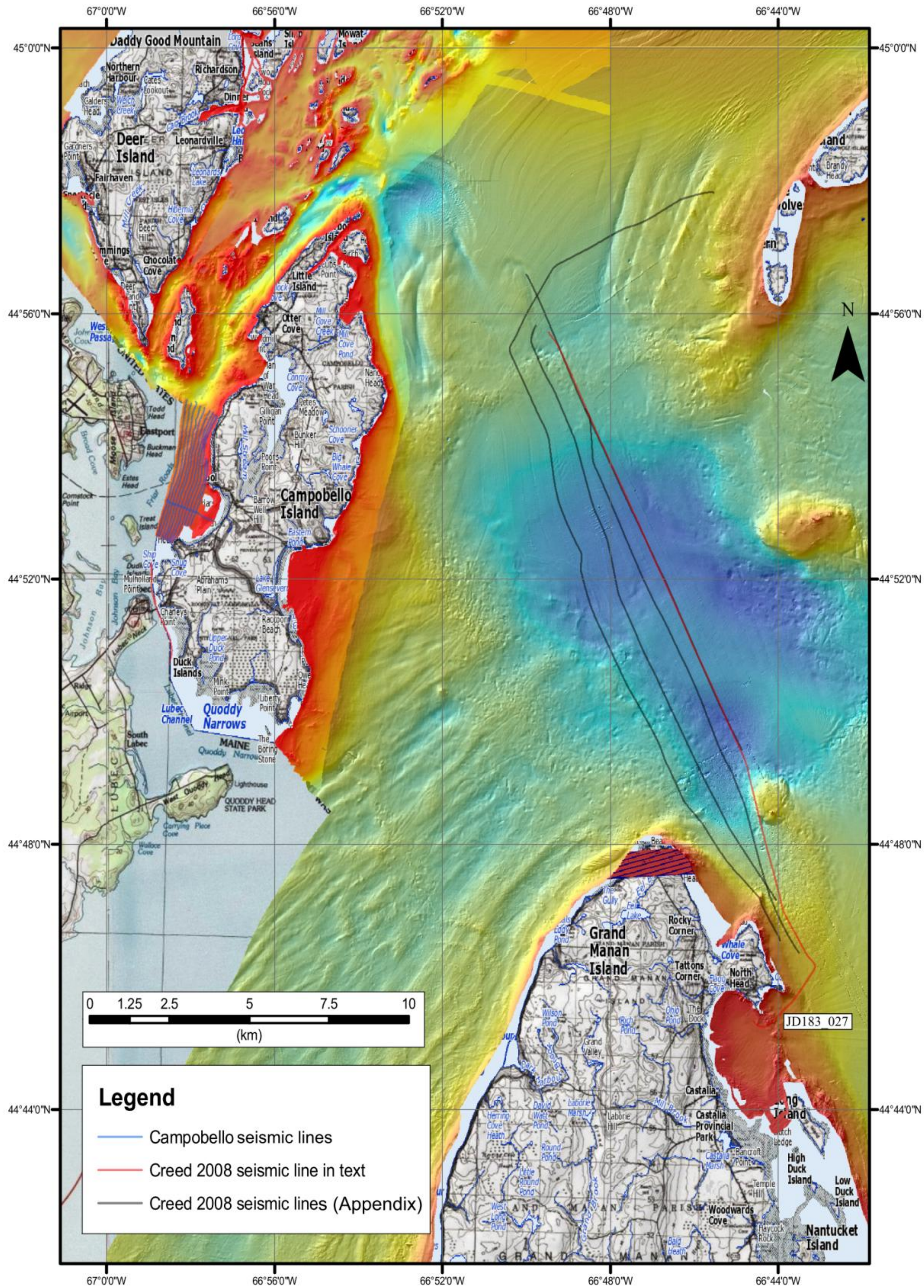

Figure 4.51: Creed 2008 and Campobello area and survey lines.

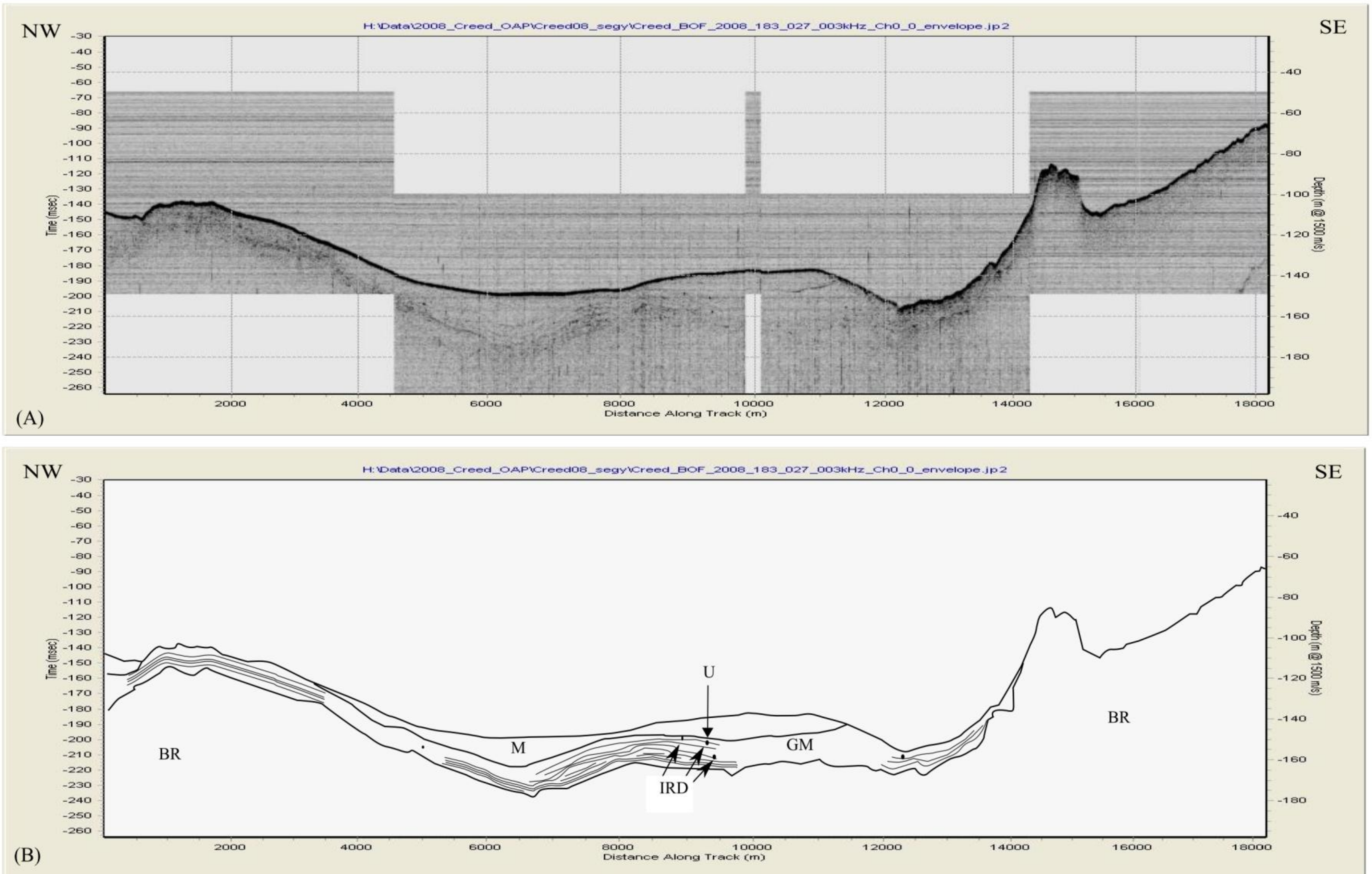

Figure 4.52: Seismic track line JD183_027 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the stratification in the GM representing alternations in the sand and silt content.

The intensity of the reflectors varies in unit 2 from strong with very distinct internal returns to moderate and weak. An unconformity (U) is seen at the base of the Holocene mud, unit 2, and the unit abruptly ends at a valley (Figure 4.52). Occasionally high points of reflection are observed, likely ice-rafted debris (IRD). The top unit, Unit 3, is documented in all the sub-bottom lines. It is interpreted as Holocene mud (M), and drapes over the lower units.

The Pennfield data collected in 2008 on the Creed survey are comprised of 30 seismic lines that generally run in a southwest to northeast direction parallel to the shoreline, but after the Wolves Islands turn into a west to east direction (Figure 4.53). These sub-bottom lines show a continuation of the same pattern as observed in the nearer shore seismic data discussed earlier. The lowest observed unit is bedrock (BR), displaying peak and valley morphology and having a very high intensity return from an irregular surface. This is overlain in a ponded fashion by Unit 2, the glacimarine unit (GM). Due to poorer resolution it is difficult to see the separation of GM-1 and GM-2 in all of the lines (Figure 4.54). Several of the survey lines go through the elongated depressions on both the western and eastern sides of The Wolves (Figure 4.53). The elongated depressions on the southwestern flank of The Wolves show a pockmark-like indentation within the Holocene mud that is draped over a bedrock valley (Figure 4.55). The depressions follow highs and lows in the bedrock depressions (Figure 4.55) and may represent elongated glacial ridges and troughs, created when glaciers actively overrode the bedrock. The depressions probably are sustained by high energy currents.

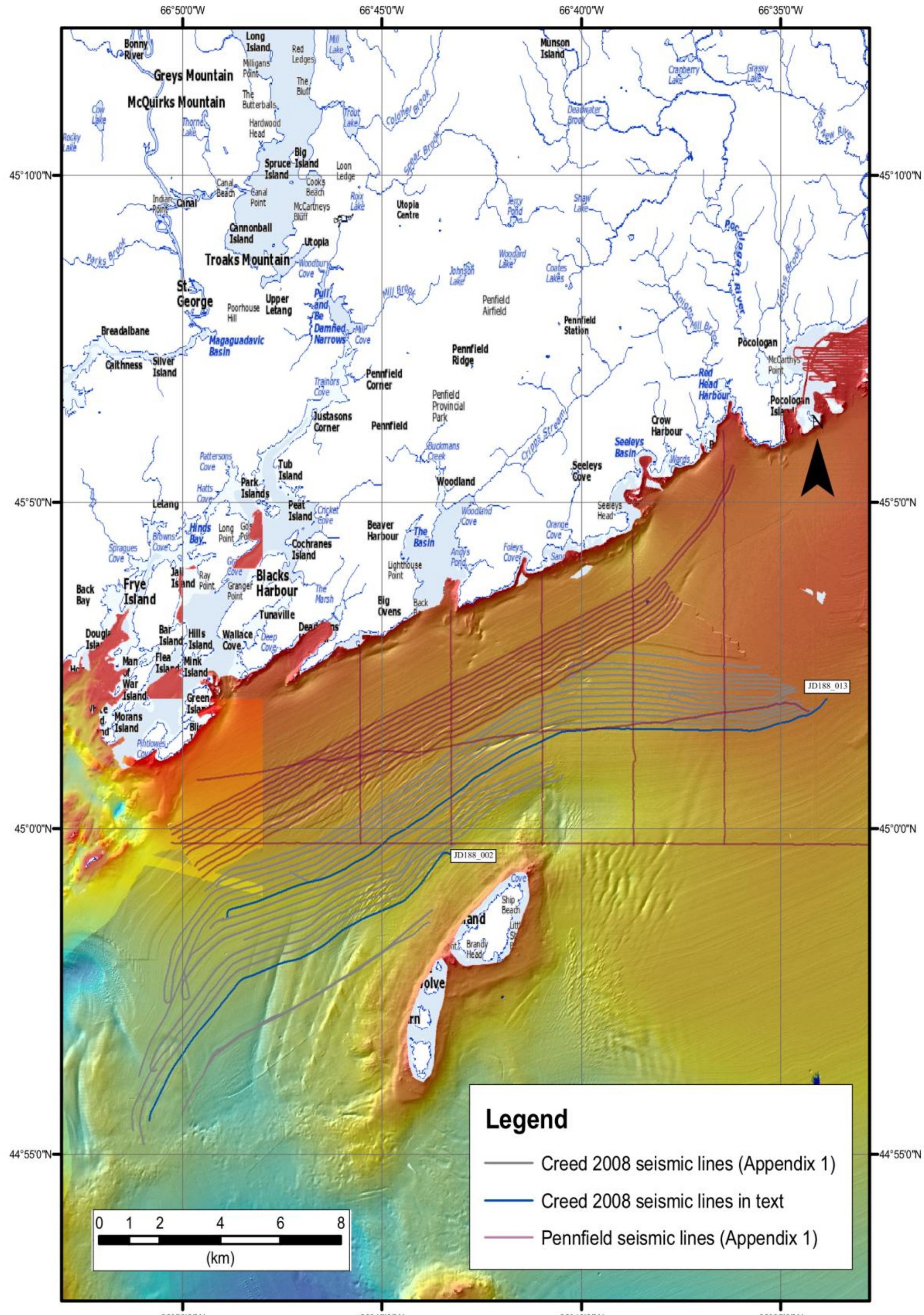

Figure 4.53: Location of Creed 2008 and Pennfield seismic lines.

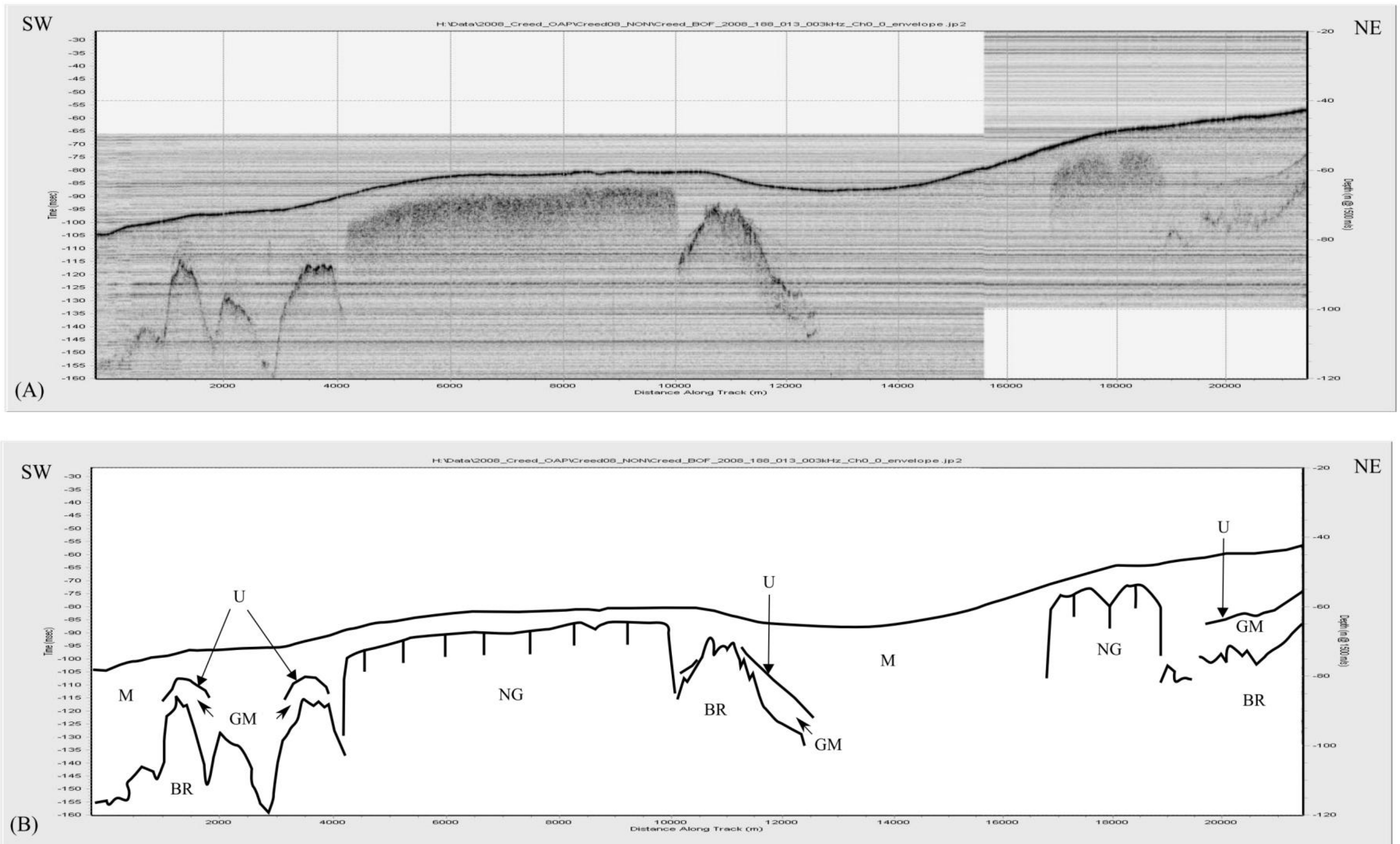

Figure 4.54: Seismic track line JD188_013 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud.

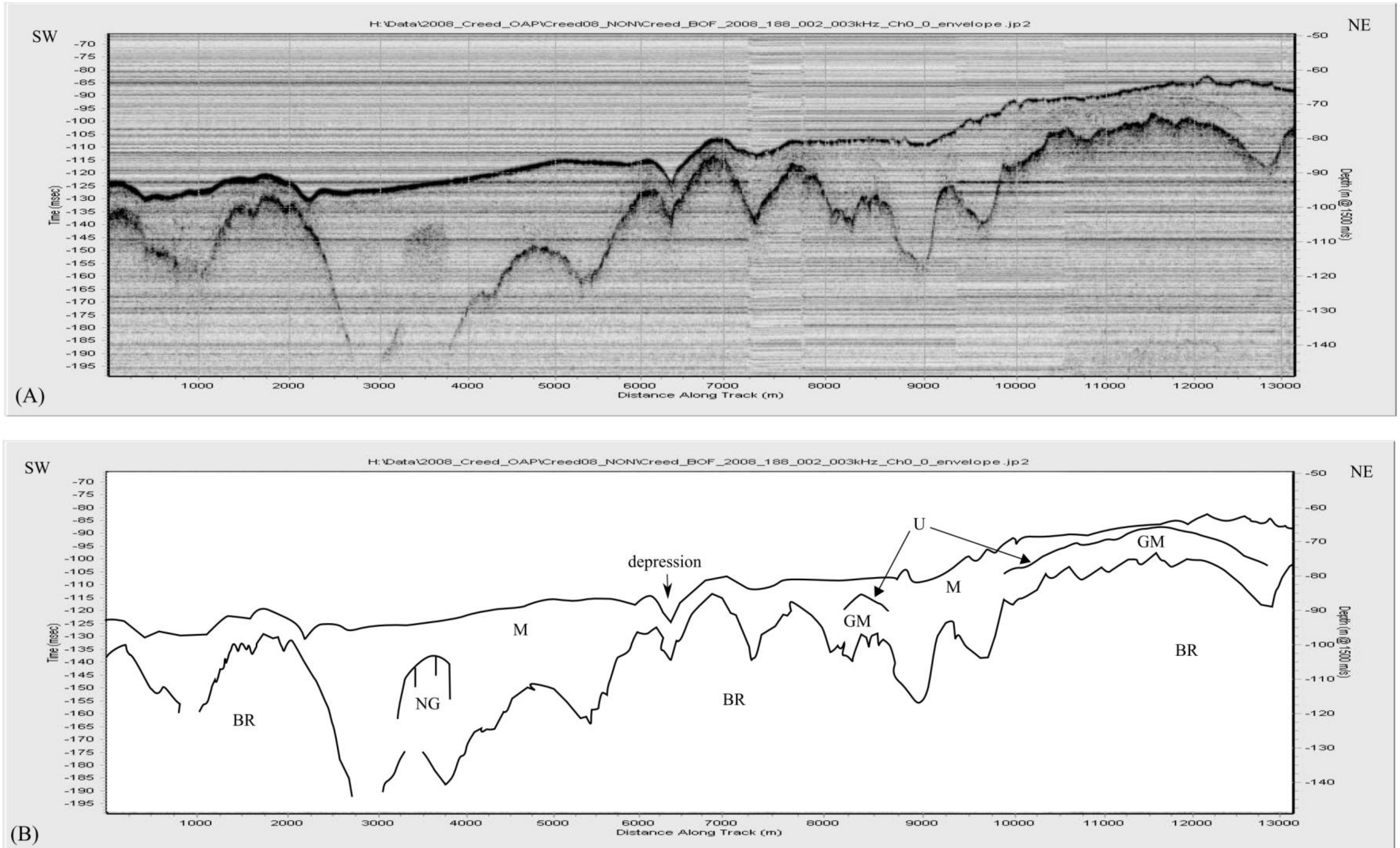

Figure 4.55: Seismic track line JD188_002 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud. The bedrock displays hummocky ridges and troughs trending north south.

The Maces Bay data collected from the Creed extend further offshore from the Maces Bay data collected by the Heron in 2008, going from Pt. Lepreau to the eastern side of the Wolves Islands. The offshore Creed survey includes 32 seismic reflection lines (Figure 4.56), with the majority of the lines trending in a southwest to northeast direction going from the Wolves Island chain to the southern tip of Pt. Lepreau. Unfortunately, there is a large gap with no data in between the Maces Heron 2008 and the Maces Creed 2008 surveys. Figure 4.57 shows the location of 6 seismic lines at the eastern end of the survey area.

The lowest observed unit is bedrock (BR), Unit 1, with a very high intensity return from the irregular surface (Figure 4.58). It can be seen over the nearshore bedrock highs, but it dips steeply below the survey window and apart from a local high at 7000 m, it is not observed throughout the majority of the line. Overlying bedrock is till (T); dipping below the survey window throughout most of the sub-bottom line except where the unit occurs near shore. The lower bounding surface generally has a more intense return than the upper bounding surface return. However, the upper bounding surface is also intense but shows more variability. At the eastern end of the survey, a till unit outcrops to the surface, as seen on Figures 4.57, 4.58 and 4.59, it appears as a rougher surfaced lobe-shaped landform than the surrounding Holocene mud. Of the six lines shown in Figure 4.57, three go over the till unit, but the most southerly line does not. As is seen in the corresponding 3 sub-bottom lines, JD184_014 (Figure 4.58), JD184_018 (Figure 4.59), and JD183_024 (Figure 4.60), which show the till unit at the seafloor, compared to JD182_003 (Figure 4.61), just south of the lobe-shaped landform the till unit is draped by another seismic unit.

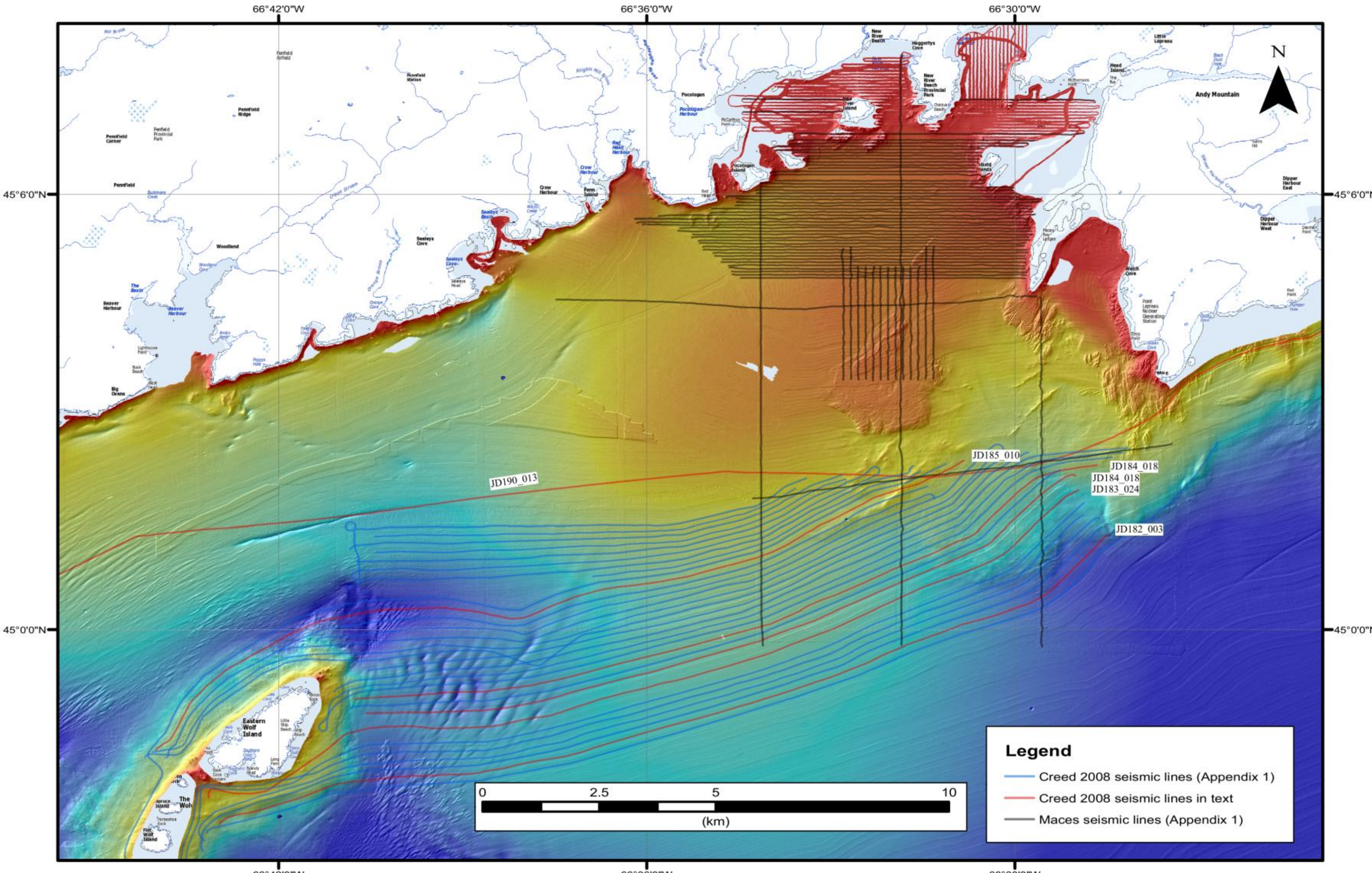

Figure 4.56: Location of Creed 2008 and Maces Bay survey lines.

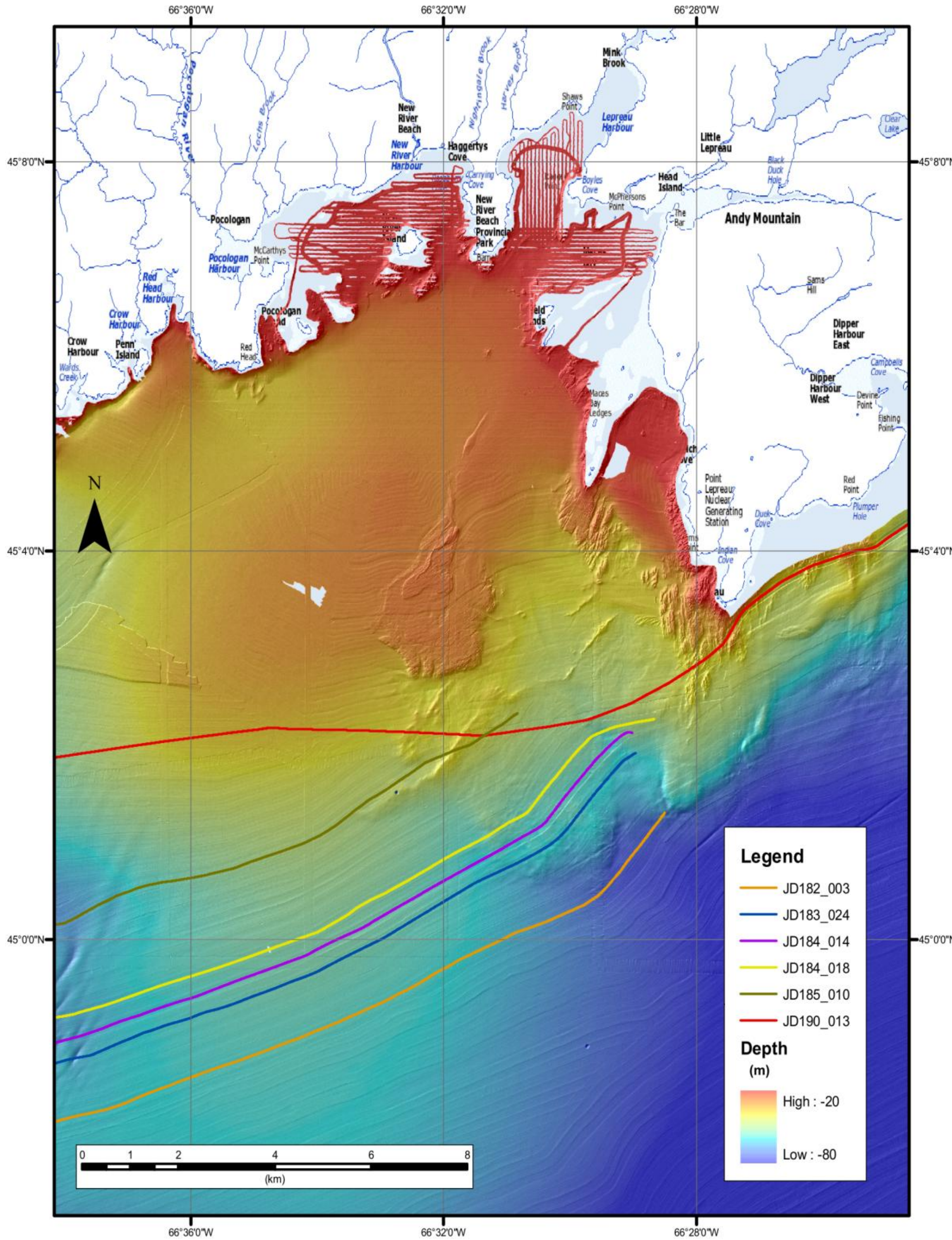

Figure 4.57: Location of Creed 2008 seismic lines in Maces Bay discussed in text.

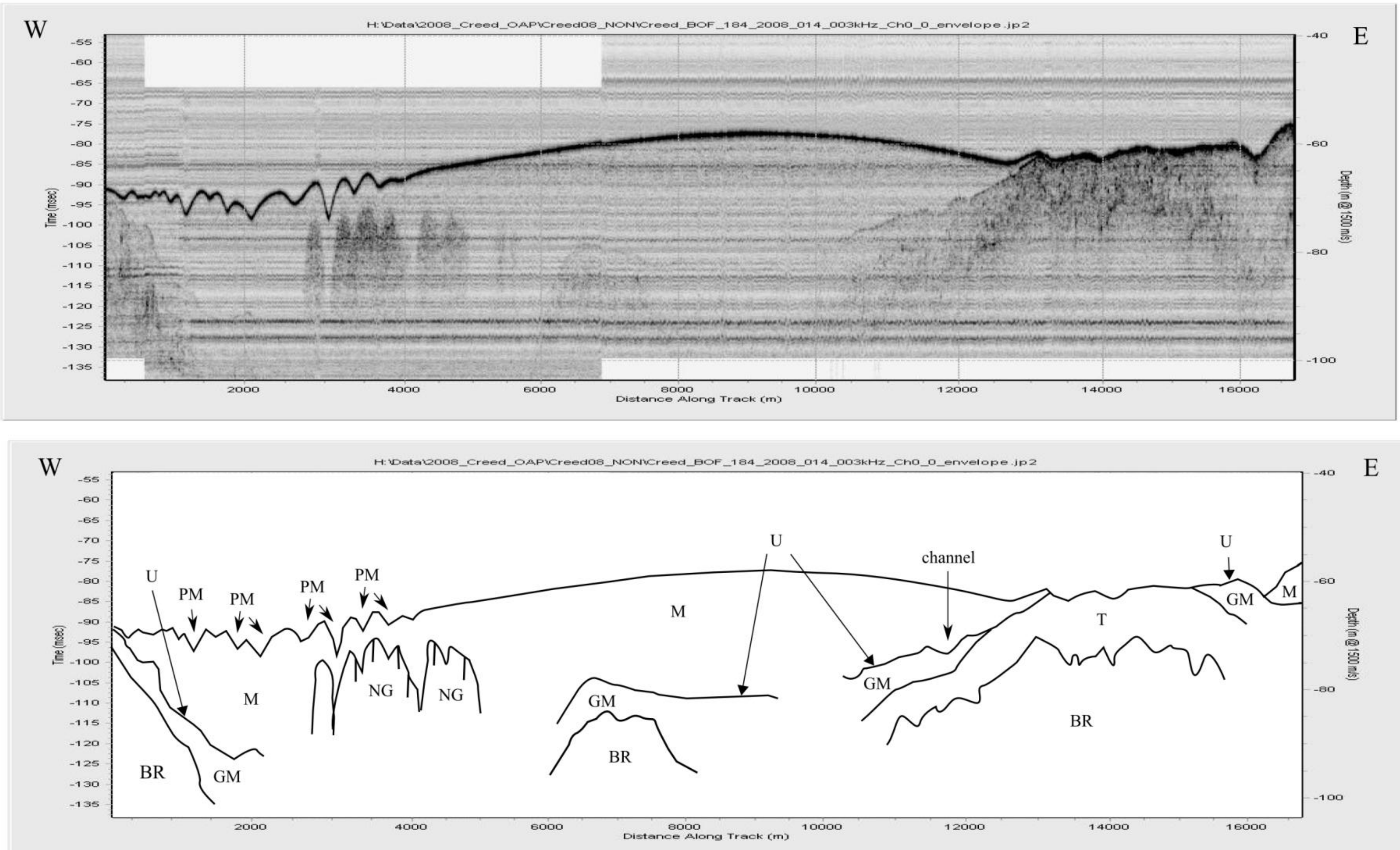

Figure 4.58: Seismic track line JD184_014 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, NG = natural gas, U = Pleistocene/Holocene unconformity, M = Holocene mud and PM = pockmarks.

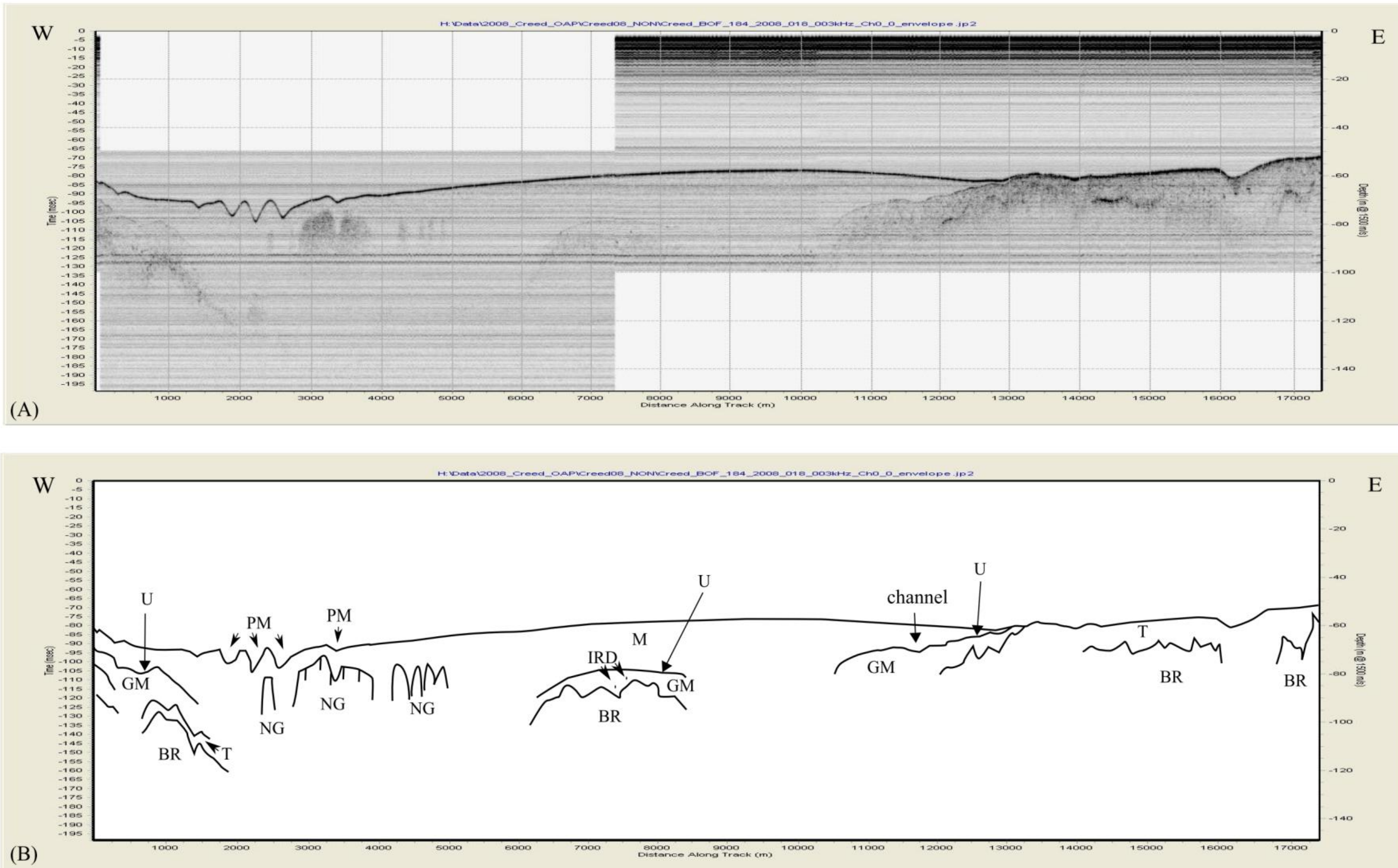

Figure 4.59: Seismic track line JD184_018 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, IRD = ice rafted debris, NG = natural gas, U = Pleistocene/Holocene unconformity M = Holocene mud and PM = pockmarks.

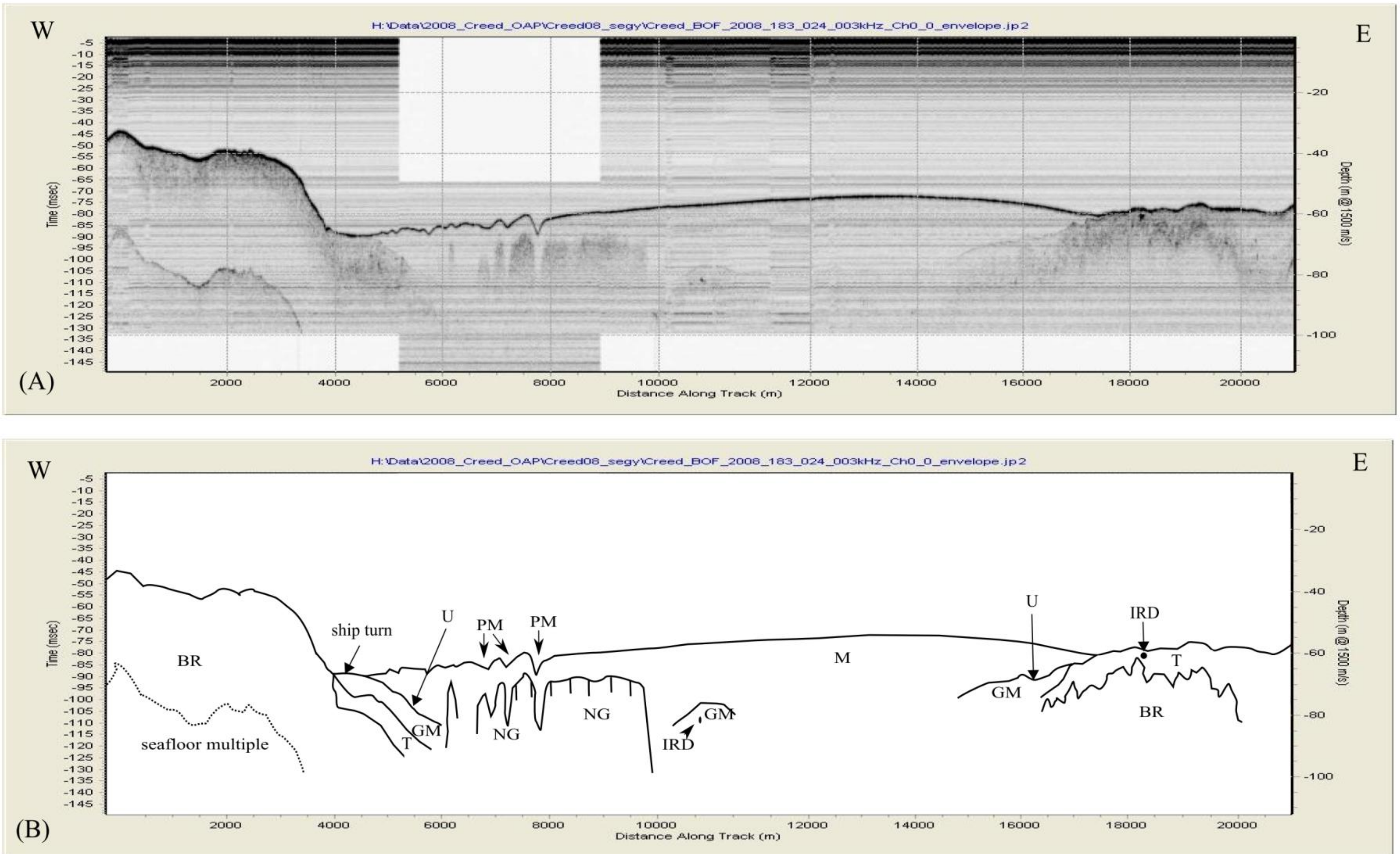

Figure 4.60: Seismic track line JD183_024 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till , GM = glacimarine sediment, IRD = ice rafted debris, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud.

On the bathymetric map (Figure 4.57) a raised ridge is observed south of the delta in Maces Bay. Figure 4.57 shows two of the seismic lines going over this landform, with Figures 4.62 and 4.63 showing seismic lines and the interpreted line drawing. Along the track lines of Figures 4.62 and 4.63 at 35000 m and 18000 m, the bedrock can be seen to be overlain by a moraine containing till. In neighbouring Maine, Bacchus (1993) found that bedrock highs may have provided a buttressing point for the glacier during deglaciation, where ice retreat was stalled temporarily, allowing for the thickening of the sand and gravel creating a moraine from eroding till in the surrounding area. As seen from the sub-bottom image the bedrock lies close to the surface, providing a buttress for the retreating glacier to stall and deposit till.

Overlying the till is Unit 3; this unit is interpreted as glacimarine (GM) and is seen in all lines of the survey. The lower bounding surface of unit 3 has a more intense return than the upper bounding surface. The upper bounding surface intensity shows more variability, from an intense return to less intense. The unit contains occasional high point reflectors interpreted as ice-rafted debris (IRD, Figure 4.60). As with the nearer shore Maces Bay survey, an unconformity (U) is recognized up to depths at 70 m below sea level. A channel like depression in the unconformity is noted, which represents fluvial deposits (e.g. Figure 4.58). These channels are infilled with Holocene mud and are consistently 200 metres wide and 2 metres deep.

Unit 4 is an acoustically transparent, dome-shaped unit (Figure 4.61) that is always present as the upper most seismic unit and is found in all the seismic lines. It is interpreted as Holocene mud (M), and is the most massive unit in the Creed 2008 survey.

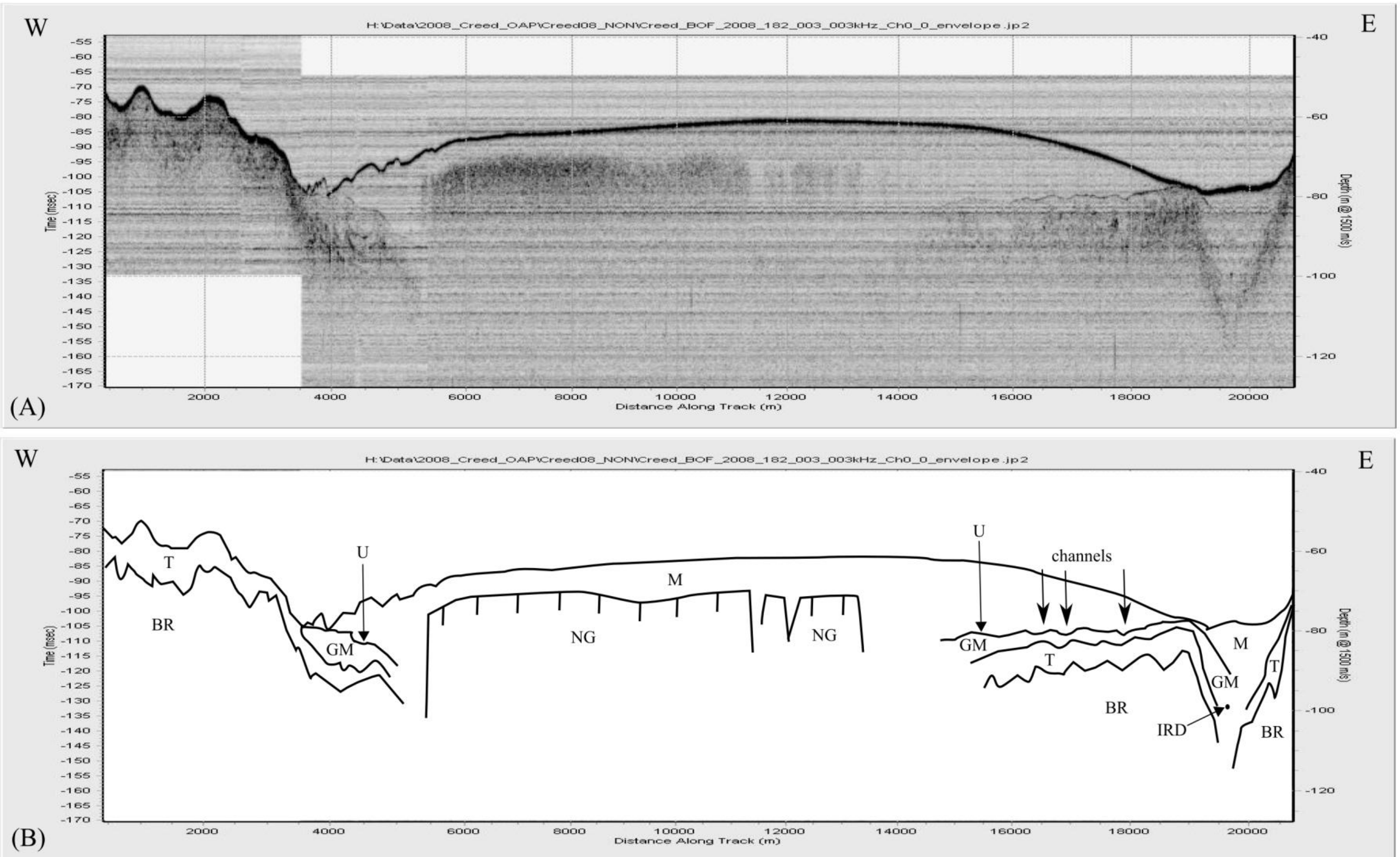

Figure 4.61: Seismic track line JD182_003 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till , GM = glacimarine sediment, IRD = ice rafted debris, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud.

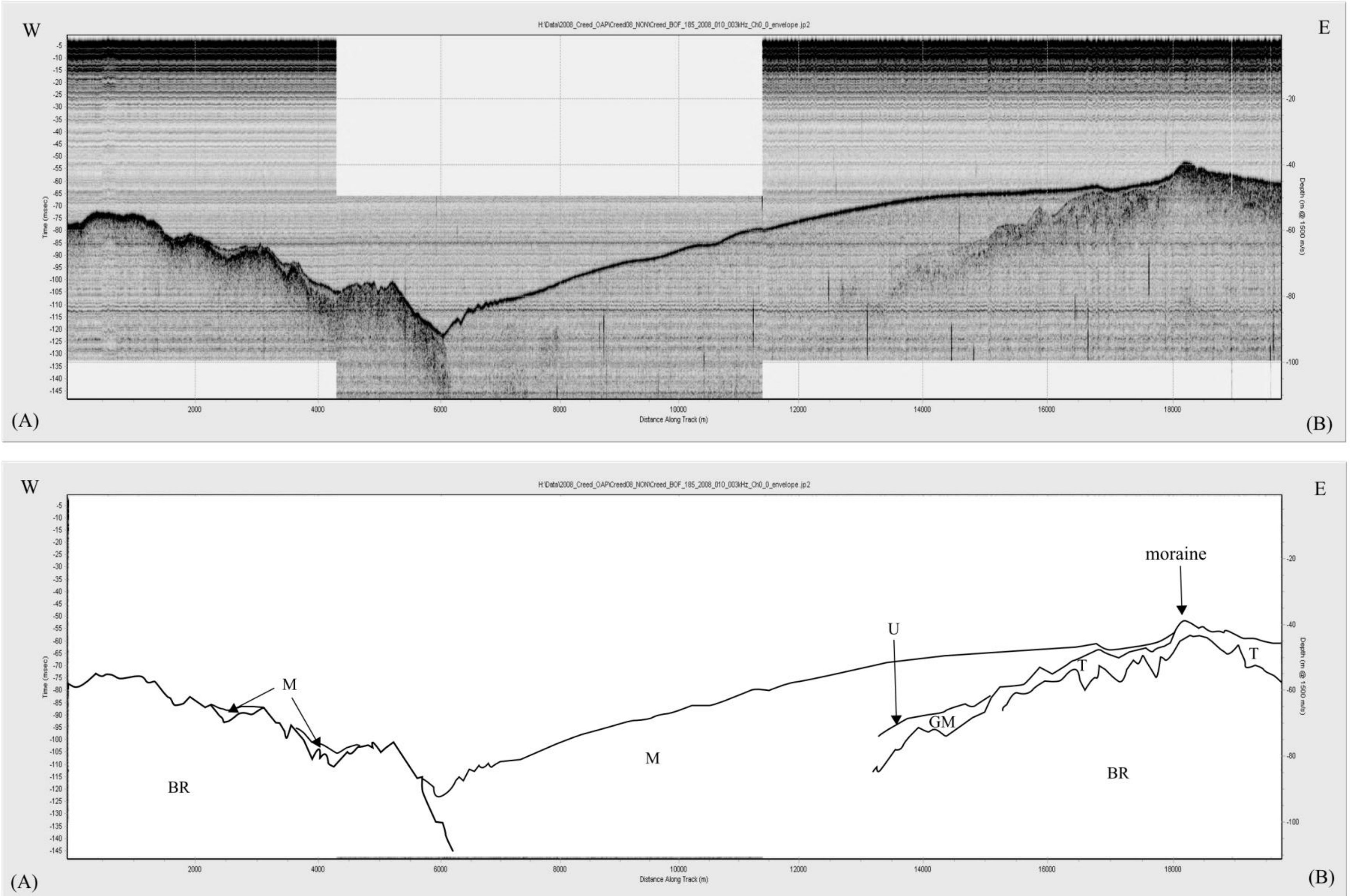

Figure 4.62: Seismic track line JD185_010 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, U = Pleistocene/Holocene unconformity and M = Holocene mud. Note the moraine at 18000 m.

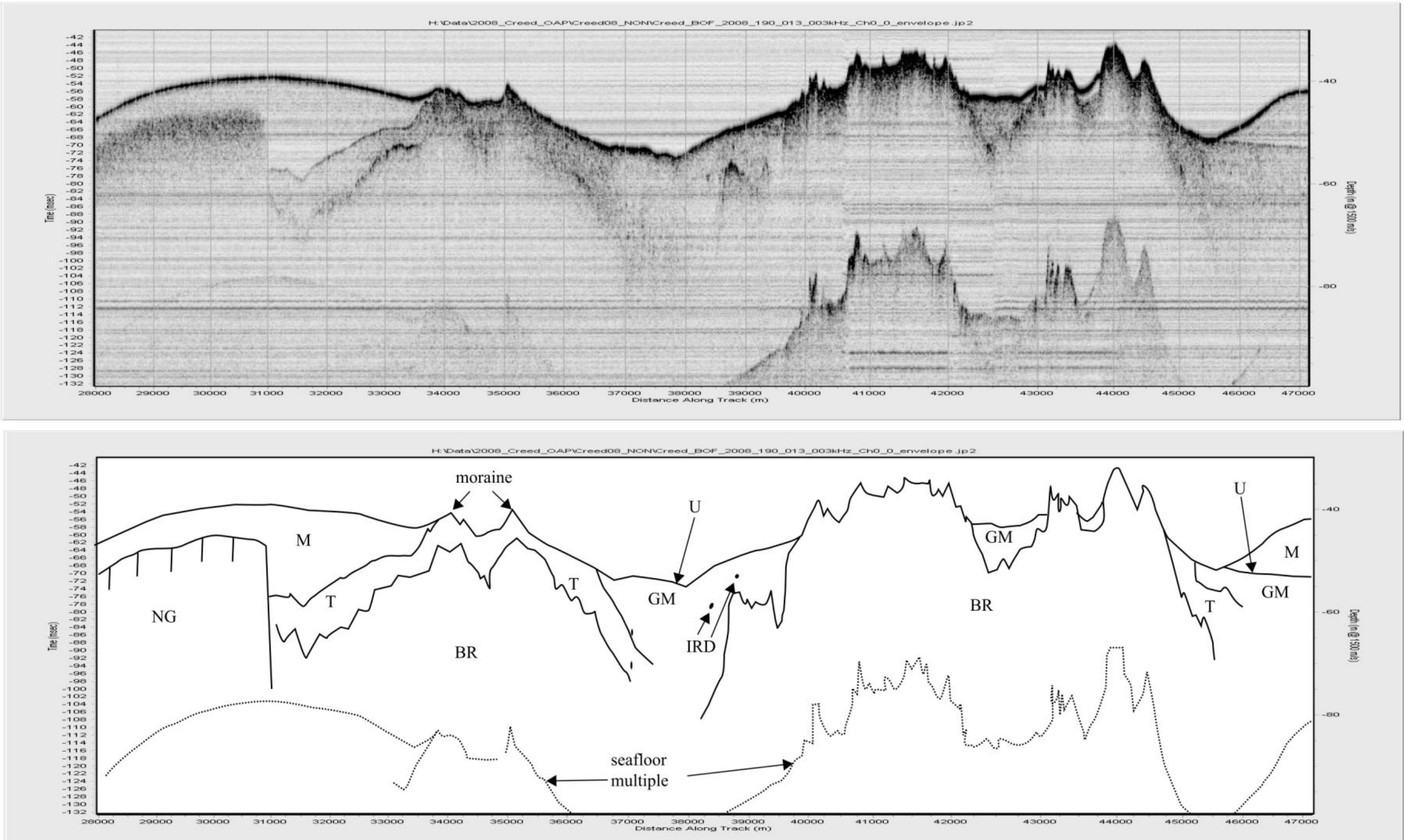

Figure 4.63: A section along the seismic track line JD190_013 for the Creed 2008 survey area located in Maces Bay. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, IRD = ice rafted debris, NG = natural gas, U = Pleistocene/Holocene unconformity and M = Holocene mud.

A draping darkened shadow with a very sharp surface return occurs which is interpreted as natural gas (NG). Offshore of Maces Bay it occurs in many lines located in the outer western flanks of the bay and off the eastern side of the Wolves islands (e.g. Figure 4.60). Figure 4.64 shows the extent of gas mapped within the Creed 2008 survey. Tracks on the northeastern side of The Wolves show elongated pockmarks within the Holocene mud layer, with natural gas below. The pockmarks show indentations directly below, indicating that they formed from gas expulsion (e.g. Figure 4.60).

The Creed 2008 survey for Chance and Blacks data are combined into one, as the survey lines are continuous through both areas (Appendix 1). The survey area totals 25 seismic lines running parallel to the shoreline in a southwest to northeast direction (Figure 4.65). The lowest observed unit is interpreted as bedrock (BR), and is found in all the lines. The BR unit is characterized by a strong, high intensity return on a highly irregular surface and displays a peak and valley morphology in the survey lines (Figure 4.66). Overlying BR is Unit 2. It is found in all the lines, and is interpreted as till (T, Figure 4.67). It is draped over the BR, and displays a moderate to high intensity of return on both the upper and lower bounding surface on an irregular surface. Occasionally high points of reflection are observed; Figure 4.68 shows an exceptionally large reflective point, likely representing large boulders. The internal configuration of the (T) unit is dark and chaotic. To the east and west of Musquash Harbour the till unit is not overlain by any sediments and forms the sea floor at the southwestern and northeastern ends of the line (Figure 4.66). Overlying the till is Unit 3, this unit is interpreted as glacimarine and is seen in all the lines of the survey. It is found in all parts of the line west of the thin gravel layer discussed in the Blacks survey, but is absent to the east of this unit.

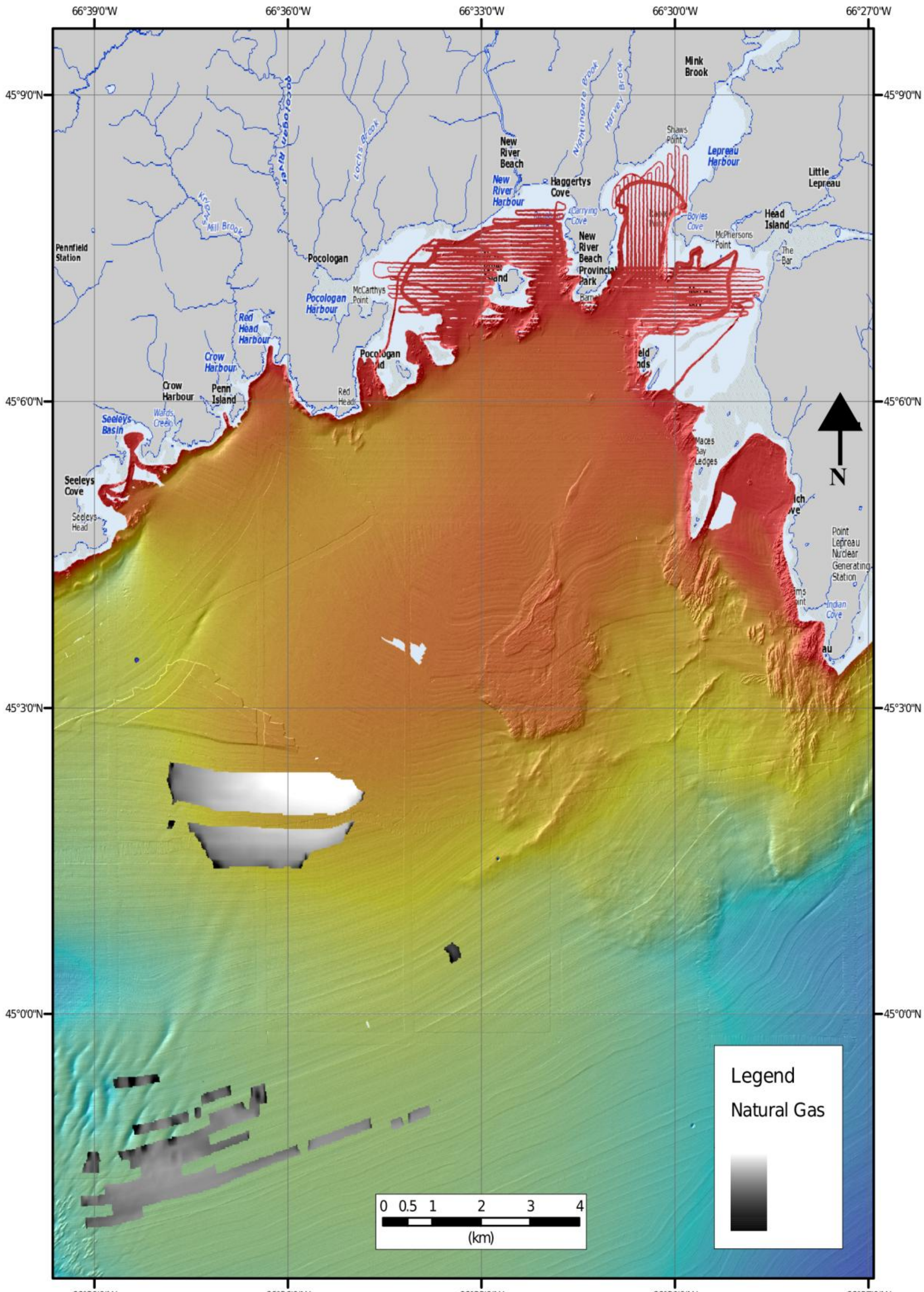

Figure 4.64: Location of natural gas in the Creed 2008 survey area.

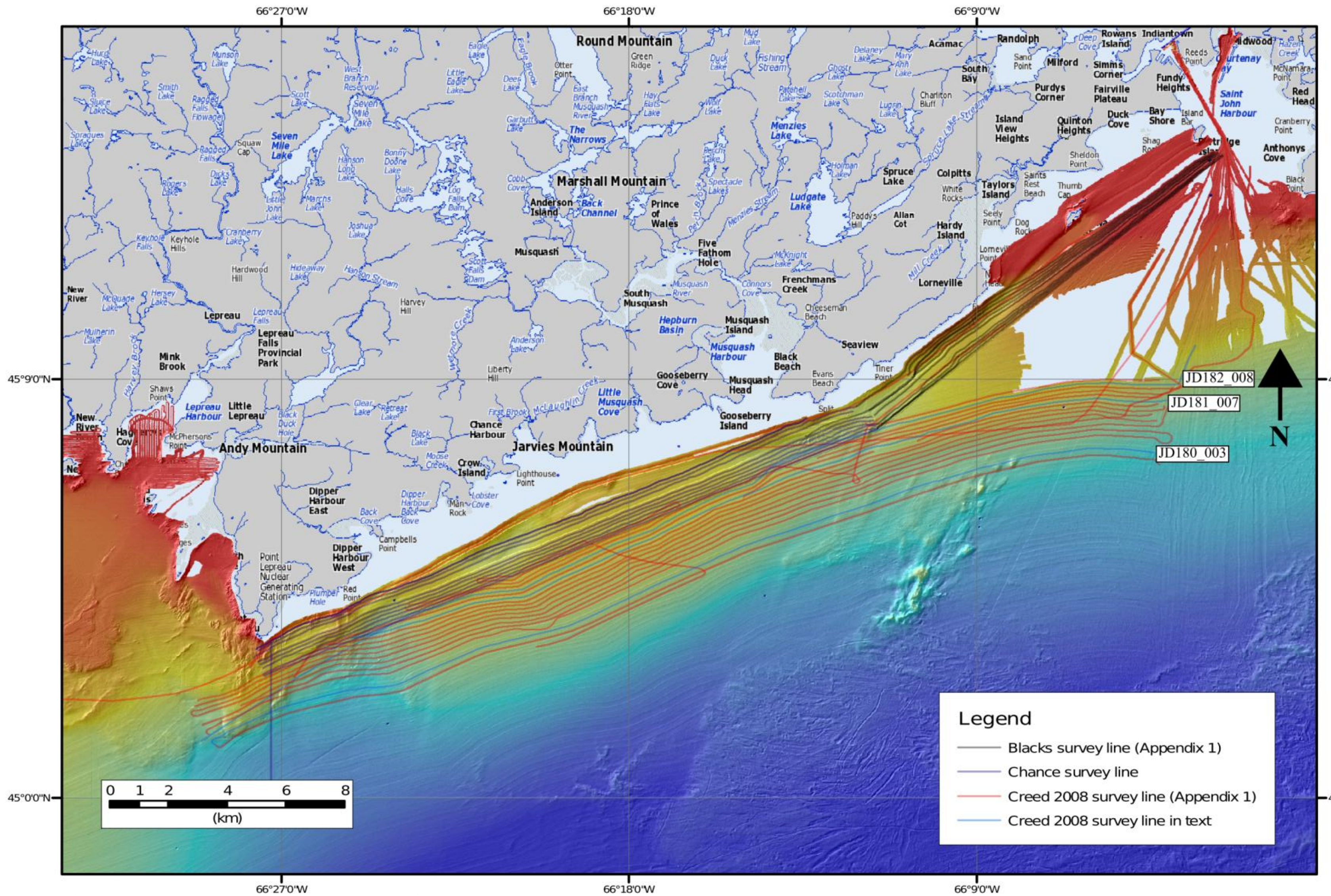

Figure 4.65: Location of the Creed 2008, Chance and Blacks survey.

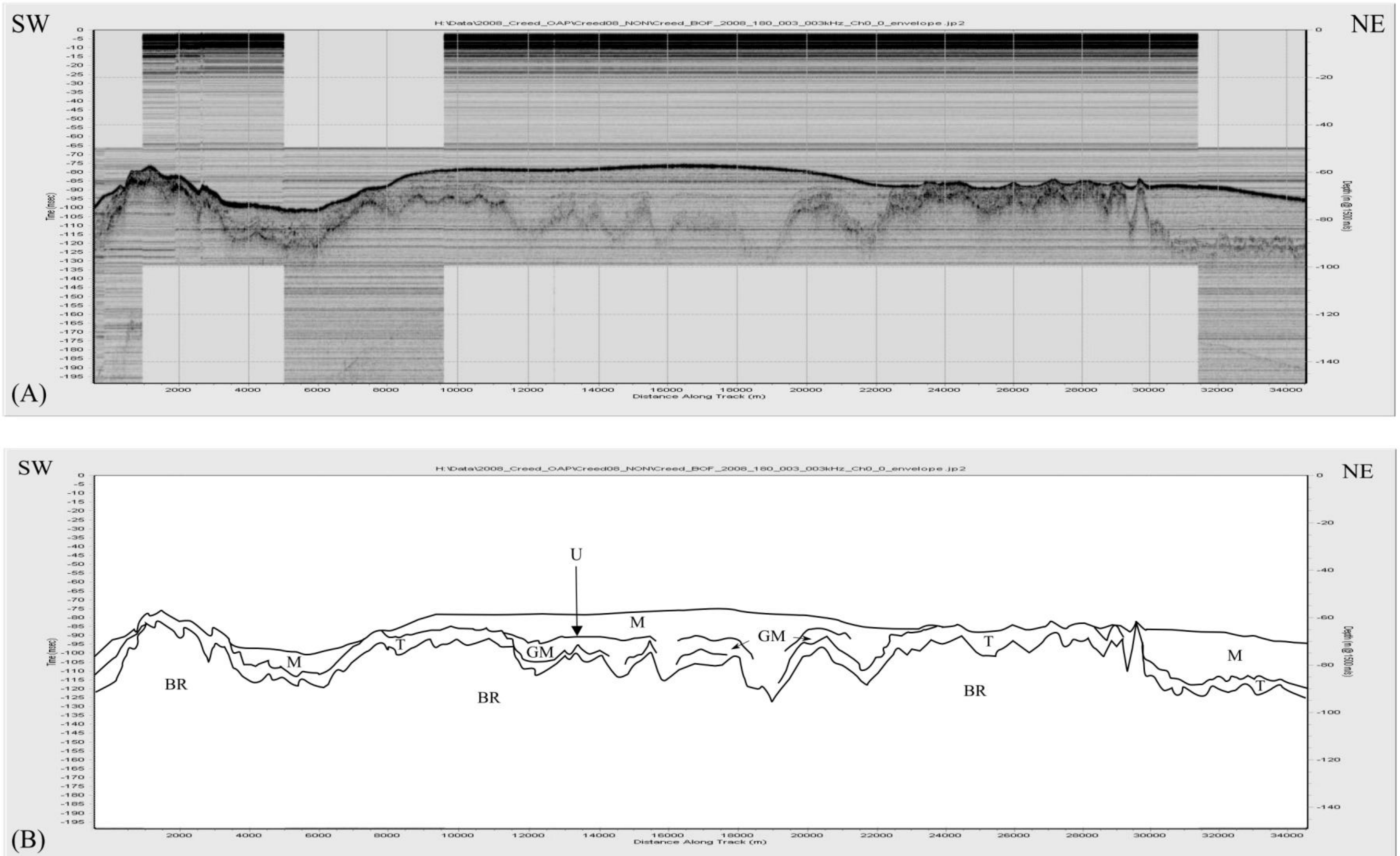

Figure 4.66: Seismic track line JD180_003 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, U = Pleistocene/Holocene unconformity and M = Holocene mud.

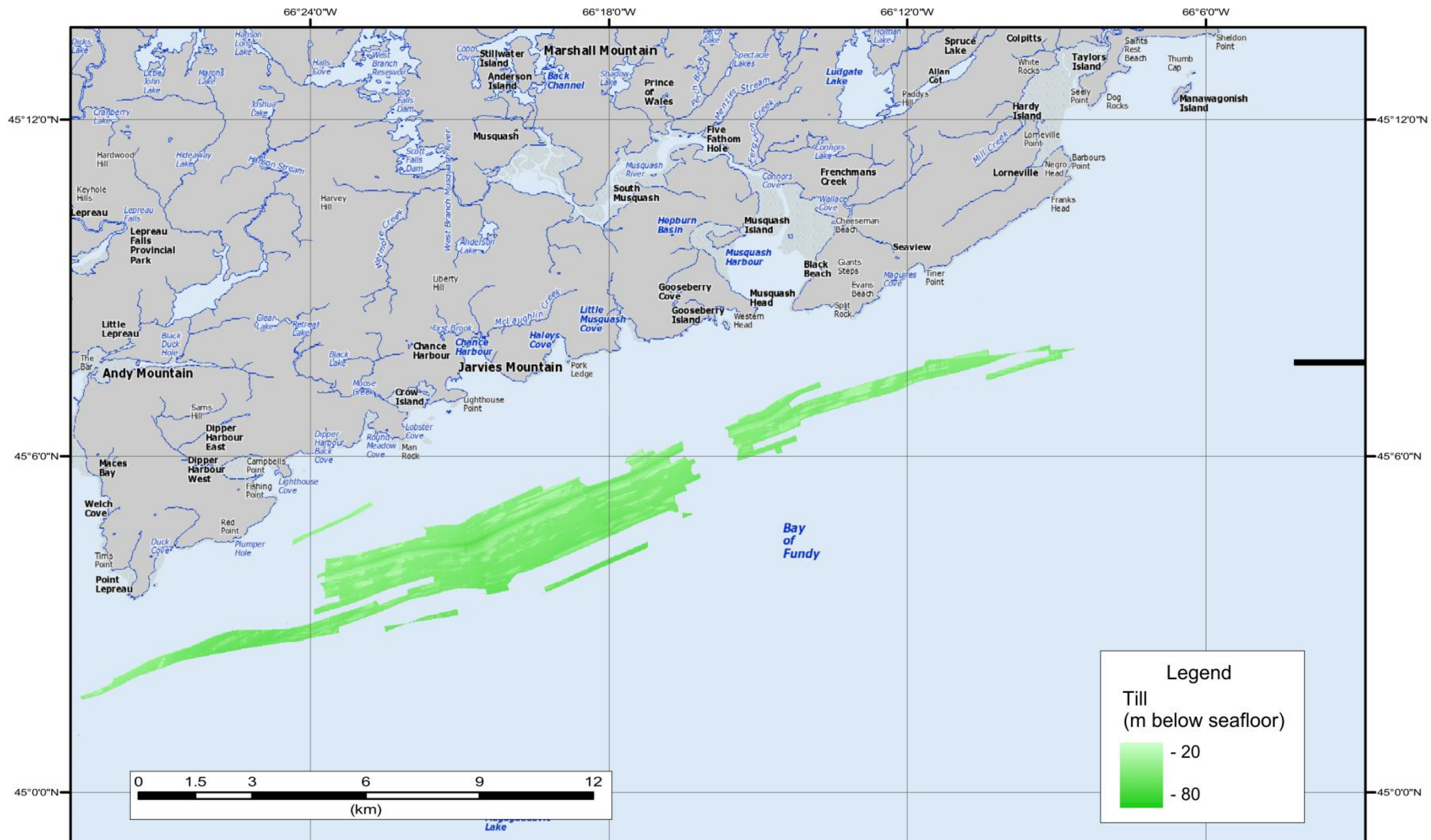

Figure 4.67: Location of the till unit mapped for the Creed 2008 survey area.

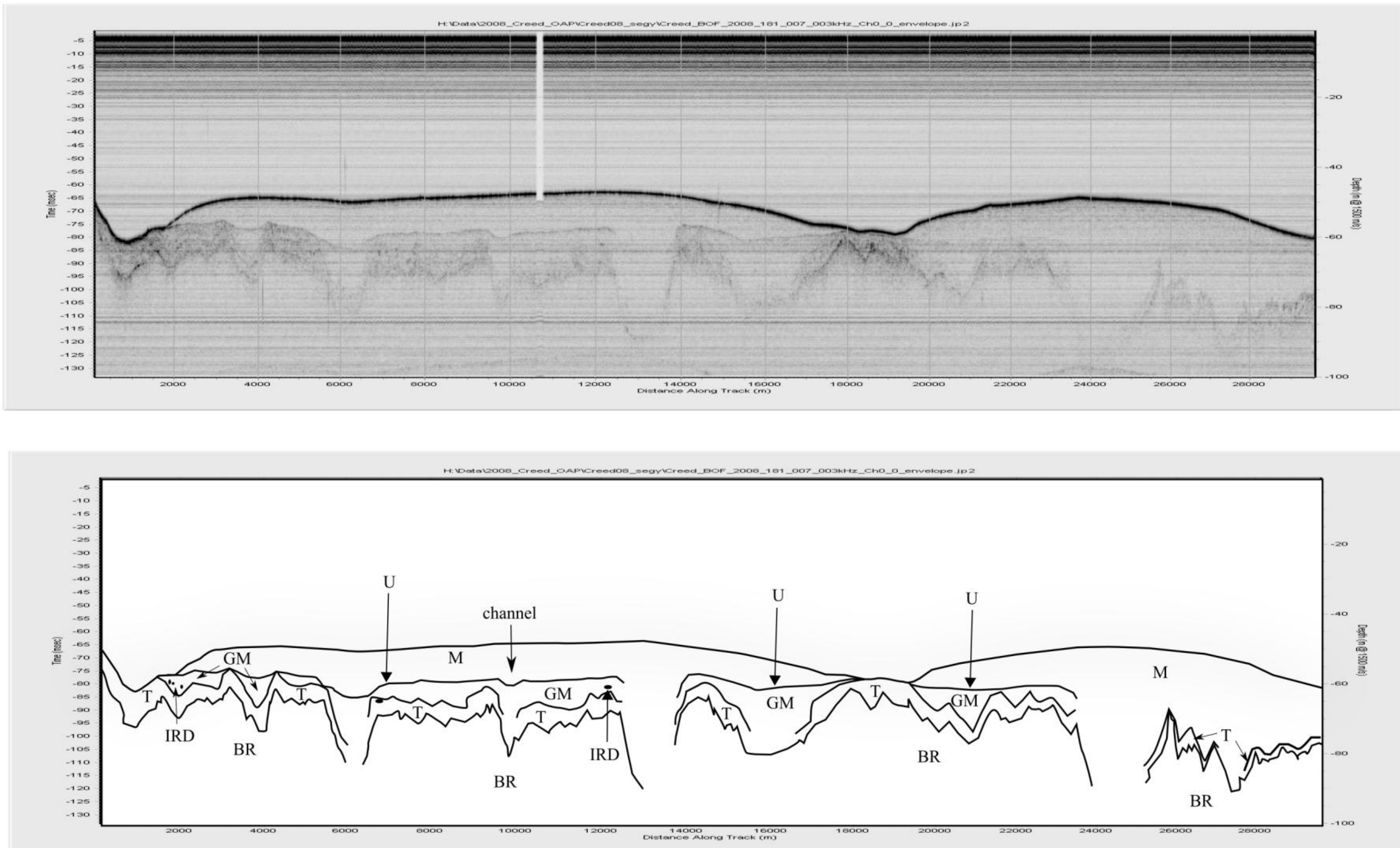

Figure 4.68: Seismic track line JD181_007 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, IRD = ice rafted debris, U = Pleistocene/Holocene unconformity, and M = Holocene mud.

The map in Figure 4.69 shows the extent of the glacimarine unit. In the till and glacimarine units a gap is observed measuring 1500 m wide; this gap is observed in the same area for both units. The upper bounding surface displays a moderate return. Unit 4 is acoustically transparent; it is a domed shape unit that is always present and found in all lines (Figure 4.66) and is interpreted as Holocene mud (M). In the bathymetric map (Figure 4.49) a lobe shape is observed. An unconformity (U) is recognized at depths between -55 m and – 65 m below the sea level. Depressions in U suggest the presence of fluvial deposits (Figure 4.70).

4.2.15 Creed 2007 Survey

The Creed 2007 survey area extends further offshore from the Creed 2008 data set from Campobello Island and Grand Manan eastward to offshore Saint John near the Blacks 2008 data (Figure 4.71). There are a total of 41 sub-bottom lines covering over 810 km^2 and totaling over 3360 km (Appendix 1). Many of the Creed 2007 sub-bottom lines include ship turns, but these sections of the sub-bottom lines were not included as they are distorted and unusable. The survey was conducted on board the CCGS Frederick G. Creed during the 2007 cruise. The majority of the lines run parallel to the shoreline in a southwest-northeast direction. The data also include 15 m resolution bathymetric and backscatter data collected in 2007.

Unfortunately the data, as discussed earlier, is of very poor quality (Figure 4.1) and will be used only for the further expansion of the discussion of location of natural gas within the Bay of Fundy.

The bathymetry of the Creed 2007 is a continuation from the Pennfield, Maces, Chance and Blacks surveys. At the eastern end of the Creed 2007 survey area the

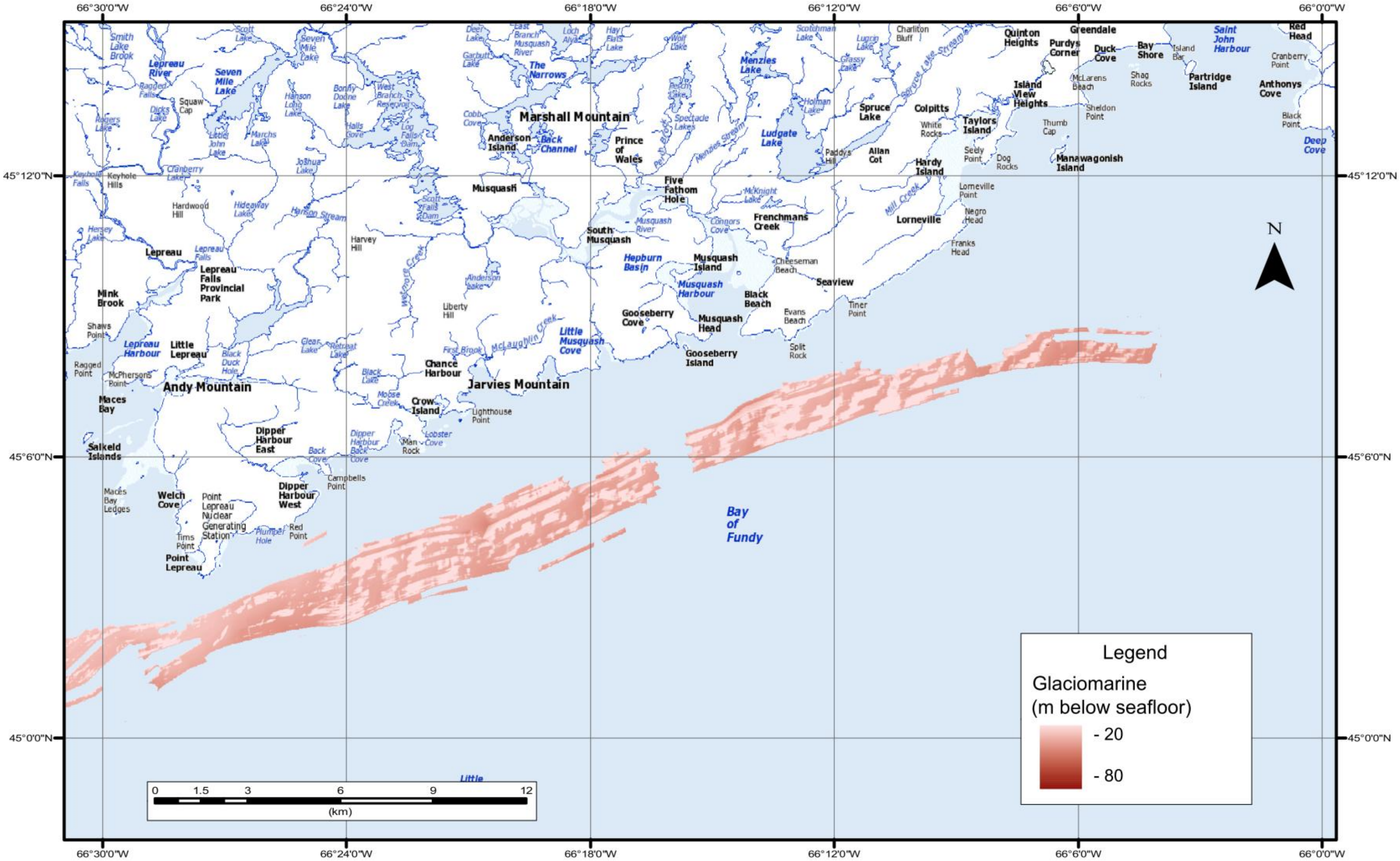

Figure 4.69: Location of the glaciomarine unit mapped for the Creed 2008 survey area.

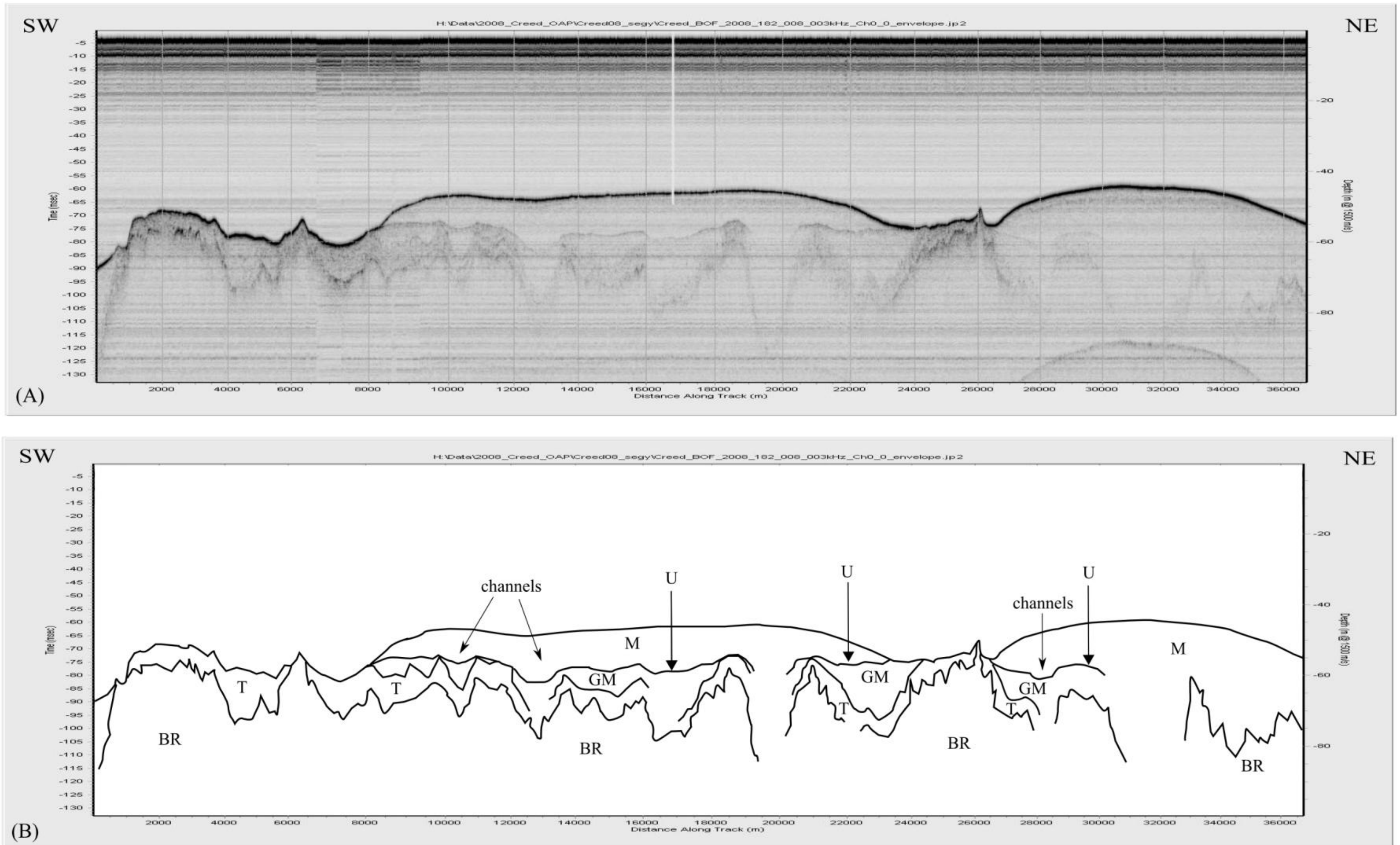

Figure 4.70: Seismic track line JD182_008 for the Creed 2008 survey area. (A) Original record with (B) the interpreted line drawing. BR = bedrock, T = till, GM = glacimarine sediment, U = Pleistocene/Holocene unconformity and M = Holocene mud.

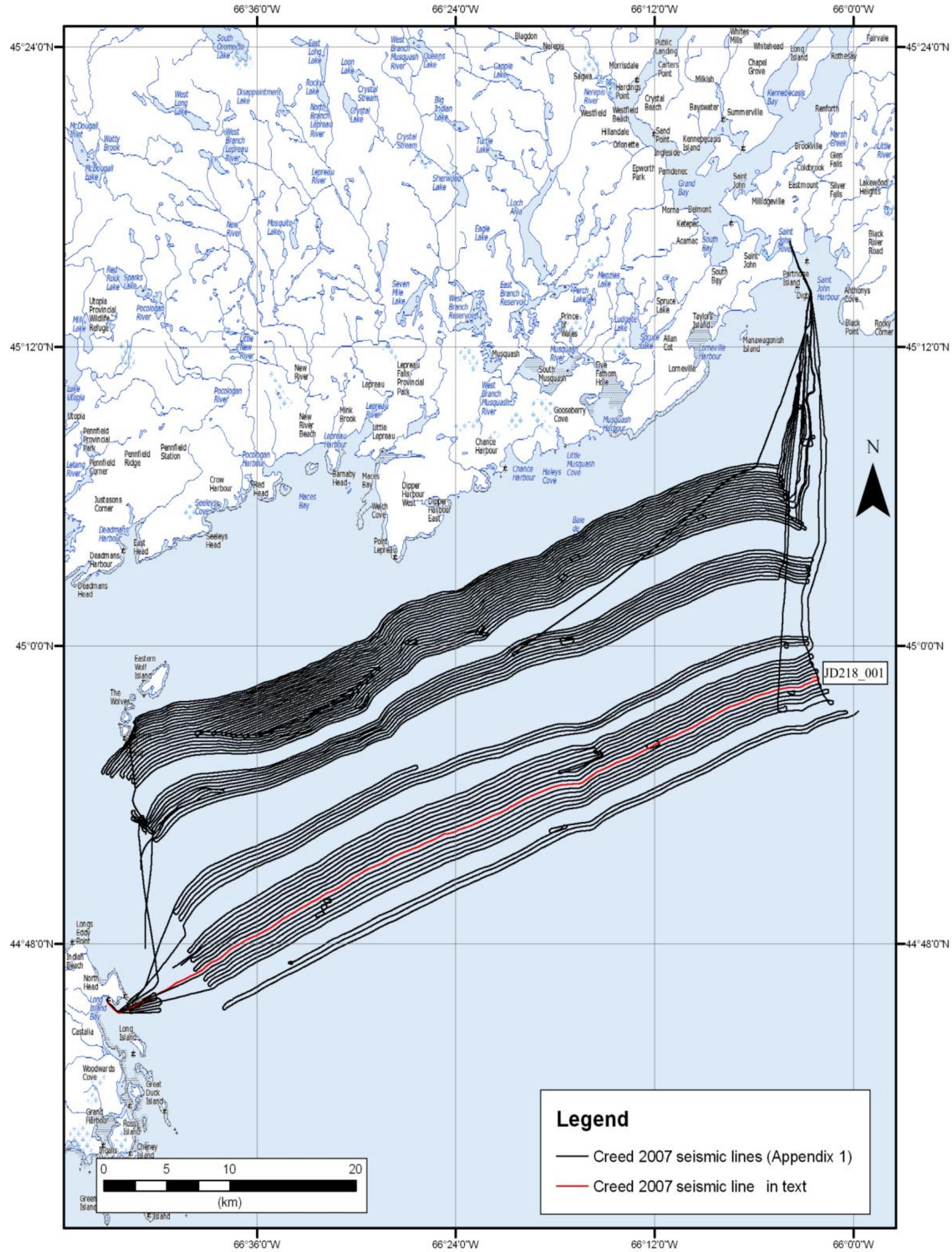

Figure 4.71: Location of Creed 2007 survey area.

seafloor is heavily scarred with iceberg scours in over 100 m deep water. Figure 4.72 shows an area of a iceberg scoured seafloor in the Bay of Fundy. A Creed 2007 survey line (JD218_001) that intersects with a 4.5 km long scour, and the corresponding sub-bottom line and location of the iceberg scour, which measures over 3 metres in depth is shown in Figure 4.73.

4.2.16 Creed 2007 Survey Stratigraphy

Although separate units can be distinguished in some lines, the majority of the sub-bottom lines from the Creed 2007 data are indistinguishable. However, due to the unique acoustic masking effect of natural gas, it is easily distinguished from the seismic data. Figure 4.74 is a map of the natural gas mapped in the study area, including the Creed 2007.

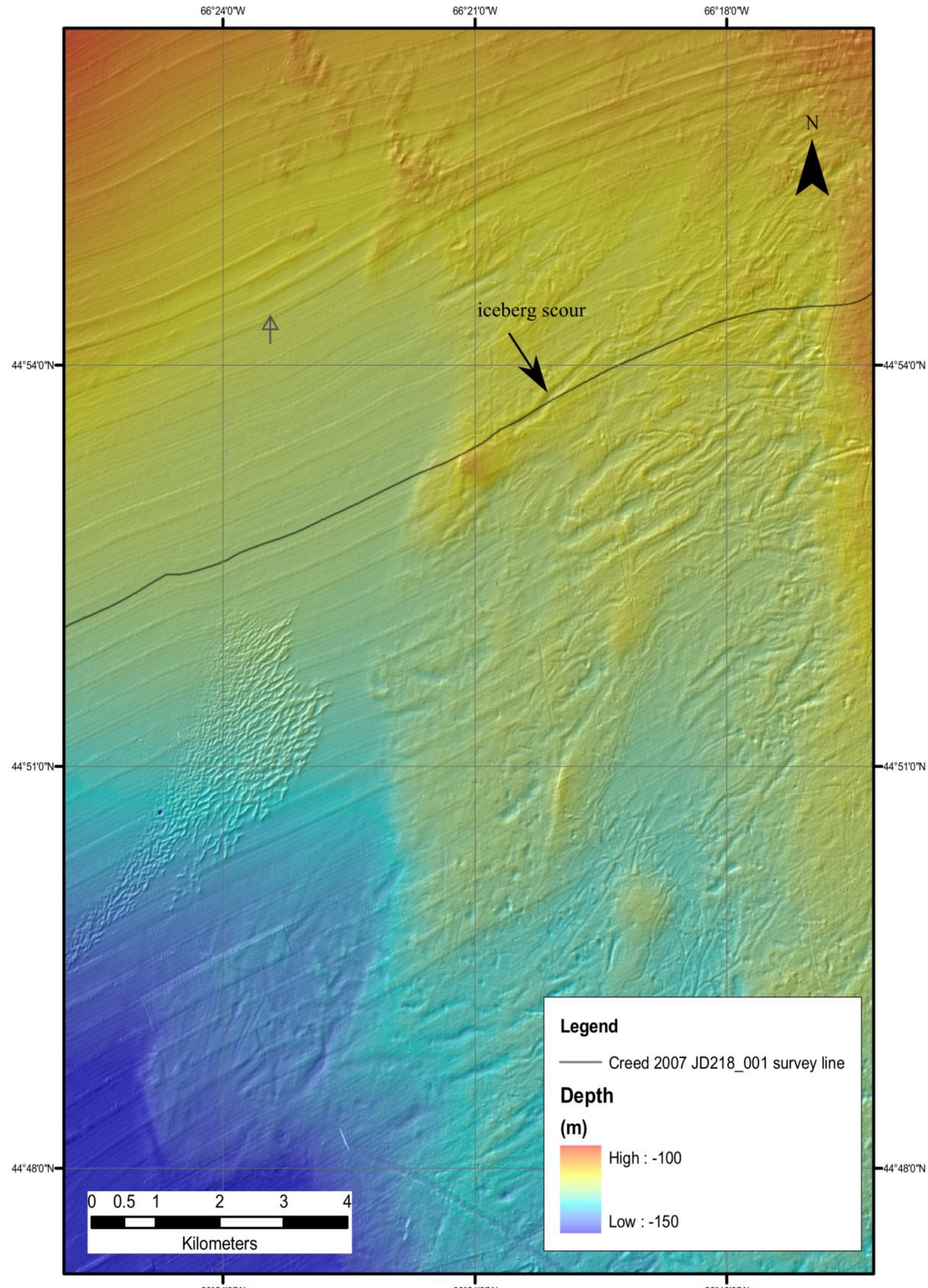

Figure 4.72: Location of Creed 2007 survey line JD218_001 and iceberg scour. Grey arrow points to artefacts due to data collection which are not landform features.

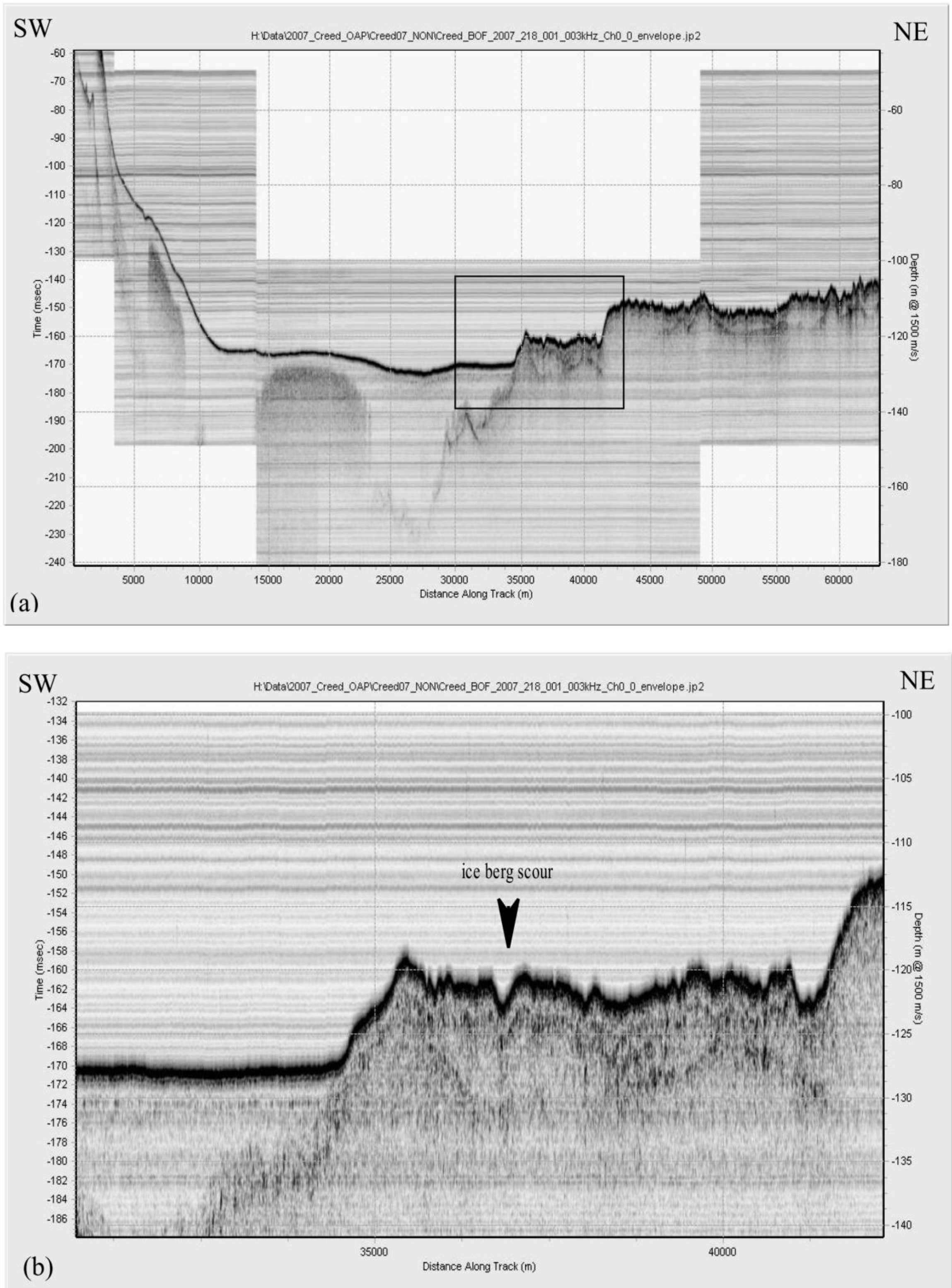

Figure 4.73: Seismic profile (a) with inset showing iceberg scour at 120 m depth.

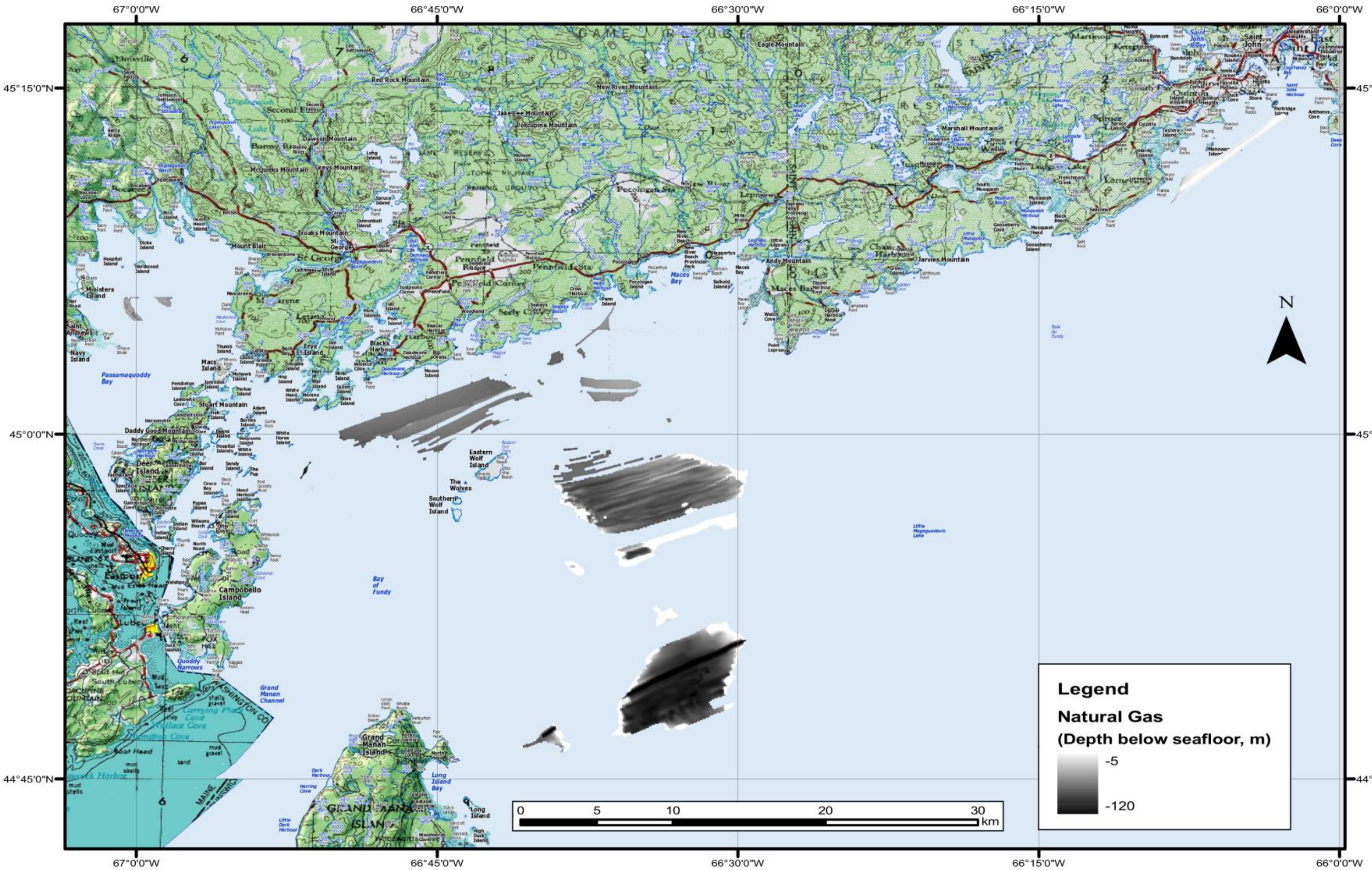

Figure 4.74: Extent of natural gas in the survey area.

Chapter 5

DISCUSSION

This investigation of the seismic stratigraphic nearshore framework of the Bay of Fundy between the St. Croix River and Saint John relies primarily on seismic reflection profiles. Chapter 4 documented the stratigraphy found within each of the six primary areas of the near shore Bay of Fundy between Grand Manan and Saint John. An examination of seismic data for the study area has revealed a stratigraphic sequence of bedrock, Pleistocene glacial sediments (till, glacimarine and glacimarine mud), followed by Holocene sediments (sand and gravel, and mud). Separating the glacial Pleistocene sediments from the modern Holocene sediments is an unconformity (U), resulting from subaerial and/or fluvial erosion during sea level lowstand. In this chapter the seismic stratigraphy is interpreted and integrated into a regional synthesis. This section will also discuss several features observed on seismic profile data in each area that are unique compared to the rest of the study area in New Brunswick.

The survey area contains 427 seismic lines totaling over 6,500 km of track lines from the research vessels CSL Heron and CCGS Frederick Creed over the 2007 to 2009 survey seasons. The area was divided into six primary locations: Campobello, Passamaquoddy Bay, Pennfield, Maces Bay, Chance and Blacks. Lines were examined using SegyJp2Viewer Geological Survey of Canada software and then compiled in ESRI ArcMap along with navigation lines for digital interpretation and mapping of related features. The units identified include the acoustic basement, bedrock, overlain occasionally by till and overlying the till and bedrock is the glacimarine deposit. In some areas, the glacimarine can be separated into two subunits. There may have been more till in the study area and more than one subunit in the glacimarine unit, however the resolution is controlled by the seismic system and the 3.5 kHz has a low resolution.

The Campobello Island survey covers a small area but shows very interesting features. The surveys include Friar Bay located between Campobello Island and Maine, and the Creed 2008 survey lines, 4 km to the east of the island. The Pineo Ridge Moraine complex in Maine is located less than 4 km from Friars Bay; an extension of the moraine on Campobello Island is less than 2 km away from the survey area and mapped as a kame moraine in New Brunswick (Kaplan 1999, Figure 4.3). The sub-bottom images show stratified internal configuration of glacimarine sediments infilling a valley in both the Friars Bay survey and the Creed 2008 survey on the southeastern side of Campobello Island (Figures 4.6 and 4.52).

The term stratified drift indicates deposition and sorting by water. Shipp (1989) further divided stratified drift in lowland Maine into three categories: 1) ice-contact stratified drift commonly associated with moraines; 2) subaerial outwash usually associated with highstand deltas; and 3) subaqueous outwash associated with moraines and deltas. The proximity to the Pineo Ridge Moraine complex suggests a major outflow channel was located at Campobello, at the time of a major stillstand of the deglaciation during the formation of the moraine. Kaplan (2007) constrained the timing of the moraine formation to between 13.4 and 12.8 ka yr B.P. At this time, as the moraine was growing, debris laden meltwater was discharging outwash deposits of stratified sediment consisting of gravel, sand and silt, and infilling the paleochannel from Friars Bay to the eastern shores of Campobello Island. Pockmarks located in the Friars Bay survey are of an unknown origin, but could be the result of gas or fluid expulsion following the rapid deposition of the stratified glacimarine sediments. There is an absence of any significant amounts of modern Holocene mud in the Friars Bay survey. This area is subjected to high energy tidal currents. Figure 4.4 shows erosion in the tightly confined channels to the northwest of the Friars Bay survey area, along the Canada-US border. This is interpreted to be the reason for the absence of modern Holocene mud within the survey area.

The Passamaquoddy surveys, 2008 and 2009, are located 16 km south of the Pineo Ridge Moraine and 15 km to the southeast of the St. George moraine. The 2008 survey has a prominent rock drumlin in the centre of the survey oriented in a southeasterly direction. The drumlin is covered in a thin veil of till (Figure 4.11); the till likely covers the top of the drumlin but due to the limitations of the 3.5 kHz echosounder it may be difficult to resolve from the bedrock. In Figure 4.13 another drumlin is identified; it is buried and not visible in the bathymetry maps. The Passamaquoddy Bay drumlins formed parallel to glacial flow, showing the active glacier flowed in a northwest to southeast direction. Located on the western flank of the rock drumlin is a concave depression. This depression is an erosional surface in the glacimarine unit measuring 400 m across and 2.6 m deep (Figures 4.13 and 4.14). It is found of depths at - 40 m and at the base of the channel - 47 m below relative sea level (Figure 4.11), and is interpreted as an unconformity (U) formed by fluvial erosion. This depression resulted from subaerial fluvial erosion during lowering of sea level, created by cut and fill deposition. Following this erosional event the area was transgressed by rising sea levels and covered over with modern marine muds.

The Passamaquoddy 2009 survey does not show any fluvial erosion but has a streamlined flute, likely composed of till. The strong upper surface reflector masks seismic units that lie below (Figure 4.12). This flute measures 585 m in length and is possibly as long as 2 km based on the linear nature of the surrounding pockmarks; it varies in width between 155 and 170 m, height between 55 and 60 m, and trends in a northwest-southeast direction (Figure 4.14). The flute represents a period of local readvance, depositing a new glacial unit overlying the glacimarine deposit. Benn and Evans (1998) describe flutes as streamlined parallel ridges or groove-formed subglacial deposits, parallel to ice flow direction, and predominately composed of lodgement till. A compact lodgement till could have kept the flute from being eroded during subsequent transgression. The flute may have originated during the Younger Dryas; when glaciers formed and were reactivated (Stea et al. 2011). Evidence for local glacier readvance in the area includes: the St. George moraine has an undeformed bottom layer of laminated marine clays, silts and fine sands overlain by glaciofluvial gravels, which have been sheared and folded by a glacial readvance toward the south-east (Seaman et al. 1993); the Sheldon Point moraine has marine fossils overlain by an overturned black, stratified, fine sand layer which is overlain by a thin diamicton, which in turn is overlain by gravelly, glaciofluvial foreset beds (Nicks 1988; Seaman 2006) and the Pineo Ridge Moraine complex possibly represents a readvance (Kaplan 1999; Kaplan 2007). This interpretation is supported by evidence for glacier readvance found at the Pineo Ridge Moraine complex, St. George moraine and Sheldon Point Moraine (Nicks 1988; Kaplan 2007; Seaman 2004).

The interpretation of the Pennfield survey shows till draping bedrock, and two types of glacimarine deposits. The lowest glacimarine unit contains draped stratified sediments with occasional ice rafted debris. The upper glacimarine unit contains uniform sediment with an unconformity (U) found at its upper surface representing fluvial erosion. Much of the survey is masked by natural gas, but a few lines do reveal more information.

The lowest glacimarine unit contains stratified sediments, which is very well illustrated in one north-south line (Fig 4.18), but otherwise only observed in the east west lines (Figure 4.19), as the rest of the seismic lines are masked by gas. This stratified unit is interpreted to be equivalent to Shipp's (1989) glacimarine-massive unit found in the Maine inner Shelf and Bacchus' (1993) proximal glacimarine unit (Table 2.3). The rhythmically bedded nature is inferred to represent alternating sand and silt content, resulting from strong seasonal fluctuations in the meltwater supply due to the close proximity to the ice margin (Bacchus 1993). In Figure 4.20 there is no stratified unit observed. This location lies furthest south in the survey and may have been too far away from the ice margin to be affected by the seasonal fluctuations in meltwater, thus receiving only finer sediments and ice rafted debris. The ice rafted debris, occasionally observed, was likely deposited by calving icebergs from the nearby ice margin, releasing debris-laden ice as it disintegrated.

The top glacimarine unit (GM-2) seen in Figures 4.19 and 4.20 is uniform and lacks any visible stratification. This unit is inferred to represent a period of deglaciation when the ice margin was further away and the effects of seasonal fluctuations were not as dominant. The glacimarine unit-2 (GM-2) is characterized by an erosional surface found at depths of – 52 m to – 89 m below sea level; the deepest values were measured at the bottom of the glaciofluvial channel (Figures 4.18 and 4.20). The unconformity (U) separating the Pleistocene glacimarine (GM) sediments from the upper Holocene mud (M) sediment results from fluvial erosion during lowstand sea level (Shipp 1989). The erosional event was of sufficient duration to permit erosion and formation of glaciofluvial incised channels. Following erosion, the area was transgressed by rising sea levels and covered over with modern marine muds.

The elongated pockmark located on the eastern side of The Wolves Islands, and between The Wolves Islands and the shoreline, is interpreted to be formed as a result of expulsion of natural gas, as seen in Figures 4.18 and 4.20 showing gas underlying a pockmark. The elongated pockmark is likely preserved and perhaps enhanced by the bottom tidal currents. The source of the gas is unknown.

Glacial and glacimarine deposits identified in the Maces Bay survey include till, two types of glacimarine sediments, sand and gravel. Two different glaciofluvial channels were eroded in sequence, recognized at depths – 35 m to – 40 m below sea level. The channels include stratified cut and fill deposits, and channels infilled by unstratified Holocene muds. Modern marine mud and lenses of sand and gravel cap the glacial sequence.

An unconformity (U) found between - 35 m and - 40 m below sea level represents a time of lower sea level of sufficient duration to permit erosion and formation of glaciofluvial incised channels. Following stillstand the current transgression began in the early Holocene and further eroded the Pleistocene deposits with current and wave action. However, these forces were not strong enough to completely erode the glaciofluvial sequence.

A prominent elongated sinuous ridge located in the centre of the bay is interpreted as an esker. It is recognized in the seismic data as a 4 metre high landform (Figure 4.26). Eskers are composed of coarser glaciofluvial sediments such as sand and gravels (Benn and Evans 1998). Due to the nature of these sediments any units below are masked within the seismic image. Eskers are commonly associated with outwash plains, deltas and moraines. Underlying the esker is a cardiod shaped feature, which is interpreted as a double lobed delta (Figure 4.24). Although there are no seismic units seen below this complex, Figures 4.26 and 4.35 show bedrock to the north and east of the esker delta. The esker-delta complex appears to be overlying a bedrock high. A possible explanation for not seeing bedrock within the seismic images below the esker-delta formation is the masking effect of gravel (Belknap and Shipp 1991). This esker delta complex would have also had the finer sediments of sand and silt winnowed away during transgression and by wave action.

The ridge located further south of the esker (Figure 4.24) is interpreted as a moraine, formed over a bedrock high. The bedrock high likely acted as a buttressing point for the glacier during deglaciation, causing the retreating ice to stall temporarily allowing for the thickening of the sand and gravel creating the moraine and discharging sand and gravel in the surrounding area (Figure 4.24). The Maces Bay moraine runs in a southwest to northeast direction (Figure 4.57). The bathymetry map, Figure 4.26, shows the moraine on the seafloor; in Figure 4.36 the moraine is seen continuing the southwest trend but buried under Holocene muds. Kaplan (2007) suggests that the Pineo Ridge

Moraine Complex (Figure 1.1) extends at least 100 km into New Brunswick on Campobello Island and likely continues into the province, possibly linking up with the Sheldon Point Moraine. The location and position of the Maces Bay moraine suggests that it may be a segment of these moraines. Figure 5.1 shows a map of all the moraines and their sub-parallel common direction.

Following the esker, delta and moraine formation, the most recent transgression submerged these features, and likely modified them during the current Holocene transgression, winnowing away finer sediments by wave and current action and forming an erosional pavement. Following this phase of deglaciation, the ice margin reached the present day shoreline, local sea levels dropped from isostatic rebound, and glaciofluvial waters eroded channels within the glacimarine deposits. This regression was followed by the second (Holocene) transgression submerging the former shoreline.

Chance and Blacks surveys mapped till draping bedrock, and two types of glacimarine deposits; the lowest unit contains draped stratified sediments with occasional ice-rafted debris and the upper units contain uniform glacimarine sediment with an unconformity (U) separating the Pleistocene sediments from the modern Holocene sediments. The lowest glacimarine unit containing stratified sediments indicates a proximal glacier supplying alternating sand and silt from seasonal fluctuations in meltwater and sediment supply. The occasional ice-rafted debris was deposited by calving icebergs near the ice margin. Overlying this unit is a more uniform glacimarine unit with no stratification observed, likely indicating sediments were deposited further from the stagnating glacier and less affected by seasonal discharge.

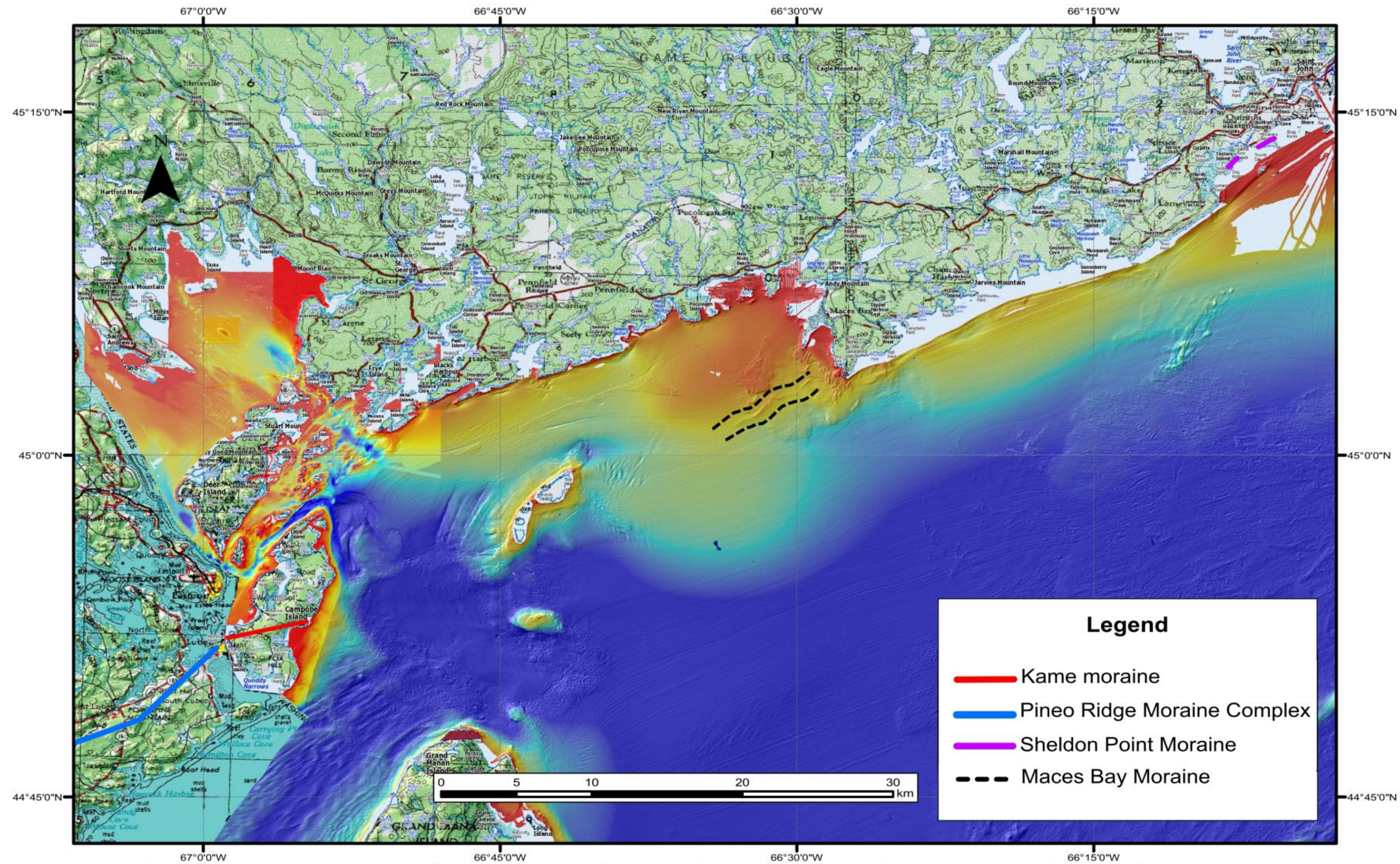

Figure 5.1. Location of Kame moraine segment on Campobello Island, Pineo Ridge moraine in Maine, Sheldon Point moraine near Saint John and the Maces Bay moraine.

An unconformity (U) underlying the Holocene mud (M) and found at depths ranging from – 60 to -70 m below the sea floor, is the result of subaerial erosion (Figures 4.40 and 4.70). This erosion likely resulted mainly from fluvial erosion during lowering of sea level. Following this erosional event the area was transgressed by rising sea levels and covered with modern marine muds.

Another erosional surface found in all the Creed 2008 lines is a 1500 m section where the till and glacimarine units are absent, found in between the 12000 m and 14000 m track line of Figure 4.68. Figures 4.67 and 4.69 shows the absence of the till and glacimarine unit, in the area located southwest of Musquash Harbour. However, the till unit may still be present overlying the bedrock, but it is not observed in the sub-bottom line due to the resolution problems. This feature could have resulted from a larger paleochannel during deglaciation, with high enough energy to erode the glacimarine, possibly down to the till or bedrock.

At the western end of the Blacks survey there is a rock drumlin trending in a southwesterly direction (Figure 4.44), indicating the former direction of active ice flow. Further northwest of the drumlin lies a thin gravel layer unit (TGL, Figure 4.46). The thin gravel layers in the Blacks survey are present at depths of 44-57 m below the sea floor; Figure 4.48 shows the extent of the TGL. Belknap and Shipp (1991) described thin gravel layers being produced by sediment gravity flows in low-lying channels, proximal to grounded ice. The proximity of the Sheldon Point Moraine to the TGL suggests that rapid deposition of sediments likely occurred during the moraine formation, leading to accumulation of unstable sediments and subsequent gravity flows creating a new TGL with each gravity flow.

The different seismic facies described from the sub-bottom seismic lines across the research area have been interpreted to show the changing environment in the Bay of Fundy during the Wisconsinan glaciation, deglaciation and the modern Holocene. The lowest lying Quaternary unit is till, found in much of the research area draping the bedrock. This till unit was deposited during glacial advance by grounded ice overriding the region during glaciation. The till unit is also found at the surface near Maces Bay, where a bedrock topographic high provided a buttressing point during deglaciation for the glacier to create a large lobe of till.

At the same time as the lobe of till was being deposited, the Maces Bay esker and delta were being formed. Overlying the till unit is a glacimarine unit, which in some locations can be divided into two sub-units. The lowest glacimarine sub-unit is stratified sediment likely composed of sand and gravel. The stratified glacimarine sediments were found in a variety of settings, but all indicating ice-proximal conditions with rising sea levels exerting seasonal fluctuations in the meltwater supply.

In Friars Bay, off Campobello Island, the stratified glacimarine sediments are ice-contact stratified drift associated with the Pineo Ridge Moraine complex, representing a major outwash channel (Figure 4.3). In the Pennfield and Chance survey areas, the stratified glacimarine sediments were produced at or near the ice front during a period of glacial retreat. The second glacimarine sediment unit observed in most of the survey areas is characterized as homogenous, showing no internal reflection, but with occasional ice rafted debris. This glacimarine unit represents deposition from a receding glacier and calving icebergs from the ice front. The icebergs calved from the glacier terminus, dropping their entrained load as they were carried in the marine environment by waves and currents. As well as dropping debris, icebergs plowed and reworked bottom sediments when their keels came in contact with the seafloor, as seen offshore of Grand Manan Island (Figure 4.73) where relict iceberg scours are found at water depths of 120m.

The entire glacimarine unit is capped by modern Holocene marine sediments. The contact between Holocene mud and the underlying Pleistocene glacimarine sediments is an erosional unconformity, providing evidence for low stand sea level following glaciation and glacimarine conditions immediately following deglaciation. The low stand sea level was the result of isostatic rebound of the recently ice free region, resulting in regression and exposure of the glacimarine unit to subaerial erosion. Evidence for fluvial erosion is recorded in the channels found in the Passamaquoddy Bay, Pennfield, Maces Bay and Chance surveys. Preservation of ice scours (Figures 4.38 and 4.72) shows areas of the Bay of Fundy that were below the lowstand sea level.

In the Maces and Chance surveys the Holocene deposits are lobate fan shaped in plan view (Figure 4.49). In side view, in the sub-bottom profiles, the Holocene mud appears as bulges (Figures 4.61, 4.40 and 4.41). The lobate fans overlie the unconformity, indicating they were deposited post glaciation in a marine environment. Because only small streams currently empty into the bay at this location the fans are interpreted to record a Holocene depocentre resulting from terrestrial melting of remnant ice masses, draining into the Bay of Fundy.

In Maces Bay stratified drift is found as glaciofluvial fill deposits in incised channels (Figure 4.32). The stratified infilled channel represents an ice proximal environment with seasonal fluxes of melt water, deposited at a time of lower sea level.

The other channels in the survey areas have an homogenous infill, showing no stratification (Figures 4.11, 4.20 and 4.40).

Passamaquoddy Bay shows evidence of a local glacial readvance with a flute composed of till trending in a northwest-southwest direction overriding the top glacimarine unit (Figures 4.12 and 4.14). The flute would have been exposed to subaerial erosion during sea level low stand and marine erosion during the Holocene transgression. However, both erosive forces were likely of a short enough duration and/or the flute's composition of till was protected it from erosion. Following lowstand sea level, the next transgression occurred and the area became inundated with rising sea levels exposing the area to erosion by tides and currents.

The gas in the survey area (Figure 4.74) could be from two sources, but its exact origin remains unknown. One source of natural gas is biogenic, caused by the anaerobic microbial breakdown of organic matter (Wildish et al. 2008; Fader 1991). Following the glacier's retreat, the period of lowered sea level and isostatic rebound resulted in the former seafloor being an exposed shelf (Figure 5.2, Rogers et al. 2006). During emergence the newly exposed coastal environment gave rise to lakes and wetlands (Figure 5.2). A rising sea level since then has submerged this area subjecting it to erosive processes actively removing sediments and depositing modern Holocene muds in the area, burying the organic matter with sediments (Figure 5.2, Kelley et al. 1998). The organic-rich Holocene muds may be a source of the natural gas mapped in the survey area. The second possible source for natural gas is thermogenic methane, which is developed at great depths by the cracking of kerogens at high temperatures and pressures in sedimentary rocks (Fader 1991).

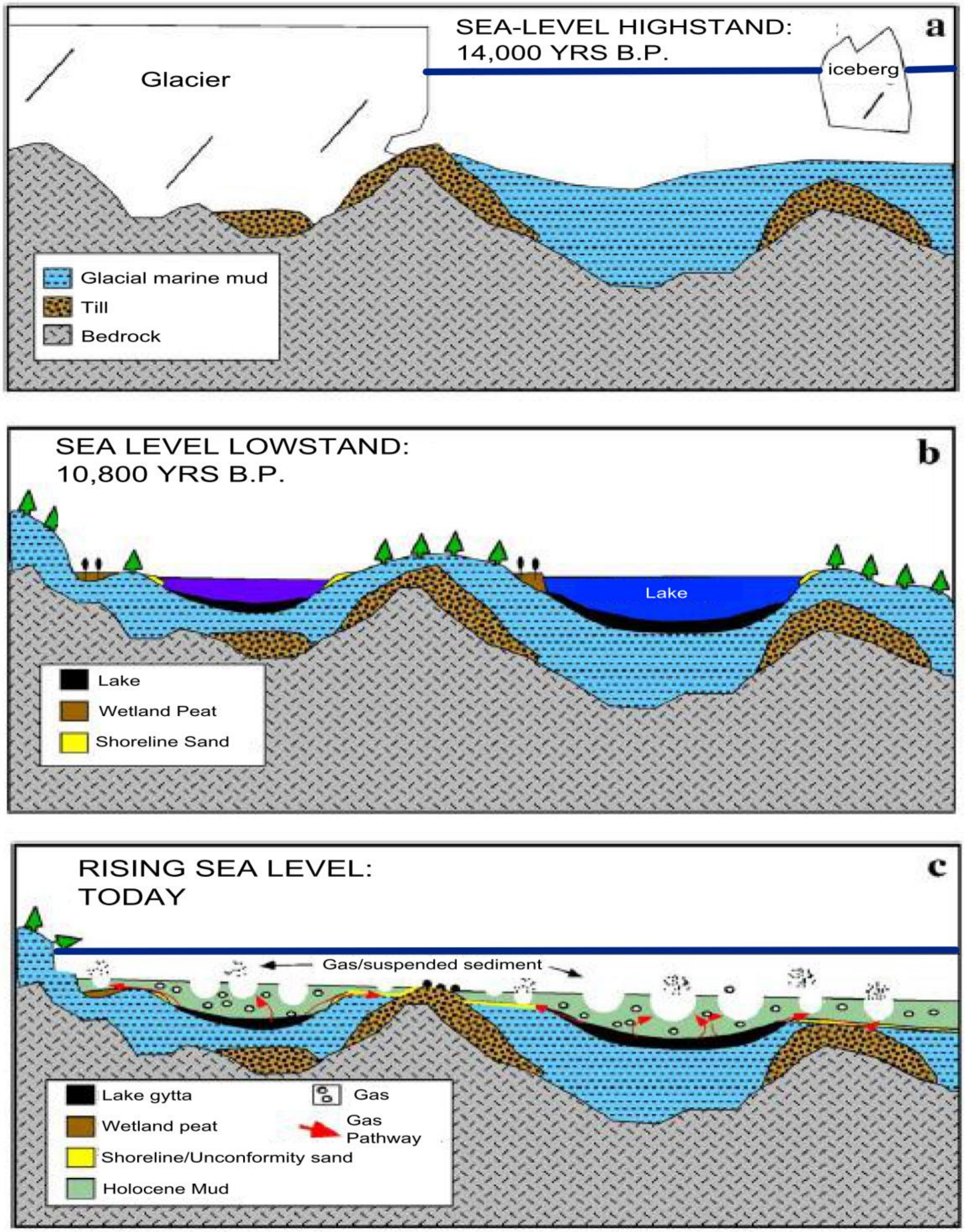

Figure 5.2. Cartoon diagram of natural gas and pockmark formation in coastal New Brunswick. (a) Glacier reaching to the present day shoreline with the first sea level transgression and deposition of till and glaciomarine mud. (b) Regression with lowstand sea level with lakes and wetlands formed. (c) Second transgression representing contemporary sea level with evolution of natural gas from buried organic matter creating pockmarks (after Rogers et al. 2006).

Chapter 6

CONCLUSIONS

This investigation of the Quaternary and stratigraphic framework of the nearshore Bay of Fundy between the St. Croix River and Saint John, New Brunswick relies primarily on seismic sub-bottom images, with bathymetry and backscatter providing supporting data. The nearshore stratigraphy from the shoreline to approximately the 100 m isobath contains 6 seismic units which include bedrock often overlain by till, and two different glacimarine units, separated by an unconformity from a Holocene marine mud, with intermittent gas units. The glacial landforms within the study area include drumlins, an esker, a delta, a moraine, thin gravel layers, glaciofluvial and outwash channels, and a ridge formed by localized readvance.

The following sequence of events is proposed to explain the Quaternary and stratigraphic framework of the nearshore Bay of Fundy. During the Wisconsinan glaciation the active ice sheet deposited till over the study area and formed drumlins under subglacial conditions. The drumlin within Passamaquoddy Bay documents glacial ice flow from northwest to southeast, whereas the drumlin offshore Musquash Harbour has a north to south direction.

At the time of deglaciation, marine waters were in contact with the receding ice terminus. Deposition occurred at the glacier front and from actively calving icebergs depositing ice rafted debris. The seafloor was reworked by floating iceberg keels. Evidence for a proximal glacier during the early part of deglaciation is the presence of stratified glacimarine sediments overriding till in the Pennfield and Chance area from the seasonal deposition of sands and gravels from a disintegrating ice margin.

Deglaciation continued, with the glacier retreating to Maces Bay. At this point, a bedrock high provided a buttressing point for the glacier, forming a moraine that deposited till into the surrounding area. As deglaciation progressed in Maces Bay an esker and delta formed in the bay. The landforms in Maces Bay represent a standstill in deglaciation. Maces Bay moraine may be a segment of a larger moraine belt that includes Maine's Pineo Ridge Moraine Complex and the Sheldon Point moraine.

The glacier changed from marine based to land based once the receding glacier reached the shoreline. Sediments reaching the marine environment changed from ice proximal, with stratified sediments, to ice distal, homogenous sediments.

Sea level was changing in conjunction with deglaciation. Transgression occurred at the same time as the glacier receded. When the glacier reached the shoreline, isostatic rebound of the previously depressed land started, exposing the glacimarine sediments to a period of subaerial erosion by river and streams. After this period of lowstand, the second transgression occurred inundating the area. The submerged recently eroded surface was exposed to further erosional forces from waves, currents and tides. However, these forces were not strong enough to completely erode the Pleistocene sediment record of glaciation and deglaciation.

The land-based glacier formed significant morainal systems in the area, examples being the Sheldon Point Moraine near Saint John, New Brunswick and the Pineo Ridge Moraine Complex in Maine near Campobello Island, New Brunswick. The proximity of the Blacks survey area to the Sheldon Point Moraine produced a series of thin gravel layers from sediment gravity flows now seen offshore. The Pineo Ridge Moraine

Complex led to the production of a major out flow channel extending from Friars Bay on the western side of Campobello Island to the eastern side of the island.

A glacial re-advance overrode the glacimarine sediments in Passamaquoddy Bay. The local glacial ice flow was northwest to southeast, depositing a ridge of sands and gravels. This may have occurred during the Younger Dryas Chronozone. This interpretation is supported by evidence for local glacier readvance found at the Pineo Ridge Moraine complex, St. George Moraine and Sheldon Point Moraine.

The six seismic units form two depositional sequences bounded by a Pleistocene/Holocene unconformity. The lower sequence of Pleistocene sediments consists of glacial and glacimarine units formed by an active glacier and a disintegrating glacier. The unconformity was formed by isostatic rebound, exposing the glacimarine sediments to a period of by aerial and fluvial erosion. The unconformity is found at depths from -40 to -89 m and represents subaerial erosion from low stand sea level during isostatic rebound.

A second transgression occurred depositing the upper sequence of Holocene modern marine muds. As sea level rose the glacimarine sediments were exposed to further erosional forces from waves, currents and the increasing influence of tides.

References

Bacchus, T.S. 1993. Late Quaternary stratigraphy and evolution of the eastern Gulf of Maine. Ph.D. thesis, Marine Studies, Oceanography, and Quaternary Studies, University of Maine, Orono, Maine, USA.

Barnhardt, W.A., Belknap, D.F. and Kelley, J.T. 1997. Stratigraphic evolution of the inner continental shelf in response to late Quaternary relative sea-level change, northwestern Gulf of Maine. *Geological Society of America Bulletin*, **109**, 612-630.

Belknap, D.F., and Kraft, J.C. 1981. Preservation potential of transgressive coastal lithosomes on the U.S. Atlantic Shelf. *Marine Geology*, **42**, 429-442.

Belknap, D.F., and Shipp, C. 1991. Seismic stratigraphy of glacial marine units, Maine inner shelf. *In* Glacial marine sedimentation: Paleoclimatic significance. *Edited by* J.B. Anderson and G.M. Ashley. Geological Society of America Special Paper 261, 137-159.

Belknap, D.F., Andersen, B.G., Anderson, R.S., Anderson, W.A., Borns, Jr. H.W., Jacobson, G.L., Kelley, J.T., Shipp, C.R., Smith, D.C., Stuckenrath, Jr. R, and Tyler, D.A. 1987. Late Quaternary sea level changes in Maine. *In* Sea Level Fluctuations and Coastal Evolution. *Edited by* D. Pilkey, J.D. Howard. Society of Economic Paleontologists and Minerologists. Special publication 41, 71-85.

Belknap, D.F., Kelley, J.T., and Gontz, A.M. 2002. Evolution of the glaciated shelf and coastline of the northern Gulf of Maine, USA. *Journal of Coastal Research*, **36**, 37-55.

Belknap, D.F., Gontz, A.M., and Kelley, J.T. 2005. Paleodeltas and preservation potential on a paraglacial coast: evolution of eastern Penobscot Bay, Maine. *In* High resolution morphodynamics and sedimentary evolution of estuaries. *Edited by* D.M. FitzGerald and J. Knight. Kluwer Academic Publishing, 335-360.

Benn, D.I., and Evans, D.J.A. 1998. Glaciers & Glaciation. Oxford University Press Inc., New York, NY.

Bloom, A.L. 1963. Late-Pleistocene fluctuations of sea-level and postglacial crustal rebound in coastal Maine. *American Journal of Science,* **261**, 862-879.

Boe, R, Rise, L., and Ottesen, D. 1988. Elongate depressions on the southern slope of the Norwegian Trench (Skagerrak): morphology and evolution. *Marine Geology*, **146**, 191-203.

Borns, Jr. H.W., Doner, L.A., Dorion, C.C., Jacobson, Jr. G.L., Kaplan, M.R., Kreutz, K.J., Lowell, T.V., Thompson, W.B., and Weddle, T.K. 2004. The deglaciation of Maine, USA. *In* Quaternary glaciations – extent and Chronology, Part II. *Edited by* J. Ehlers and P.L. Gibbard. Elsevier Science Publishers B.V., 89-109.

Bostock, H.S. 1970. Physiographic regions of Canada. Geological Survey of Canada, Map 1254A.

Broster, B.E., Allaby, G.M. and Pronk, A.G. 2004. Lithology and geochemical dispersal in till: Petitcodiac area, New Brunswick. *Atlantic Geology*, **40**, 169-188.

Burke, K.B.S. 2004. Historical seismicity in the Central Highlands, Passamaquoddy Bay, and Moncton regions of New Brunswick, Canada, 1817-1961. *Seismological Research Letters*, **75**, 419-431.

Caron, V., Nelson, C.S., and Kamp, P.J.J. 2004. Transgressive surfaces of erosion as sequence boundary markers in cool-water shelf carbonates. *Sedimentary Geology*, **164**, 179-189.

Cooper, L. Gulf of Maine Symposium. 2009. June 10 2010. [Available from http://www.rargom.org/Symposium2009/images/maine_gulf_map.jpg 15June2011].

Cumming, L.M. 1968. Geology of Passamaquoddy Bay region, Charlotte County, New Brunswick. Geological Survey of Canada Paper 64-26.

Dashtgard, S.E., White, R.O., Butler, K.E., and Gingras, M.K. 2007. Effects of relative sea level change on the depositional character of an embayed beach, Bay of Fundy, Canada. *Marine Geology*, **239**, 143-161.

Davies, T.A., Austin Jr., J.A., Lagoe, M.B., and Milliman, J.D. 1992. Late Quaternary sedimentation off New Jersey: New results using 3-D seismic profiles and cores. *Marine Geology*, **108**, 323-343.

Department of Fisheries and Oceans. 2010. WebTide Tidal Prediction Model [Available from http://www.pac.dfo-mpo.gc.ca/science/oceans/data-donnees/webtide/index-eng.htm 10June2010]

Desplanque, C., and Mossman, D.J. 2001. Bay of Fundy Tides. *Geoscience Canada*, **28**, 1-11.

Desplanque, C., and Mossman, D.J. 2004. Tides and their seminal impact on the geology, geography, history, and socio-economics of the Bay of Fundy, eastern Canada. *Atlantic Geology*, **40**, 1-130.

Dickinson, P.J. 2008. Geomorphological processes and the development of the lower Saint John River human landscape. Ph.D. thesis, Department of Geology, University of New Brunswick, Fredericton, New Brunswick.

Fader, G.B.J. 1991. Gas-related sedimentary features from the Eastern Canadian continental shelf. *Continental Shelf Research*, **11**, 1123-1153.

Fader, G.B.J. 1997. Effects of shallow gas on seismic reflection profiles. *In* Glaciated continental margins: An atlas of acoustic images. *Edited by* T.A. Davies, T. Bell, A.K. Cooper, H. Josenhans, L. Polyak, A. Solheim, M.S. Stoker, and J.A. Stravers. Chapman & Hall, New York, NY. 29-30.

Fader, G.B.J. 2005. Glacial, post glacial, present and projected sea levels, Bay of Fundy. Atlantic Marine Geological Consulting Ltd. Halifax, Nova Scotia.

Fader, G.B.J., King, L.H., and MacLean, B. 1977. Surficial geology of the eastern Gulf of Maine and Bay of Fundy. Marine Sciences Paper 19. Geological Survey of Canada Paper 76-17.

Flaherty, G.F. 1989. The Reversing Falls: The present outlet of the Saint John River in southern New Brunswick. Popular Geology paper 89-1, Minerals and Energy Division, Department of Natural Resources and Energy, Fredericton, New Brunswick, Canada.

Folger, D.W., O'Hara, C.J., and Robb, J.M. 1975. Maps showing bottom sediments on the continental shelf of the northeastern United States: Cape Ann, Massachusetts to Casco Bay, Maine. US Geologic Survey, Misc. Invest Series map I-839.

Gadd, N.R. 1973. Quaternary geology of southwestern New Brunswick with particular reference to Fredericton area. Geological survey of Canada, Paper 71.

Gates, O. 1984. The geology of the Passamaquoddy Bay area, Maine and New Brunswick. Maine Geological Survey, Department of Conservation, Open-File 84-10.

Gontz, A.M. 1999. Evolution of seabed pockmarks in Penobscot Bay, Maine. M.Sc. thesis, Department of Geological Sciences, University of Maine, Orono, Maine, USA.

Grant, D.R 1970. Recent coastal submergence of the Maritime Provinces, Canada. *Canadian Journal of Earth Sciences*, **7**, 676-689.

Hachey, H.B. and Bailey, W.B. 1952. The general hydrography of the waters of the Bay of Fundy. Manuscript Reports of Biological Stations, Fisheries Research Board of Canada, Report 455.

Hovland, M, Gardner, J.V., and Judd, A.G. 2002. The significance of pockmarks to understanding fluid flow processes and geohazards. *Geofluids*, **2**, 127-136.

Josenhans, H.W., King, L.H., and Fader, G.B.J. 1978. A side-scan sonar mosaic of pockmarks on the Scotian Shelf. *Canadian Journal of Earth Sciences*, **15**, 831-840.

Kaplan, M.R. 1999. Retreat of a tidewater margin of the Laurentide ice sheet in eastern coastal Maine between ca. 14 000 and 13 000 ^{14}C yr B.P. *Geological Society of America Bulletin*, **111**, 620-632.

Kaplan, M.R. 2007. Major ice sheet response in eastern New England to a cold North Atlantic region, ca. 16-15 cal ka BP. *Quaternary Research*, **68**, 280-283.

Kelly, J.T., Barnhardt W.A., Belknap D.F., Dickson S.M., Kelley, A.R. 1989. The Seafloor Revealed. Maine Geological Survey, Department of Conservation. Open File 96-9.

Kelley, J.T., Dickson, S.M, Belknap, D.F., Barnhardt, W.A., and Henderson, M. 1994. Giant sea-bed pockmarks: Evidence for gas escape from Belfast Bay, Maine. *Geology*, **22**, 59-62.

Kelley, J.T., Dickson, S.M., and Belknap, S.M. 2005. Moosehead Lake and the Tale of Two Rivers. [Available from http://www.maine.gov/doc/nrimc/mgs/explore/surficial/sites/jun05.htm 15May2011].

Knebel, H.J. and Scanlon, K.M. 1985. Sedimentary framework of Penobscot Bay, Maine. *Marine Geology*, **65**, 305-324.

Kiewiet de Jonge, E.J.C. 1951. Glacial water levels in the St. John River Valley. Ph.D. thesis, Department of Geography, Clark University, Worcester, Massachusetts, USA.

King, L.H. 1972. Relation of plate tectonics to the geomorphic evolution of the Canadian Atlantic provinces. *Geological Society of America Bulletin*, **83**, 3083-3090.

King, L.H., and Fader, G.B.J. 1986. Wisconsinan glaciation of the Atlantic continental shelf of southeast Canada. Geological Survey of Canada, Bulletin 363.

King, L.H. and Maclean, B. 1970. Pockmarks on the Scotian Shelf. *Geological Society of America Bulletin*, **84**, 3141-3148.

Leavitt, H.W., and Perkins, E.H. 1935. Glacial geology of Maine. Maine Technology Experiment Station Bulletin 30, **2**, 107-108.

Lee, H.A. 1957. Surficial geology of Fredericton, York and Sunbury Counties, New Brunswick. Geological Survey of Canada, Paper 56-2.

Legget, R.F. 1979. Glacial geology of Grand Manan Island, New Brunswick. *Canadian Journal of Earth Sciences*, **17**, 440-452.

Lurton, X. 2002. An introduction to underwater acoustics: principles and applications. Praxis Publishing Ltd, New York, NY.

Maine Geological Survey. 2007. Maine's Coastal Marine Geology. [Available from http://www.maine.gov/doc/nrimc/mgs/about/index.htm May 2010]

Mayle, F.E., Levesque, A.J., and Cwynar, L.C. 1993. Accelerator-mass-spectrometer ages for the Younger Dryas event in Atlantic Canada. *Quaternary Research*, **39**, 355-360.

Mayle, F.E., and Cwynar, L.C. 1995. A review of multi-proxy data for the younger dryas in Atlantic Canada. *Quaternary Science Reviews*, **14**, 813-821.

Mcleod, M.J. 1979. The geology of Campobello Island, southwestern New Brunswick. M.Sc. thesis, Department of Geology, University of Fredericton, Fredericton, New Brunswick, Canada.

McLeod, M.J., Johnson, S.C., and Ruitenberg, A.A. 1994. Geological map of southwestern New Brunswick. New Brunswick Department of Natural Resources and Energy, Mineral Resources Branch, Map NR-5.

Miller, B.V., Barr, S.M., and Black, R.S. 2007. Neoproterozoic and Cambrian U-Pb (zircon) ages from Grand Manan Island, New Brunswick: implications for stratigraphy and northern Appalachian terrane correlations. *Canadian Journal of Earth Sciences*, 44, 911-923.

Murphy, J.B., Sernandez-Suarez J., Keppie, J.D., and Jeffries, T.E. 2004. Contiguous rather than discrete Paleozoic histories for the Avalon and Meguma terranes based on detrital zircon data. *Geology*, **32**, 585-588.

NBDNR. 2006 Surficial Geology of New Brunswick. Natural Resources of New Brunswick [Available from http://www.gnb.ca/0078/minerals/Surficial_Mapping_Glacial_Geology-e.aspx March 2010]

Nicks, L. P. 1988. The study of the glacial stratigraphy and sedimentation of the Sheldon Point moraine, Saint John, New Brunswick. M.Sc. thesis, Department of Geology, Dalhousie University, Halifax, Nova Scotia, Canada.

NRCAN. 2009. Natural Resources Canada. Reference Maps. [Available from http://atlas.nrcan.gc.ca/site/english/maps/reference November 2009]

Ocean Mapping Group. 2003. Passamaquoddy Bay Pockmarks. [Available from http://www.omg.unb.ca/Projects/PassamaquoddyBay/PassamaquoddyBay Pockmarks.html September 21 2010].

Oliveira Jr., A.M.O., and Hughes Clarke, J.E. 2007. Recovering wide angular sector multibeam backscatter to facilitate seafloor classification. *In* United States Hydrographic Conference. Norfolk. May 2007

Ostericher Jr. C. 1965. Bottom and sub-bottom investigations of Penobscot Bay, Maine. U.S. Naval Oceanographic Office, Technical Report 173.

Parrott, R.D., Todd, B.J., Shaw, J., Hughes Clarke J.E., Griffin, J., MacGowan B., Lamplugh, M., and Webster, T. 2008. Integration of multibeam bathymetry and LiDAR surveys of the Bay of Fundy, Canada. *In* Proceedings of the Canadian Hydrographic Conference and National Surveyors Conference. Paper 6-2.

Pirazzoli, P.A. 1991. World atlas of Holocene sea-level change. Elsevier Oceanography Series. Elsevier, New York, NY.

Pecore, S.S., and Fader, G.B.J. 1990. Surficial geology, pockmarks, and associated neotectonic features of Passamaquoddy Bay, New Brunswick, Canada. Geological Survey of Canada, Open File 79-20.

Powell, R.D. 1981. A model for sedimentation by tidewater glaciers. *Annals of Glaciology*, **2**, 129-134.

Pratt, R.M. and Schlee, J. 1969. Glaciation on the continental margin off New England. *Geological Society of America Bulletin*, **80**, pp. 2335-2342.

Rampton, V.N. 1984. Surficial geology, New Brunswick. Geological Survey of Canada. Map 1594A (scale 1:500 000).

Rampton, V.N., Gauthier, R.C., Thibault, J. and Seaman, A.A. 1984. Quaternary geology of New Brunswick. Geological Survey of Canada, Memoir 416.

Rast, N., Burke, K.B.S., and Rast, D.E. 1979. The earthquakes of Atlantic Canada and their relationsship to structure. *Geoscience Canada*, **6**, 173-180.

Rogers, J.N., Kelley, J.T., Belknap, D.F., Gontz, A., and Barnhardt, W.A. 2006. Shallow-water pockmark formation in temperate estuaries: A consideration of origins in the western Gulf of Maine with special focus on Belfast Bay. *Marine Geology*, **225**, 45-62.

Schnitker, D. 1974. Postglacial emergence of the Gulf of Maine. *Geological Society of America Bulletin*, **85,** pp. 491-494.

Scott, D.B., and Greenberg, D.A. 1983. Relative sea-level rise and tidal development in the Fundy tidal system. *Canadian Journal of Earth Sciences*, **20**, 1554-1564.

Seaman, A.A. 2004. Late Pleistocene history of New Brunswick, Canada. *In* Quaternary glaciations – extent and Chronology, Part II. *Edited by* J. Ehlers and P.L. Gibbard. Elsevier Science Publishers B.V., pp 151-167.

Seaman, A.A. 2006. A new interpretation of the late glacial history of central New Brunswick: The Gaspereau ice centre as a Younger Dryas ice cap. *In* Geological investigations in New Brunswick for 2005. *Edited by* G.L. Martin. New Brunswick Department of Natural Resources; Minerals, Policy and Planning Division, Mineral Resources Report 2006-3, 1-36.

Seaman, A.A., Broster, B.E., Cwynar, L.C., Lamothe, M., Miller, R.F., and Thibault, J.J. 1993. Field guide to the Quaternary geology of southwestern New Brunswick. New Brunswick Department of Natural resources and Energy, Open File Report 93-1.

Semchur, C. 2011. Rivers of New Brunswick. [Available from http://fr.wikipedia.org/wiki/Fichier:Rivers_of_New_Brunswick_topographic_map-fr.svg. 18July2011].

Shaw, J., Piper, D.J.W., Fader, G.B.J., King, E.L., Todd, B.J., Bell, T., Batterson, M.J., and Liverman, D.G.E. 2006. A conceptual model of the deglaciation of Atlantic Canada. *Quaternary Science Reviews*, **25**, 2059-2081.

Shaw, R.R., Todd, B.J., Brushett, D., Parrott, D.R., and Bell, T. 2008. Late Wisconsian glacial landsystems on Atlantic Canadian shelves: New evidence from multibeam and single-beam sonar data. *Boreas*, **38**, 146-159.

Shipp, C.R. 1989. Late Quaternary sea-level fluctuations and geologic evolution of four embayments and adjacent inner shelf along the northwestern Gulf of Maine (Volume I and II). Ph.D. thesis, Marine Studies, Oceanography, and Quaternary Studies, University of Maine, Orono, Maine, USA.

Shipp, C.R., Belknap, D.F., and Kelley, J.T. 1991. Seismic stratigraphic and geomorphic evidence for a post-glacial sea-level lowstand in the Northern Gulf of Maine. *Journal of Coastal Research*, **5**, 341-364.

Shipp, C.R., Belknap, D.F., and Kelley, J.T. 1999. A submerged shoreline on the inner continental shelf of the western Gulf of Maine. *In* Studies in Maine Geology. *Edited by* R.D. Tucker and R.G. Marvinney, Maine Geological Survey, Augusta, Maine, **5**, 11-28.

Stea, R.R. 2004. The Appalachian glacial complex in Maritime Canada. *In* Quaternary glaciations – extent and Chronology, Part II. *Edited by* J. Ehlers and P.L. Gibbard. Elsevier Science Publishers B.V., pp 213-232.

Stea, R.R., Piper, D.J.W., Fader, G.B.J., and Boyd, R. 1998. Wisconsinan glacial and sea-level history of Maritime Canada and the adjacent continental shelf: A correlation of land and sea events. *Geological Society of America Bulletin*, **110**, pp. 821-845.

Stea, R.R., Fader, G.B.J., Scott, D.B., and Wu, P. 2001. Glaciation and relative sea-level change in Maritime Canada. *In* Deglacial History and Relative Sea-Level Changes, Northern New England and Adjacent Canada. *Edited by* T.K. Weddle and M.M. Retelle. Geological Society of America Special Paper 351, pp. 35-49.

Stea, R.R., Seaman, A.A., Pronk, T., Parkhill, M.A., Allard, S., and Utting, D. 2011. The Appalachian glacier complex in Maritime Canada. *In* Quaternary glaciations – extent and Chronology, A Closer Look. *Edited by* J. Ehlers and P.L. Gibbard. Elsevier Science Publishers. 631-659.

Stumpf, A.J., Broster, B.E., and Seaman, A.A. 1997. Lithological and geochemical dispersal in till: McAdam area, New Brunswick. *Atlantic Geology*, **33**, 31-42.

Stuiver, M. and Borns, Jr. H.W. 1975. Late Quaternary marine invasion of Maine: its chronology and associated crustal movement. *Geological Society of America Bulletin*, **86**, pp. 99-104.

Swift, D.J.P., and Lyall, A.K. 1968. Origin of the Bay of Fundy, and interpretation from sub-bottom profiles. *Marine Geology*, **6**, 331-343.

Tagg, A. R., and Uchupi, E. 1966. Distribution and geologic structure of Triassic rocks in the Bay of Fundy and the northeastern part of the Gulf of Maine. U.S. Geological Survey Professional paper 550-B.

Tanoli, S.K., and Pickerill, R.K. 1988. Lithostratigraphy of Cambrian-Lower Ordovician Saint John group, southern New Brunswick. *Canadian Journal of Earth Sciences*, **25**, 669-690.

Thibault, J., Seaman, A.A., and Mott, R.J. 1985. Geomorphic features associated with deglaciation and marine submergence in southwest New Brunswick. Geological Association of Canada and Mineralogical Association of Canada Annual Meeting, Excursion 7 Guidebook, Fredericton New Brunswick, Canada.

Trites, R.W., and Garrett C.J.R. 1983. Physical oceanography of the Quoddy region. *In* Marine and Coastal Systems of the Quoddy Region, New Brunswick. *Edited by* M.L.H. Thomas. Canadian Special Publication of Fisheries and Aquatic Sciences, pp. 64.

Trefethen, J.M. Leavitt, A.H., Miller, R.N., and Savage, C. 1947. Preliminary report on Maine clays. Report of the State Geologists, Maine Development Commission, Augusta, Maine, 10-46.

Uchupi, E. 1966. Structural framework of the Gulf of Maine. *Journal of Geophysical Research, **71**, 3013-3028.*

Uchupi, E., Bolmer, S.T. 2008. Geologic evolution of the Gulf of Maine region. *Earth-Science Reviews*, **91**, 27-76.

USGS. 2007. United States Geological Society. Sidescan-Sonar Imagery [Available from http://woodshole.er.usgs.gov/pubs/of2007-1150/html/sidescan.html May 2011]

Wade, J.A., Brown, D.E., Traverse, A., and Fensome, R.A. 1996. The Triassic-Jurassic Fundy Basin, eastern Canada: regional setting, stratigraphy and hydrocarbon potential. *Atlantic Geology*, **32**, 189-231.

Wightman, D., and Cooke, H.B.S. 1978. Postglacial emergence in Atlantic Canada. *Geoscience Canada*, **5**, 61-65.

Wildish, D.J., Akagi, H.M., McKeown, D.L., and Pohle, G.W. 2008. Pockmarks influence benthic communities in Passamaquoddy Bay, Bay of Fundy, Canada. *Marine Ecology Progress Series*, **357**, 51-66.

APPENDIX I

Seismic profiles used for the database

Line #	Start of line		End of line	
	Latitude	Longitude	Latitude	Longitude
Campobello				
140_140	44.89656833 N	66.96018361 W	44.88028806	66.96668056
140_135	44.88002639 N	66.96848972 W	44.89703194	66.96079083
140_134	44.90788889 N	66.95644889 W	44.88039194	66.96914861
140_133	44.87584722 N	66.97498750 W	44.90796556	66.95716417
140_131	44.90876861 N	66.95785639 W	44.87610972	66.97522750
140_130	44.87623944 N	66.97586917 W	44.90899861	66.95921222
140_124	44.90972778 N	66.96131500 W	44.87653861	66.97639333
140_123	44.87670472 N	66.97757889 W	44.91014806	66.96295972
140_121	44.91142111 N	66.96481083 W	44.87693722	66.97814778
140_115	44.87759306 N	66.97909444 W	44.91101639	66.96646111
140_114	44.91180361 N	66.96825583 W	44.87774528	66.97986472
140_150	44.88664056 N	66.97595778 W	44.88133639	66.95777861
Passamaquoddy 2008				
148_1609	45.08199222 N	67.01068806 W	45.08200694	66.97070167
148_1558	45.08288000 N	66.97085167 W	45.08272250	67.01077306
148_1546	45.08348806 N	67.01052861 W	45.08342111	66.97075389
148_1534	45.08428139 N	66.97093694 W	45.08414528	67.01017667
148_1522	45.08498778 N	67.01060167 W	45.08488389	66.97065528
148_1511	45.08570444 N	66.97091222 W	45.08560028	67.01061944
148_1459	45.08642639 N	67.01084583 W	45.08629083	66.97074528
148_1448	45.08712556 N	66.97052528 W	45.08704194	67.01111278
148_1436	45.08796639 N	67.01021194 W	45.08778111	66.97051861
148_1425	45.08863667 N	66.97059222 W	45.08841306	67.01041806
148_1413	45.08933417 N	67.01008556 W	45.08918889	66.97075722
148_1402	45.08995139 N	66.97094556 W	45.08991472	67.01025306
148_1350	45.09076472 N	67.01062639 W	45.09066222	66.97079667
148_1339	45.09143778 N	66.97056528 W	45.09136028	67.01063722
148_1327	45.09214278 N	67.01056167 W	45.09207889	66.97079778
148_1316	45.09293556 N	66.97073028 W	45.09278222	67.01033000
148_1305	45.09351556 N	67.00876417 W	45.09350611	66.97074750
148_1252	45.09435611 N	66.97054778 W	45.09422444	67.01059833
148_1241	45.09490361 N	67.01079222 W	45.09497472	66.97062083
148_1229	45.09575417 N	66.97054639 W	45.09564611	67.01082278
148_1218	45.09643750 N	67.01093278 W	45.09640417	66.97077972
148_1206	45.09721000 N	66.96994028 W	45.09706194	67.01091139
148_1154	45.09795444 N	67.01015611 W	45.09785139	66.96959667
148_1142	45.09869472 N	66.96998389 W	45.09854222	67.01050806
148_1131	45.09939139 N	67.01042639 W	45.09933472	66.96987639
148_1119	45.10001444 N	66.96966028 W	45.09994028	67.01095528
Passamaquoddy 2009				
160_2104	45.04486444 N	67.00250722 W	45.05561278	67.00868972

160_2059	45.04565944 N	67.01050417 W	45.04787583	67.00330417
160_2055	45.04828944 N	66.99820889 W	45.04584806	67.00582944
160_2037	45.03793639 N	67.02732139 W	45.04699056	66.99695889
160_2026	45.04721917 N	66.99757083 W	45.03820917	67.02794833
160_2015	45.03854611 N	67.02776083 W	45.04763861	66.99745444
160_2005	45.04800000 N	66.99786056 W	45.03887861	67.02810944
160_1953	45.03920722 N	67.02800333 W	45.04834750	66.99775861
160_1942	45.04863667 N	66.99829083 W	45.03947194	67.02871444
160_1931	45.03988056 N	67.02866611 W	45.04902833	66.99825250
160_1920	45.04928056 N	66.99878389 W	45.04017389	67.02922278
160_1908	45.04055389 N	67.02902111 W	45.04993528	66.99870056
160_1858	45.04993528 N	66.99916444 W	45.04078278	67.02952778
160_1845	45.04353889 N	67.03051028 W	45.05271667	67.00041639
160_1834	45.05013333 N	66.99956444 W	45.04121139	67.02963833
160_1822	45.04155000 N	67.02966667 W	45.05069111	66.99912528
160_1811	45.05100694 N	66.99956472 W	45.04186278	67.03012028
160_1759	45.04218194 N	67.03022111 W	45.05132583	66.99964861
160_1748	45.05155917 N	67.00011139 W	45.04250194	67.03052389
160_1737	45.04311972 N	67.02970778 W	45.05197583	67.00011750
160_1719	45.05234250 N	67.00062306 W	45.04318556	67.03097000
160_1708	45.05301194 N	67.03051417 W	45.05301194	67.00062972
160_1656	45.05329417 N	67.00079167 W	45.04407417	67.03161833
160_1644	45.04468333 N	67.03136139 W	45.05346333	67.00126722
160_1632	45.05375083 N	67.00193333 W	45.04484306	67.03194944
160_1620	45.04535778 N	67.03175889 W	45.05431028	67.00141639
160_1609	45.05459639 N	67.00178472 W	45.04547972	67.03244639
160_1558	45.04596833 N	67.03215667 W	45.05505083	67.00143528
160_1546	45.05518750 N	67.00245556 W	45.04618806	67.03277556
160_1534	45.04649972 N	67.03260444 W	45.05567694	67.00219361
160_1523	45.05584083 N	67.00265306 W	45.04685056	67.03339833
160_1511	45.05671528 N	67.00273667 W	45.04739278	67.03370278
160_1459	45.05667278 N	67.00281528 W	45.04742194	67.03360500
160_1448	45.04813167 N	67.03305583 W	45.05693056	67.00305639
160_1436	45.05679000 N	67.00344167 W	45.04804528	67.03491472
Pennfield				
141-1106	45.00025806 N	66.83796333 W	45.06394000	66.63856972
141_1317	45.09265139 N	66.60337722 W	44.99901139	66.83672972
142_1114	44.99812611 N	66.83441222 W	45.06110389	66.63742861
142_1209	45.06155556 N	66.63671972 W	45.09167944	66.60264500
149_1137	45.04597389 N	66.75930972 W	44.99636778	66.75901306
142_1940	45.06313333 N	66.63152056 W	44.99712139	66.83266861
143_1119	44.99528417 N	66.83290333 W	45.06237583	66.63125444
144_1107	44.99378972 N	66.83136722 W	45.06179111	66.63012083
144_1812	45.06048194 N	66.62911250 W	45.01209083	66.77372806

145_1135	44.99078611 N	66.82817000 W	45.05919944	66.62834639
145_1808	45.05866167 N	66.62656250 W	44.98986861	66.82615167
146_1113	44.98780083 N	66.82504778 W	45.05749778	66.62512417
149_1317	45.09251306 N	66.60349944 W	44.99911583	66.83639139
149_1206	44.99630722 N	66.72107750 W	45.05574639	66.72124667
149_1240	45.06529139 N	66.68342583 W	44.99600694	66.68301250
149_1315	44.99613000 N	66.64494111 W	45.07562778	66.64522389
146_1230	45.10165389 N	66.60750806 W	44.99702861	66.60690806
138_1404	45.03020472 N	66.57184972 W	45.01252722	66.82659944
146_1838	44.99622139 N	66.53227806 W	44.99656472	66.83726639
Maces				
146_1408	45.12205000 N	66.54461583 W	45.12194917	66.57389861
146_1527	45.12167389 N	66.51479583 W	45.10730139	66.47859250
146_1438	45.11928694 N	66.54686861 W	45.11891139	66.57375556
145_1554	45.11377667 N	66.55985861 W	45.11408889	66.48148556
144_1714	45.11242389 N	66.56059639 W	45.11237778	66.49486944
144_1656	45.11159056 N	66.49648028 W	45.11159861	66.56115389
144_1637	45.11073194 N	66.56587722 W	45.11095944	66.49816750
144_1621	45.08144750 N	66.49739139 W	45.11001778	66.56516111
144_1607	45.10939222 N	66.56035861 W	45.10960556	66.50910389
144_1552	45.10870222 N	66.50912722 W	45.10881861	66.55928417
144_1539	45.10814611 N	66.55729806 W	45.10813278	66.50997250
144_1526	45.10730139 N	66.51016639 W	45.10750861	66.55639833
144_1513	45.08506806 N	66.57917056 W	45.10670833	66.51088778
144_1458	45.10587639 N	66.51000722 W	45.10598694	66.55610000
144_1443	45.10506917 N	66.56626194 W	45.10523972	66.51063806
144_1427	45.10408722 N	66.50857639 W	45.10453194	66.56573444
144_1354	45.10331722 N	66.50812472 W	45.10318806	66.56494778
144_1338	45.10249944 N	66.56905111 W	45.10281139	66.50827222
144_1319	45.10150444 N	66.50045944 W	45.10177000	66.56879306
144_1300	45.10098806 N	66.57050639 W	45.10094611	66.50112000
144_1241	45.09989722 N	66.49872806 W	45.10026972	66.56990500
144_1220	45.09929222 N	66.57841028 W	45.09927278	66.49878472
143_1438	45.09890222 N	66.49298778 W	45.09907528	66.57557250
143_1417	45.09828472 N	66.57569222 W	45.09833528	66.49354167
143_1346	45.10094611 N	66.50112000 W	45.09743722	66.57575389
143_1324	45.09671500 N	66.58219444 W	45.09665389	66.49570833
143_1259	45.09602861 N	66.49636306 W	45.09576056	66.58239972
143_1232	45.09318778 N	66.60102083 W	45.09294722	66.49658139
141_1217	45.09442361 N	66.60318611 W	45.09442500	66.49628417
141_1245	45.09374139 N	66.49648333 W	45.09356889	66.60254889
142_1223	45.09318778 N	66.60102083 W	45.09294611	66.49676722
142_1251	45.09225778 N	66.49613833 W	45.09226056	66.59092861
142_1317	45.09166944 N	66.59138500 W	45.09154417	66.49597833

142_1342	45.09086694 N	66.49536361 W	45.09081222	66.58711139
142_1407	45.09907528 N	66.57557250 W	45.09018611	66.49661806
142_1431	45.08960861 N	66.49614417 W	45.08939111	66.58571694
142_1455	45.08886167 N	66.58572861 W	45.08883833	66.49755000
142_1519	45.08819028 N	66.49753528 W	45.08796278	66.58466444
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142_1606	45.08662056 N	66.49671917 W	45.08653056	66.58174833
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142_1844	45.08161444 N	66.57652278 W	45.08144750	66.49739139
142_1906	45.08068444 N	66.49738056 W	45.08079389	66.57417306
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146_1748	45.13177972 N	66.53069722 W	44.99695889	66.53083056
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160_1505	45.08325750 N	66.54271444 W	45.05749250	66.54250278
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160_1530	45.08328861 N	66.53831278 W	45.05731917	66.53859333
160_1542	45.05748167 N	66.53645056 W	45.08331917	66.53655806
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160_1609	45.05760861 N	66.53223556 W	45.08327111	66.53247611
160_1621	45.08327722 N	66.53045778 W	45.05754583	66.53037750
160_1633	45.05751250 N	66.52830222 W	45.08333500	66.52839972
160_1646	45.08340667 N	66.52595056 W	45.05746444	66.52634167
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Chance				
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134_1158	45.13524889 N	66.19542889 W	45.04811833	66.45873861
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181_001	45.26587972 N	66.06275917 W	45.06006389	66.37921889
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181_004	45.03524806 N	66.48627250 W	45.13995972	66.06904528
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181_008	45.14461056 N	66.07003167 W	45.10815028	66.25042778
181_009	45.12205056 N	66.21850222 W	45.14713556	66.06579111
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187_019	45.03467278 N	66.55706861 W	45.00990889	66.68091750
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188_005	45.01429639 N	66.67894083 W	44.93837083	66.84481833
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218_001	44.75341389 N	66.73818444 W	44.97921722	66.03925333
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219_001	44.78439333 N	66.68314861 W	44.93109500	66.25774806
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223_001	44.75456222 N	66.73602444 W	44.82417028	66.67614694
223_002	44.87365139 N	66.56415944 W	44.76466444	66.72289083
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229_001	45.04444306 N	66.08419472 W	44.89843472	66.65015444
230_001	44.85065861 N	66.71654028 W	44.88812611	66.69543083
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237_002	44.93994917 N	66.63300611 W	44.97473722	66.49197583
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240_002	45.03656806 N	66.39980806 W	44.76195944	66.72235278

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Atlantic Geoscience Society 37th Colloquium and Annual Meeting, Fredericton, NB, February 11-13, 2011. *Surficial sediments and Quaternary stratigraphy of Maces Bay, Bay of Fundy.*

Atlantic Geoscience Society 35th Colloquium and Annual Meeting, Moncton, NB, February 6-7, 2009. *Acoustic mapping of the Bay of Fundy between Maces Bay and Passamaquoddy Bay.*