University of Alberta

# The Origin of the Morphology of the Eastern Scotian Shelf, Atlantic Canada

By

©

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A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree Master of Science

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The undersigned certify that they have read, and recommended to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled "The Origin of the Morphology of the Eastern Scotian Shelf, Atlantic Canada" Submitted by Lisa Marguerite Sankeralli in partial fulfillment of the requirements for the degree of Master of Science in Geomorphology.

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# ABSTRACT

The morphology of the eastern Scotian Shelf, Atlantic Canada is distinct from the western Scotian Shelf. The eastern Scotian Shelf appears to be dissected by complex patterns of incisions of varying scales. Debate centres on whether this morphology is a product of glacial modification of a pre-glacial fluvial morphology, subglacial meltwater erosion (catastrophic or non-catastrophic), ice erosion alone, or a combination of each.

A zone classification is presented which subdivides the exposed eastern Scotian Shelf incisions into four zones (A, B, C, and D) based on location, size, incision fill, plan-form patterns and orientations. Each zone appears to have been affected by processes unique from the other zones in addition to common processes.

Lithostratigraphy and biostratigraphy suggests open marine and ice shelf conditions from the Mid-Wisconsinan to the Holocene, with no evidence of input from sudden discharges of large quantities of freshwater. Large inputs of ice from ice streams were likely necessary to support a temperate ice shelf in a relatively open marine environment such as the Scotian Shelf. It is concluded the eastern Scotian Shelf morphology is a product of glacial modification, by both ice and meltwater, of a pre-glacial fluvial morphology and that Ice-streaming had a strong influence.

## ACKNOWLEDGMENTS

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*It is to my family and to Gordon Fader that I dedicate this thesis.*

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# CHAPTER 1: INTRODUCTION

## 1.1: INTRODUCTION

The complex morphology of the eastern Scotian Shelf, offshore Atlantic Canada, presents both a difficult terrain for engineering activities such as pipeline and telecommunication route selection, and a rich and varied seabed habitat for fisheries resource development and management (Figs. 1.1, 1.2).

The eastern Scotian Shelf is morphologically complex when compared to the rest of the Scotian Shelf. It appears to be extensively dissected by networks of buried and exposed incisions of varying widths and depths. This is in contrast to the remainder of Scotian Shelf which exhibits broad open basins between shallow banks. The origin of this rough and irregular morphology is cause for much debate: Were the incisions formed by water or by ice? In a sub-aerial or subglacial environment? What are the relative ages of the incisions and their associated processes of formation? (The term incision refers to linear or sinuous erosional forms, isolated or part of a system, which possesses a generally constant width and a length greater than its width. This term is adopted from Wingfield, 1990, and is preferred over other terms used to describe similar features—i.e. channel, scour, valley—which imply a mode or origin. Such terms will only be used when referring to the work of others, and when an implied mode or origin is intended.)

Some buried incisions appear to be filled with Tertiary sediments and are, therefore, likely representative of fluvial valleys formed as part of a Tertiary age drainage system that

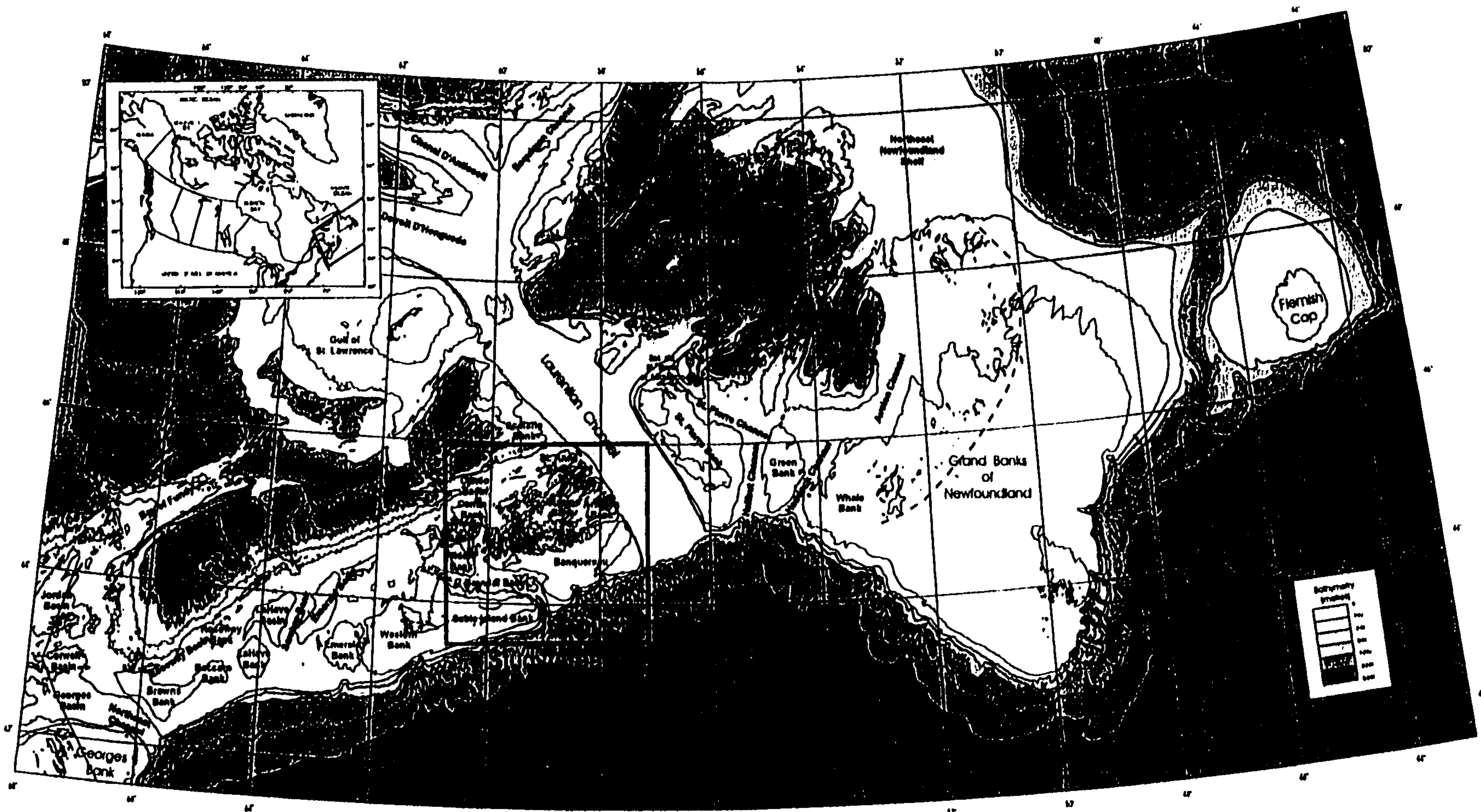
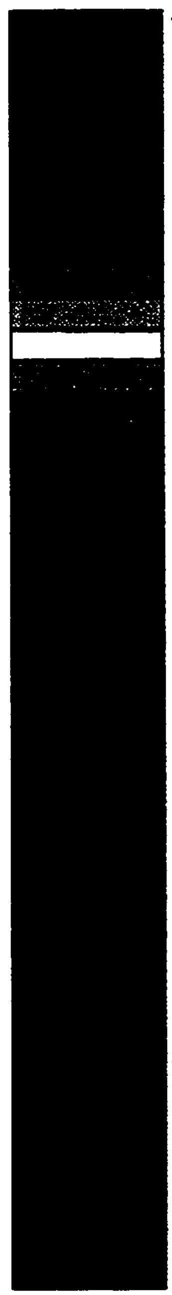



Figure 1.1 Regional map of Atlantic Canada showing the eastern Scotian Shelf study area. The dashed line on the Grand Banks of Newfoundland marks the seaward boundary of exposed incisions).

Figure 1.2 Bathymetry of the eastern Scotian Shelf, Atlantic Canada. This image of the study area was created using GRASS software at the Bedford Institute of Oceanography, Geological Survey of Canada (Atlantic).
Parameters as follows:
Grid: 250 x 250 meters
Vertical Exaggeration: 100 times
Illumination: 25 degrees
Azimuth: 320 degrees (from the northeast)

Scale in Meters



flowed across the Scotian Shelf. However, many incisions exposed at the surface, in addition to others buried beneath the eastern Scotian Shelf banks are cut into Cretaceous, Tertiary, and Quaternary sediments, and are partially to completely filled with Quaternary sediments. All of the eastern Scotian Shelf incisions are located below maximum estimates of Late Wisconsinan glacial low sea-level stand (110-120 m bsl; King and Fader, 1986; Fader, 1989). This suggests that the Pleistocene phase of incision erosion took place in a subglacial environment.

The origin of the eastern Scotian Shelf morphology is highly controversial. It was initially proposed that the rough and irregular morphology of the eastern Scotian Shelf is a product of Tertiary fluvial system evolution which was subsequently modified by glacial processes (MacLean and King, 1971; King et al., 1974; King and MacLean, 1976). It was later proposed that portions of the buried eastern Scotian Shelf incisions were excavated by ice in a manner similar to fjords (Amos and Knoll, 1987—Banquereau), whereas others proposed that they were formed by catastrophic outbursts of subglacial meltwater (Boyd et al., 1988—Sable Island Bank).

## 1.2: PURPOSE OF STUDY

The origin of deep incisions found on glaciated continental shelves, such as those representative of the eastern Scotian Shelf will be explained. Were such features formed by erosion by catastrophic releases of subglacial meltwater, non-catastrophic meltwater release, or by meltwater at all? To what extent did direct glacial ice erosion play a role in their excavation? Answering questions such as these may contribute greatly to the understanding of glacial processes in continental shelf environments, where the great continental ice sheets terminated.

The origin of the morphology of the eastern Scotian Shelf is considered in detail in this thesis. The approach taken involves addressing the issue by applying different theories of incision formation to the eastern Scotian Shelf morphology, and hence tests different hypotheses. The outburst-flood hypothesis is considered which requires interpreting the incisions as channels and scours, products of subglacial meltwater erosion (section 5.4.2). Chapter 4 presents a zonal classification of the eastern Scotian Shelf incisions. Based on this classification the incisions are considered according to their different characteristics and proposed modes of origin are presented for incisions within each distinct zone (section 5.4.3). It is concluded that the later approach is the one which best fit the question of the origin of the morphology of the eastern Scotian Shelf.

## 1.3: PREVIOUS WORK ADDRESSING THE ORIGIN OF THE EASTERN SCOTIAN SHELF INCISIONS

To date, few major works exist which directly address the origin of incisions on the eastern Scotian Shelf (MacLean and King, 1971; King et al., 1974; King and MacLean, 1976; Boyd et al., 1988; McLaren, 1988; Amos and Knoll, 1987; Amos and Miller, 1990; Loncarevic et al., 1992; Stea, 1995). With the exception of Loncarevic et al. (1992), these works are focused on local areas (Fig. 1.3), and do not address the possibility of genetically related incisions, or numerous incision systems on the eastern Scotian Shelf, of similar and differing scales, potentially formed by multiple processes and at different times.

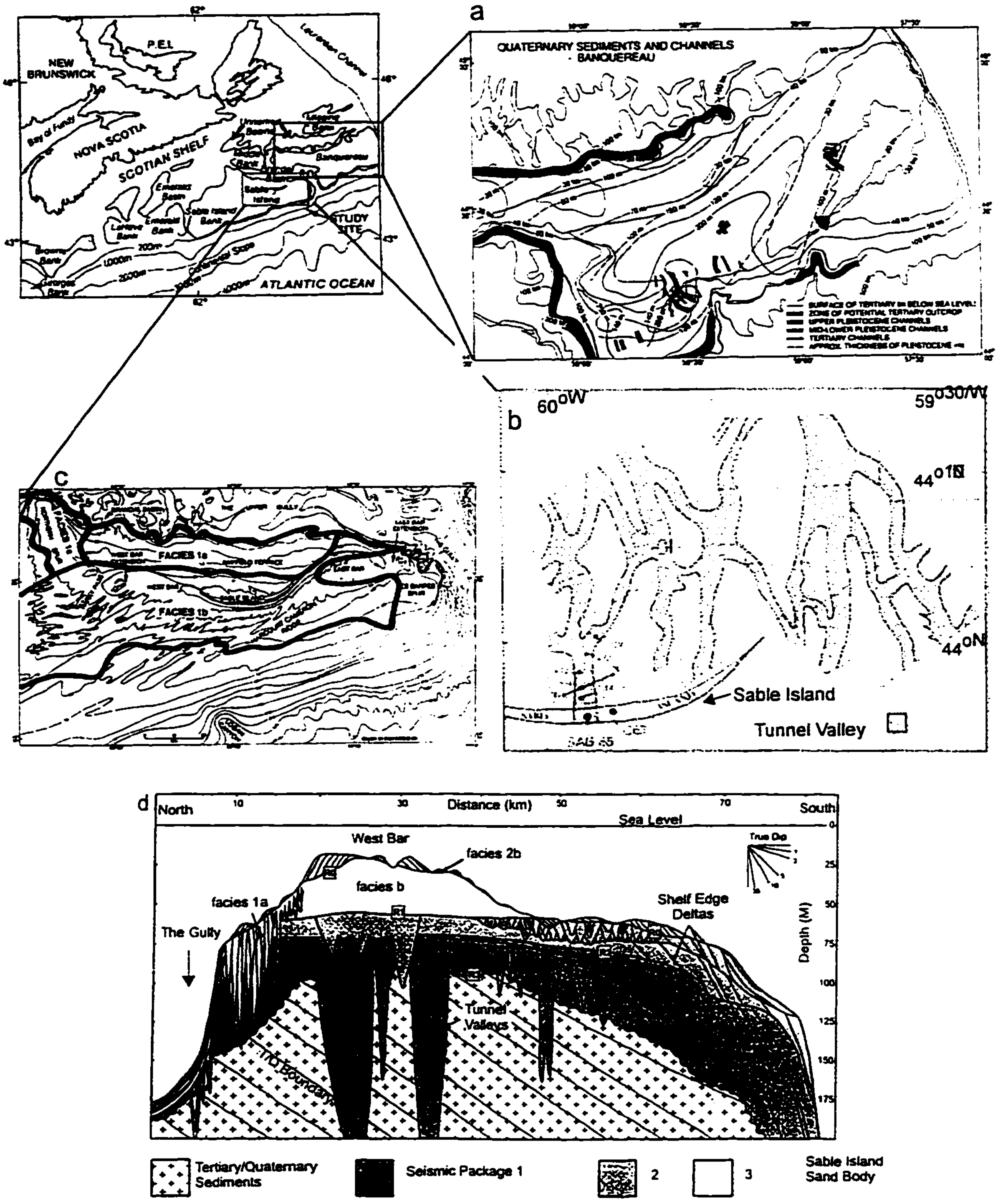



Figure 1.3 Mapped buried incisions of the eastern Scotian Shelf. a) Banquereau (Amos and Knoll, 1987; b) Sable Island Bank (Boyd et al., 1988); c); Sable Island Bank - Sable Island Sand Body (McLaren, 1988); d) Stratigraphy of Sable Island Sand Body (McLaren, 1988; note location of R3, R2, and R1 discussed in text). Modified from Amos and Knoll (1987), Boyd et al. (1988), and McLaren (1988).

Earlier works on the general morphology and geology of the Scotian Shelf (King et al., 1974; MacLean and King, 1971; King and MacLean, 1976), addressed the origin of the eastern Scotian Shelf incisions. Based on detailed analysis of seismic reflection data and correlation with samples (core, borehole, and grab samples), bedrock units and major regional unconformities were mapped (MacLean and King, 1971; King et al., 1974; King and MacLean, 1976). As a result of these observations it was proposed that the eastern Scotian Shelf morphology is a result of fluvial erosion during the Early and Late Tertiary with subsequent modification by glacial processes (MacLean and King, 1971; King et al., 1974; King and MacLean, 1976). These authors believed the present physiography of the eastern Scotian Shelf to be representative of an exhumed surface inherited primarily from the Early Tertiary erosional phase. This is in contrast to later works which place stronger emphasis on glacial erosion (ice and/or meltwater) as a mode of origin (i.e. Amos and Knoll, 1987; Boyd et al., 1988; Loncarevic et al. 1992; Stea, 1985).

Four regional unconformities were mapped on the outer eastern Scotian Shelf banks (Sable Island Bank and Banquereau) associated with the Pleistocene (King et al., 1974; Amos and Knoll, 1987; Boyd et al., 1988; McLaren, 1988; Amos and Miller, 1990; Fig. 1.3d). The lower most unconformity, R3, occurs on the bedrock surface and marks the onset of Pleistocene glaciation (King et al., 1974; Amos and Knoll, 1987; Boyd et al., 1988; McLaren, 1988; Amos and Miller, 1990). The middle Pleistocene unconformity, R2, occurred during the Pleistocene (within Pleistocene sediments). R1 is a second unconformity within Pleistocene sediments above R2, and McLaren identified the Pleistocene-Holocene boundary as R0 (Amos and Knoll, 1987; Boyd et al., 1988; McLaren, 1988; Amos and Miller, 1990; Fig. 1.3d). R3 is interpreted as a glacial erosional surface, and R2 and R1 are interpreted as marine transgression erosional surfaces. Incisions of varying scales originate between and/or in association with these unconformities and will be discussed according to their stratigraphic position from the lowermost to the uppermost.

1. Incisions Between R3 and R2

Incisions between R3 And R2 are the largest in scale. Amos and Knoll (1987; Fig. 1.3a) identified a prominent incision originating in Quaternary sediments between R3 and R2, and truncating R3. The incision is approximately 200 m deep and 2.5 km wide, and contains a sedimentary unit described as boulders intermixed with sand and fines which is densely packed (based on drill core depths of 126-190 m below the seabed). This unit was interpreted as glacial debris.

Boyd et al. (1988; Fig. 1.3b) mapped a system of buried incisions beneath Sable Island Bank from industry seismic records (2 second sweep), originating between R3 and R2, and truncating R3. The interconnected network of incisions displays bifurcation and rejoining. Larger incision bases are >300 m deep and average 2-3 km in width, however some are in excess of 400 m deep with widths up to 4 km. The incisions are steep walled, and the axis of a larger incision displays a thalweg of variable depth, and no trend of deepening in any particular direction. Tributary incisions display bases shallower than those of the major incisions. Based on core descriptions and seismic stratigraphy, incision fill was described as consisting of a lower unit displaying a chaotic or incoherent acoustic character, separated by an unconformity and capped by an upper, broader unit interpreted as glaciomarine sediments displaying channel cut and fill towards the upper portion, and an opaque lower portion.

A core through one of the incisions located beneath Sable Island penetrated the capping unit described above recovered a $^{14}C$ date of 37 210 ± 400 yr BP at 72 m (RIDDL #639). Based on this date, and the view of King and Fader (1986), who proposed oxygen isotope stage 4 as the last major glacial cycle to affect the Scotian Shelf, it was postulated that the youngest possible age of origin for the Sable Island buried incisions was the Early Wisconsinan (Boyd et al., 1988). This does not account for the possibility that the dated material may have been reworked from older deposits, or more recent dates than those discussed by King and Fader (1986), indicating that ice was grounded in Emerald Basin during the Late Wisconsinan (Gipp and Piper, 1989). Still, strong land and marine evidence exists which indicates that the Early Wisconsinan (OIS 4) and the Illinoian (OIS 6), were times when glaciation was most extensive over Nova Scotia and on the Scotian Shelf (this evidence is presented in Chapter 2, section 2.3.3a).

Boyd et al. (1988) are advocates of the outburst-flood hypothesis. They proposed that a catastrophic release of subglacially stored meltwater evolved from unstable sheet flow to channel cutting, to explain the origin of the buried incision in their study area. They interpret the basal acoustically incoherent incision fill unit as a meltwater deposit. This interpretation is in contrast to that of Amos and Knoll (1987) who interpreted a basal acoustically incoherent incision fill unit as glacial debris and proposed glacial excavation as the means by which the large scale Quaternary incisions of Banquereau were formed. The basal acoustically incoherent unit interpreted as glaciofluvial deposits by Boyd et al. (1988) was not sampled, while the interpretation of the acoustically incoherent incision fill unit as glacial debris by Amos and Knoll (1987) was based on well log descriptions. Amos and Miller (1990) identified buried incisions (3 km wide and 225 m deep beneath Cohasset bore hole) beneath

Sable Island Bank in the same stratigraphic location, but did not discuss their origin.

The buried incisions mapped by Boyd et al. (1988) were interpolated in areas not covered by seismic survey lines. It may be that if these incisions were exposed, the apparent anabranching pattern would more closely resemble the exposed, isolated linear incisions, displaying cross-cutting relationships to the north of Sable Island Bank and Banquereau (Fig. 1.2). Boyd et al. (1988) propose that a regional unconformity which cuts across the surface of interfluves between incisions suggests regional high-volume discharge (sheet flow) which subsequently localized into incisions. They did not consider the possibility of advancing ice eroding the unconformity.

Boyd et al. (1988) suggested that because incisions buried beneath Sable Island are cut far below sea level (some >400 m bsl), the most likely driving force for meltwater discharge would have been pressure from overlying ice as meltwater catastrophically escaped from subglacial reservoirs. They suggest Brandal Basin, just north of Sable Island Bank, and the area north of Banquereau as possible reservoirs. The eastern Scotian Shelf bathymetry (Fig. 1.2) clearly shows that Brandal Basin and the area north of Banquereau are extensively dissected by incisions, and it is unknown whether the incisions are genetically related to the incisions buried beneath Sable Island Bank and Banquereau. If the incisions are genetically related, these areas would not have acted as reservoirs, but would have also been subjected to meltwater erosion.

Calculations by Boyd et al. (1988) indicate that potential meltwater storage in Brandal Basin would not be enough to erode the entire incision system mapped beneath Sable Island Bank. Further mapping may also reveal that the system is much more extensive than was mapped by Boyd et al. (1988), thus requiring additional meltwater sources. However, if the incised basin areas north of Sable Island Bank and Banquereau were sources of meltwater, a steep hydraulic gradient would be required to drive water out of the combined depressions with enough force to erode the buried Sable Island incisions. This would require ice over these areas to have been substantially thicker than ice over Sable Island Bank. Such a situation is not likely as ice floating over subglacial lakes would tend to thin and have lower surface profiles (Shoemaker, 1991; Ellis-Evans and Wynn-Williams, 1996).

Amos and Knoll (1987) observed that incisions buried beneath Banquereau are similar in geometry to Baffin Island Fjords. Based on this comparison, they suggest that the incision fill deposits they interpreted as glacial debris indicate ice was in contact with the incision. The sampled incision was the only location where the boulder rich glacial deposit was observed, hence, Amos and Knoll (1987) suggested that Late Wisconsinan ice did not cover Banquereau, but was restricted to fjords (the incisions) which were shaped and infilled by glaciers. They did not consider the possibility of fast moving, erosive ice streams occupying the incisions while slower moving, less erosive ice (possibly frozen to the bed) covered the remainder of the bank and interfluves. It may be that the incisions contain boulder rich glacial deposits as a result of deposition by faster moving ice transporting large quantities of debris from farther distances, while the areas between the incisions may possess locally derived debris from the underlying Tertiary and Cretaceous sediments (sandstone and siltstones), and glaciomarine sediments, which were subsequently reworked by glaciofluvial processes and marine transgression. Additional considerations may be that the deposits within the incisions post-date the incision-forming erosional event and are not genetically related, and/or that the incisions are glaciofluvial in origin if Boyd et al. (1988) are correct in their hypothesis.

Neither of these studies had enough core data to conclude absolutely the nature of the origin of the incision fill deposits.

## 2. Incisions Between R2 and R1

Smaller buried incisions, stratigraphically overlying the incisions discussed above have been mapped on the banks of the eastern Scotian Shelf between R2 and R1 (Amos and Knoll, 1987; McLaren, 1988; Amos and Miller, 1990). Amos and Knoll (1987) identified a regionally extensive barren gravel/sand deposit on Banquereau with 'channel-like' features in association with R2. This deposit was interpreted to be derived from the transport and sorting of glacial outwash which was subsequently winnowed during low sea level stands (60-70 m bsl) between 20 000 and 26 000 yr BP which also produced R2 (Amos and Knoll, 1987). These dates are based on the timing of Late-Wisconsinan glacial advance over the banks proposed by King and Fader (1986), and hence ice-proximal conditions. One $^{14}C$ date of 19 480 ± 250 yr was derived from a shell fragment obtained from the overlying unit.

Amos and Miller (1990) identified a regionally extensive barren gravelly/sand deposit on Sable Island Bank and correlated it using core samples with the gravelly/sand deposit of Banquereau. However, they interpreted the unit to have been deposited during late Mid-Wisconsinan low sea-level stand from 28 000 to 32 000 yr BP. These dates were derived from $^{14}C$ age determinations from bounding sediments, and do not correlate with those proposed for Banquereau (Amos and Knoll, 1987). Incisions associated with this unit were identified and interpreted to be tidal in origin, formed during low sea level stand. This interpretation was based on assemblages of foraminifera suggestive of open marine conditions. Seismic survey lines were not dense enough to establish a plan view of the incision patterns identified on 2-D seismic reflection records for Banquereau or Sable Island Bank. Such knowledge of the incision patterns would help in establishing if the incisions associated with the barren gravelly/sand unit are glaciofluvial in origin as proposed by Amos and Knoll (1987), or tidal in origin proposed by Amos and Miller (1990).

### 3. Incisions Associated With R1

Amos and Knoll (1987) observed 10 m deep incisions associated with a regional unconformity (R1) between 60 and 110 m bsl (within maximum estimates of Late Wisconsinan low sea level stand; King and Fader, 1986; Fader, 1989). This unconformity is underlain by a widespread unit of sorted gravels (Channel Gravel) cut by small channels up to 20 m in relief. The Channel Gravel unit separates glaciomarine deposits (Emerald Silt) from overlying sand and gravel deposits produced by the reworking of till and glaciomarine deposits during marine transgression (Sable Island Sand and Gravel). Amos and Knoll (1987) interpreted this unit as a subaerial glaciofluvial deposit, formed during Late Wisconsinan low sea-level stand by the transport and reworking of glacial outwash derived from an ablating ice sheet situated immediately north of Banquereau. The overlying unconformity was interpreted to be a transgressive erosional surface (Amos and Knoll, 1987).

Amos and Miller (1990) identified a regional unconformity (R1) and wide spread Channel Gravel unit directly underlying the unconformity on Sable Island Bank. These observations were correlated using core samples to the same unconformity (R1) and Channel Gravel unit identified on Banquereau (Amos and Knoll, 1987; Amos and Miller, 1990). Incisions identified at the base of the Channel Gravel member on Sable Island Bank are up to 10 m deep and 500 m wide.

The interpretations for the incisions associated with the Channel Gravel member for Sable Island Bank and Banquereau are not in agreement; Sable Island Bank was interpreted to possess tidal channels, while Banquereau was interpreted to possess glaciofluvial channels. As with the conclusions for the barren gravelly/sand unit, these interpretations are based on 2-D seismic profiles, and therefore, a plan view of channel patterns cannot be established with which to verify these interpretations. Amos and Knoll (1987), and Amos and Miller (1990) are in agreement that the Channel Gravel member and associated unconformity (R1) separate the Emerald Silt formation from the Sable Island Sand and Gravel formation, and marks a late glacial low sea-level stand. Dates derived from sediments bounding R1 on Banquereau provided a time bracket of 8000 and 16 000 yr BP, and dates derived from Sable Island Bank narrowed this bracket to between 11 000 and 13 000 yr BP for the time of R1 development.

4. Incisions above R1

McLaren (1988) studied incisions buried within the Sable Island Sand Body of Sable Island Bank (Fig. 1.3c). These incisions occur stratigraphically above R1 and appear to cross-cut incisions, or 'tunnel valleys,' similar in dimension to those mapped by Boyd et al. (1988), Amos and Knoll (1987), and Amos and Miller, (1990), between R3 and R2. McLaren sub-divided the Sable Island Sand Body into two sedimentary sequences. The lower sequence, containing facies 1a, 1b, and 1c, is thought to be glacial in origin, while the upper sequence, containing facies 2a and 2b, is thought to be Holocene in origin (McLaren, 1988) and will not be discussed here (Fig 1.3d).

Facies 1a incisions, located in water depths greater than 80 m, display maximum depths of 95-100 m and widths of 3-4 km (similar to incisions identified between R2 and R3 by Amos and Knoll, 1987, and Amos and Miller, 1990), no preferred gradient along their length (though they terminate upslope), a predominate north-south trend, and an abrupt termination north of Sable Island. McLaren (1988) suggested that they are Late Wisconsinan subglacial meltwater channels responsible for transporting sediment-laden meltwater to the ice margin. This process resulted in a 40 m thick sand deposit on the central area of Sable Island Bank, interpreted to be a moraine. Facies 1a incision fill consists of cut and fill sequences displaying a transparent, amorphous or stratified character (McLaren, 1988). McLaren (1988) suggested that stratified incision fill is representative of episodic deposition, while amorphous fill is representative of a single catastrophic cut and fill event. This interpretation acknowledges the possibility of non-catastrophic meltwater processes being periodically punctuated by jõkulhlaup type events, and does not favour one process over the other as the primary incision forming process.

Facies 1b incisions are smaller in scale than those found in facies 1a. They display depths of 10-15 m and widths of 200-550 m, similar in dimension to the incisions associated with the Channel Gravel member of Amos and Knoll (1987), and Amos and Miller (1990). These incisions occur in isolated areas distal to the termination point of facies 1a incisions in water depths less than 60 m (McLaren, 1988; Fig. 1.3c). It was suggested that the smaller scale incisions of facies 1b may be either distal time-equivalents of those located north of Sable Island Bank in facies 1a, or a result of separate erosional events. McLaren (1988) further suggested that facies 1b is representative of a moraine derived from sediments transported seaward through the subglacial channel network located north of Sable Island in facies 1a.

Facies 1c incisions (Fig. 1.3c) lie in a restricted area, stratigraphically above the subglacial sequence (facies 1a and facies 1b). These incisions are located in water depths of 35-40 m, and display depths of 10-15 m and widths of 200-250 m. Facies 1c incisions themselves were interpreted as glaciofluvial in origin, representing the subaerial component of a delta (McLaren, 1988).

King (1993) interpreted the entire Sable Island Sand Body sequence discussed above as glacial in origin, related to Late Wisconsinan ice (King 1993; 1994; 1996), representative of a moraine formed at a tide water margin. This is in agreement with McLaren (1988), though McLaren proposed three ice advances. King (1994) interpreted the existence of ice on the eastern Scotian Shelf as late as 12-14.5 ka though he acknowledges this is highly speculative. Thicker ice was proposed for the eastern Scotian Shelf based on the concepts of Boyd et al. (1988), McLaren (1988), and King (1993) who attribute incisions to subglacial meltwater processes requiring steep hydraulic gradients. If larger incisions (below R2) are related to ice streaming, such hydraulic gradients would not be necessary and therefore thick ice would not be necessary.

Amos (pers. comm., 1988) suggested that the R1 reflector interpreted by McLaren (1988) may be incorrect, and that facies 1a incisions may be located below R1 rather than above it (Fig. 1.3d). This interpretation may allow for correlation of these incisions with the incisions mapped by Amos and Knoll (1987), and Amos and Miller (1990) identified below R1. Amos (pers. comm., 1988) further suggests that the smaller facies 1b and 1c incisions may not be glaciofluvial in origin, but tidal. A further interpretation may be that the incisions are gas escape features for which there is increasing evidence (Amos, pers. comm., 1998). As with the other studies concerning incisions on the eastern Scotian Shelf, McLaren's interpretations were based on 2-D seismic surveys. Additional information on the plan-form pattern of the incisions may require reinterpretation of their origin. McLaren (1998) also interprets incision fill as relevant to the formation of the incision, however, this may not be the case. The fill may be post-incision formation.

## 5. Additional Interpretations for the Origin of the eastern Scotian Shelf Incisions

Loncarevic et al. (1992) describe the morphology of the eastern Scotian Shelf as a whole. This paper gives an overview of alternate theories of formation for the more deeply incised incisions (i.e. the exposed incisions, and the deeply eroded buried incisions described by and Amos and Knoll, 1987 and Boyd et al., 1988). They conclude that the incisions are likely glacial in origin based on the fact that similar incisions along the eastern margin of the Laurentian Channel reach depths up to 1300 m bsl, deeper than Pliocene or Quaternary drops in sea level.

Loncarevic et al. (1992) believe the eastern Scotian Shelf incisions were formed by sub-glacial meltwater streams under high pressure, however they remain uncertain as to whether or not the process was non-catastrophic of catastrophic. They observe cross-cutting relationships between the exposed incisions north of Sable Island Bank and Banquereau, and suggest that this was a result of numerous incision cutting events under different glacial ice configurations.

Loncarevic et al. (1992), described two types of incision fill: (1) thick deposits resembling Scotian Shelf basin fill sequences (Scotian Shelf Drift, Emerald Silt, LaHave Clay; section 2.3.1); and (2) an irregularly stratified unit directly overlying Tertiary bedrock with variable amounts of capping LaHave Clay. It was proposed that type 1 fill, correlated with larger incisions, may be associated with pre-Late Wisconsinan ice advances, while type 2 fill, correlated with smaller and possibly younger incisions, were flushed of earlier sediment deposits, and may be related to final ice-contact during a late Late Wisconsinan ice advance. These interpretations acknowledge ice contact with the incisions, yet meltwater was stressed as the incision forming erosive agent. The incisions may have been formed independent of the fill, however, the presence of Scotian Shelf Drift within the incisions indicates ice was in contact with the incisions, and therefore ice cannot be discounted as a possible contributing incision forming agent.

Stea (1995) suggested that the lack of incisions covering Misaine Bank supports the concept that an offshore ice divide was situated there (Ice Flow Phase 3; section 2.3.3). The northeast-southwest trending incisions northeast of Misaine Bank, and the north south trending incisions situated south of Misaine Bank represent ice flow away from such an ice divide. Stea (1995) does not distinguish whether ice or meltwater was responsible for the formation of the incisions. The eastern Scotian Shelf bathymetry image (Fig. 1.2) clearly shows that Misaine Bank does not 'lack' incisions, but is extensively dissected by incisions, though at a much smaller scale than the incisions surrounding the bank.

### 1.3.1: Summary

In summary, five processes have been proposed for the formation of identified buried and exposed Pleistocene eastern Scotian Shelf incisions:

1. Catastrophic subglacial meltwater outburst floods (Boyd et al., 1988)
2. Subglacial meltwater excavation; non-catastrophic and/or catastrophic? (McLaren, 1988; Loncarevic, et al., 1992)
3. Ice excavation (Amos and Knoll, 1987)
4. Subaerial: Glaciofluvial (McLaren, 1988; Amos and Knoll, 1987)

5. Subaerial: Tidal (Amos and Miller, 1990)

Loncarevic et al. (1992) considered all incisions covering the eastern Scotian Shelf and remained undecided as to whether the incisions were formed by catastrophic or non-catastrophic subglacial melt water processes. Stea (1995) did not address a particular choice for mode of formation, though, he suggested incisions surrounding Misaine Bank are representative of extensions from an ice divide situated over Misaine Bank. Boyd et al. (1988), and Amos and Knoll (1987) were the only authors who directly proposed a mode of origin for the larger scaled incisions below R2 (>300 m deep and > 2 km wide); one chose meltwater, and the other chose ice.

A lack of 3-D seismic survey data over the banks resulted in interpolation of buried incision patterns. In addition, samples from the deep incisions between R2 and R3 are lacking. Interpretations of the origin of the incisions are therefore, based strongly on 2-D seismic survey data and are subject to modification upon the recovery of additional data. Studying exposed incisions removes the necessity to interpolate incision patterns because they are clearly visible in the bathymetry. For instance, incisions over Misaine bank cannot be interpreted as subaerial tidal scour (Amos and Miller, 1990) as the tributary/distributary incision plan-form pattern clearly does not support such an interpretation.

This study expands on the work of Loncarevic et al. (1992) and concentrates on all exposed incisions covering the eastern Scotian Shelf.

# CHAPTER 2: STUDY AREA

## 2.1: THE EASTERN SCOTIAN SHELF STUDY AREA

The eastern Scotian Shelf study area (Fig. 2.1), is bounded to the north by Cape Breton Island, 46° 10'N (UTM 511480N, Zone 20), to the south by 43° 25'N (UTM 4824400N, Zone 20), to the west by 61° 05'W (UTM 649400E, Zone 20), and to the east by the Laurentian Channel, 57° 05'W (UTM 977800E, Zone 20). It is comprised of the widest and most complex portion of the Scotian Shelf and covers an area approximately 320 km x 290 km. The banks included in this area are Canso Bank, Misaine Bank, Middle Bank, Sable Island Bank, and Banquereau. Canso Bank, Misaine Bank, and Middle Bank are smaller banks towards the central region of the shelf. Sable Island Bank and Banquereau, large banks along the outer shelf, are separated by The Gully, a spectacular submarine canyon over 83 km long, 15 km wide, and reaching depths in excess of 3000 m (MacLean and King, 1971). The area north of Sable Island Bank and south of Middle Bank has been named Brandal Basin, though this area is dissected by exposed incision networks and does not resemble the major basins to the west (i.e. Emerald Basin and La Have Basin).

The eastern Scotian Shelf is complex and of particular interest due to the extensive system of buried and exposed incisions which dissect the entire area. Limited previous work in the area has mapped regions of buried incisions, and addressed possible modes of formation for the buried and exposed incisions (section 1.3; Fig. 1.3).

Water depths vary from sea level where Sable Island is subaerially exposed, to over 400 m bsl in the deeper incisions. Surface water circulation

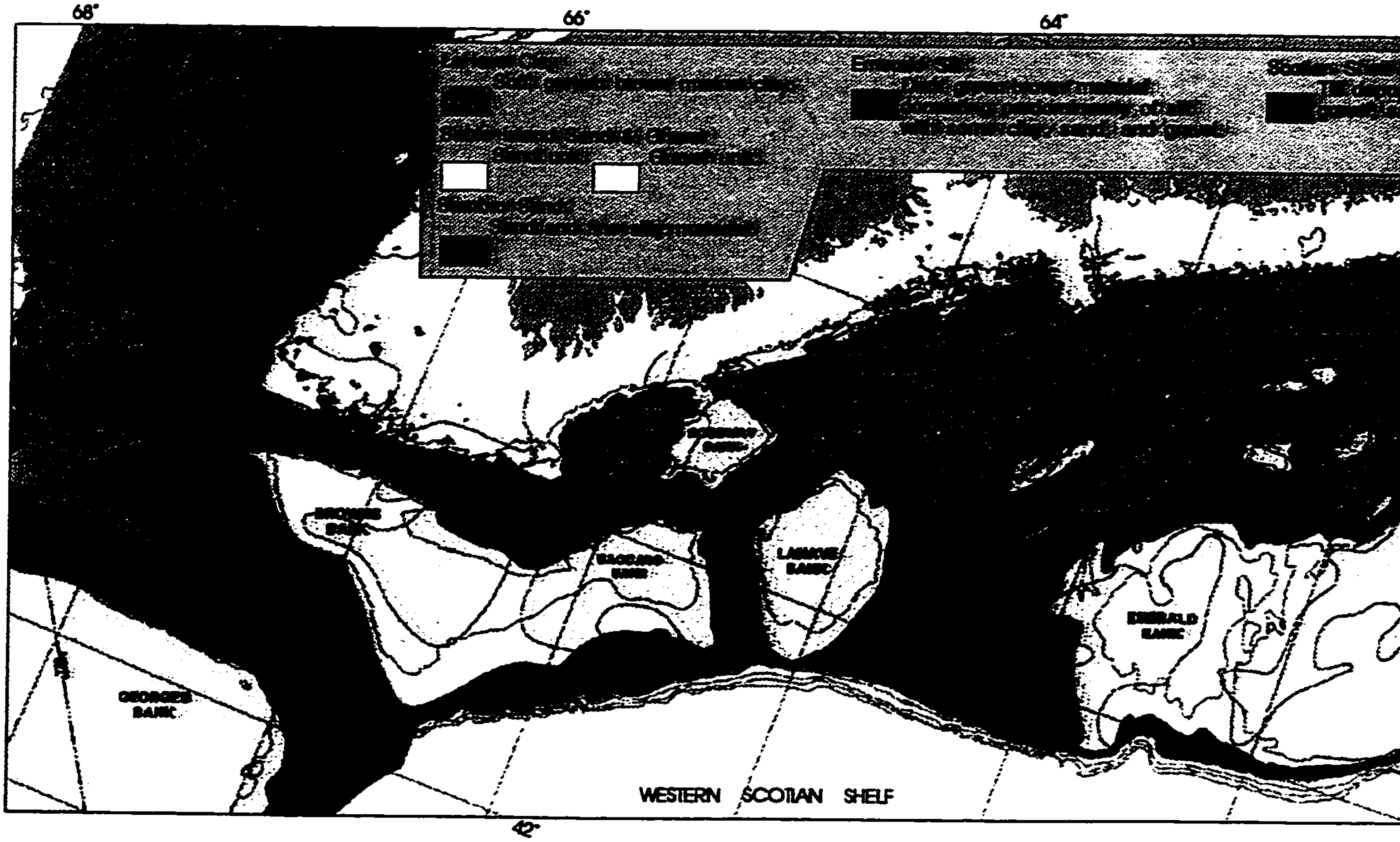



Figure 2.1 Quaternary geology of the Scotian Shelf, Atlantic Canada (Modified from King and Fader, 1986).

62°
60°
58°
46°
45°
43°
44°
Study Area
SABLE ISLAND BANK
SABLE ISLAND
BANQUEREAU
EASTERN SCOTIAN SHELF


Fader, 1986).

involves polar water from the western Arctic Ocean and temperate water from the North Atlantic Drift, a northern extension of the tropical Gulf Stream (Piper et al., 1990). Due to a complex morphology, surface water circulation on the Scotian Shelf is complex, though the general circulation is cyclonic including southwest movement of the Nova Scotia current along the inner shelf, and northeast flow of warmer water entrained by the North Atlantic Drift along the continental Margin (Piper et al., 1990). Tidal currents are strong on the outer banks due to accelerated flow as tidal waves move from deeper water into shallower water on the banks. These currents are minimal on the central and inner shelf (Amos and Knoll, 1987).

## 2.2: GENERAL MORPHOLOGY AND BEDROCK GEOLOGY OF THE SCOTIAN SHELF

Theories on the origin of the morphology of the eastern Scotian Shelf are highly dependent upon interpretations of the pre-glacial morphology, and bedrock geology. This is because some of the present day morphology was likely inherited from a pre-glacial landscape, therefore, what we see is not entirely glacial in origin.

### 2.2.1: General Morphology

The Scotian Shelf, extending west to east from Georges Basin and the Northeast Channel to the Laurentian Channel (Fig. 2.1), is approximately 160 000 km$^2$ in area and dips gently seaward at approximately 1°-2°. It varies in width from 125 km in the west to 230 km in the east, with the shelf edge break occurring at depth of about 140-180 m bsl (King and MacLean, 1976).

The general morphology of the Scotian Shelf consists of plateaus and mesas separated by incisions and basins which reflect erosion by both pre-Quaternary fluvial processes and glacial processes (King and Fader, 1986; King and MacLean, 1976; Grant, 1989). The Scotian Shelf has been subdivided into three regions: (1) a rough inner shelf (extends from ~25 km seaward to 100-120 m bsl); (2) a central/middle shelf composed of longitudinal and transverse depressions with adjacent banks (width varies from 20 km to 140 km west to east respectively. The west is characterized by broad basins, and the area east of 61°W is characterized by incisions of variable size and orientation; and (3) an outer shelf composed of shallow banks with intervening saddles (parallel to the shelf edge; 50-75 km wide), (King and MacLean, 1976; King and Fader, 1986). The inner Scotian Shelf area is included in the Atlantic Uplands of the Appalachian Region consisting primarily of igneous and metamorphic rocks, while the middle and outer shelf are part of the submerged Atlantic Coastal Plain Region, underlain by Mesozoic and Cenozoic Strata (Williams et al., 1972).

Initially, it was believed that the overall morphology of the Scotian Shelf is bedrock controlled, and representative of a coastal plain environment displaying varying degrees of maturity resulting from subaerial erosion, with glacial landforms and deposits only modifying this (King and Fader, 1986). The interpretation that the middle bank depressions are coastal plain lowlands and pre-glacial in origin (King et al., 1974; King and Fader, 1986), is challenged by the idea that they represent significant localized glacial scour of a pre-glacial landscape (Grant, 1989). More recent work on the Scotian Shelf suggests more extensive glaciation (i.e. Mosher et al., 1989; King, 1994; King, 1996).

The inner Scotian Shelf was further subdivided by Stea (1995) who proposed a 'Zonal Concept'. The inner Scotian Shelf, essentially an extension of the land area into the offshore, was subdivided into five terrain zones based on bedrock type and structure, patterns of glacial erosion and deposition, and episodes of sea-level rise and fall (Fig. 2.2): (1) Scotian Shelf End-Moraine Complex (30-40 km offshore; King et al., 1972); (2) Basin Zone (>145 m bsl; Stea, 1995); (3) Outcrop Zone (80-120 m bsl; Forbes et al., 1991); (4) Morainal Zone (80-120 m bsl; Stea et al., 1993 a, b); and (5) Truncation Zone (90 m bsl to shoreline; Forbes et al., 1991; Stea et al., 1993 a, b). The Scotian Shelf End Moraine Complex and Basin Zone are areas of glacial deposition, while the Outcrop Zone is lacking in surficial sediments. East of the Outcrop Zone is the Morainal Zone which consists of till ridges overlying bedrock, including the Truncated Emerald Silt Subzone, an area of outcropping Emerald Silt with a planar erosional surface (Stea et al., 1994; Stea, 1995). Finally, the Truncation Zone, which is further divided into: (1) the Valley Subzone; (2) the Transition Subzone; (3) the Platform Subzone; and (4) the Estuarine Subzone, is a region of muted topography and planar erosional surfaces truncating bedrock and surficial deposits (Stea et al., 1993 a, b; Stea et al., 1994; Stea, 1995; see Stea, 1995 for a detailed description of each zone). Some of these zones may apply to the inner shelf areas of the eastern Scotian Shelf study area.

### 2.2.2: Bedrock Geology

The Scotian Shelf comprises a large portion of the Scotian Basin, part of an accreted wedge of Mesozoic-Cenozoic sediments deposited during the development of the North Atlantic ocean basin (Wade and MacLean, 1990). These sediments were deposited on the eastern flank of the Appalachian Orogen which is composed of the Meguma Group, low grade Cambro-Ordovician metasedimentary rocks (Wade and MacLean, 1990).

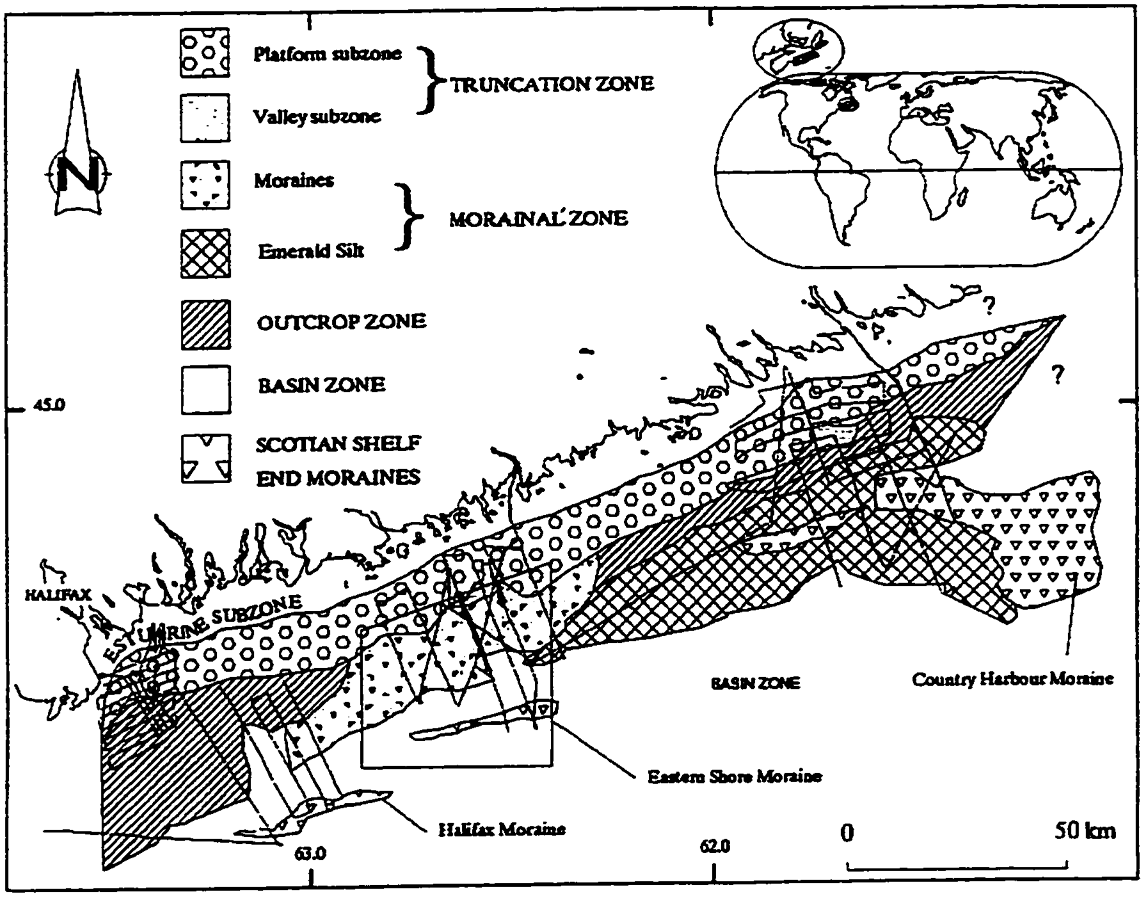



Figure 2.2 Subdivision of the terrain zones which define unique bottom morphology and seismostratigraphy on the inner Scotian Shelf. From Stea (1995). (See Stea, 1995 for details; Black lines represent tracklines used for Stea, 1995, study.)

Bedrock types outcropping on the Scotian Shelf, and immediately underlying the Quaternary sediments include those from the Meguma Group (Acoustic Basement) in the vicinity of the inner shelf, and Cretaceous and Tertiary sediments extending from the erosional edge of these sediments to the shelf edge, in the vicinity of the middle and outer shelf (King and MacLean, 1976; Fig. 2.3). These bedrock types would have been the most influential control on glacial erosional features. The inner shelf areas affected by crystalline bedrock would not be eroded as easily as Tertiary and Cretaceous sediments. This would be reflected by differences in the inner shelf bathymetry compared with middle and outer shelf bathymetry.

The Gully Group overlies basement, Triassic, Jurassic, and Early Cretaceous rocks, and is composed of the uppermost sedimentary strata underlying the Quaternary sediments. The Gully Group, represents deep water sedimentation form the late Cretaceous until the Tertiary, and consists of the Dawson Canyon Formation (grey marine shale or mudstone, occasional thin beds of siltstone, sandstone and limestone), the Wyandot Formation (chalks, grading upslope into calcareous shales and mudstones), and the Banquereau Formation (mudstone grading upwards into sandstone and conglomerate) which range in age from Late Cretaceous to Tertiary. They represent Late Cretaceous global transgression, followed by sea-level fluctuations during the mid-Tertiary, and pre-Pleistocene regression (Wade and MacLean, 1990). The Banquereau Formation consists of all pre-glacial sediments above the Wyandot Formation (Late Cretaceous). Banquereau Formation shows seaward progradation (McIver, 1972; Wade and MacLean, 1990) and is affected by shelf edge unconformities, and subaqueous unconformities which likely resulted from sea-level fluctuations. Most importantly, are the multiple incision cut and fill sequences which probably formed subaerially during the Tertiary (King and

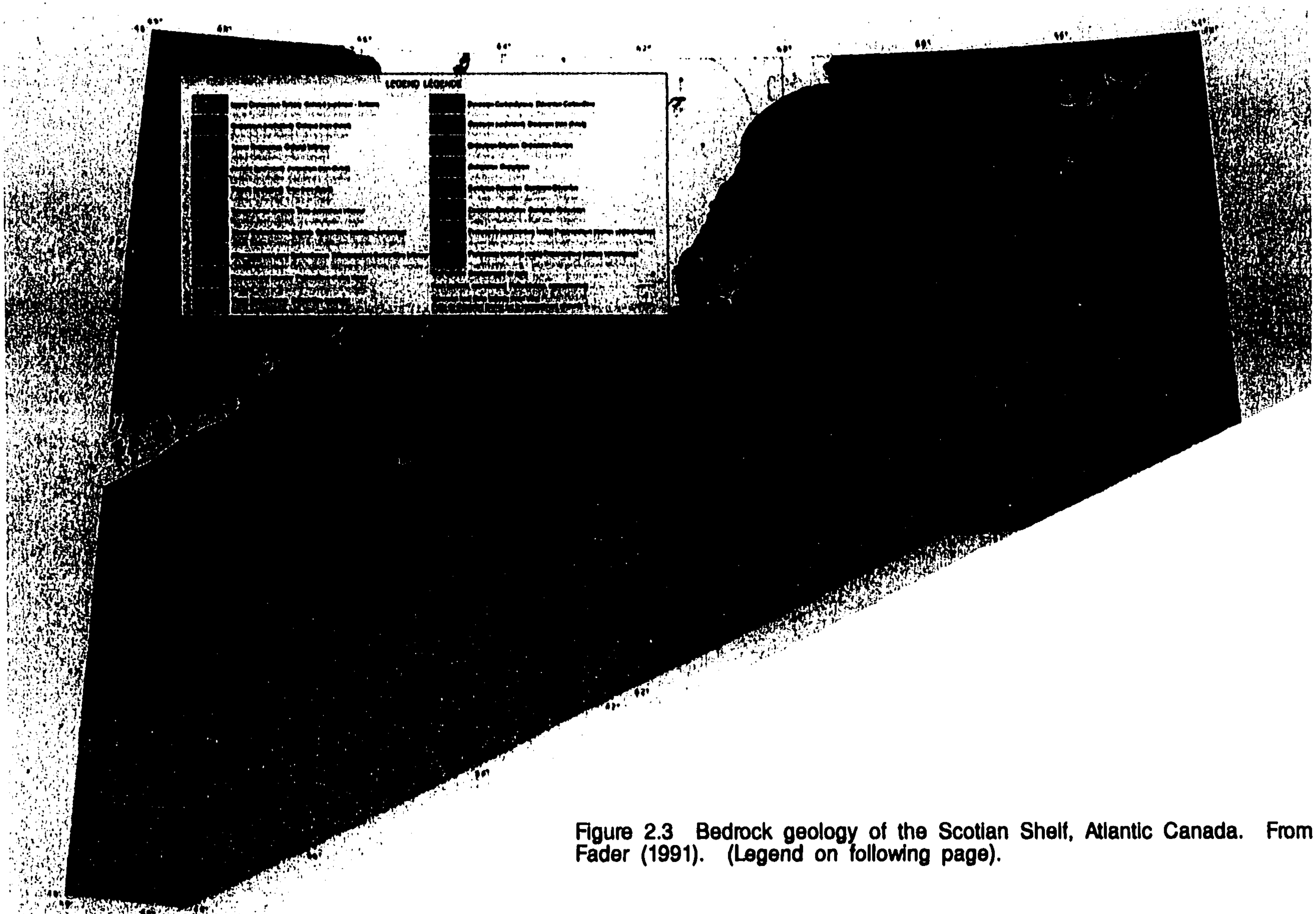


Figure 2.3 Bedrock geology of the Scotian Shelf, Atlantic Canada. From Fader (1991). (Legend on following page).

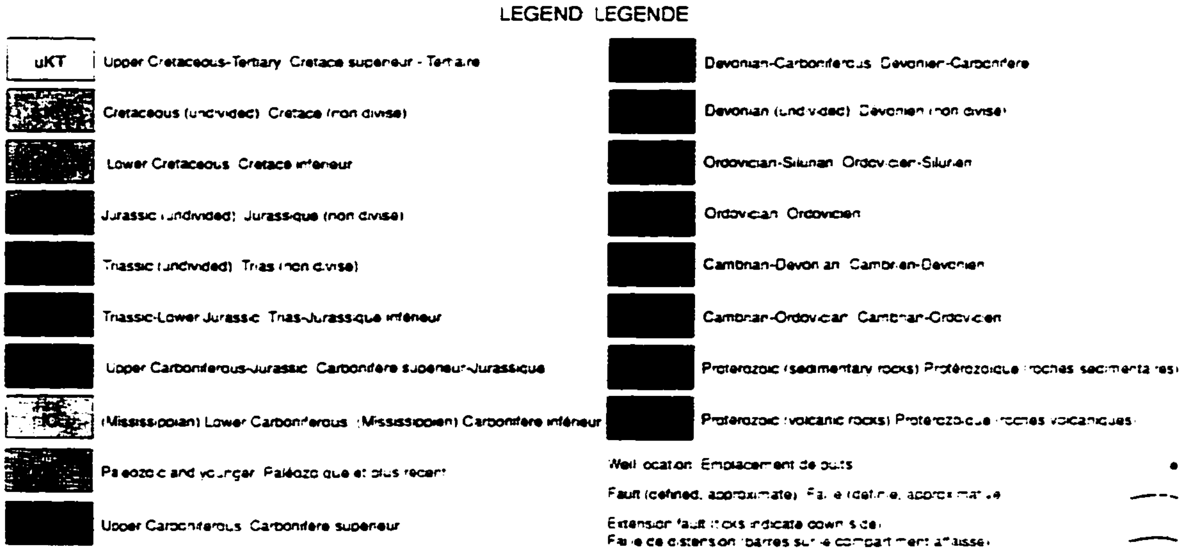



**Figure 2.3 Legend to bedrock geology map. From Fader (1991).**

MacLean, 1976; Wade and MacLean, 1990). These cut and fill sequences likely had a strong influence on glacial processes. The morphology of the Appalachian area along the inner shelf likely began development as a result of drainage system evolution beginning in the Late Jurassic and Early Cretaceous times (King and MacLean, 1976).

Tertiary strata form a discontinuous veneer over Cretaceous strata. Over most of the Scotian Shelf, Cretaceous and Tertiary rocks lie directly on Meguma Group rocks with Cretaceous sediments generally outcropping in the basin areas (King and MacLean, 1976). Tertiary shelf-edge erosion cuts into the Wyandot Formation to the south, and Tertiary incisions have cut into both the Dawson Canyon and Wyandot formations, particularly to the east where the Wyandot Formation is the thickest, and also shows evidence of an eastward shifting depocentre which resulted in southeastward progradation during the Late Cretaceous (Wade and MacLean, 1990).

### 2.2.3: Unconformities

Four major unconformities on the Scotian Shelf were mapped by King et al. (1974). They are: (1) the Early Cretaceous unconformity between Jurassic and Cretaceous strata at the western end of the shelf in the Georges Basin area (it is not known whether the unconformity is restricted to the basin margins or if it extends into the basin); (2) the Late Cretaceous unconformity between the Wyandot Formation and basal Banquereau strata which extends over most of the Scotian Shelf (it is not known if this unconformity is shelf wide or restricted to basin margins and structural highs); (3) the Early Tertiary unconformity expressed as incision cut and fill which extends across the whole Scotian Shelf and divides the Banquereau Formation into two units; and (4) the Late Tertiary-

Pleistocene unconformity, also expressed as incision cut and fill in some areas.

The Early Tertiary and Late Tertiary-Pleistocene unconformities are the most important pre-glacial influence on the present-day morphology of the Scotian Shelf. King et al. (1974) suggest that the Early Tertiary and Late Tertiary-Pleistocene unconformities were primarily subaerial and fluvial in origin, although some deeper incisions may be the result of submarine canyon development along the shelf edge, and were subsequently modified by glacial processes. The Late Tertiary-Pleistocene unconformity generally follows the pattern of an Early Tertiary drainage system (unconformity). The origin of the unconformities may be related to: (1) eustatic sea level lowering; and/or (2) tectonism related to variations in the height and width of the oceanic ridge systems resulting in sea level variations (King et al., 1974; Fig. 2.4).

## 2.3: THE QUATERNARY GEOLOGY OF THE SCOTIAN SHELF

Pleistocene glaciations were the last major geologic events to affect the Scotian Shelf. Much of what is observed on the Scotian Shelf today is the result of glacial erosion and deposition on a pre-glacial continental shelf/coastal plain environment. In some areas, particularly the banks, these deposits were modified and redeposited by post-glacial and Holocene processes, including marine transgression following the last glacial cycle. A detailed account of the Quaternary stratigraphy of the eastern Scotian shelf is provided first. This is the basis on which the interpretations for this study are founded.

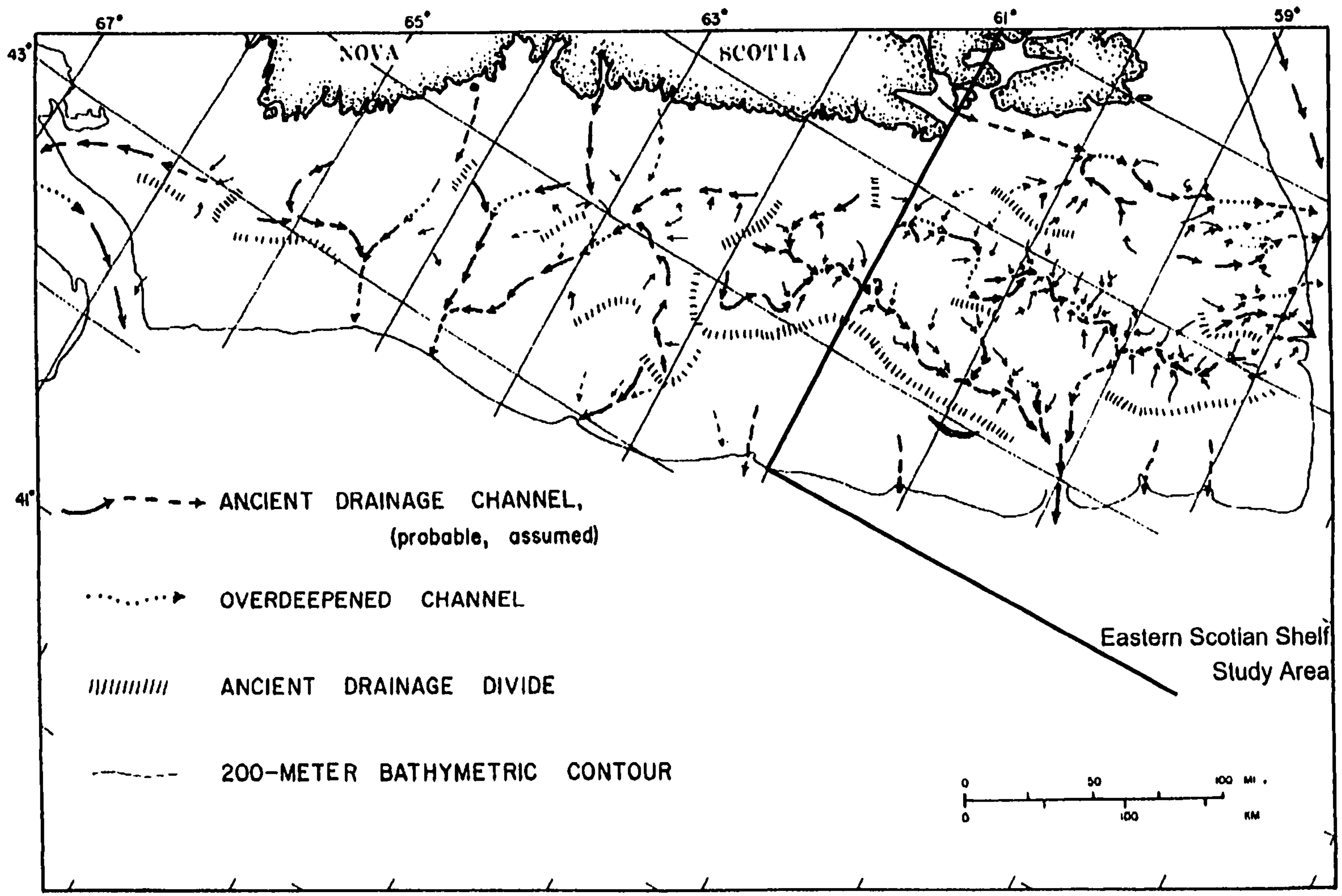



Figure 2.4 Postulated Tertiary fluvial drainage system crossing the Scotian Shelf at the Late Tertiary-Pleistocene unconformity. From King et al. (1974).

### 2.3.1: Terminology

The following is a description of terms which will be used for this study. In order to account for alternative origins for deposits displaying an incoherent acoustic signature on seismic records, the following interpretive terms will be used: (1) if a deposit is believed to be subglacial in origin the term till will be used; (2) if a deposit is believed to be related to glacial processes, but not necessarily subglacial (i.e. ice marginal deposits including such features as debris flows, push moraines, end moraines, till deltas, ice contact and grounding line fans etc.) the term glacial diamict will be used; (3) if no interpretation can be made, the term incoherent acoustic facies will be used.

The terms 'till tongue' and 'lift-off moraine' will be used to describe all features displaying the same seismic character and geometries as described in the discussions for till tongues and lift-off moraines, sections 2.3.3a and 2.3.3b respectively. Though the names may indicate a genetic origin, they are merely used for descriptive purposes unless otherwise stated. The terms till tongue and lift-off moraine are well recognized throughout literature pertaining to the Scotian Shelf and this will retain some consistency when discussing similar geomorphic features.

### 2.3.2: Quaternary Stratigraphy of the Scotian Shelf

The Quaternary stratigraphy of the Scotian Shelf which is described in this section forms the basis of all seismic record interpretations from this study. Here, the formal names, lithologies, and acoustic characteristics of each well established unit found on the Scotian Shelf are described in detail.

King (1970) first described and formalized the Quaternary stratigraphy of the Scotian Shelf (Fig. 2.1; Table 2.1). The first comprehensive study on the Quaternary geology of the Scotian Shelf, Atlantic Canada, was that of King and Fader (1986). This work, based on seismostratigraphic interpretation utilizing combined sediment sampling and acoustical techniques, was subsequently modified in a more recent work (King, et al., 1991). The following overview of the Scotian Shelf Quaternary stratigraphy is based on King and Fader (1986), and King et al., 1991) with the exception of referenced papers. Table 2.1 summarizes the following description. The sediments are presented in chronological order.

*Scotian Shelf Drift (Till)*

Scotian Shelf Drift is an acoustically incoherent unit which conformably overlies the bedrock surface and underlies the subsequent succession of surficial deposits. It generally occurs as a continuous blanket of uniform thickness (10-15 m), although moraines contain thicker deposits (>100m). It is found to depths of 500 m bsl on the upper continental slope. Deposits are sparse on the inner shelf and shallow bank areas as a result of erosion by Late Pleistocene-Holocene marine transgression.

Scotian Shelf Drift is characterized as a very dark grayish-brown, massive, cohesive, poorly sorted mixture of sand, silt, clay, and gravel (diamicton) ranging in size from granules to boulders. The unit is devoid of clasts and very fine grained in some areas. Wedge shaped bodies of diamicton or "till tongues (section 2.3.3a)" are located in the stratigraphic transition zone between Scotian Shelf Drift and Emerald Silt on the distal side of moraines and on the flanks of basins and depressions. Similar features have also been recognized on the Scotian Slope ( Piper, 1988; Mosher et al., 1989). The

| Formation | Lithostratigraphy | Thickness | Seismostratigraphy |
| --- | --- | --- | --- |
| LaHave Clay | Greyish brown, soft, silty, clay grading into clayey silt; confined mainly to basins and depressions of the shelf; derived by winnowing of glacial sediments on banks and transported to basins; time equivalent to Sable Island Sand and Gravel, and Sambro Sand on banks. | 0-70 m | Generally transparent without reflections. Some weak continuous coherent reflections in base of section becoming stronger in nearshore sandy facies and on Grand Banks of Newfoundland. |
| Sable Island Sand and Gravel | Fine to coarse, well sorted sand grading into subrounded to rounded gravels; unconformity overlies Emerald Silt and Scotian Shelf Drift, and derived from these deposits through reworking during Holocene transgressions above 120 m present depth; time equivalent to LaHave Clay in basins. | 0-50 m generally veneer | Highly reflective seabed; generally closely spaced continuous coherent reflections if deposit is of sufficient thickness to resolve. |
| Sambro Sand | Silty sand grading locally to gravelly sand and well sorted sand; deposited sublittorally with respect to the Pleistocene shoreline below 120 m present depth; time equivalent to basal LaHave Clay and upper Emerald Silt, Facies B. | 0-20 m generally veneer | Similar to Sable Island Sand and Gravel. |
| Emerald Silt Facies B | Darkish geryish brown, poorly sorted clayey and sandy silt with some gravel; poorly developed rhythmic banding; proglacial in origin. | 0-40 m | Medium to low amplitude continuous coherent reflections, and some degree of ponded sedimentational style. |
| Emerald Silt Facies A | Darkish greyish brown, poorly sorted clayey and sandy silt, some gravel; well developed rhythmic banding; subglacial in origin; time equivalent to parts of Scotian Shelf Drift. | 0-100 m | High amplitude continuous coherent reflections, highly conformable to substrate irregularities. |
| Emerald Silt Facies C | Not well sampled | 0-100 m | Discontinuous coherent reflections; transitional between Emerald Silt Facies A and glacial till |
| Scotian Shelf Drift | Very dark greyish brown, cohesive glacial till composed of poorly sorted sandy clay and silt with variable gravel. | 0-100 m | Incoherent reflections, sometimes with scattered point source reflections; variations in grey tones may represent consolidation differences. |

Table 2.1 Quaternary stratigraphy of the Scotian Shelf. Formations listed in sequence as they are commonly found from the uppermost sediments to the bedrock surface. Modified from King and Fader (1986).

surface of the Scotian Shelf End Moraine Complex (30-40 km off-shore, up to 120 m high) displays evidence of relict iceberg furrows (~2-3 m deep and up to 200 m wide). Scotian Shelf Drift found in Scotian Shelf basins displays a hummocky/ridge-like morphology interpreted as lift-off moraines (section 2.3.3b). Finally, King (1969) mentioned the existence of bedrock controlled moraine ridges south of the main moraine complex, and Stea et al. (1992; 1994), describe a zone of morainal ridges on the inner shelf, north of the Scotian Shelf End Moraine Complex.

*Emerald Silt Formation*

The Emerald Silt Formation consists of three distinct facies (A, B, and, C) and varies in thickness from a few meters to over 100 m, with the thickest deposits occurring in basins and as channel fill. Facies descriptions for the Emerald Silt Formation are as follows:

*Emerald Silt Facies A (Ice-Proximal Glaciomarine)*

Emerald Silt Facies A, the lowermost unit of the Emerald Silt Formation, displays high amplitude, continuous coherent reflections on seismic records. These reflections are smooth, parallel, and closely spaced sometimes maintaining their integrity over 100 km. It's maximum observed thickness is 80 m. It was deposited at the same time as parts of the Scotian Shelf Drift with transitional, abrupt, and onlapping contacts. Emerald Silt Facies A is conformable with the underlying irregular till and bedrock surface, and in some areas it interbeds with the Scotian Shelf Drift (till tongues).

Emerald Silt Facies A is a dark grayish-brown, poorly-sorted, clayey to sandy silt unit, with variable amounts of gravel and some evidence of dropstones (point source reflections). It displays strong rhythmic banding (alternating layers of silt and clay) with individual couplets ranging in thickness from 1 to 3 cm. Facies A contains a low abundance of foraminifera and some in-situ mollusks, interfingers with till tongues, has closely spaced continuous coherent reflectors, and is draped across the subsurface relief. All of these factors lead to an interpretation of this unit as an ice-proximal glaciomarine deposit. The lower half of Emerald Silt Facies A lacks shells, microfauna, and bioturbation, indicating a quiet depositional environment, possibly beneath a pinned ice shelf, while the upper half of the section shows evidence of iceberg scouring indicating open water conditions. Modes of formation for this unit include deposition from overflow plumes at a grounding line and/or rain out from the undersurface of an ice shelf.

*Emerald Silt Facies B (Ice-Distal Glaciomarine)*

Emerald Silt Facies B displays medium to low amplitude, continuous coherent reflections on seismic records. The reflections are smooth, parallel, and closely spaced, but do not group into dense bands and cannot be traced for long distances as they tend to fade laterally. Facies B directly overlies Facies A in shelf basins and depressions and is only occasionally found on the banks. It may have originally been more extensive, but was eroded and reworked during the late Wisconsinan marine transgression of the banks. Facies B displays a conformable character in the lower part, and a ponded character in the upper part and has a maximum observed thickness of 40 m.

Emerald Silt Facies B is a dark grayish-brown, poorly sorted, clayey to sandy silt, containing variable amounts of sand and gravel displaying weak to strong rhythmic layering. Shells and shell fragments are common, as are bioturbation, foraminifera, and dropstones, all of which are more abundant in Facies B than in Facies A. The contact of Facies B with underlying Facies A is distinct with a decrease in reflection amplitude, whereas its contact with the overlying LaHave Clay (marine clay) is conformable to unconformable. Emerald Silt Facies B is characterized by an increase in planktonic foraminifera, the abundance of dropstones, fauna, and bioturbation when compared to Facies A. These characteristics are interpreted to reflect a decrease in the influence of sediment plumes and an increase in ice-rafting with distance from the ice margin. The abrupt contact of Facies B with underlying Facies A is thought to indicate a rapid retreat of the grounding line, or possibly the removal of an ice shelf following deposition of Facies A–a change from a floating ice shelf to drifting ice in an open marine environment. This is further supported by a reduction in the rate of sedimentation of Facies B. The change in character of Emerald Silt Facies B from a conformable to a ponded style of sedimentation has been interpreted as marking a transition from ice-proximal to ice-distal conditions.

*Emerald Silt Facies C*

Emerald Silt Facies C displays discontinuous, coherent reflections on seismic records. It is a sparse deposit sometimes observed in transitional areas between Emerald Silt Facies A and Scotian Shelf Drift. Emerald Silt Facies C has a texture similar to Facies A and B, and displays strongly laminated beds which are sometimes steeply dipping. This unit reaches a maximum observed thickness of 20 m and is interpreted to be a subglacial deposit due to its stratigraphic location between Scotian Shelf Drift (till) and Emerald Silt Facies A.

*Sambro Sand*

Sambro Sand is highly reflective with closely spaced continuous coherent reflections. Acoustic resolution can be poor with this deposit due to its highly reflective nature. It is a silty sand, grading locally to gravelly sand, and well-sorted sand. It was deposited in association with the Late Wisconsinan low sea level stand, approximately 120 m bsl, at the same time as the upper portion of Emerald Silt Facies B and the basal portion of LaHave Clay. Sambro Sand was likely formed by the reworking of Emerald Silt and Scotian Shelf Drift along basin margins, while bank areas were subject to subaerial erosion and deposition of Sable Island Sand and Gravel.

*Sable Island Sand and Gravel*

Sable Island Sand and Gravel is acoustically similar to Sambro Sand. It is a well-sorted fine to coarse sand coarsening to subrounded and rounded gravels. It unconformably overlies Emerald Silt and Scotian Shelf Drift and was likely derived from these deposits through reworking during Holocene marine transgression above 120 m bsl. Sable Island Sand and Gravel was deposited on the banks during the Late Pleistocene/Holocene at the same time Sambro Sand was deposited along basin margins and LaHave Clay was deposited in basins.

*LaHave Clay*

LaHave Clay is characterized by a discontinuous transparent seismic signature, and is a grayish-brown, soft silty clay, grading into clayey silt. This deposit is confined mainly to the shelf basins and depressions. It was likely derived by the winnowing of glacial sediments from the banks, with subsequent transportation to basin areas. LaHave Clay was deposited during the Holocene at the same time as Sable Island Sand and Gravel and Sambro Sand.

### 2.3.3: A Discussion on Moraine Types and Their Relevance to Glacial Ice Terminus Configurations on the Scotian Shelf

The existence of an ice shelf or tidewater margin on the Scotian Shelf is important for determining the origin of the Eastern Scotian Shelf morphology. It has been proposed that extensive channelization (incisions) may be a product of meltwater erosion at tidewater termini (McLaren, 1988; King, 1993; King, 1994; King, 1996). The results of this study will show that this may not necessarily be true.

Conclusions regarding the existence of an ice-shelf or tidewater margin on the Scotian Shelf must consider necessary conditions for an ice-shelf to exist. Powell (1984; 1988) explains that temperate marine ending glaciers with floating termini are unlikely to exist. He explains this as a response of temperate ice at pressure melting point with a low tensile strength due to intercrystalline water films. This facilitates easier fracture propagation and allows ice to calve readily. In addition, rapid spreading of temperate ice could cause surface and basal crevasses to join also causing rapid calving (Weertman, 1973; Rasmussen and Meier, 1982). Others disagree with this conclusion and believe temperate ice can form ice-shelves (King and Fader, 1986; Alley, et al., 1989), however specific protective conditions are necessary and are not well understood. Alley et al., (1989) point out that ice shelf disintegration may be reduced by lateral constraint, protection from bottom melting, and high discharge from outlet glaciers and ice streams.

It is unclear whether or not ice was always temperate on the Scotian Shelf, as a new interpretation by Stea et al. (1994) calls for the existence of an area of frozen bed conditions which resulted in the preservation of what they consider to be a possible relict bedrock surface (Outcrop Zone; Fig. 2.2).

Extensive ice shelves exist in polar regions such as the ice shelves of Antarctica. Some predominantly temperate ice shelves exist (i.e. Jakobshavns Isbræ, Greenland), however these ice shelves are small, supported by rapid discharge from large outlet glaciers and ice streams, and generally lack stability (Fahnestock, 1996). In addition, the ice shelf extending off Jakobshavns Isbræ is within the protective environment of a fjord, unlike the open and unprotected environment of the Scotian Shelf.

Moraines may hold critical information concerning the configuration of ice termini on continental shelves, and hence their stability which is important to proposed origins of the eastern Scotian Shelf incisions. King and Fader (1986), proposed models for the origin of two different types of moraines thought to have formed under an ice-shelf environment on the Scotian Shelf: (1) till tongues (wedge shaped features associated with regional, subglacial, ice-shelf moraines; i.e. the Scotian Shelf End Moraine Complex), and (2) lift-off moraines (hummocks identified in basin areas).

#### *2.3.3a: Till Tongues*

The till tongue model proposed by King and Fader (1986), and King et al. (1991), is used to explain lateral extensions of wedge-shaped bodies of glacial diamict (Scotian Shelf Drift?) interbedded with a conformable stratified ice-proximal glaciomarine deposit (Emerald Silt Facies A). Till tongues can occur in simple wedge or complex stacked succession forms. They are thought to have been formed as part of a grounding line system associated with a floating terminus near the margins of marine ice-sheets. Till tongues tend to be confined to the flanks of basins on the Scotian Shelf and on the distal (seaward) side of the Scotian Shelf Moraine Complex (King and Fader, 1986). However, later surveys identified similar features at the edge of continental shelves and on upper continental slopes (King, et al., 1991; Mosher, et al., 1989; Piper, 1988). Till tongues are composed of a root (area of thickest glacial diamict, usually associated with Scotian Shelf Drift deposits), and a feather edge (area where the till tongue thins and is interbedded with glaciomarine deposits), and appear to be restricted to deep water regimes (~75 m-1000 m bsl). However those in water depths greater than 600 m bsl may have been influenced by large debris flow influence in their origin (King. et al., 1991). Till tongues average in width from 10 to 30 km, with maximums up to 300 km, and from 100 m to over 25 km in length, and in the root area they are from a few meters to 100 m thick (King and Fader, 1986; King et al., 1991). The rate of thickening with distance varies from a maximum of 400 m over 700 m to 8 m over 3000 m (King and Fader, 1986). Meltwater channels are generally not observed in association with these features, although there is a possibility that channelized subglacial distributary systems may have been identified at the base of some till tongues (King et al., 1991), though at a much smaller scale than those observed on the eastern Scotian Shelf. The surfaces of many till tongues are covered with lift-off moraines (section 2.3.3b).

According to King and Fader (1986), and King, et al. (1991), the formation of till tongues relates to grounding-line fluctuations of a thin, neutral to negatively buoyant, low gradient ice sheet (Fig. 2.5a). This model implies the existence of an ice-shelf by explaining the alternating till tongue and glaciomarine deposits in terms of successive advances and retreats of a grounding line (King and Fader, 1986; King et al., 1991). This interpretation is challenged, as many researchers believe these enigmatic features may have been formed by other processes. For example: (1) Powell (1984; 1988) described features similar to till tongues as push/squeeze and debris flow deposits at tidewater glacier grounding lines; (2) Boulton (1986) considered till tongue features to be a result of push moraines formed in association with ice contact fans; (3) Piper (1988) and Mosher et al. (1989) observed similar features on the Scotian Slope and attributed them to debris flows and/or bulldozed till; (4) Vorren et al. (1989) related these features to grounding-line fans; (5) Alley (1989) invoked formation by subglacial deformation causing a pro-glacial till delta; and finally, (6) Syvitski (1991) related them to debris flows or outwash fans, although he states that moraines are tidewater glacier features, ignoring the possibility that these features may also be associated with the grounding line between an ice-sheet and an ice-shelf.

King et al. (1991), modified the original till tongue model (King and Fader, 1986; Fig. 2.5a), which explained till tongue formation as a purely subglacial phenomenon. This was upon the recognition that Emerald Silt Facies A (section 2.3.1), and ice berg furrows on the surface of some moraines, provide evidence for open-water conditions during till tongue formation. They proposed the sharp contact between Emerald Silt facies A and B represented the disintegration of a widespread ice-shelf. Evidence for open-water conditions during the formation of the upper half of Emerald Silt Facies A (see Emerald Silt Facies A description, section 2.3.2), caused King et al. (1991) to propose that the contact between Emerald Silt Facies A and B may indicate an abrupt retreat of the grounding-line of a proximal ice-sheet. Nevertheless, they still maintain that a pinned ice-shelf existed during deposition of the lower portion of Emerald Silt Facies A, which shows evidence of deposition in an

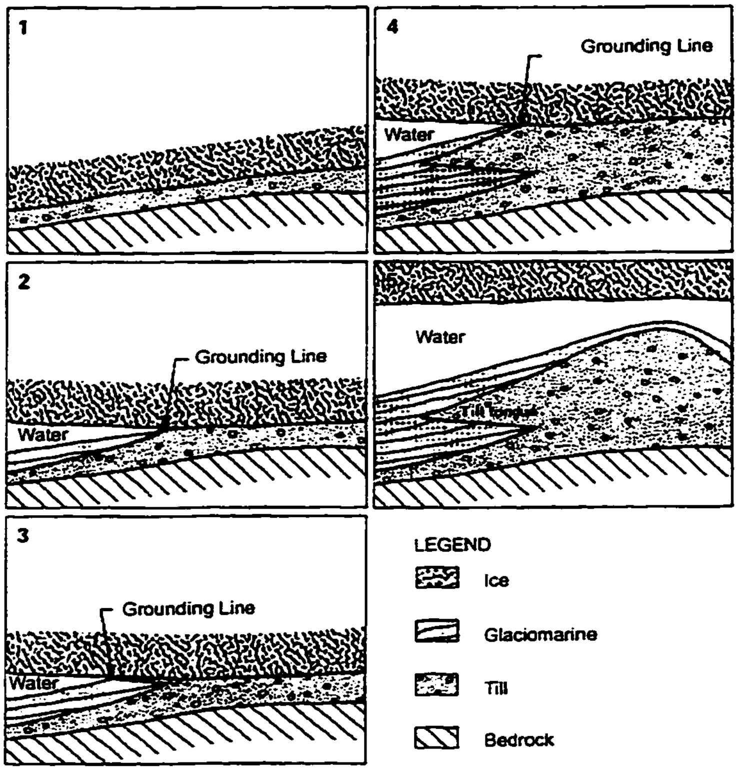



Figure 2.5a The original model for till tongue formation as a result of a migrating grounding line. In the original model it was hypothesised that the till tongue deposits were composed of till deposited by direct ice contact. From King et al. (1991), after King and Fader (1986).

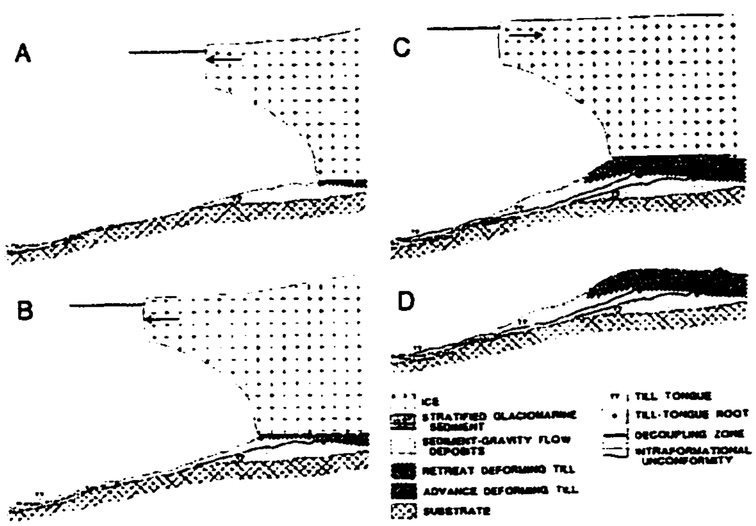



Figure 2.5b Modified till-tongue model which incorporates alternate modes of formation or modification including sediment-gravity flow (debris flow) deposits, and deforming till deposits. Stages A and B represent deposition of till-tongues by advancing ice, stage C represents early retreat with deposition in the decoupling zone, and stage D shows till-tongues as a part of the foreset development of a till delta. From King (1993).

extremely quiescent environment (See Emerald Silt Facies A description, section 2.3.1).

The modified till tongue model (King et al., 1991), acknowledged alternatives proposed for the formation of till tongue stratigraphy. They recognized the potential importance of a deformable bed (Alley et al., 1989) for till tongue formation, and that distal portions (feather edge) of till tongues may not be entirely subglacial in origin. For instance, slumping was recognized as probable at the distal end of till tongues possibly forming pro-glacial aprons as a result of gravity flows (Fig. 2.5b). In addition, if a slope is steep (i.e. the continental slope region), till tongues, as described in the original model, would likely not form as mass failure would become the dominant process. Mass movement was not considered as the primary mode of till tongue formation on the Scotian Shelf as features associated with shearing and rotational movement are not observed to exist (King et al., 1991), though such features would be difficult to resolve on seismic records and through sampling. This was further emphasized by Stea (1995), who showed a contrast in the morphology of mudflows (scarp at the head of the failure, evacuation zone, and toe bulge at the end) and till tongues. However, Sejrup et al. (1996), interpreted features on the Norwegian Shelf, displaying till tongue morphology, as debris flows, and Piper (1990), described core sediments as a product of debris flow deposits which were derived from Emerald Silt Facies A. King (1993), interpreted the composition of till tongues in terms of gravity flow deposits, above which lies a subglacial deposit of deforming till (as per Alley et al., 1989) derived from the decoupling zone at the base of the ice (Fig. 2.5b). He interpreted till tongues as part of the foreset development of a till delta with the till tongue root area being the topset portion of a delta. This revised model allows for ice advance into deeper water at shelf edges by providing a platform of thicker sediments (thus lowering relative sea level) in the form of stacked till successions (King, 1993; Fig. 2.5b).

From this review, it is clear that till tongue origin is highly controversial, and many factors must be considered for the origin of this type of stratigraphy. However, the fact that this type of stratigraphy is associated with Emerald Silt Facies A (ice proximal glaciomarine deposits) indicates that ice was in a delicate state of floatation or near floatation, and the grounding line may have been prone to fluctuations during the time of till tongue formation. The identification of lift-off moraines (2.3.3b) on the surface of till tongues lend further evidence to the ungrounding of an ice sheet near the time of till tongue formation.

### *2.3.3b: Lift-off Moraines*

Lift-off moraines are generally identified as mounds of retreat till (Scotian Shelf Drift) observed on the upper surface of till blankets at the base of deep basins and in incisions to the east (King and Fader, 1986; King et al., 1991; this study). They occur as parallel to subparallel ridges which often bifurcate and are formed in association with ice-proximal glaciomarine deposits (Emerald Silt Facies A). Their morphology is characterized by heights which vary from a few meters to 20 m, and widths from 20 to 150 m, at a spacing of 30-400 m (King et al., 1991).

King and Fader (1986), explain that 'lift-off' moraines are formed simultaneously in fields during ice recessional phases when a grounded ice-sheet lifts off the seabed. They conclude that structural controls in the form of fractures or crevasses in the ice above, into which meltwater and till can collect, control the formation of lift-off moraines (Fig. 2.6). However, King et al. (1991),

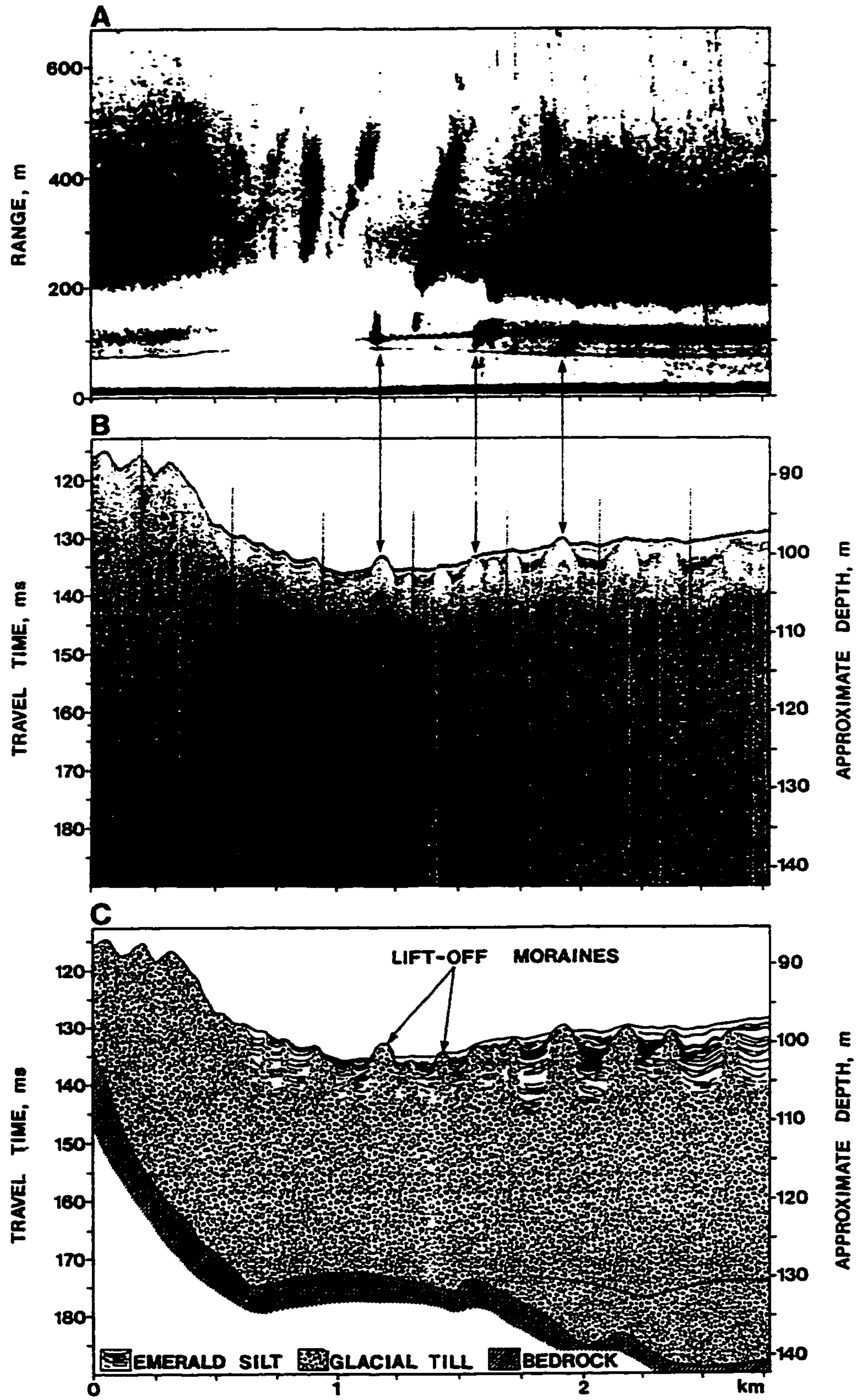



Figure 2.6 Lift-off moraines. A) Sidescan sonar image of lift-off moraine ridges as seen at the sea bed; B) Huntec DTS seismic profile; C) interpretation of Huntec DTS profile. From King and Fader (1988).

later express uncertainty as to whether or not lift-off moraines were formed individually or simultaneously. It is not clear if Emerald Silt Facies A is interbedded with the moraine ridges, or if the appearance is an artifact due to loss of seismic resolution along steep flanks (King et al., 1991).

Syvitski (1991), disputed the King and Fader (1986), explanation for lift-off moraines, and proposed several other explanations for these features; push-morainal banks, Rogen moraines, De Geer moraines, and/or push-squeeze deposits (i.e. Powell, 1988), all formed by active ice. An alternative interpretation to simultaneous deposition of Emerald Silt and diamict is time transgressive deposition. It is possible for this style of sedimentation to occur between ridges by the settling out of sediment from meltwater overflow plumes as ice retreats from each moraine. This would be similar to development of De Geer moraines. King and Fader (1986), pointed out, there is a lack of deformation within the glaciomarine deposits (Emerald Silt Facies A), an expected product of a push mechanism of formation.

Vogt (1997), explained features resembling lift-off moraines in the Norway basin and Eastern Iceland Plateau as a result of diapirism or dewatering processes, requiring a sediment density inversion (lower units are less dense than upper units). Vogt (1997), observed features on seismic reflection records which appear almost identical in form and acoustic signature to lift-off moraines, however they are much larger in scale (500-2000m spacing, >50 m high), located in water depths >3000 m, and associated with different types of sediments (Miocene biosiliceous ooze overlain by Pliocene-Pleistocene glacigenic clays). Aside from scale, water depth, and sediment type, one important distinguishing difference between the lift-off moraines of King and Fader (1986), and the diapirism/dewatering features of Vogt (1997), is evident on sidescan sonar images. Vogt's (1997), features display a low-relief hummocky morphology, while lift-off moraines as per King and Fader (1986), display a linear ridge-like morphology (Fig. 2.6). No hummocky morphology associated with these features has been documented for the Scotian Shelf. There are some areas where glaciomarine deposits show evidence of deformation, possibly resulting from dewatering processes, though such areas are restricted in extent and not necessarily associated with lift-off moraines.

Boulton et al. (1996), identified features resembling lift-off moraines near the island of Coraholmen, Spitsbergen. They were observed as till ridges on land and linear ridges in shallow water off shore displaying dimensions up to 300 m wide, 2 m high, and closely spaced between 5-20 m apart. The dimensions of these features are more comparable to lift-off moraines observed by King and Fader (1986), than the features described by Vogt (1997). Similar to King, and Fader (1986), Boulton et al. (1996), attributed the origin of the narrow rectilinear ridges 'crevasse-intrusion' ridges to the process of subglacially deformed material intruding into overlying glacial ice crevasses. Crevasses are typical of surging glaciers undergoing extensional flow, and Boulton et. al. (1996), explain that such ridges can be left after a surging glacier stagnates. Sharp (1985) previously identified similar features and associated them with surging glaciers. The features observed by Boulton et al. (1996), were observed on-land and off-shore, and were supported by documentation of modern glaciers displaying debris intruding into bottom crevasses.

### 2.3.3c: Relationship of Moraines and Incisions

King (1993; 1994), proposed that the presence of channels (incisions), formed by excess meltwater discharge, in moraines may be used to distinguish between deposits from ice-shelf environments and tidewater glaciers, significant incisions being associated with the latter. King (1994), bases this explanation on incisions of a smaller scale (seldom exceeding 20 m in depth) than the more deeply incised incisions which cut into bedrock (i.e. Amos and Knoll, 1987; Boyd et al., 1988; this study).

McLaren (1988), recognized larger incisions up to 100 m deep and 3 km wide on Sable Island Bank and attributed them to subglacial meltwater erosion along an ice terminus resulting in deposition of the Sable Island Sand Body. She interpreted the Sable Island Sand Body as a moraine related to a tidewater margin. Interpreting incisions as being associated with tidewater margins may not apply to the larger scale incisions extending over 450 m bsl. Terminal moraines do not appear to be associated with all of the incisions observed on the eastern Scotian Shelf, though this is likely, in part, a result of erosion of moraines by marine transgression.

Interpreting moraines as a product of excess meltwater discharge at a tidewater terminus assumes they are all meltwater features. Alternative models of ice streaming proposed for the formation of some incisions (section 5.4.3c) requires a reinterpretation of the ice margin configuration.

### 2.3.4: Overview of the Quaternary History of the Scotian Shelf

This overview of the Quaternary History of the Scotian Shelf is primarily based on the first synthesis for glacial events on the Scotian Shelf done by King and Fader (1986), and a work in progress which correlates land events to off-shore events, and in so doing reviews much of the work done previously (Stea et al., in press). For further details, the reader is directed to: King and Fader (1986); Piper et al. (1990); Stea (1995); King (1994); King (1996); Stea et al., in press; and references therein. Table 2.2 is a summary of the Pleistocene

| | | | STAGE | SUB-STAGE | OXYGEN ISOTOPE STAGE | APPROX. AGE (ka) |
|---|---|---|---|---|---|---|
| QUATERNARY | HOLOCENE | | | | 1 | |
| | | | | | | 10 |
| | | | | | | 13 |
| | PLEISTOCENE | late | WISCONSINAN | Late | 2 | |
| | | | | | | 23 |
| | | | | | | 32 |
| | | | | Middle | 3 | |
| | | | | | | 65 |
| | | | | Early | 4 | |
| | | | | | | 80 |
| | | | SANGAMONIAN | | 5 | |
| | | | | | | 130 |
| | | middle | ILLINOIAN | | 6 | |
| | | | | | 7 | |
| | | | | | 8 | |
| | | | | | 9 | |
| | | | ? | | ⋮ | |
| | | | | | 15 | |
| | | | | | | 790 |
| | | early | | | | |
| | | | | | | 1650 |
| TERTIARY | PLIOCENE | | | | | |

GSC

Table 2.2 Quaternary time scale and nomenclature. From Fulton (1984) time scale used in Canada (Fulton, 1989).

### *2.3.4a: Chronology of Pleistocene Events*

Piper et al. (1987), believe that the main phase of slope canyon cutting occurred during the Late Pliocene and Early Pleistocene, and that this correlates with major Scotian Shelf incisions. In addition, Piper and Nomark (1989), recognize high sedimentation rates and increased slope gullying during the Early Pleistocene. In the Sable Island area, the Plio-Pleistocene boundary was identified at 262 m below the sea bed (Hardy, 1975; Piper et al., 1990), and the Tertiary-Quaternary boundary was identified at 220 m (Boyd et al., 1988). Both likely correlate to the late Tertiary-Pleistocene Unconformity (R3), identified by King et al. (1974). An early to Late Pliocene boundary was identified at 420m bsl (Gradstein and Agterberg, 1982; Piper et al., 1990), which may also be important when considering the origin of the eastern Scotian Shelf incisions if they are in part inherited from these unconformities.

Evidence for Pre-Wisconsinan glacial episodes on the mainland of Nova Scotia and the Scotian Shelf was largely removed by Wisconsinan ice, thus direct and datable evidence for pre-Wisconsinan glaciations is sparse and/or indistinguishable. Evidence on the Scotian Shelf for pre-Late Wisconsinan glaciations is difficult to observe as sediments at depth from earlier glacial episodes are difficult to resolve due to resolution limitations of marine seismic reflection techniques (section 3.2.2), and difficulties in sampling sediments at depth. It appears that the Scotian Shelf was affected by pre-Wisconsinan, and Early (75-65 ka; OIS 4; note: OIS refers Oxygen isotope stage), Middle (65-23 ka; OIS 3), and Late (23-10 ka; OIS 2) Wisconsinan glaciations. However, unlike the majority of Canada for which the Late Wisconsinan was the period of most extensive glaciation during the Wisconsinan (OIS 6—Illinoian—had ice volumes comparable to that of the Late Wisconsinan; Piper et al., 1990), it appears the Early-Mid Wisconsinan was the most extensive Wisconsinan event to cross the Scotian Shelf.

1. Pre-Wisconsinan

No pre-Wisconsinan age glacial sediments have yet been dated on the Scotian Shelf (Piper et al., 1990). Therefore, pre-Wisconsinan Glaciation of the Scotian Shelf has mainly been inferred from deeper marine seismic and sediment core records from the Scotian Slope and further seaward, and from sparse evidence on land. Piper et al. (1994), present what is perhaps some of the most important off-shore data supporting extensive pre-Wisconsinan glaciations on the Grand Banks of Newfoundland. Through analyzing the sediments of a core from the J-anomaly Ridge (3877 m bsl), representative of 0.9 Ma, Piper et al. (1994), proposed that OIS 14 to OIS 12 (Middle Pleistocene) were phases of extensive erosion removing considerable amounts of Cretaceous and Tertiary bedrock. In addition, they propose that the major excavation of the Laurentian Channel occurred during Late OIS 12. The Middle Pleistocene, OIS 11 to 6 (~Illinoian) also showed evidence of extensive glacial erosion (Piper et al., 1994), and it is believed that OIS 6 supplied the Fogo Seamounts (off Grand Banks, NF) with the largest amount of Pleistocene sediment (Alam and Piper, 1977). Mosher et al. (1989), inferred that the lowest-wedge shaped acoustically incoherent unit (till tongue) on the Scotian Slope, in the area if Verrill Canyon, was deposited during OIS 6 (Illinoian). Land evidence for pre-Wisconsinan glaciations in Nova Scotia is based on sparse occurrences of conglomerates composed of weathered and iron-cemented drift and outwash, and pre-Wisconsinan interglacial (dated at 75 000-120 000 yr BP) sediments resting on probable Illinoian tills (Grant, 1989; Stea et al., in press). On land, as many as four till sheets with distinct provenances were identified above Sangamon organic deposits.

2. Early-Mid Wisconsinan

Evidence for extensive Early-Mid Wisconsinan (OIS 4/3) glaciation on the Scotian Shelf stems from both on-land records and off-shore records. Stea et al. (in press.) discuss the land based Ice Flow Phase Model (Stea, 1982; Stea, 1984; Stea and Fink, 1984; Stea et al., 1987; Stea et al., 1992; Stea, 1995), and correlate it to the off-shore. The Ice Flow Phase Model consists of four phases, Phase 1 (eastward, southeastward flow) represents the Early-Middle Wisconsinan, and the remainder, Phases 2, 3, and 4, represent the Late Wisconsinan (Fig. 2.7). Lawrence Town/Red Head Till is easily identified because of its distinct red colour derived from Carboniferous redbeds in the Gulf of St. Lawrence. This till is the thickest and most extensive till unit found in Nova Scotia, thereby indicating that the Early-Middle Pleistocene glaciation which crossed Nova Scotia was extensive. This unit has been tentatively dated between 38 and 75 ka (Piper et al., 1990), and is assumed to be older than Late Wisconsinan glacial deposits based on its stratigraphic position. Stea et al. (in press) do not directly acknowledge a relation of the red tills to Phase 1, however an early grey till (Hartlen till) which underlies the red tills has been identified as a product of Phase 1.

In the off-shore, corresponding to Phase 1, Stea et al. (in press), present a summary of evidence which lends further support to the implication that the Early Wisconsinan was the most extensive Wisconsinan glacial event to affect the Scotian Shelf. The most important evidence includes: 1) Significant amounts of red muds (distally derived from the Gulf of St. Lawrence) found on

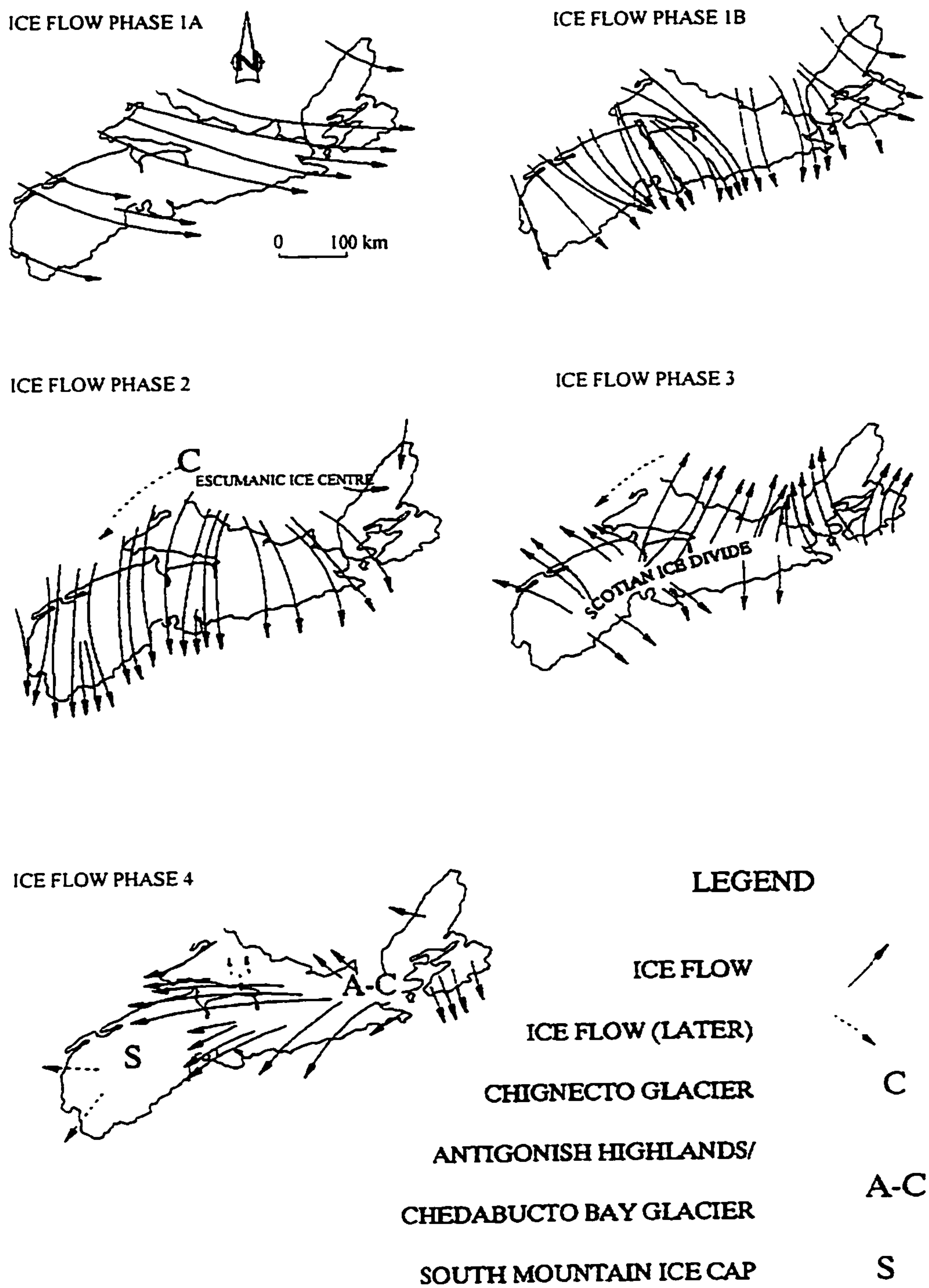



Figure 2.7 Four phase Ice Flow Phase Model for glaciation on the mainland of Nova Scotia. Modified from Stea (1995).

the Fogo Seamounts (Alam et al., 1983); 2) A wedge of acoustically incoherent material on the Scotian slope which lies between a pre-Wisconsinan deposit and a dated late Wisconsinan deposit (Mosher et al., 1989; It should be noted that the wedges below the dated Late Wisconsinan wedge have not been absolutely dated, and therefore the possibility that their origin could be due to some Late Wisconsinan events should not be overlooked); 3) A change in sedimentation style along the southeast Canadian continental margin delivering coarse sediment to the continental slope (Piper and Normark, 1989); 4) Stage 1 and possibly part of Stage 2, times of maximum ice cover on the Scotian Shelf of the King and Fader (1986) five stage model for glaciation on the Scotian Shelf (Fig. 2.8; to be discussed later); 5) Evidence of an older/earlier till believed to be Middle Wisconsinan in age in the near shore zone of the south shore of Nova Scotia (Piper et al., 1990); 6) A core from Sable Island through the upper sediments of a buried incision yielded a date of 37.2 ka (Boyd et al., 1988).

3. Late Wisconsinan

Following the extensive Early-Middle Wisconsinan glaciation, the Scotian Shelf was affected by weaker Late Wisconsinan influences, though the ice cover on the Scotian Shelf was still extensive, reaching the outer banks. The Late Wisconsinan was characterized on land by Ice Flow Phases 2, 3, and 4 (Stea, 1995; Stea et al., in press), of which Phase 2 (southward and southwestward flow; 18-21 ka; Fig 2.7) was the strongest. Stea et al. (in press) identify the red Lawrence Town till (18-21 ka BP) with Phase 2 and evidence of eroded Carboniferous redbeds has been found as far off-shore as the outer banks on the Scotian Shelf. The deposits from this time of glaciation outcrop on

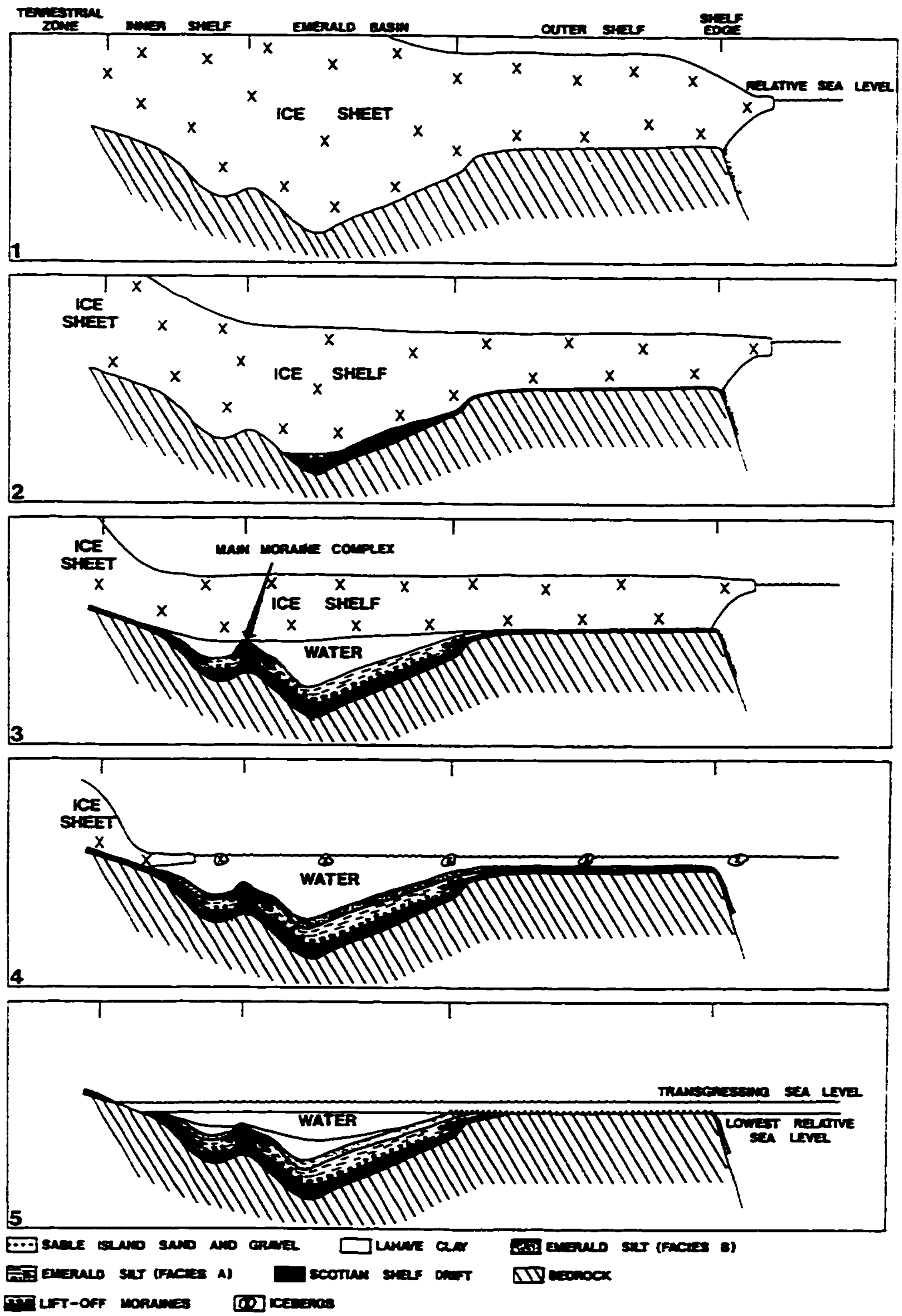



Figure 2.8 Five stage model of glaciation on the Scotian Shelf. From King and Fader (1986).

the Scotian Shelf. They are the most extensive, and easily observed by marine techniques.

The 5 stage model proposed by King and Fader (1986; Fig. 2.8), still stands as a general model for glaciation on the Scotian Shelf, though there are problems with the original dates. Stages 3 and 4, and possibly Stage 2 of this model likely correspond to Ice Flow Phase 2. Stage 2 was an ice recessional phase, where ice lifted off in basins, but remained grounded on the banks. It is represented by deposition of an early till, development of lift-off moraines at the surface of this till deposit within basins, and deposition of early glaciomarine deposits during the formation of the lift-off moraines (King and Fader, 1986). Early Stage 2 was likely the time of development of the outer shelf moraines found along Sable Island Bank and Banquereau (McLaren, 1988; King, 1993; King, 1994; King, 1996), and King (1996) correlates Phase 2 with ice reaching its terminal position along the Scotian slope at about 21 ka. Lift-off of thinning ice in Emerald Basin has been given a revised date of 18 ka (Gipp and Piper, 1989; King and Fader, 1989; original date from King and Fader, 1989: 50-45 ka).

Stage 3 (King and Fader, 1986; Fig. 2.8), was a phase of continued ice recession marked by deposition of Emerald Silt Facies A and development of the Scotian Shelf End Moraine Complex. Till tongue development occurred during this time as the buoyancy lines of the ice shelf fluctuated. Ice was characterized as floating ice shelves over basins, and grounded ice over the banks. The exposed incisions of the eastern Scotian Shelf also display lift-off moraines in their sedimentary sequence. This may indicate a partially buoyant ice shelf in this area, pinned only along the high points between the incisions and along the banks. Ice flow Phase 3 (Fig. 2.7), involved the development of an ice divide over Nova Scotia, and corresponds with this stage. This is indicated by ice berg scouring in the basins of the middle shelf representing calving ice (Fader, 1991; Stea et al., in press), and lithological correlation between tills on land and the Scotian Shelf End Moraine Complex (Stea et al., in press; Stea, 1995). Tills from this phase are of local provenance, and may correspond with the thin veneer of 'young tills' mentioned by Grant (1989). The Scotian Shelf End Moraine Complex may be multigenerational, marking an ice terminus during different ice flow phases. Stea et al. (in press), proposed that the Scotian Shelf End Moraine Complex is composed of numerous tills from different events separated by erosional breaks (i.e. Hartlen and Lawrence Town tills may be at the base of the end moraine complex). Grounding at the Scotian Shelf End Moraine Complex has been dated at 15-16 ka in the northeast, and 16.8-18 ka to the south (Stea et al., in press; Gipp, 1989; Gipp and Piper, 1989; King, 1996).

Stage 4 (King and Fader, 1986; Fig. 2.8), which probably corresponds with the recession during Ice Flow Phase 3 (Fig. 2.7), was a time of retreat of the ice-shelf to near the present shore line with possible minor readvances, though Ice Flow Phase 4 marks a major readvance (Stea et al., in press). According to King and Fader (1986), Stage 4 was characterized by deposition of Emerald Silt Facies B and the maximum lowering of sea level. The outer shelf low sea-level stand has been dated at about 17-15 ka (Amos and Miller, 1990; King, 1994), and Emerald Silt (Facies A) lying above the Scotian Shelf End Moraine Complex has been given a date of about 14.5 ka with an idea that the ice margin reached the present coast line at about 14 ka (Stea et al., in press).

Ice Flow Phase 4 (Fig. 2.7), which included development of the Scotian Ice Divide which extended offshore (Stea et al., in press), likely corresponded with Stage 5 (Fig. 2.8). Stage 5 was a period of lowest sea level followed by transgression with final disintegration and removal of ice influences off-shore

(King and Fader, 1986). This time was marked on land as the reactivation of a number of small ice caps during a late-glacial cooling event and possibly by the formation of the local till veneers of Grant (1989), and it is represented by De Geer moraines along the inner shelf Morainal Zone (Stea et al., in press; Stea, 1995). The timing of this event may be somewhere between 13.5-12.7 ka (Stea et al., in press; Stea et al., 1996). Ice is assumed to have been gone by 11 ka, and sea level reached its lowest along the inner shelf (-70 m) at about 11.6 ka, followed by rapid transgression from 11.6 ka to 11 ka and slower transgression from 11 ka to 9 ka (Stea et al., in press; Stea et al., 1994).

Recent works regarding the off-shore discuss the influence of a Younger Dryas glacial event (King, 1994; Stea, 1995; King, 1996; Stea et al., in press). The Younger Dryas, characterized by a climatic cooling event around 11-10 ka, is represented by an ice flow Phase 5 (Stea et al., in press; not represented on Fig. 2.7). Changes in land and off-shore assemblages of pollen and forams support this notion, but off-shore evidence suggests Younger Dryas ice influences were minimal on the shelf, and the Younger Dryas was characterized by increased sea-ice cover, ice bergs and, storminess in the off-shore (Stea et al., in press, and references therein).

The term 'early post-glacial' is used for sediments older than the Holocene and which display interglacial characteristics (Piper et al., 1990). Such sediments range in age from 15 ka in the off-shore to 12 ka in the near-shore (Piper et al., 1990). Ice margin conditions, as exemplified by low salinity water, are believed to have persisted up to 11 ka in Emerald Basin and Canso Basin (Piper et al., 1990; Mudie, 1980; Scott et al., 1984b). This gives a range from 15 ka to about 10 ka for ice disintegration in the Maritimes (Piper et al., 1990), though overall retreat was initiated from 21 ka (King, 1996).

The concept of local ice caps on the bank areas of the Scotian Shelf has not been addressed in great detail, however this should not be overlooked, particularly where there is evidence on land for such ice configurations to have existed. Seismic data observed during the course of this study yielded some stratigraphic evidence for a possible local ice cap on Canso Bank. This evidence consisted of till tongues extending off the flanks of Canso Bank, buried incisions, thick till sequences, and thick sequences of intermixed contorted glaciomarine deposits and incoherent facies (till?), possibly indicative of fluctuating grounding and floating ice conditions. With this in mind, it is possible that other off-shore banks such as Middle Bank, Sable Island Bank, and Banquereau may have supported local ice caps during various glacial episodes.

### *2.3.4b: Sea Level*

Relative sea level fluctuations resulted from isostatic depression (ice sheet loading), peripheral crustal bulge, and changes in world ocean volume resulting from glacial ice accumulation and retreat (Piper et al., 1990). The Late Wisconsinan low sea-level stand for the Scotian Shelf, dated at 15 ka (King and Fader, 1986), has been identified at 110-120 m bsl (King, 1967; King, 1970; King and Fader, 1986; Fader, 1989), and is in general agreement with Fairbanks (1989) estimated global sea level lowering of 121±5 m during the last glacial maximum. The Scotian Shelf low sea level stand is based on a widespread unconformity, in places marked by terraces and subaerial dissection of the bedrock, a lack of till and glaciomarine sediments above the unconformity, and the distribution of Sable Island Sand and Gravel (Fader, 1989; Piper et al., 1990). It has been proposed that the consistent depth of the terrace and the distribution of Sable Island Sand and Gravel and Sambro Sand indicate little crustal warping (isostatic depression or forebulge migration) occurred, or that the crust had recovered prior to the low sea level stand along the middle and outer shelf. However, younger evidence for low sea-level stand has been traced northwards in shallower water, indicating both the landward movement of post-glacial transgression and evidence of isostatic rebound (Piper, et al., 1990).

Transgression and regression events which followed the Late Wisconsinan low sea-level stand lead to both the development of barrier complexes on the outer banks, and the removal of glacial sediments in the inner shelf region, while marine muds (LaHave Clay) were deposited in the basins of the middle shelf (Piper et al., 1990).

Stea et al. (1994), identified a low sea-level stand along the inner shelf at 65 m bsl which has been dated at 11 650-11 250 $^{14}C$ yr BP. A new sea level curve, was developed based on this information which acknowledges greater crustal depression due to greater ice loads nearer to the ice centres, and sea levels which correspondingly become shallower.

Clusters of ice berg scours were observed on sidescan data from a recent survey of the Misaine Bank area (January, 1998, CCGS Teleost). Whether or not these ice berg scours are modern or relict may have implications regarding earlier estimates of low sea level stand at 110-120 m bsl. As these scours occur in dense clusters it is thought they may not be modern (Fader, pers. comm., 1998) If the scours are from glacial times, it is an indication sea level was not as low as the previous estimate, and the Sable Island Sands and Gravels may be a deposit which was reworked in water depths as much as 20 m bsl (Fader, pers. comm., 1998). Holtedahl (1989) attributed sands and gravels to depths of 40 m bsl as a result of currents and waves.

One final consideration regarding sea level fluctuations, particularly during the Early Wisconsinan, and possibly the Illinoian which was considered to have been the time of most extensive glaciation on the Scotian Shelf and Nova Scotia, is the affect of ice-water gravitational attraction. Clark (1976), proposed that this type of gravitational attraction could have been responsible for as much as an 85 m apparent rise in sea level. This, combined with isostatic depression indicates that sea levels may have been higher during the times that the Scotian Shelf was covered with grounded ice. Based on maximum estimates of low sea level stand (110-120 m bsl), it is clear that all phases of Pleistocene channel incision took place in a subglacial environment.

## 2.4: SUMMARY

This chapter provides background information on the physiography and glacial history of the Scotian Shelf. This will be important for chapters 4 and 5 in which interpretations and discussions regarding the origin of the morphology of the eastern Scotian Shelf are presented.

The present physiography of the eastern Scotian Shelf appears to have been influenced by pre-glacial fluvial activity (Cretaceous and Tertiary; sections 2.2.2, 2.2.3) and may in part represent a preserved landform assemblage. The most extensive glaciation of the eastern Scotian Shelf appears to have been during the pre and early-mid Wisconsinan and this may have been the most important time of glacial modification of the morphology of the eastern Scotian Shelf (incision formation; section 2.3.4). Late Wisconsinan ice covered the whole eastern Scotian Shelf and resulted in the uppermost succession of Pleistocene deposits (sections 2.3.2, 2.3.3). Section 2.3.2 described these sediments in detail and form the basis of the siesmostratigraphic interpretations for chapters 4 and 5.

# CHAPTER 3: METHODS

Analysis of the eastern Scotian Shelf morphology is based primarily on two types of data housed at the Bedford Institute of Oceanography (BIO), Geological Survey of Canada (Atlantic) (GSCA), Dartmouth, Nova Scotia:

1. bathymetry
2. high resolution shallow marine seismic reflection records

## 3.1: EASTERN SCOTIAN SHELF BATHYMETRIC IMAGE

Figure 1.2, showing the morphology of the eastern Scotian Shelf, was produced by manipulating bathymetric data (depth soundings) with a GIS software package called GRASS (public domain software developed by the U.S. Army Corps of Engineers). The depth soundings were collected by the Canadian Hydrographic Service (CHS) via echosounders from the CSS Baffin, between the years of 1979 and 1990 (Loncarevic et al., 1992). The data were then processed into digital format, each point consisting of x, y, and z (easting, northing, and depth) coordinates. A digital elevation model (DEM) of the eastern Scotian Shelf was created by importing the digital data set of depth soundings into GRASS and further processing it through an imaging package written by R.C. Courtney of the GSCA. The data were converted to a 250 m x 250 m grid using a program which applies a tension and spline method to generate an interpolated surface (Ocean Mapping On-line help). The gridded data set was then assigned a scaled colour range chosen to maximize the resolution of morphologic features between 20-400 m bsl. Finally, the colour image was combined with a shaded relief image (shadowgram) with a vertical exaggeration of 100x, and an illumination source set at an inclination of 25° and an azimuth of 320° (from the northeast). These parameters were chosen to maximize the resolution of the morphology of the sea floor.

## 3.2: HIGH RESOLUTION SHALLOW MARINE SEISMIC REFLECTION RECORDS

The original high resolution shallow marine seismic records interpreted for this study include over 6000 km of tracklines providing data collected using the Huntec Deep Tow System (DTS), airgun, sleeve gun, and sparker systems. These records were collected during various Bedford Institute of Oceanography, Geological Survey of Canada (Atlantic), cruises between the years 1967 to 1996 (Figure 3.1).

### 3.2.1: Seismic Reflection Theory

Simply stated, collection of seismic data involves the acquisition of measurements of seismic wave energy (the elastic strain of the earth) resulting from an induced stress (a seismic source). A reflection results at any boundary at which the acoustic impedance (the speed at which a seismic wave propagates through a unit) of various lithologic units differs. The impedance contrast is the difference between the acoustic impedance of the bounding units—the greater the contrast, the stronger or 'brighter' the reflection seen on the output record. This process is dependent on the velocity at which a seismic wave propagates through a medium, and hence on its density.

Seismic waves are grouped into two types; body waves and surface waves. Body waves are further subdivided into compressional waves

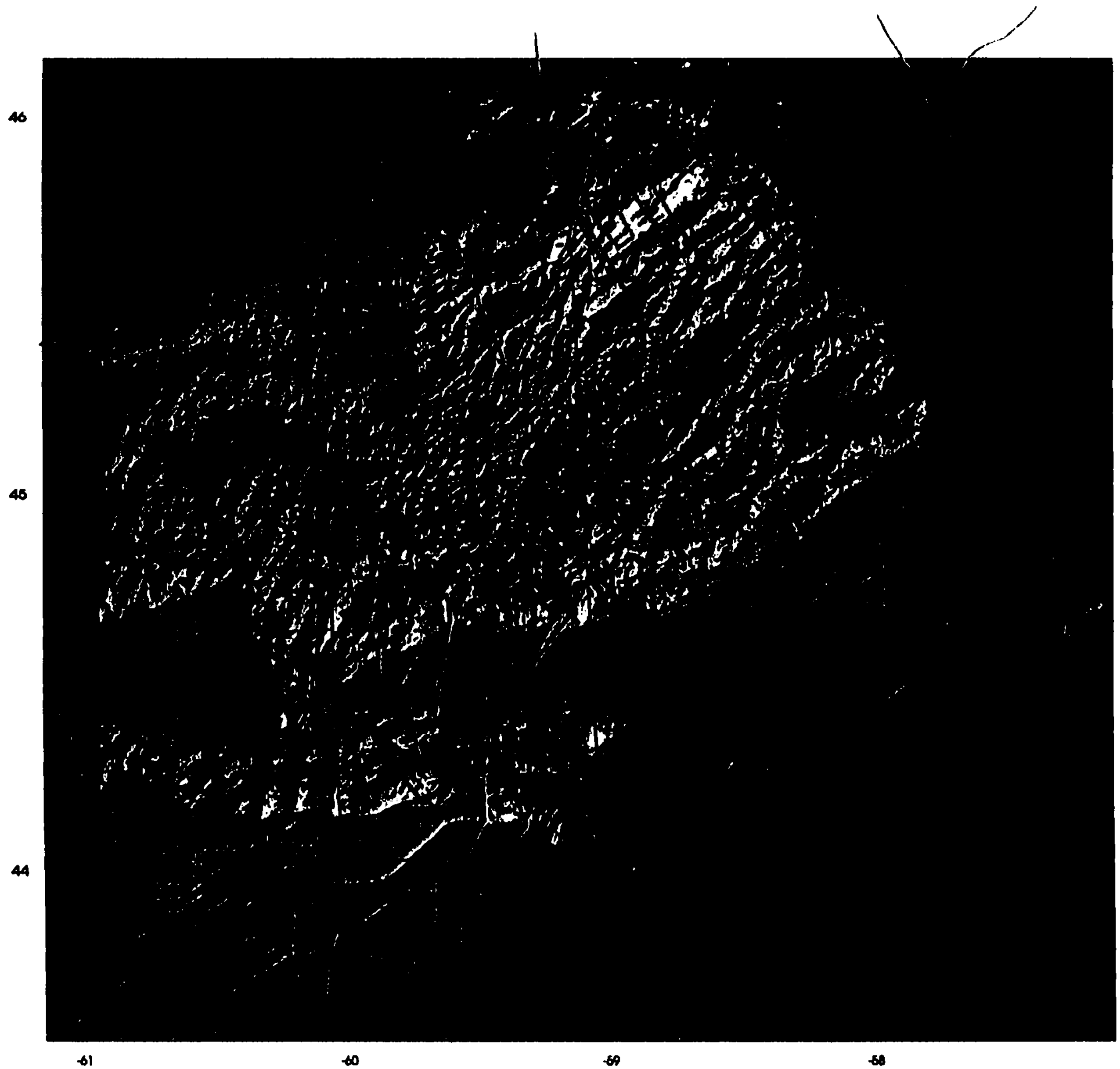



Figure 3.1 Eastern Scotian Shelf bathymetry showing tracklines of shallow marine seismic reflection data used in this study. (All data are from Geological Survey of Canada (Atlantic), Bedford Institute of Oceanography cruises between 1967 and 1996.)

(longitudinal, primary waves/P-waves) which propagate through an elastic solid by compressional and dilational strains in the direction of wave travel, and shear waves (transverse, secondary waves/S-waves) which propagate in a direction perpendicular to the direction of wave travel (Kearey and Brooks, 1991; Fig. 3.2a). Surface waves propagate along the surface boundaries separating units possessing different elastic properties, and are subdivided into Rayleigh waves (elliptical particle movement perpendicular to the surface and containing the direction of propagation), and Love waves (oscillatory particle movement parallel to the surface and perpendicular to the direction of wave motion) (Kearey and Brooks, 1991; Sheriff and Geldart, 1982; Fig. 3.2b). Seismic surveying is primarily concerned with compressional, or P-waves.

### *3.2.1a: Seismic Wave Rays*

The two most important types of seismic rays resulting from seismic energy inputs are reflected and refracted rays. Seismology utilizes both reflected and refracted rays for different types of surveys, the difference being that reflection surveys generally measure ray travel paths which are vertical, while refraction surveys measure ray travel paths which are horizontal. The hydrophone or geophone is placed further from the seismic source for refraction surveys than for reflection surveys, and the systems are attuned to receive the respective waves (refracted waves arrive sooner than reflected waves; Kearey and Brooks, 1991).

Reflected rays result when energy from a source wave hits a boundary with a different acoustic impedance (density) from the bounding formations, and the energy is reflected or returned from this surface (Fig. 3.3). Reflection

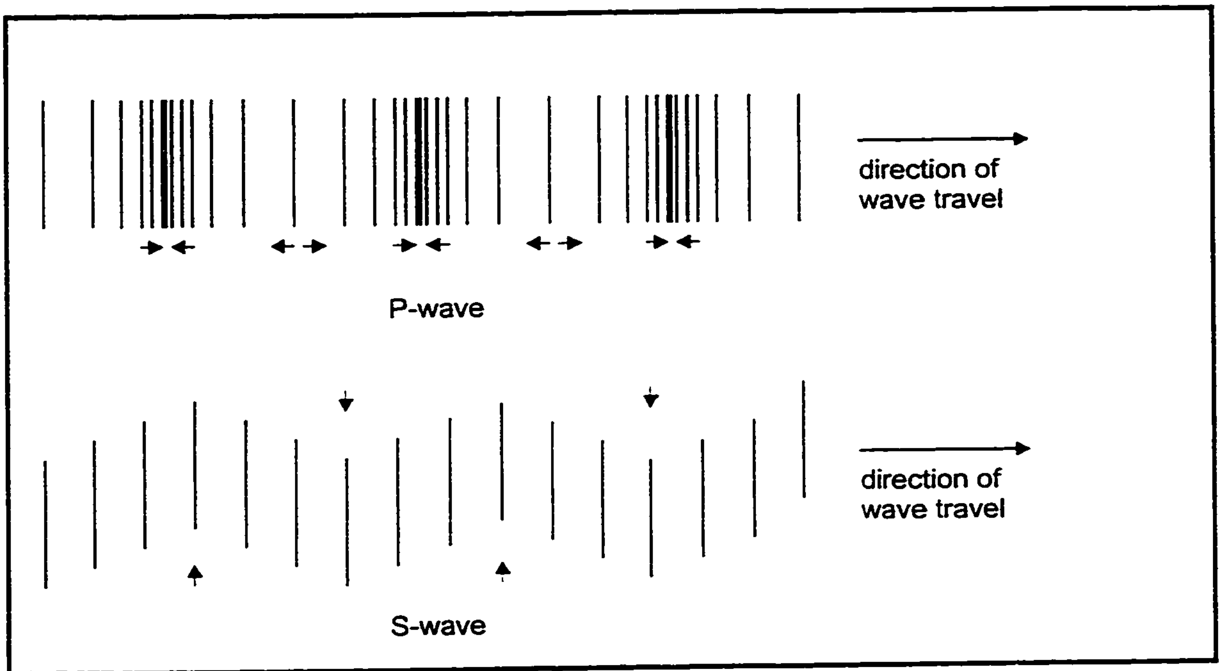



Figure 3.2a Body Waves

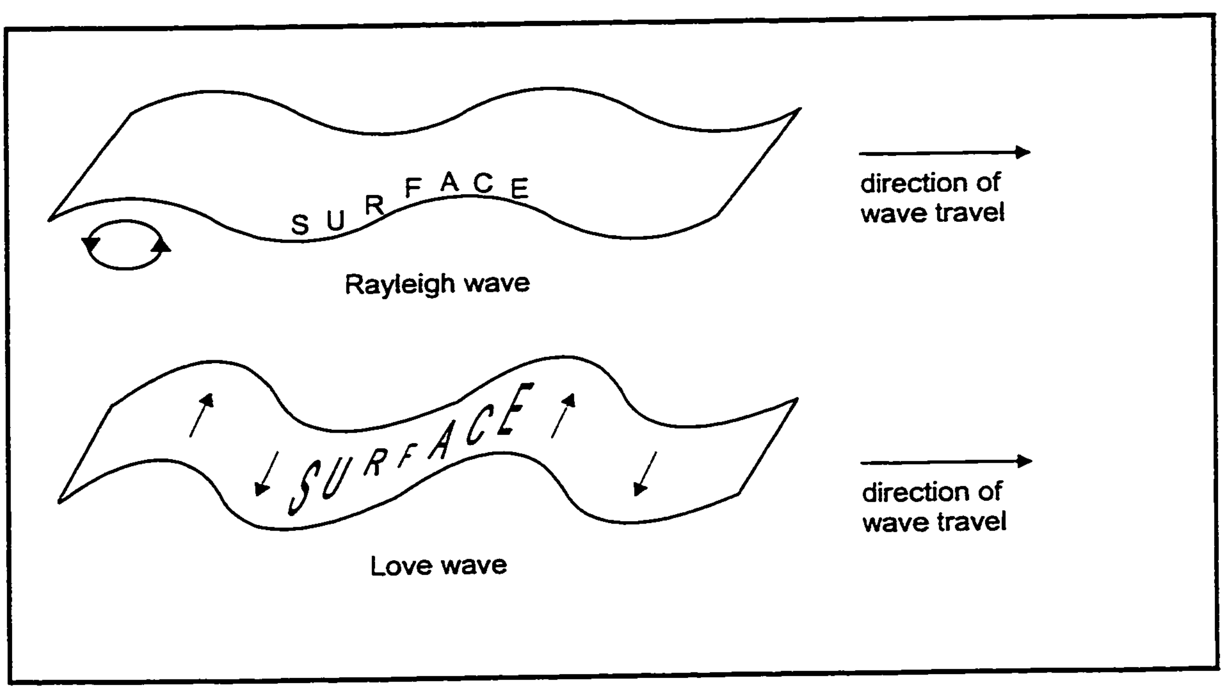



Figure 3.2b Surface waves

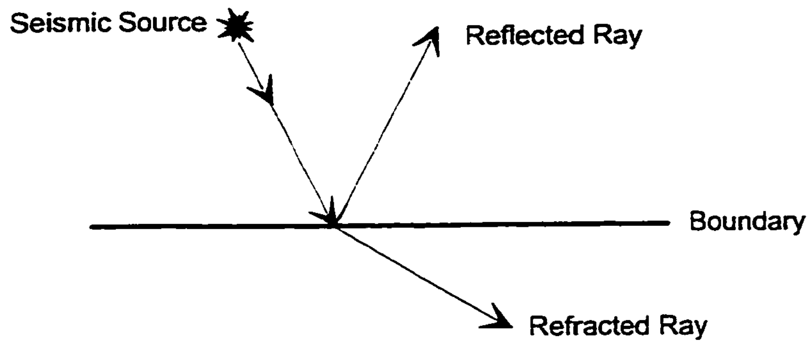



Figure 3.3 Reflected and refracted wave rays.

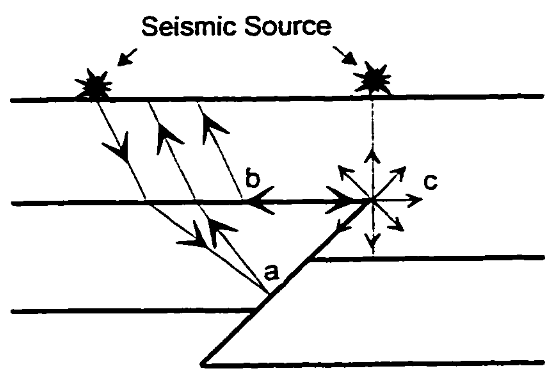



Figure 3.4 Reflected refraction rays (a and b), and diffracted rays (c).

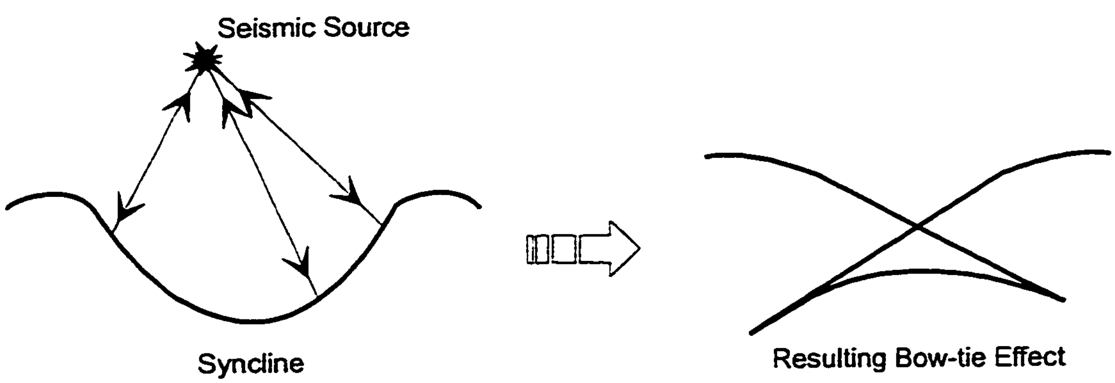



Figure 3.5 Bow-tie effect resulting from diffraction along a syncline.

seismic surveying is concerned with recording the properties of reflected body waves.

Refracted rays result when wave energy from a source wave continues through an acoustic impedance contrast boundary, with a changed direction of propagation, into a unit with a different density (or acoustic impedance). In reflection seismic surveys, refracted rays can result in noise on the output record in the form of reflected refractions which occur when a refracted wave terminates along an oblique boundary and is subsequently reflected. Reflected refractions are difficult to distinguish from primary reflections (Fig. 3.4a, b).

Diffracted rays are a third type of ray important to interpreting seismic reflection data. Diffracted rays (radial scattering of incoming seismic energy) result when seismic source waves hit an abrupt discontinuity (i.e. fault plane), and the end of such a reflector acts as a point source, thus scattering energy in all directions (Fig. 3.4c). The output on a seismic record will be a hyperbolic trace with the diffraction source at the apex (Stoker et al., 1997). Diffracted waves can result in a 'bow-tie' event which occurs when the radius of curvature of a reflector exceeds the curvature of the source wave (i.e. channels, canyons, synclines), thus reflections are generated from more than one point, including the flanks and the base of a synclinal feature. The bow-tie is a result of hyperbolic reflections from three discrete reflection points (Stoker et al., 1997; Kearey and Brooks, 1991; Fig. 3.5). Such a reflector may result in a misleading interpretation of a channel being shallower than its true depth, or possessing a V-shaped bottom (Stoker et al., 1997). Also associated with diffraction is scattering, which occurs when the point reflector has a smaller radius of curvature than the wavelength of the source wave (i.e. boulders and gravel). The result on the output record resembles a chaotic signature with few hyperbolic reflections (Stoker et al., 1997).

### *3.2.1b: Noise*

Noise on a seismic record is grouped into two types: coherent and incoherent. Noise encompasses events other than the primary event or events from which desired information can be derived. Coherent noise can be followed across numerous traces, while incoherent noise, also known as random noise, is dissimilar, unpredictable, and cannot be followed along traces. Some sources for coherent noise include source generated noise (i.e. bubble pulse), some resulting in multiple reflections (to be discussed in section 3.2.1c), surface waves, refractions, reflected refractions (discussed in section 3.2.1a), and diffractions (also discussed in section 3.2.1a). Some sources for incoherent noise include electrical sources, small structural irregularities, and environmental sources (wind, waves) (Stoker et al., 1997; Sheriff and Geldart, 1982).

If the term signal is used to qualify primary reflections or any event on the seismic record providing desired information, all other seismic energy displayed on the record is considered noise. The quality of a seismic output record is dependent on the signal-to-noise ratio (S/N), the larger the ratio, the better the record (Sheriff and Geldart, 1982).

### *3.2.1c: Multiple Reflections*

Multiple reflections result when seismic wave energy is reflected more than once, after hitting secondary reflection horizons. The larger the impedance contrast of the surface, the stronger the multiple. Multiples are divided into short-path multiples—reflections arriving shortly after the primary reflection—resulting in an extension of the primary signal, and long-path multiples—

reflections with a long travel path compared to the primary reflection—resulting in a distinct event on the seismic record (Fig. 3.6).

In marine seismic surveys, long-path multiples, called sea bed multiples, result when energy reflected from the sea bed is reflected back to the sea surface and back to the sea bed again, thereby producing one or more multiples. The timing of the sea bed multiple is dependent on the depth of water. The shallower the water, the sooner the arrival. These multiples can cause difficulty interpreting seismic records produced in shallow water where the vertical acoustic window (area on the seismic record which is not obscured by noise) becomes smaller, obscuring reflections at depth.

The sea-surface reflection multiple which hinders records produced by deep-tow systems (i.e. Huntec DTS) is produced when source energy goes to the sea surface and is reflected back to the hydrophone. This type of multiple meanders across the record as the system is adjusted in variable water depths (Stoker et al., 1997). With the exception of the sea surface reflection on DTS records, short-path multiples do not significantly obscure seismic records as the long-path multiples do.

### 3.2.2: Data Acquisition

Marine seismic reflection profiling is done via multi-channel or single-channel systems. The multi-channel system, predominantly used by the hydrocarbon industry, records the results from a source array (group) via multiple channels which digitally record and process the data prior to output. This system allows for various methods of processing and filtering the data which ultimately reduce the influence of noise in the records. The single-channel system, predominantly used for high resolution marine seismic

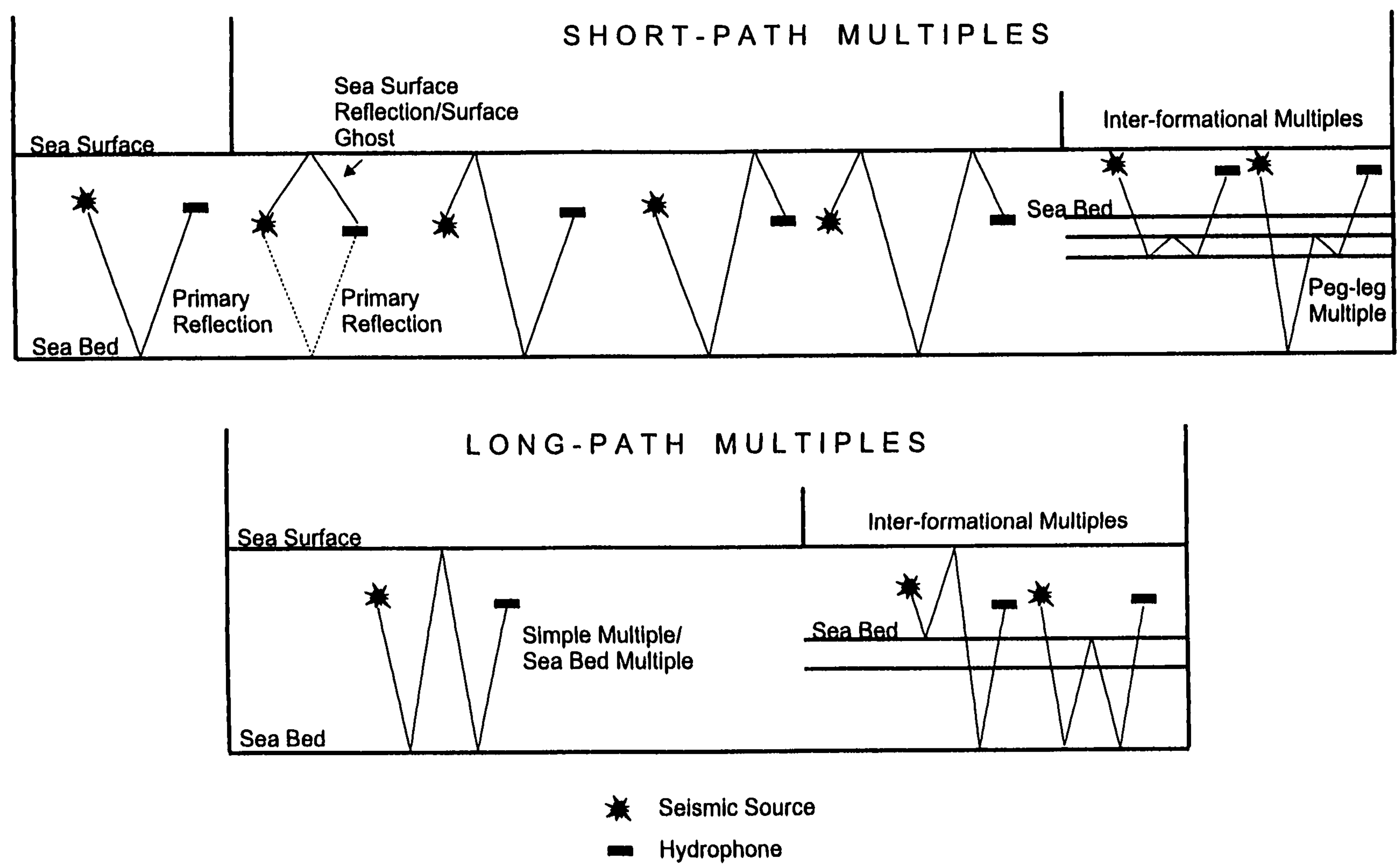



Figure 3.6 Most common short-path multiples and long-path multiples resulting in noise on seismic records.

reflection profiling, records the results from a source via a single channel which are digitally recorded or translated directly into paper records (analogue record). This type of system can filter out noise if the data are collected digitally, however, analogue records do not allow for a great deal of filtering, and hence multiples and other types of noise remains as a significant problem, particularly in shallow water where sea-bed multiples are common. Single-channel surveys can only be done in a marine environment which allows for continuous acoustic profiling as the seismic source and detectors (hydrophones) are continuously moved forward. Figure 3.7 shows the general set up of a shallow marine seismic survey. In general, a seismic source system produces a seismic signal which is received by a hydrophone system (i.e. NSRF tapered eel array), from which the signal is passed through an amplifier, filtered, and finally displayed on graphic recorders operating at a chosen sweep rate. The high resolution marine seismic reflection records used for this study are of the single-channel analogue type.

The band width of a seismic source refers to the frequency range it emits. The higher the frequency, the higher the resolution of the records, but depth penetration is sacrificed. Broad band systems generate a greater range of frequencies which can be recorded resulting in more detailed seismic records. Boomer, airgun, sleeve gun, and sparker sources are all high resolution, broad band systems, however, their frequency ranges differ, and hence their resolution and depth penetration capabilities differ. The boomer data yield the highest resolution records, however, depth penetration is compromised and is not as great as that of the airgun, sleeve gun, and sparker data which compromise resolution for penetration. Using these records in combination allows for identification of detail lost in the airgun, sleeve gun, and sparker data, and sediments at depth not observed through boomer data (Airgun

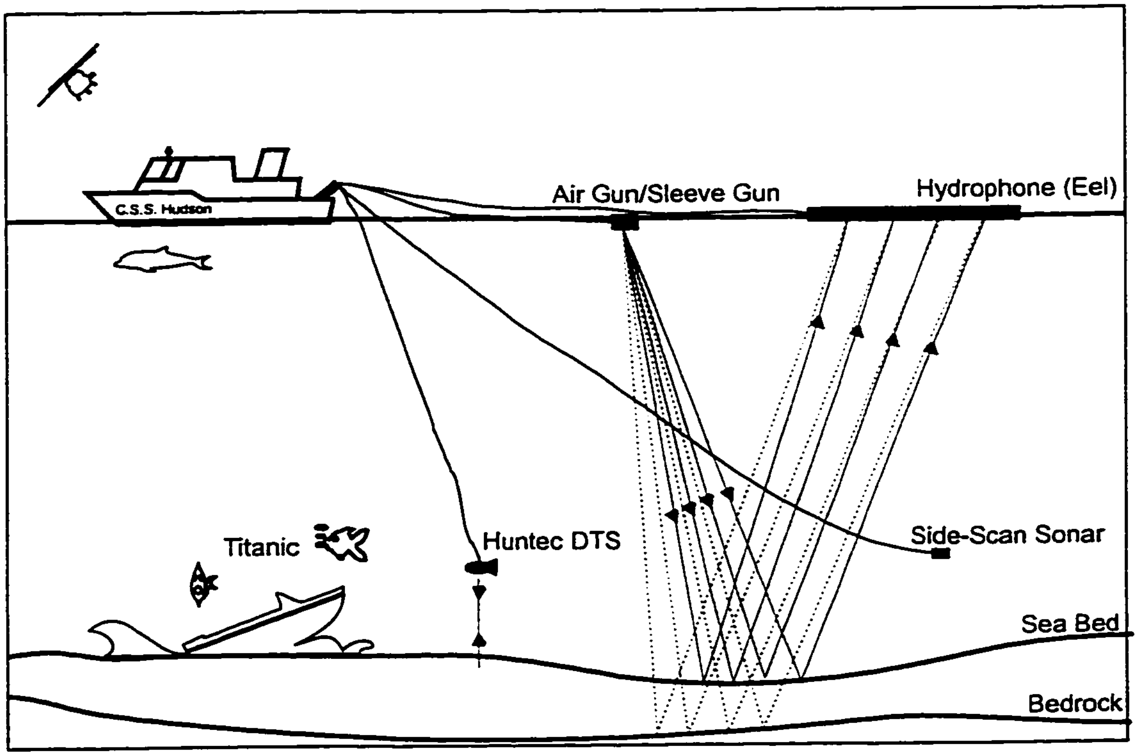



Figure 3.7 Basic survey set-up for a high-resolution, shallow marine seismic reflection survey.

systems can be attuned to deep crustal and hydrocarbon surveys for which depth penetration is a critical factor. Such low resolution, low frequency records using a much stronger seismic input are not considered here.)

Though other sources of acoustic energy exist, the seismic reflection records interpreted for this study include high resolution Huntec DTS (boomer), airgun, sleeve gun, and sparker data (note: A 3.5 KHz Acoustic Profiler which is mounted in the ship's hull, records high resolution bathymetry and is usually included in the database for various cruises):

*Huntec DTS (Boomer)* A boomer, or electro-dynamic acoustic energy source, operates by generating an electromagnetic field which sets up eddy currents in a rigid aluminum plate attached by a spring-loaded mounting to a resin block embedded with a spiral coil (Kearey and Brooks, 1991). The eddy currents generate a reverse polarity field which causes the plate to rapidly repel. This generates a compressional wave in the water (Kearey and Brooks, 1991; McKeown, 1975). The Huntec DTS contains both the acoustic source and hydrophone in one body which is towed at depths up to 305 m rather than near the sea surface thus reducing the influence of wave action, and increasing signal to noise ratios. The system is also heave compensated, removing noise resulting from the motion of the ship from the records (Boyce, pers. comm., 1998). In addition, the boomer source does not produce a bubble pulse, allowing for geology near the sea bed to be observed, further increasing the signal to noise ratio (McKeown, 1975).

General Specifications (as per McKeown, 1975; Hutchens, et al., 1976, and Boyce, pers. comm., 1998):

Bandwidth: 400 Hz to 8-7 KHz

Resolution: <0.3 m

Penetration: >100 m over soft bottoms, and up to 33 m over acoustically hard bottom layers

Noise: sea bed multiples, sea-surface reflections (the strength of the sea-surface reflection is stronger in the external hydrophone output record than on the internal hydrophone output record.

*Airgun/Sleeve gun* Airguns and sleeve guns release compressed air into the water in the form of a high-pressure bubble. Chamber sizes vary according to the desired energy outputs and frequency characteristics of a survey. A larger chamber will produce a bigger bubble with lower frequencies than a smaller gun (Pieuchot, 1984). The characteristic bubble pulse of a sleeve gun is more omnidirectional than with an airgun, therefore the bubble pulse in the water is reduced and cleaner, producing less noise on the records (Chapman, pers. comm., 1998).

General Specifications (Chapman, pers. comm., 1998):

Air/Sleeve Gun: chamber size: 10-40 in$^2$; pressure: 2200 psi; shot rate: 0.5-4 sec depending on water depth

Bandwidth: 0.02 2 KHz (function of chamber size)

Resolution: >2 m (function of bubble pulse)

Penetration: to bedrock surface (through bedrock if Tertiary or Cretaceous sediments, or to acoustic basement)

Noise: seabed multiples (long-path multiples), short-path multiples, bubble pulse (seismic wave generated as the bubble loses energy and oscillates which obscures the records at the sea bed surface up to 10-15 m; Sheriff and Geldart, 1982), closing of the airgun chamber

*Sparker* Sparkers, like boomers convert electrical energy into acoustic energy. A sparker discharges a large amount of electrical energy into salt water (does not work in fresh water) producing an acoustic pulse. The number and spacing of tips, and the energy supplied to them determines the power and acoustic spectrum of the sound (Chapman, pers. comm., 1998). Little data collected via the sparker system was observed for this study. The sparker system does not have the resolution of the Huntec system, or the power and penetration of the airgun and sleeve gun systems.

General Specifications (Chapman, pers. comm., 1998):

| | |
|---|---|
| Sparker: | volts: 3.5-4 kV; shot rate: 1-4 sec depending on water depth |
| Bandwidth: | 0.08-2 KHz |
| Resolution: | <2 m |
| Penetration: | to bedrock surface depending on the power of the source (through bedrock if Tertiary or Cretaceous sediments, or to acoustic basement) |
| Noise: | seabed multiples (long-path multiples), short-path multiples, bubble pulse |

For each type of record, a horizontal scale was determined by the speed of the vessel during the survey. Vertical scale was based on an assumption that surficial sediments are penetrated at a rate of 750m/1 second sweep. Scales were adjusted to reflect the various sweep rates (i.e. 1/4 second, 1/2 second, 1 second etc.) used in recording the data.

Navigation was provided by LORAN-C, a system which uses intersecting low frequency radio patterns for positioning with an accuracy of 100-200 m (Loncarevic et al., 1992). Currently, differential GPS (Global Positioning System), which utilizes satellites for continuous position fixes, is used for navigation providing accuracy of +/- 5m (Loncarevic et al., 1992; Johnston, et al., 1995). The type of system used for navigation will affect the accuracy of fix locations on the seismic data collected from various years.

### 3.2.3: Data Interpretation

As discussed above (section 3.2.1), it is very important to understand the potential types of noise associated with seismic reflection data acquisition in order to interpret accurately the output records (most significant types of noise were discussed in section 3.2.2). In addition, it is also important to recognize the resolution limitations of each system and, when possible, to use more than one type of record acquired by systems with differing resolution capabilities in order to interpret accurately the lithological units which appear on the seismic reflection records. Interpretations over various areas of the eastern Scotian Shelf were limited by a lack of complimentary data acquisition from more than one type of acoustic source system (some areas were limited to only high resolution, shallow penetration boomer data, while other areas were limited to lower resolution, deeper penetration airgun or sparker data). This problem was not associated with every cruise.

#### *3.2.3a: Seismic Stratigraphy*

Seismic profiles are time-related and dependent upon acoustic impedance contrasts, therefore they are not true geological cross-sections, but representative of geological cross-sections (Stoker et al., 1997). Interpreting seismic profiles is dependent on the reflection character, configuration, and external geometry. Reflection character is described by its amplitude (acoustic impedance contrast between strata), frequency (dependent on bed thickness), and continuity (continuity of stratal surfaces) (Stoker et al., 1997). Reflection configurations are broadly classed into stratified (i.e. parallel, ponded, divergent, clinoformal; representing quiet and dynamic depositional environments), chaotic (diamicton-dominated deposits), and reflection-free (massive uniform lithologies) (Stoker et al., 1997). Finally, the external geometry gives information on the environment of deposition (i.e. sheet, sheet drape, wedge, bank, lens, mound, fan, channel fill, slope-front fill, and basin fill (Mitchum et al., 1977a; Stoker et al., 1997). Reflection geometry, character, and configuration form the basis of seismic facies analysis (determining the environment of deposition; Kearey and Brooks, 1991; Fig. 3.8).

Seismic sequence analysis involves the recognition of conformable or concordant seismic reflections (depositional sequences) and bounding unconformities or correlative conformities (Stoker et al., 1997; Kearey and Brooks, 1991). Reflections are most often produced where stratal surfaces or unconformities occur and are, therefore, taken to represent boundaries between depositional sequences, or in other words, where concordant or near-concordant reflections terminate against discordant reflections or adjacent seismic sequences (Kearey and Brooks, 1991). The major defining boundaries of seismic sequences include erosional, toplap, onlap, downlap, and concordant (Kearey and Brooks, 1991). Onlap, downlap, and toplap characterize a non-depositional unconformity, while truncation indicates an erosional unconformity (Stoker et al., 1997; Fig. 3.8).

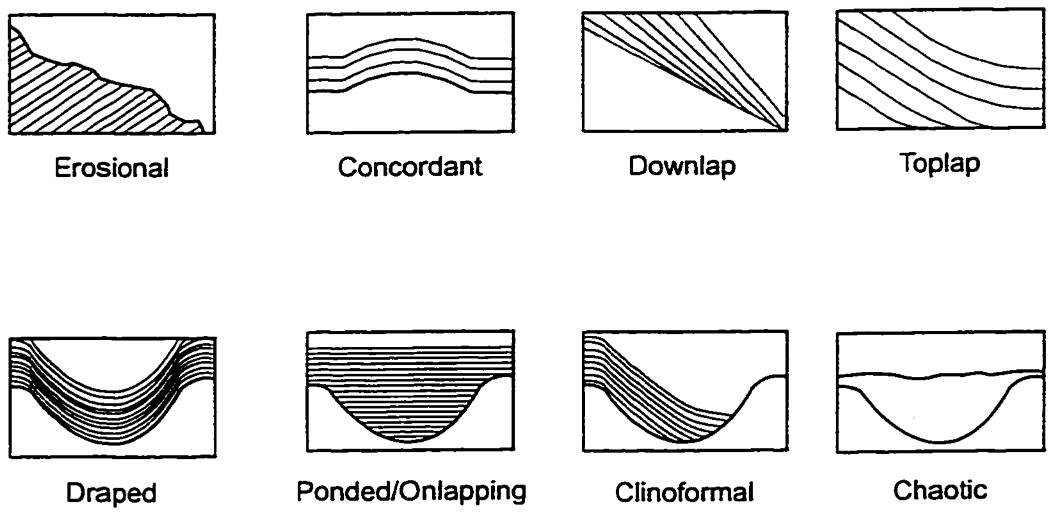



Figure 3.8 A few of the most commonly observed seismic reflection boundaries and configurations.

### *3.2.3b: Seismic Stratigraphy of the Scotian Shelf*

As per section 2.2.1, the Quaternary stratigraphy of the Scotian Shelf consists of the following formations (section 2.2.1 provides a detailed description of each):

- Scotian Shelf Drift
- Emerald Silt, facies A
- Emerald Silt, facies B
- Emerald Silt, facies C
- Sambro Sand
- Sable Island Sand and Gravel
- LaHave Clay

King and Fader (1986) distinguished each formation in terms of its lithologic and acoustic character (Table 2.1). This classification, based on groundtruthing and seismic signature, is in wide use, and provides the foundation on which interpretations of the Scotian Shelf surficial sediments are based.

## 3.3: DATA ANALYSIS

Tracklines from each cruise for which seismic data were analyzed were plotted on mylar overlays displaying the date and time of each line in thirty minute intervals. Tracklines for which the accompanying seismic data were marked by fix numbers rather than date and time were revised to display fix numbers for output on the mylar overlays. This was done by going back to the original bathymetry maps used in the field for logging fix locations, picking the location off according to the recorded time on the tracklines stored in the AGC data base and converting the corresponding day and time back to the original fix number. The mylar overlays displaying the tracklines were produced at the same scale as an accompanying bathymetric image of the eastern Scotian Shelf. By placing the mylar overlays on the bathymetric image, a fairly accurate correlation between seismic data and bathymetry was possible for mapping purposes. Seismic data were analyzed and notes of features such as the channels of this study, and mapping of surficial and channel fill sediments, were recorded on the mylar overlays. This mapping provided the primary foundation on which the results and conclusions of this thesis are based. The mylar overlays are equivalent to field notes and, therefore, are not included in the thesis.

# CHAPTER 4: RESULTS

## 4.1: INTRODUCTION

This chapter presents the results of this study. A zoned classification of the exposed incisions of the eastern Scotian Shelf is presented and described. This classification forms the basis for the discussions concerning the origin of the morphology of the eastern Scotian Shelf presented in Chapter 5.

### 4.1.1: Incisions Identified on the Eastern Scotian Shelf

Buried incisions have been identified through seismic reflection profiling techniques, on the outer eastern Scotian Shelf under Sable Island Bank and Banquereau, (King et al., 1974; King and Fader, 1986; Amos and Knoll, 1987; Boyd, et al., 1988; McLaren, 1988; Amos and Miller, 1990). Some of these incisions are incised into Tertiary and Cretaceous sedimentary bedrock exceeding depths of 400 m (from flank to base of channel), with widths averaging 2-3 km. Buried incisions, some reaching in excess of 110 m in depth, have also been identified under Middle Bank and Canso Bank (this study). Due to data penetration limitations, many of these incision bases extend beyond observable depths, and it is unknown if the incisions exceed 400 m in depth. A similar situation occurs on the northern edge of Banquereau where incisions >125 m in depth have been identified (this study). Additionally, Misaine Bank and areas bordering the banks of the eastern Scotian Shelf, are extensively dissected by complex patterns of exposed incisions, although those covering Misaine Bank are of a much smaller scale than those bordering the bank.

Incision scales vary greatly over the eastern Scotian Shelf and display both interconnected networks and isolated linear incisions with cross-cutting relationships in some areas. Incisions represent different generations as identified by multiple incision cut and fill sequences observed on the Scotian Shelf banks (Amos and Knoll, 1987; McLaren, 1988; Amos and Miller, 1990; King et al., 1994). Explanations proposed for the origin of the eastern Scotian Shelf incisions include subglacial meltwater erosion (catastrophic and steady-state), glacier erosion, and pre-glacial fluvial erosion (King and MacLean, 1976; Amos and Knoll, 1987; Boyd, 1988; McLaren, 1988; King et al., 1994; Stea, 1995; section 1.3).

### 4.1.2: Exposed Incisions as an Analogue to Buried Incisions

This study concentrates on the origin of the exposed incisions of the eastern Scotian Shelf, though the conclusions may also apply to buried incisions. Other than the complex system of incisions observed on the surface of Misaine Bank, most of the exposed incisions are similar in width and depth (~3-5 km wide, ~150-350 m deep), to those studied by Boyd et al. (1988) buried beneath Sable Island Bank, and Amos and Knoll (1987) buried beneath Banquereau. These incisions are particularly intriguing because of their enormous size.

In areas where incisions are buried, seismic reflection resolution and detail of incision fill are compromised in order to penetrate through the hard surface sediments (sand and gravel in the case of the Scotian Shelf banks) covering the incised morphology (section 3.2.2). This problem was encountered by Boyd et al. (1988) who used lower resolution (2 sec sweep) industry seismic data to determine incision fill lithology and morphology for the incisions buried beneath Sable Island Bank. The orientation and contents of incisions studied by Boyd et al. (1988) were interpolated in areas not covered by seismic survey lines. If these incisions were exposed and not buried, they may resemble the larger scale exposed incisions revealed in the Scotian Shelf bathymetry (Fig. 1.2; excluding those covering Misaine Bank) which appear to be linear and display cross-cutting relationships. In deeper water areas, over partially filled incisions, acoustic interference from sea bed multiples was reduced on the profiles. The acoustic window between multiples is larger, and hard surface sands and gravels characteristic of the banks are absent. Therefore, the eastern Scotian Shelf is of particular interest due to the fact that many incisions are incompletely filled with sediments. This results in the ability of high resolution systems, such as the Huntec DTS, to provide details of incision fill sediments unavailable in lower resolution records used to determine the fill of completely buried incisions. As a result of the detailed sedimentology obtained through high resolution seismic profiles for the exposed/incompletely filled incisions of the eastern Scotian Shelf, only these incisions will be considered in detail in this study. Depositional and lithological information interpreted from these incisions can be used as an analogue and applied to buried incisions with similar characteristics, but lacking in detailed, high resolution, high S/N, seismic records.

## 4.2: METHODS USED FOR CLASSIFYING THE EXPOSED INCISIONS OF THE EASTERN SCOTIAN SHELF

### 4.2.1: Basis of the Zone Classification

Based on interpretation of high-resolution marine seismic reflection data collected from over 6000 km of tracklines (Huntec DTS, airgun, sleeve gun, and sparker systems; Fig. 3.1) and analysis of bathymetric data (Fig. 1.2), areas of exposed incisions on the eastern Scotian Shelf have been recognized as being distinct in character. The exposed incisions have been classed into four primary zones from the inner shelf to the outer shelf; Zone A, Zone B, Zone C (including Subzones $C^1$ and $C^2$), and Zone D (Fig. 4.1). This classification is based on incision size, plan-form patterns, orientation, location, and fill characteristics. Incision cross-profiles are highly variable and do not contribute to this zonal classification. A discussion of the significance of incision cross-profiles is presented in section 4.4.

Figure 4.2 shows the location of the seismic profiles (Figs. 4.3 to Fig. 4.21) which are discussed in the zone classification for the exposed incisions of the eastern Scotian Shelf. Incision sediments are described according to acoustic character, and interpreted according to the discussions in section 4.2.1a (Seismostratigraphy). These interpretations follow the acoustic descriptions in brackets. The seismic profiles (Figs. 4.3 to 4.21) are labeled according to the interpretations (see section 4.2.1a)

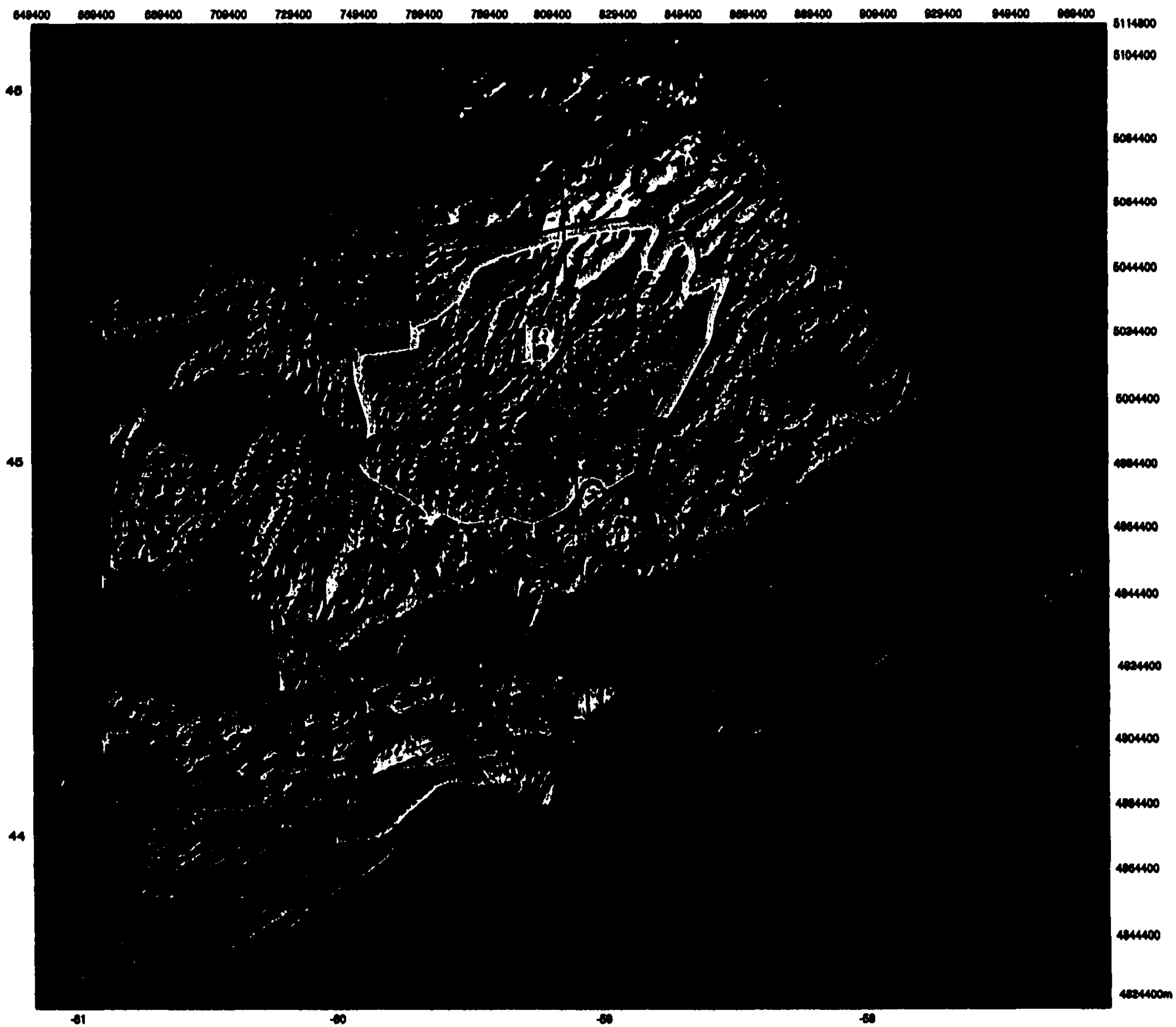
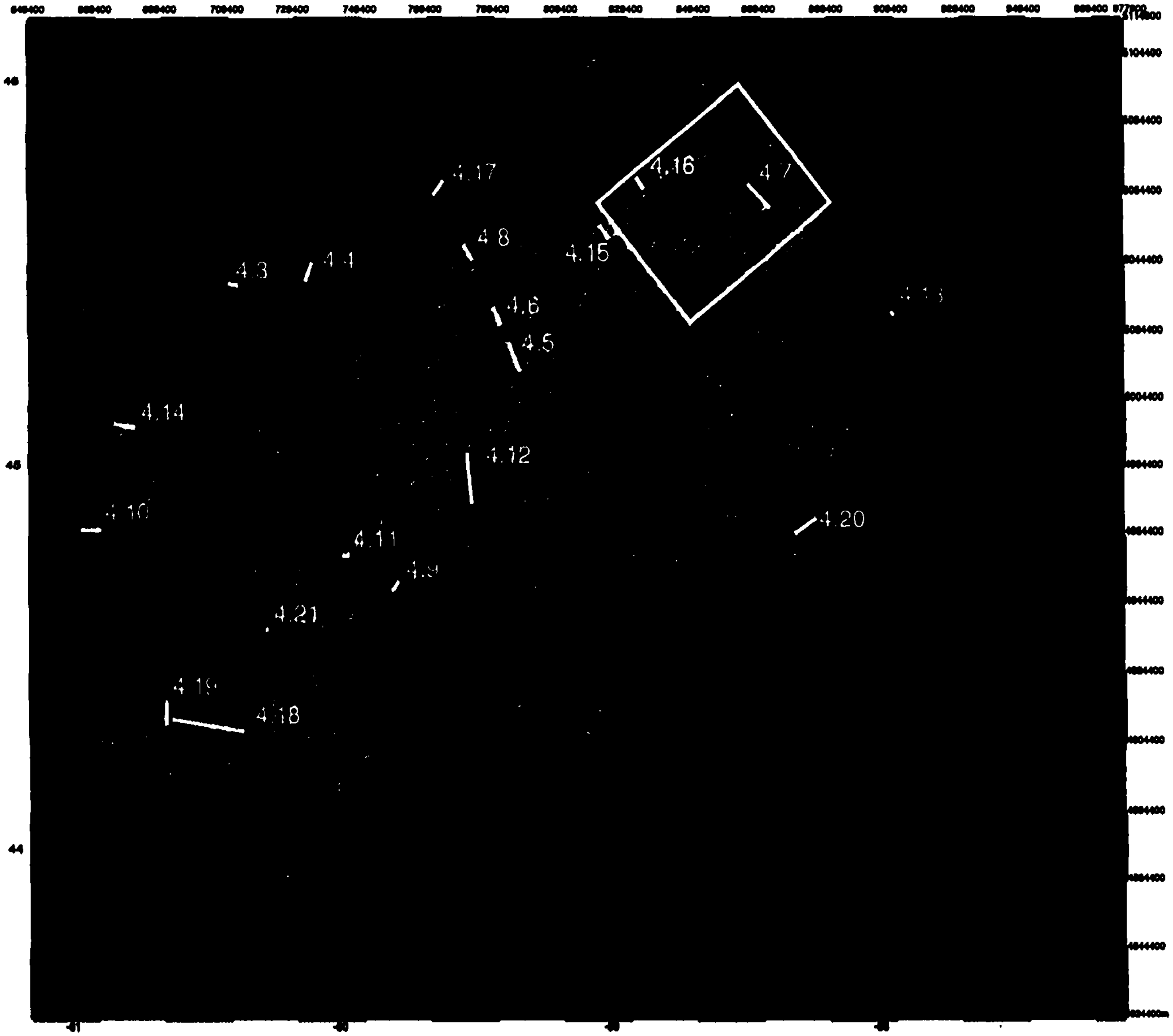



Figure 4.1 Zone classification for the exposed channels of the eastern Scotian Shelf, Atlantic Canada.



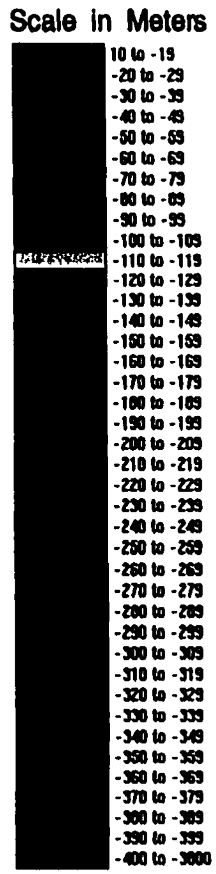




Figure 4.2 Location of seismic reflection profiles (figures 4.3 - 4.21). (Compare boxed area in this figure to boxed area in figure 5.1)

#### 4.2.1a: Seismostratigraphy

This section introduces the descriptions and interpretations used for the seismic profiles presented in the zone classification results. Alternative interpretations for a particular seismic signature are provided here. However, interpretations which best fit the seismic signature and geometry recorded on the seismic profiles are used in the text (in brackets) and for labeling the figures.

Fill towards the base of the exposed incisions of the eastern Scotian Shelf appears to be highly variable as revealed by the limited quantities of seismic reflection records. No cores have penetrated into the basal deposits of the more deeply incised Zone C and Zone D incisions (Fig. 4.1). Thus, the available airgun and sleeve gun data provide the only means for determining the nature of the incision fill beyond the resolution of the Huntec DTS data. Airgun/sleeve gun data for Subzone $C^2$ are limited, the majority of data being derived from the west of the Subzone. In Zone B, where the incisions are shallow, Huntec DTS data are lacking. Details of the incision fill in this zone were obscured by the bubble pulse of the available airgun data, and therefore cannot be resolved.

A detailed description of the Quaternary sediments of the Scotian Shelf, including acoustic character and lithology, is provided in Chapter 2 (section 2.3.2; Table 2.1). This is the foundation on which most of the interpretations for the seismic profiles were based. The following overview gives alternative interpretations for various seismic signatures.

Acoustically incoherent units can occur as extensive blankets overlying bedrock and outcropping at the sea floor in some areas. These units have been well sampled where they outcrop at the sea floor (grab samples and core; King and Fader, 1986). Sediment sampling combined with acoustic signature and geometry support the interpretation of regional till deposits (see Scotian Shelf Drift, section 2.3.2). Other interpretations for mounds of acoustically incoherent units include moraines, lift-off moraines (described in section 2.3.3b), and till tongues (described in section 2.3.3a).

Unsampled incision fills with acoustically incoherent seismic signatures are common. Therefore, interpretations are derived solely from the signature on the seismic records. The acoustically incoherent deposits may be the result of a number of depositional processes including; (1) glacial (till and glacial diamict; see section 2.3.1 for explanation of the use of these terms), (2) secondary deposition (i.e. debris flow deposits, side wall failure, slumping), (3) glaciomarine diamicts, or (4) deposition in a high-energy, variable environment (subglacial meltwater deposits). Where a genetic interpretation could not be made with confidence (based on acoustic signature and geometry), the deposit is referred to as acoustically incoherent in the text and on the figures.

Some incisions contain stratified fill displaying variable acoustic amplitudes which are generally stronger than those that characterize the Emerald Silt formation (section 2.3.2). Such units appear to be predominantly horizontally stratified and display evidence of both ponding and onlapping. No samples have been recovered form these stratified basal units of incision fill, and interpretations rely on seismic stratigraphy. The seismic signature of these deposits is characteristic of marine sedimentation, represented by relatively uniform rates of deposition during sea-level rise. The higher acoustic amplitude of these sediments suggests lithification and these sequences resemble Tertiary and Cretaceous bedrock signatures as described by McIver (1972), King et al, (1974), and King and McLean (1976), which are based on correlation between seismic reflection records, and core from exploration wells, land geology, well data, and gravity, magnetic, and seismic refraction data.

Therefore, the incisions infilled by these sediments are interpreted to be Tertiary in age, thereby eliminating a glacial origin. The identification of Tertiary age incisions (fluvial valleys) is important, as buried incisions observed in residuals may not be glacial in origin and the Pleistocene glacial unconformity may be above such incisions.

Some incisions contain horizontally stratified ponded to onlapping sediments with a weak signature compared to the signature assumed to represent Tertiary sediments. This may be real, or an artifact of the data collection. Because these sediments are found towards the base of some incisions, they do not possess the conformably blanketing geometry of Emerald Silt facies A and B, therefore, they cannot be interpreted as glaciomarine with confidence. Sediments displaying ponded/onlapping stratified reflections suggest deposition in a more dynamic environment controlled by current activity than sediments displaying draped stratified reflections (Stoker et al., 1997). Possible interpretations for these sediments include: (1) glaciomarine; (2) interglacial marine sedimentation; (3) Tertiary sediments; (4) sedimentation in local subglacial lakes. Sediments showing stacked sequences with differing acoustic signatures may represent alternating deposits of glacial diamict, glaciomarine and normal marine deposits. Stratified sediments for which an interpretation with confidence cannot be made will be referred to as stratified sediments in the text and on the figures.

### *4.2.1b: Age Determination*

The age of the eastern Scotian Shelf incisions was assessed by both seismic signature (i.e. Tertiary fill corresponds to Tertiary fluvial erosion) and the identification of unconformities. Unconformities identified on the Scotian Shelf and outer eastern Scotian Shelf banks (Banquereau and Sable Island Bank) were presented in sections 1.2 and 2.2.3.

The R3 unconformity, which is on the bedrock surface signifies the onset of glaciation and is attributed to subglacial erosional processes (King et al., 1974; Amos and Knoll, 1987; McLaren, 1988; Boyd et al., 1988; Amos and Miller, 1990; Fig. 1.3d). Incisions below the R3 unconformity can be assumed to be older than at least the Late Wisconsinan. Without dates or core, it is not possible to determine the age of sediments infilling the basal portion of incisions. Many incisions may largely be inherited from the Tertiary (as indicated by Tertiary sediments filling some of the incisions) with R3 superimposed. However the R3 unconformity is difficult to identify within incision fill. Only cores from which dates can be derived and lithologies can be verified can help in solving this problem. Unfortunately, this data does not exist and the identification of R3 is the best method for determining the age (Pleistocene or Tertiary) of the incisions.

## 4.3: ZONE CLASSIFICATION FOR THE EXPOSED INCISIONS OF THE EASTERN SCOTIAN SHELF

The zone classification will be presented in the sequence in which the zones appear on the eastern Scotian Shelf (i.e. from the inner shelf to the outer shelf); Zone A, Zone B, Zone C (composed of Subzone $C^1$ and Subzone $C^2$), and Zone D (Fig. 4.1).

### 4.3.1: Zone A

In Zone A (Fig. 4.1), incisions occur close to the inner shelf and consist of incompletely filled Pleistocene incisions, blanketed by glaciomarine sediments (Emerald Silt) with an erosional unconformity at the surface of the glaciomarine sediments (Emerald Silt; Figs 4.3 and 4.4) Zone A incisions cut into both outcropping acoustic basement of the Appalachian Region in the north of the zone, and thin Tertiary and/or Cretaceous bedrock towards the south of the zone. Portions of Zone A are not covered in the digital bathymetric map, and thus it is unclear whether some features on the airgun seismic records represent small, shallow surface incisions or undulating bedrock morphology. Zone A incisions generally display a north-south orientation, except to the west (north of Canso Bank) where they appear to trend toward the southwest.

The fill of many of the shallow incisions in the northern part of the area cannot be identified because resolution is low within the bubble pulse of the airgun seismic reflection profiles. However, towards the northwest of Zone A, there is a series of incisions <60 m deep and <1 km wide which appear to be filled with stratified deposits, some displaying clinoforms (Fig. 4.3). This type of seismic character is generally believed to be associated with prograding sedimentary systems. When the pattern is sigmoidal as represented in Figure 4.3, the depositional process may represent: (1) a low-energy regime with a low sediment supply; (2) rapid basin subsidence; or (3) rapid sea level rise resulting in aggradational topset successions (Stoker et al., 1997). These sediments may be representative of infilling during post-glacial sea level raise. incisions of similar dimensions in the same area contain what appears to be horizontally stratified glaciomarine sediments (Emerald Silt). Possible sediment

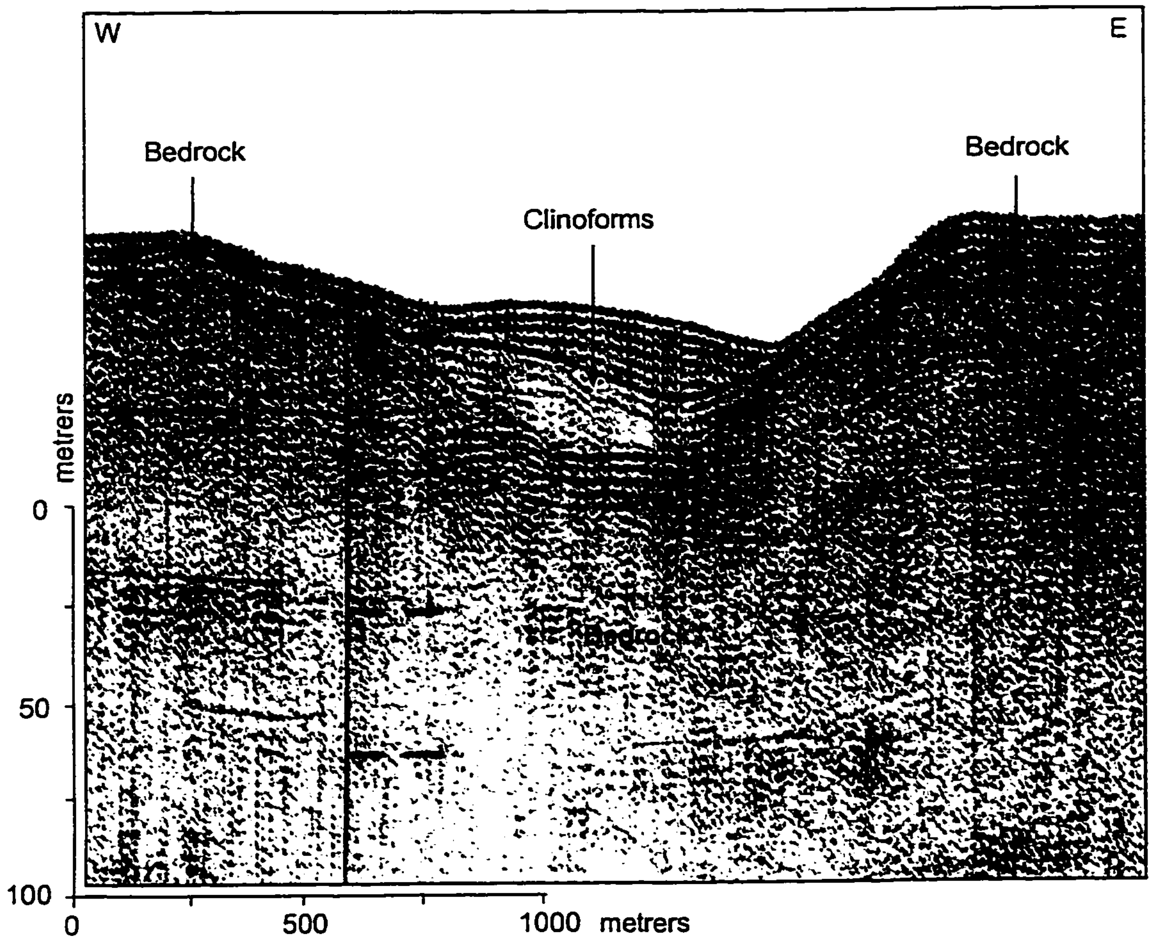



Figure 4.3. Airgun profile of an incision in Zone A. This incision displays sigmoidal clinoforms as incision fill.

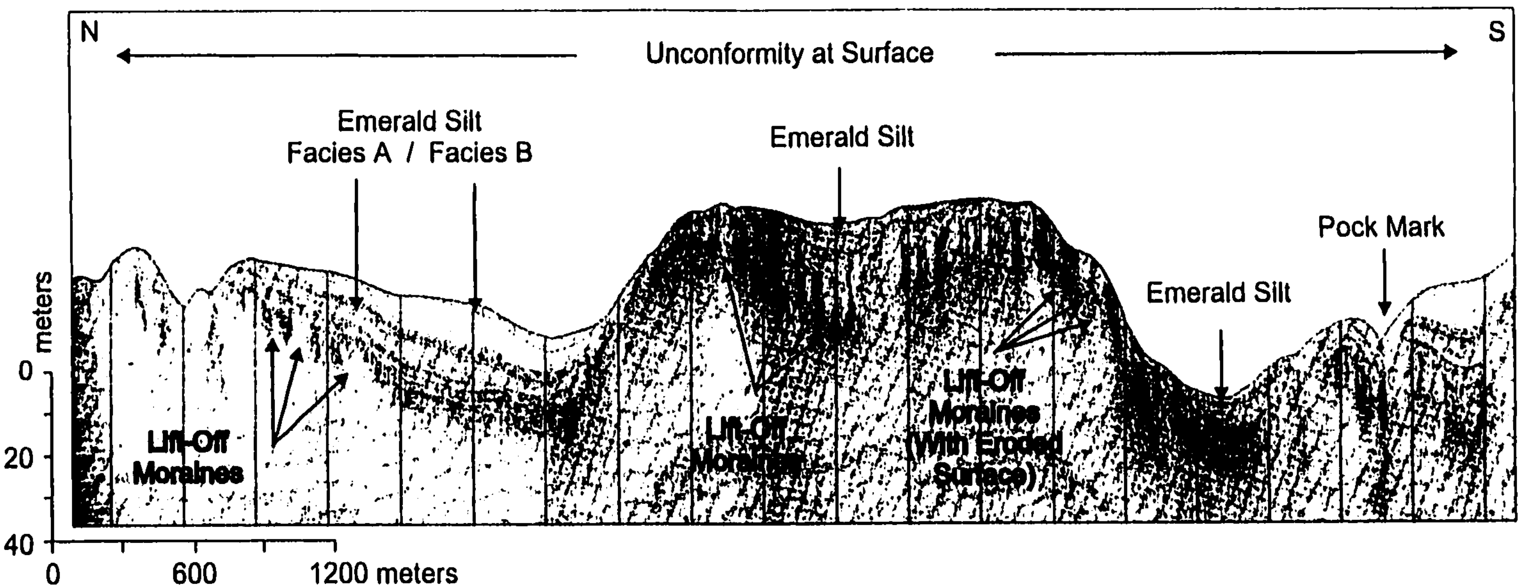



Figure 4.4 Huntec DTS profile of incisions in Zone A. Note how surface incisions related to the regional unconformity are preferentially eroded into softer glaciomarine sediments (Emerald Silt), while the overall topography reflects the existence of a more resistant undulating basal surface (bedrock and Scotian Shelf Drift; bedrock surface is unresolved in this profile).

sources include glaciomarine sedimentation, or reworked glaciomarine sediments, glacial till, and/or glaciofluvial sediments.

Some incisions towards the south of Zone A, near Canso Bank and Misaine Bank, are larger than those to the north, and reach maximum observed depths of 120 m and widths of 1 km. There appears to be an acoustically incoherent fill (till) at the base of some of these channels, overlain by stratified glaciomarine deposits (Emerald Silt). These sequences are typical of many Zone C incisions and are described more completely in section 4.3.3.

A regional unconformity occurs across the surface of this region, over Canso Bank and down through the area between Canso Bank and Misaine Bank. This unconformity may be related to the R1 reflector identified on Sable Island Bank and Banquereau (Amos and Knoll, 1987; McLaren, 1988; Amos and Miller, 1990) and was likely formed during late glacial and Holocene marine transgression as sea level rose and ocean circulation began to take on its present character. Differential erosion of the surface glaciomarine sediments (Emerald Silt) in this area left incoherent acoustic units (interpreted as till and lift-off moraines based on geometry and seismic character; see section 2.3.3b) as resistant high surfaces (Fig. 4.4).

Two linear ridges (R) between Canso Bank and Misaine Bank are evident in the bathymetry (Fig. 4.1), however the composition of these ridges (bedrock and/or glacial diamict) has not been determined. Part of these ridges has been mapped as Cretaceous Bedrock (King and MacLean, 1976). These ridges mark the southern boundary of Zone A, though in this area incisions in Zone A are transitional to Subzone $C^2$.

### 4.3.2: Zone B

Zone B (Misaine Bank; Fig. 4.1) is underlain by few buried incisions whereas the surface is heavily dissected by a system of exposed and partially infilled incisions varying in depth from <10m to 100m. Zone B incisions display more of a distributary or tributary system (channels converging to or diverging from a common point ) than an anabranching system (channels bifurcating and rejoining). It is difficult to follow a single incision displaying bifurcating and rejoining, a characteristic of anabranching systems. It remains unknown whether the apparently small number of buried subglacial incisions is real, or a result of poor seismic survey coverage and penetration.

Zone B incisions are relatively shallow when compared to the incisions of other zones (Figs. 4.3, 4.5), but there is some evidence of buried Pleistocene and reworked Tertiary age incisions/valleys (Fig 4.6). Some buried incisions in Zone B appear to be filled with Tertiary bedrock, which was subsequently eroded during the Pleistocene (Fig. 4.6). These, and other buried Pleistocene incisions are filled with acoustically incoherent sediments at the base, and stratified deposits (Emerald Silt) towards the surface (Fig. 4.5). Fill in the shallow surface incisions remains largely unknown as high-resolution Huntec DTS data are lacking, and near surface detail is obscured by the bubble pulse of the available airgun data.

A recent sidescan sonar survey (January, 1998, CCGS Teleost), used to search for the bow section of the January 1998, Flare shipwreck, recovered the only high resolution data for the Misaine Bank area. Incision fill for the small surface incisions (Fig. 4.5) still remains largely unknown, but the sidescan sonar data revealed that the incision areas (low areas/depressions) are covered with

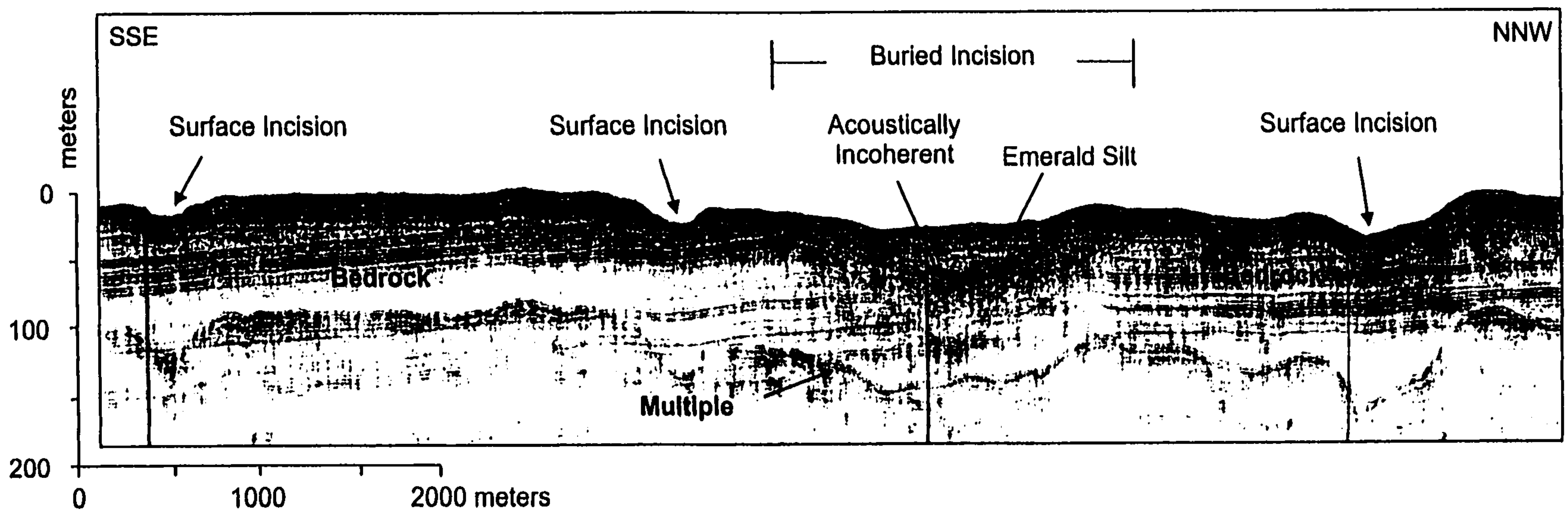



Figure 4.5 Airgun profile across Zone B (Misaine Bank). This zone is underlain by very few buried incisions. The surface is dissected by shallow incisions (<100 m deep), some of which are partially infilled with Quaternary sediments.

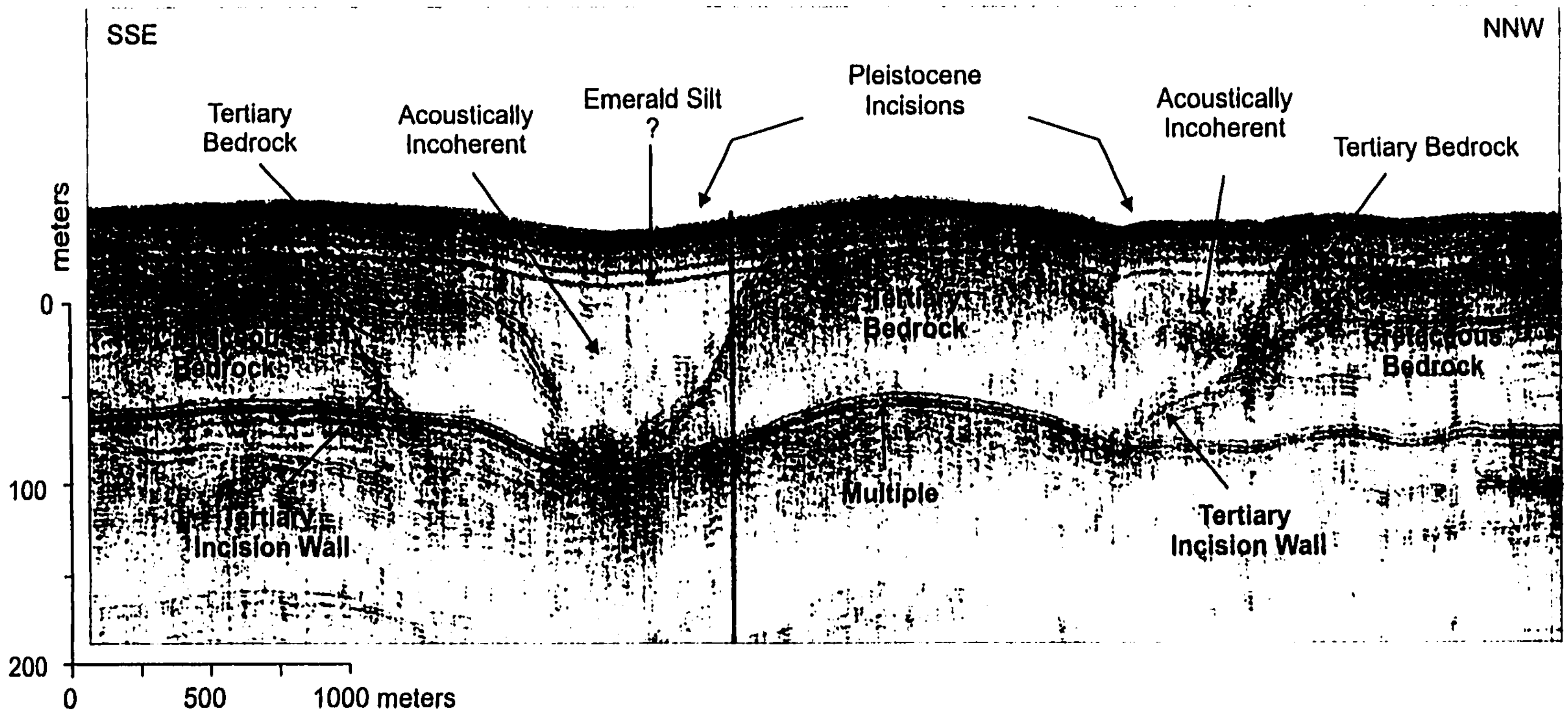



Figure 4.6 Airgun profile of buried Incisions in Zone B. This profile depicts a buried Tertiary incision Filled with Tertiary sediments with two superimposed Pleistocene incisions.

sand at the sea bed, whereas the higher surfaces are covered with gravels at the sea bed.

In the modern marine environment, sand tends to be deposited in bathymetric lows and may be the result of reworking of Emerald Silt deposits in the low areas and deposition of winnowed sands from the higher surfaces. The gravels may represent a lag deposit on the high surfaces produced by winnowing of the finer sediments from glacial diamict deposits. Extensive areas of iceberg scours were also identified on the seabed in this zone which may have important implications regarding estimates of glacial low sea-level stands depending on whether the scours are relict or modern (section 2.3.3b).

### 4.3.3: Zone C

Zone C, composed of Subzones $C^1$ and $C^2$, consists of a series of deeply incised incisions, some of which reach in excess of >300m deep (Fig. 4.1). Some incisions are completely buried, but most are exposed and incompletely filled. The incisions east and north of Misaine Bank (Subzone $C^1$) are oriented northeast-southwest, similar to the shallow incisions on the surface of Misaine Bank. Those to the south (Subzone $C^2$), have a north to south orientation (slightly skewed to the southeast or southwest) and appear to cross-cut the southern edge of the northeast-southwest trending Misaine Bank (Zone B) incisions in the area marked BC (Fig. 4.1). Unlike the south-west oriented Subzone $C^1$ incisions, Subzone $C^2$ incisions are more linear and isolated in character and do not appear to have tributary/distributary extensions.

Most of the uppermost sediments of Zone C incisions consist of depositional sequences typical of the Scotian Shelf basins to the west (From bottom to top: acoustically incoherent unit—Scotian Shelf Drift/glacial diamict;

stratified, conformably draping to ponded acoustic units—ice proximal and ice distal glaciomarine deposits of Emerald Silt Facies A and B; transparent acoustic unit—marine mud deposits of LaHave Clay, see section 2.3.2 and 4.2.1a for explanation of interpretations). Most sediments at depth, below those described above, are of an unknown age, and acoustic character varies from stratified to acoustically incoherent.

Subzone $C^1$ incisions, for which airgun data are available, appear to contain variable amounts of a basal acoustically incoherent unit covered by thick deposits of conformably draping glaciomarine sediments (Emerald Silt; Fig. 4.7). Most of these incisions have predominantly U-shaped profiles (Figs. 4.7, 4.8).

Two incisions observed in Subzone $C^1$ show that similar U-shaped cross-profile profiles are filled with contrasting materials (Fig. 4.8). One incision is almost completely filled with acoustically incoherent sediments, while the other is filled with stratified sediments (Emerald Silt or Tertiary bedrock). This observation indicates that incision fill is highly variable, even for incisions in close proximity and of similar shape. An unconformity (UC, Fig 4.8) extends over the surface of the incision containing stratified sediments and this indicates that this incision is older than the adjacent incision containing acoustically incoherent sediments. Because thick deposits of Emerald Silt are generally not found close to the inner shelf, reflections from the stratified channel fill are of a higher amplitude than usually observed for Emerald Silt, and an unconformity overlies the surface, this incision may be interpreted as Tertiary in age, filled with Tertiary sediments. However, the possibility of glaciomarine sediments from an earlier glacial event infilling the incision cannot entirely be discounted without samples to support interpretations. If the incision containing stratified sediments is Tertiary in age, it would represent a Tertiary fluvial valley and the

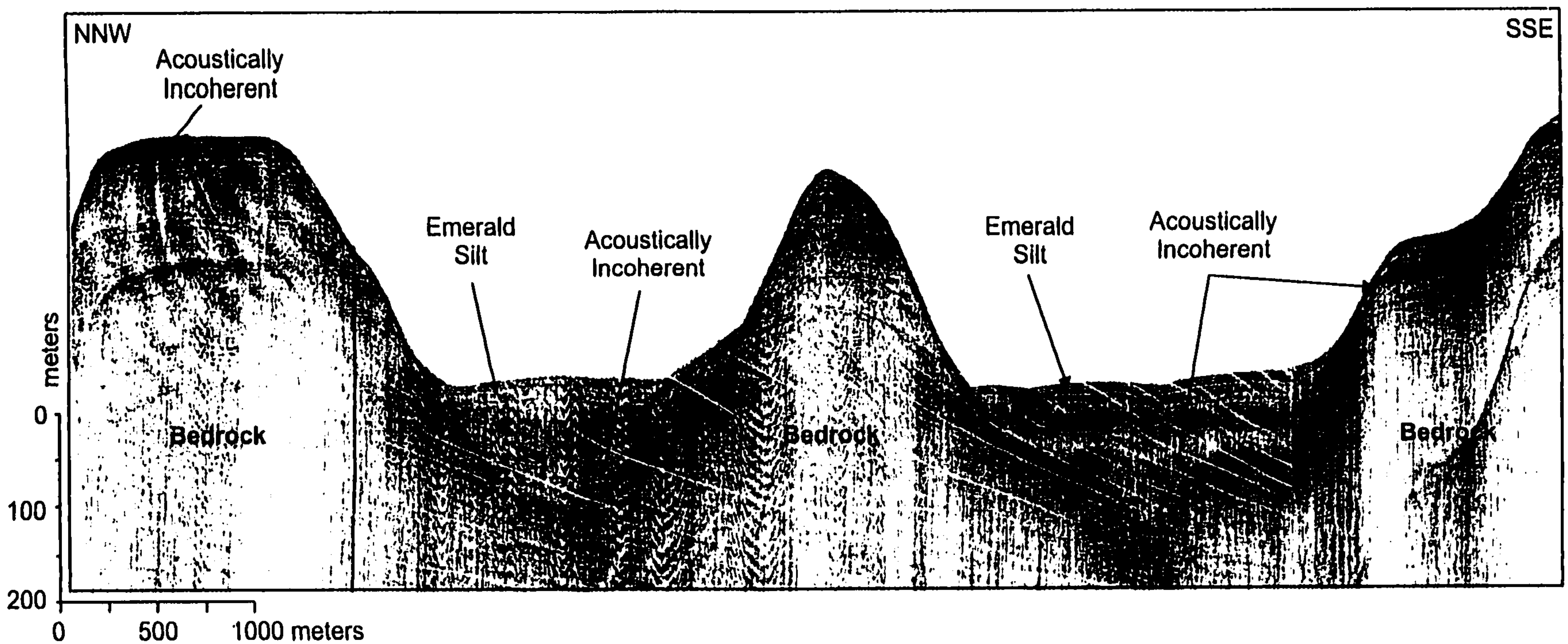



Figure 4.7 Airgun profile of incisions in Subzone C'. These incisions are part of an area of large incompletely filled incisions trending in a northeast-southwest direction. These deep incisions show a u-shape profile in the cross-section. The stratigraphic infill sequence consists of a basal incoherent unit (till) overlain by thick Emerald Silt sequences.

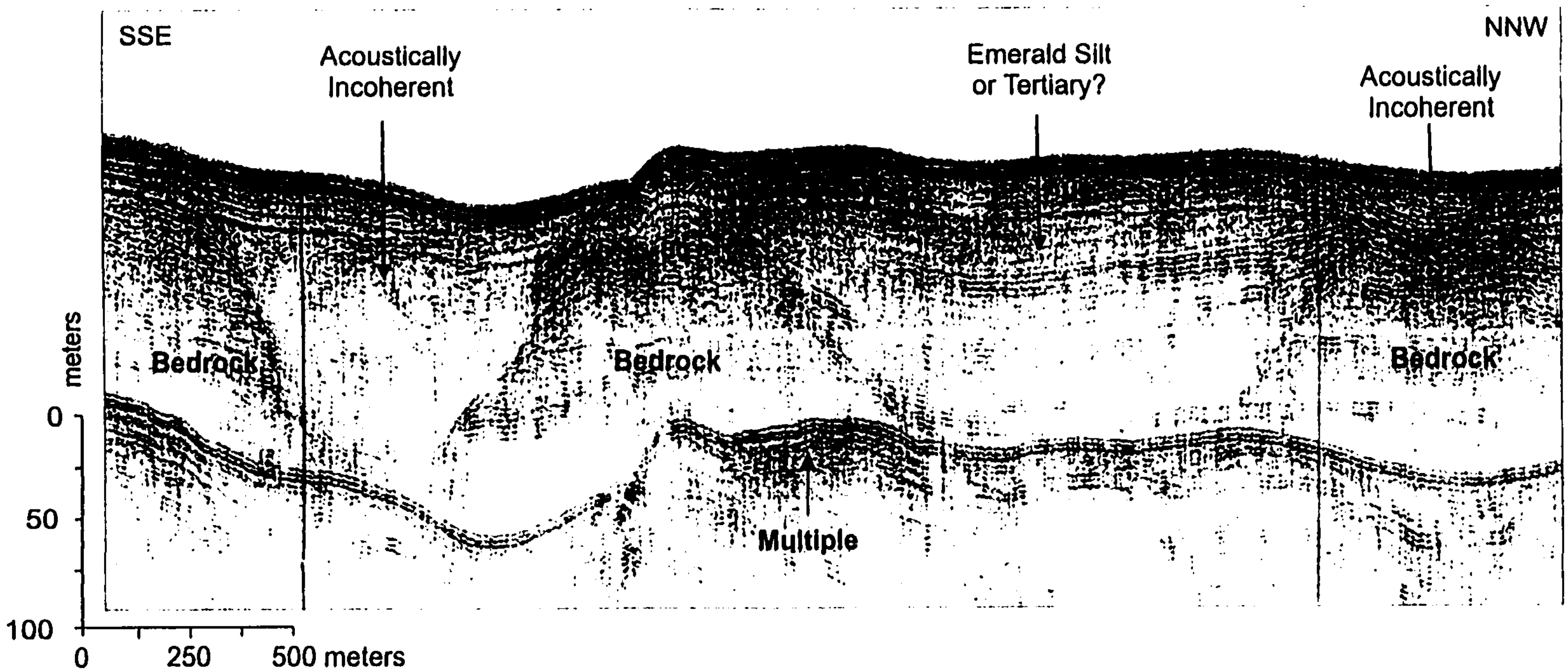



Figure 4.8 Airgun profile of two adjacent incisions in Subzone C'; one almost completely infilled, and one buried. Though these incisions are in close proximity, their fill is different and exemplifies unique depositional environments. One incision has an acoustically incoherent fill, while the other displays stratified deposits UC marks the surface of an unconformity which can be followed over the surface of the incision filled with stratified sediments indicating it may be of an older generation than the incision with acoustically incoherent fill.

adjacent incision may be the result of excavation of a Tertiary fluvial valley during glaciation.

Some Subzone $C^2$ incisions contain sediment sequences similar to those described for Subzone $C^1$, while others contain ponded to onlapping horizontally stratified deposits at depth (Fig. 4.9). The incision cross-sections appear to be parabolic in form. Figure 4.9 shows an incision filled predominantly with high amplitude reflections, horizontally stratified sediments interpreted as Tertiary sediments. Resolution is lost towards the surface of the incision, but it appears that conformably stratified glaciomarine sediments (Emerald Silt) blanket the surface of the incision. This incision appears to represent a Tertiary fluvial valley which was modified by glaciation towards the surface. It is located within the area of a Tertiary fluvial system previously mapped by King et al. (1974; Fig. 2.4).

An airgun profile from Subzone $C^2$ (Fig. 4.10) shows incisions cut into bedrock. A reflector representing a Quaternary unconformity related to subglacial erosion can be followed along the bedrock surface (uc). Incisions and sediments below this unconformity are within Tertiary/Quaternary pre-glacial formations, and therefore, would not be related to a glacial origin. A reflector occurs approximately at the middle of the eastern incision in the profile (r). This reflector separates an overlying acoustically incoherent unit (till) from an underlying incision fill of an undetermined origin. The flanks of the incision are bounded by an acoustically incoherent unit (till) to the east and bedrock to the west. This incision may be interpreted as a Tertiary fluvial valley in origin with glacial modification occurring above (r), or as completely glacial in origin with (r) separating two events. The unconformity cutting across the surface of the incision to the west is distinct. The stratified seismic signature of the incision fill closely resembles the signature of the bedrock, and appears to have been

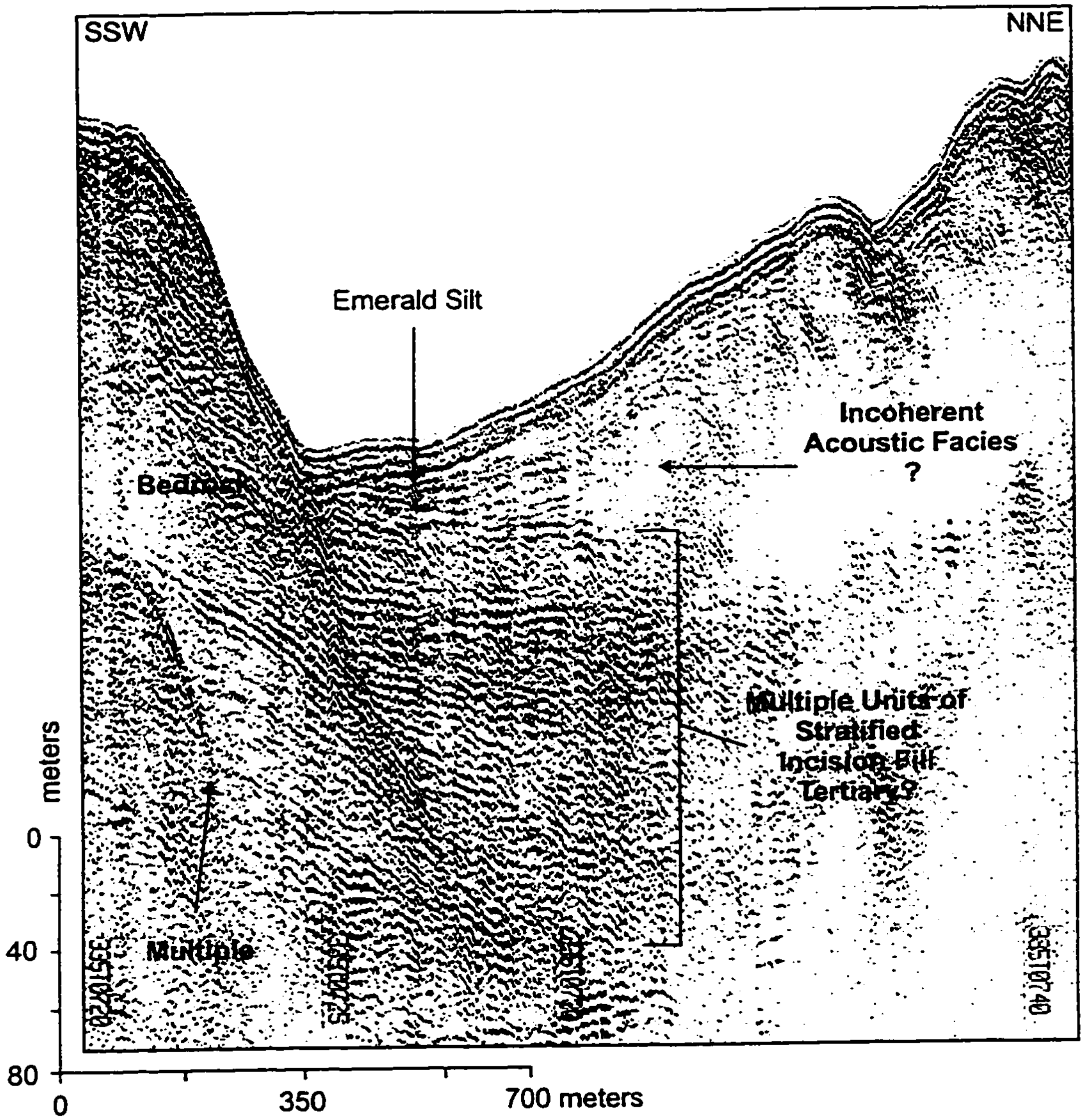



Figure 4.9 Arigun profile of Subzone $C^2$ incision showing incision fill at depths beyond the resolution of Huntec DTS data. Airgun coverage of Subzone $C^2$ is sparse, therefore, generalizations cannot be made regarding the nature of incision fill beyond Huntec DTS resolution. Note the contrast in basal fill between this profile Figure 4.7. Figure 4.7 shows an incoherent acoustic reflector basal unit with stratified deposits above, while this profile shows stratified deposits to the maximum depth of resolution. It resembles the lowest portion of incision fill of the buried incisions in Brandal Basin, Figure 4.12. These stratified sediments may be related to glacial or Tertiary infill. Based on the intensity of the reflectors, a Tertiary origin is favoured.

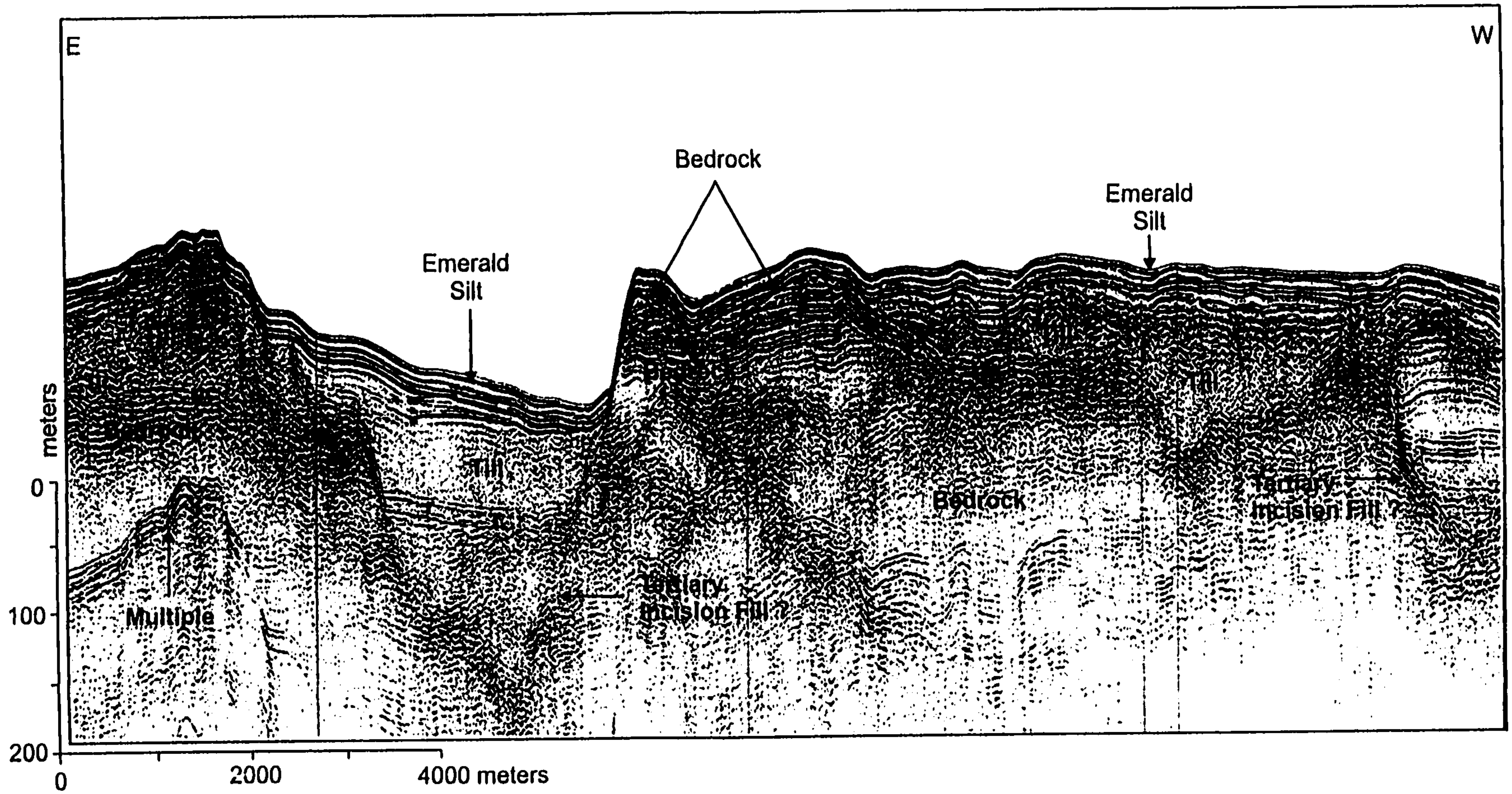



Figure 4.10 Sleeve Gun profile of incisions in Zone C². This profile depicts a Pleistocene unconformity (uc) which extends across the surface of a buried incision to the west. This incision is likely Tertiary, however, a subglacial incision related to a Pre Wisconsinan glacial event cannot be omitted as a possible origin. The incompletely filled incision to the east is bounded near the surface by a thick till deposit to the east, and bedrock outcrop to the west. This profile is a good depiction of incisions of more than one generation and complex bedrock reflections.

resistant to the erosional event producing the overlying unconformity. This incision is, therefore, interpreted as Tertiary in age containing Tertiary deposits. An alternative interpretation could be that the sediments are interglacial marine sediments if the incisions were eroded during earlier Pleistocene glacial episodes. Or, they may represent alternating deposits of glacial diamict, glaciomarine, and normal marine deposits. Such a sequence was observed by Sejrup et al. (1996) in the Norwegian Channel. There is not enough data or information to provide a definite interpretation of the age or origin of this incision.

High resolution Huntec DTS data over Zone C, including Subzones $C^1$ and $C^2$ (Fig. 4.11) indicates that the majority of incisions contain an acoustically incoherent unit which often displays a hummocky relief at the surface (glacial diamict/till with lift-off moraines; see discussion in section 2.3.3b), blanketed by conformably draping stratified glaciomarine sediments (Emerald Silt). The glaciomarine sediments are generally thicker in incisions than on the flanks of adjacent high ground. Patchy deposits of Holocene muds (LaHave Clay) often overly this sequence, but are not present in Figure 4.11. This sequence is typical of the upper incision sediments obscured by the bubble pulse or not well resolved in the airgun data (i.e. Fig 4.9 and, and the incision to the east in Fig. 4.10).

Figure 4.12 and figure 4.13 are high resolution Huntec DTS profiles across incisions located in Subzones $C^1$ and $C^2$ respectively and provide further insight into the character of surficial incision fill deposits. Both incisions depicted in the profiles extend off banks (Artimon Bank and Canso Bank respectively), and both display evidence of incoherent acoustic facies (till) overlain by glaciomarine sediments (Emerald Silt). Figure 4.12, an incision profile off Artimon Bank, adjacent to the Laurentian Channel, displays

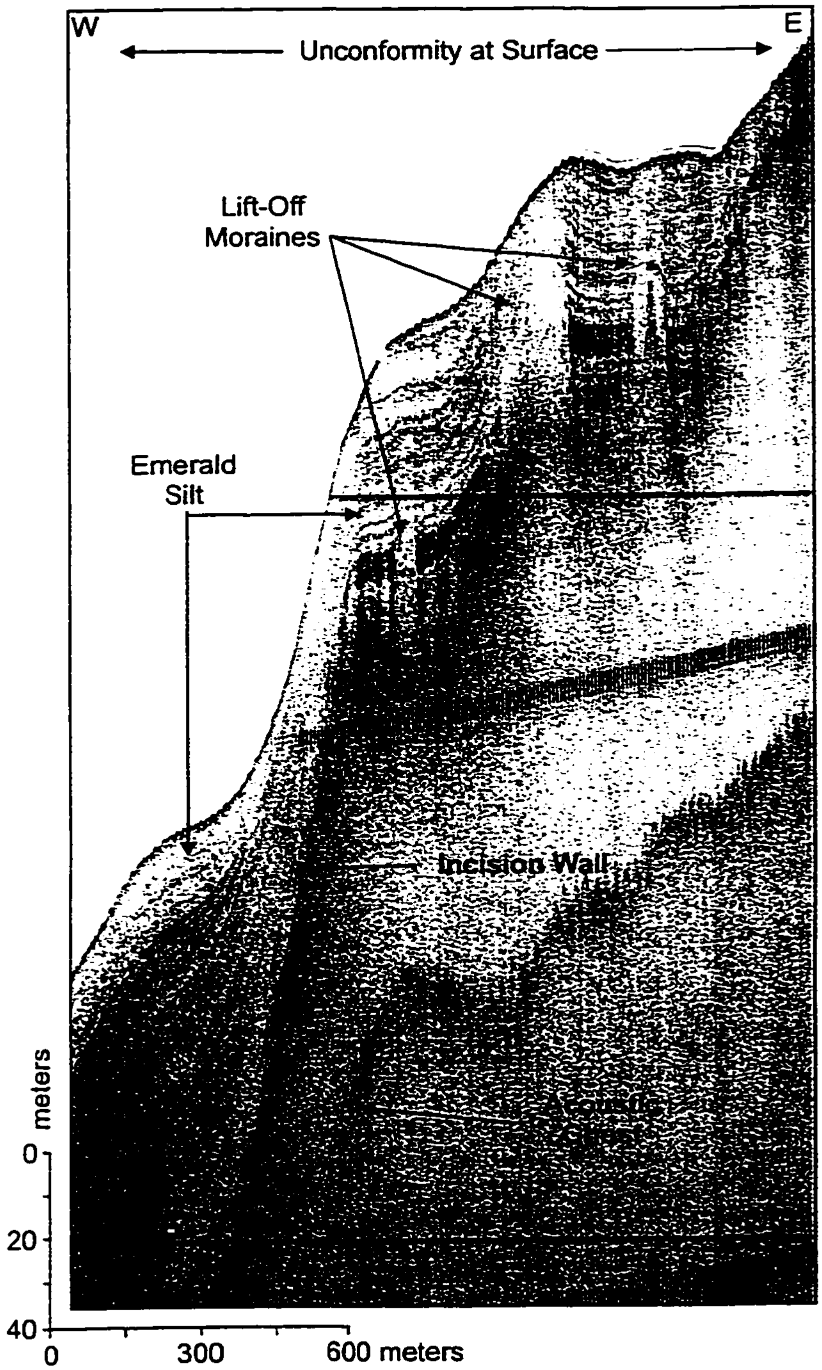



Figure 4.11 Huntec DTS profile of a Zone C incision flank. This high resolution profile shows the near surface deposits typical of Zone C incision fill; Glacial drift and lift-off moraines draped with glaciomarine Emerald Silt deposits which are generally much thicker in the lowest part of the incisions. Ponded LaHave Clay (marine mud) deposits are often found as a surface unit, however it is not seen in this profile.

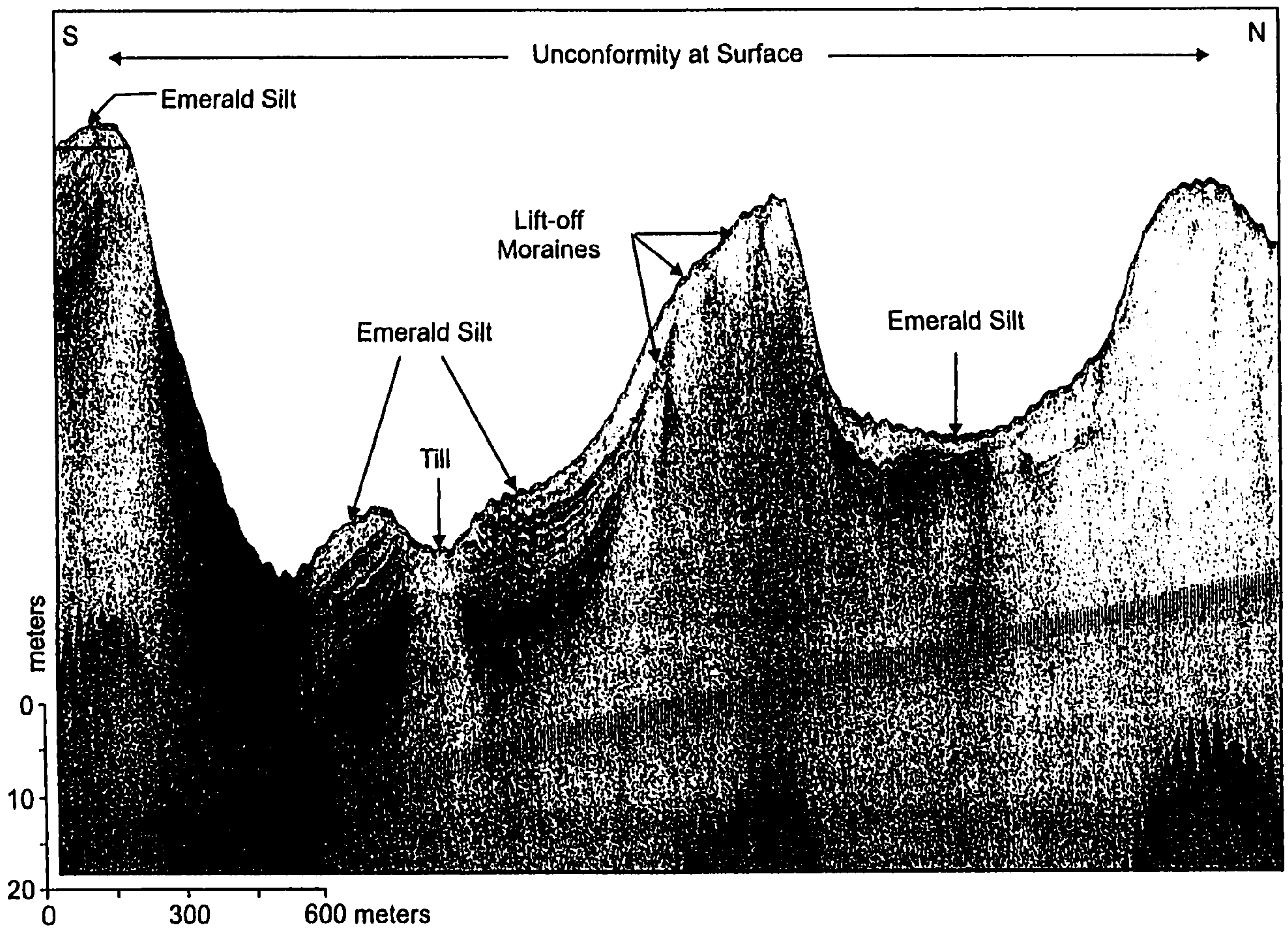



Figure 4.12 Huntec DTS profile at edge of shelf, along west flank of the Laurentian Channel. This profile shows an erosive unconformity at the surface accentuating the surface expression of the incision morphology. This unconformity may be a result of strong currents in this area, or slope failure along the flank of the Laurentian Channel.

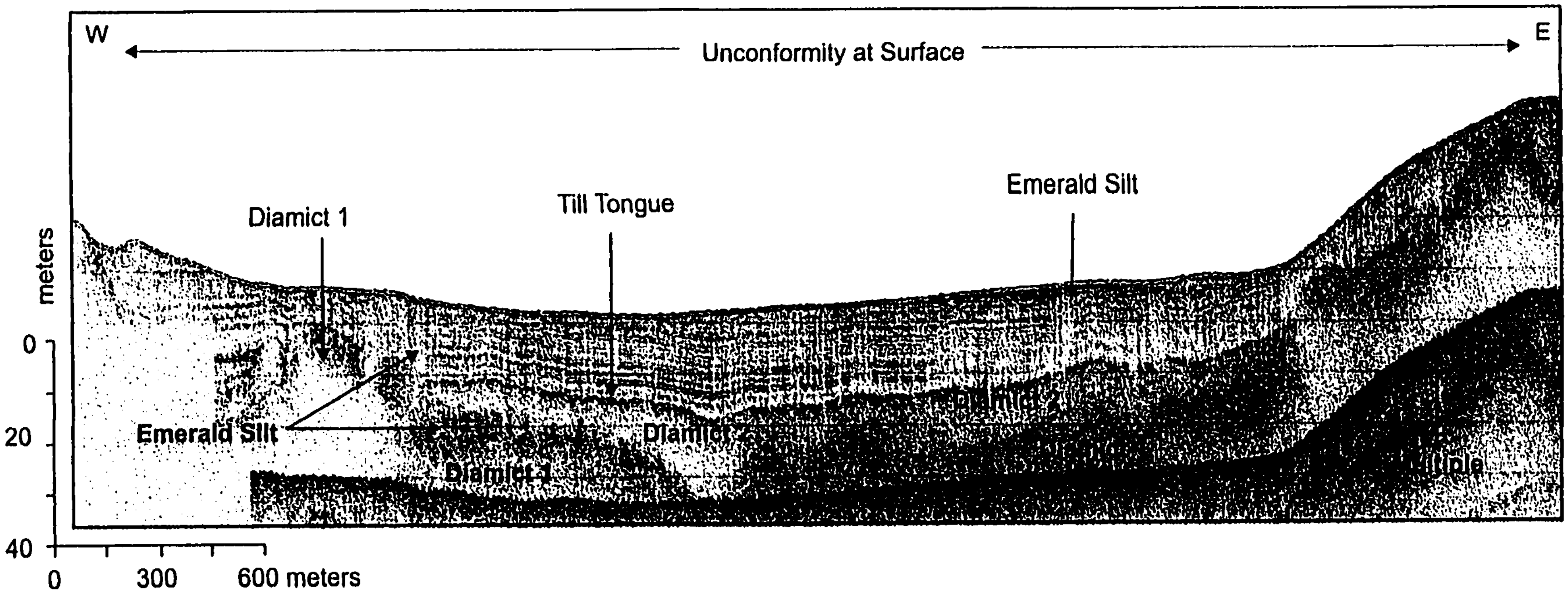



Figure 4.13 Huntec DTS profile of Subzone C² incision on flank of Canso Bank. This depression off Canso Bank contains a till tongue stratigraphy which is representative of a fluctuating, partially floating ice margin.

acoustically incoherent material in mound forms (till, lift-off moraines?) draped by glaciomarine sediments (Emerald Silt) with an extensive erosional unconformity at the surface. Figure 4.13, an incision profile off Canso Bank, displays a depositional sequence of a basal acoustically incoherent facies (till/glacial diamict; diamict unit 1) A second acoustically incoherent facies (diamict unit 2; till tongue?) is bounded by stratified glaciomarine sediments (Emerald Silt). This sequence is representative of till tongue stratigraphy (see section 2.3.3a) indicating a possible readvance of ice from a local ice cap over Canso Bank. Both of these high resolution profiles, in addition to figure 4.11, display an erosional unconformity at the sea bed. This unconformity is characteristic of many of the Zone C incisions.

Figure 4.14, in Subzone $C^2$, is the only seismic profile which was intended to follow the length of an incision in Zone C, although It is unknown if the line is entirely confined to the incision itself, or if it covers some of adjacent surfaces as well. Some morphology may also represent an irregular thalweg down the length of the incision, but there are not enough data to accurately determine if this is true. This seismic line is down the length of one of the large north-south trending incisions which originates in the area where Subzone $C^2$ incisions cross-cut Zone B channels (BC; Fig. 4.1). The section displays an undulating, irregular long-profile with bedrock highs, mounds of acoustically incoherent sediments (till and glacial diamict; moraines), ponded, to conformably draping, overlying stratified glaciomarine deposits displaying variable acoustic signatures (Emerald Silt), and patches of acoustically transparent sediments (LaHave Clay). The Subzone $C^2$ incision depicted in figure 4.14 displays similar features which characterize fjords (i.e.: Syvitski, 1987). They include basins with acoustically incoherent basal units, overlain by glaciomarine deposits, and bedrock controlled sills with hummocky acoustically

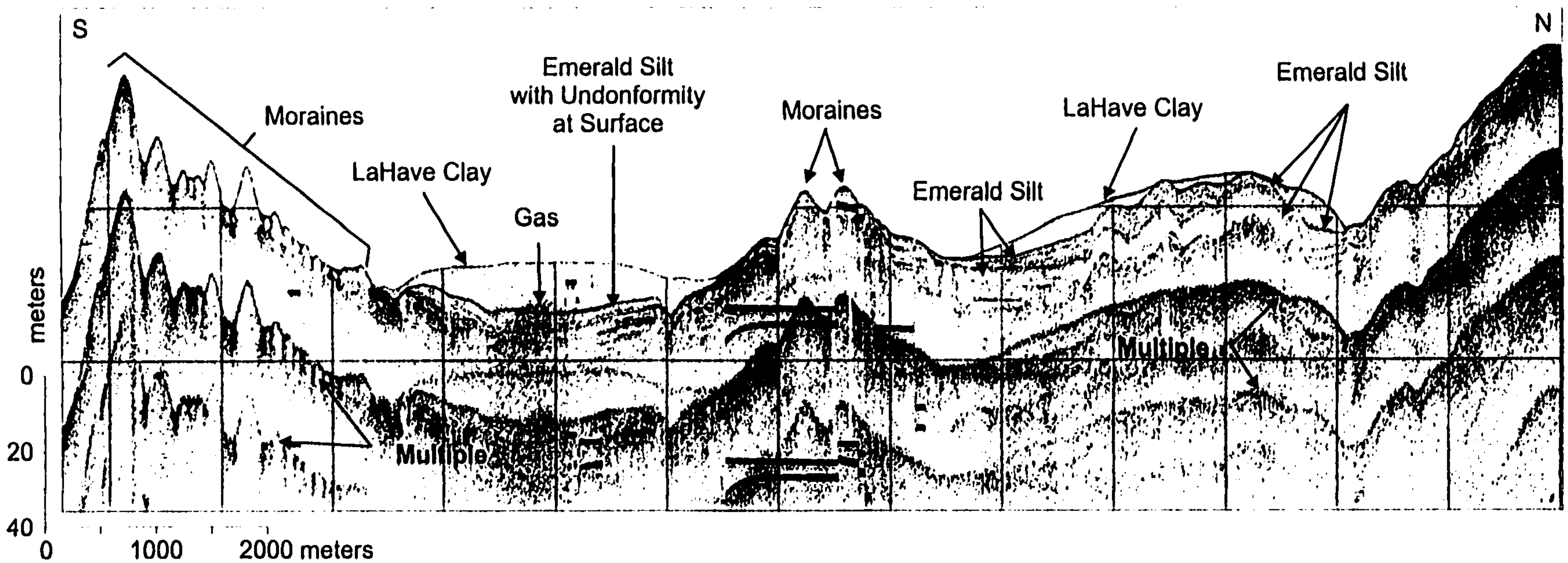



Figure 4.14 Sparker profile down length of a incision in Subzone $C^2$. This profile is the only profile along the length of an incision and displays a morphology reminiscent of a fjord (i.e. sills and basins along the length). Note the transitionary phases or Emerald Silt from ponded to draping. The overall morphology is bedrock controlled.

incoherent deposits on the surface which are likely moraines and may represent gounding lines.

Figure 4.15 is located in Zone B, in an area which is transitional to Subzone $C^1$ incisions. Figure 4.16 is located towards the north of Subzone $C^1$ and figure 4.17 is also located Subzone $C^1$ (St. Anns Basin). Figure 4.15 is from a different zone but is presented together with figure 4.16 and 4.17 as they may be genetically related as part of a Tertiary fluvial system. Figure 4.15 is an airgun profile which reveals a Pleistocene incision eroded into horizontally stratified sediments displaying some onlap sediments filling a larger incision. The stratified incision fill displays a higher amplitude acoustic reflection character which is indicative of Tertiary sediments and is similar to the sequence found in Figure 4.6, also in Zone B. This profile is interpreted to represent a Tertiary fluvial valley with a superimposed Pleistocene incision. An acoustically incoherent unit (till) bounds the south-southeast flank of the Pleistocene incision and overlies the Tertiary sediments. The Pleistocene incision is filled with horizontally stratified deposits of a lower amplitude reflection than the Tertiary fill and is interpreted as glaciomarine sediments (Emerald Silt). Figure 4.16 provides strong evidence for a Pleistocene incision eroded into a Tertiary fluvial valley. The stratified sediments interpreted as Tertiary in age have a high amplitude reflection character which strongly resembles the bounding bedrock walls. The seismic signature for these sediments is clearly distinct from the conformably draping glaciomarine sediments (Emerald Silt) at the sea bed. The bedrock incision flanks are covered with thick acoustically incoherent (till) deposits which appear to occupy the base of the Pleistocene incision.

Figure 4.17 is a high resolution Huntec DTS profile which displays a relatively small V-shaped incision filled with stratified sediments at its base, and

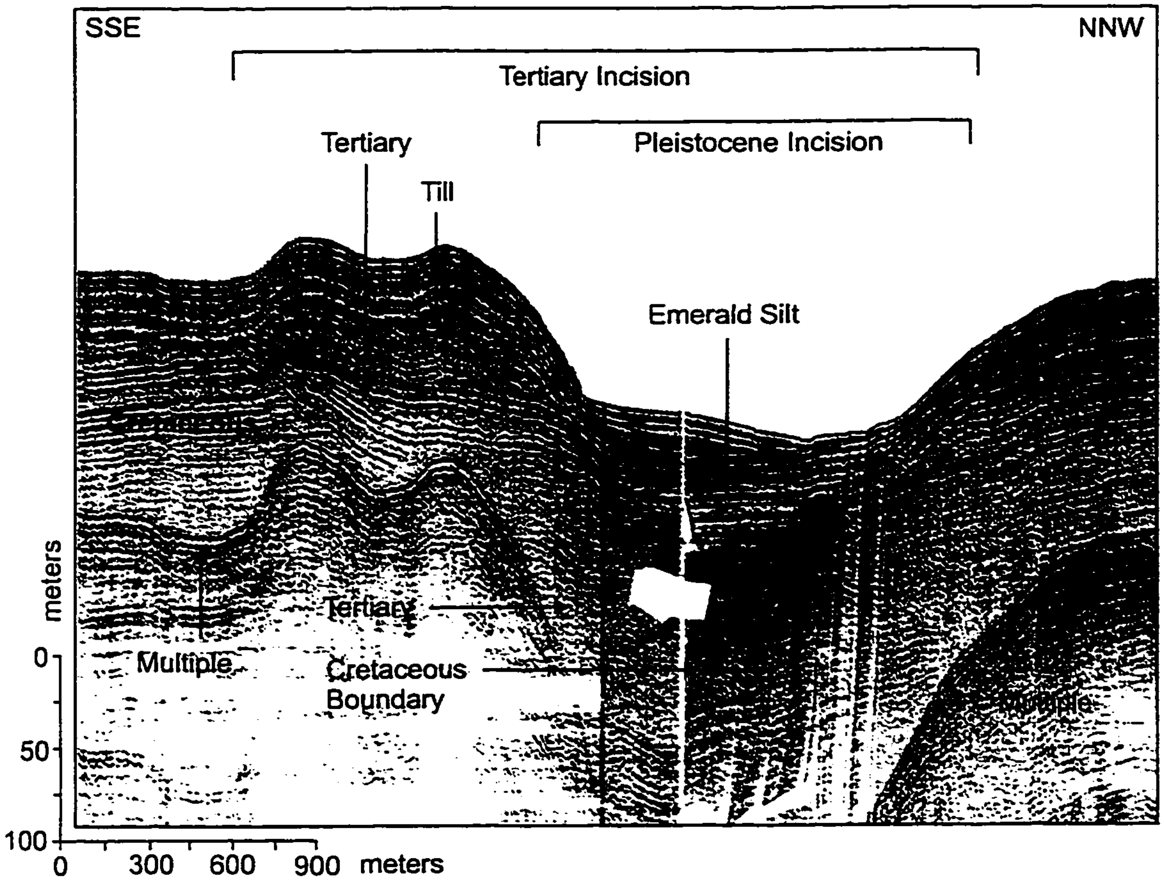



Figure 4.15 Airgun profile representative of an incision which extends between Subzone C' and Zone B. This profile is located in Zone B and shows a Pleistocene incision cut into Tertiary bedrock which is infilling an incision cut into Cretaceous bedrock.

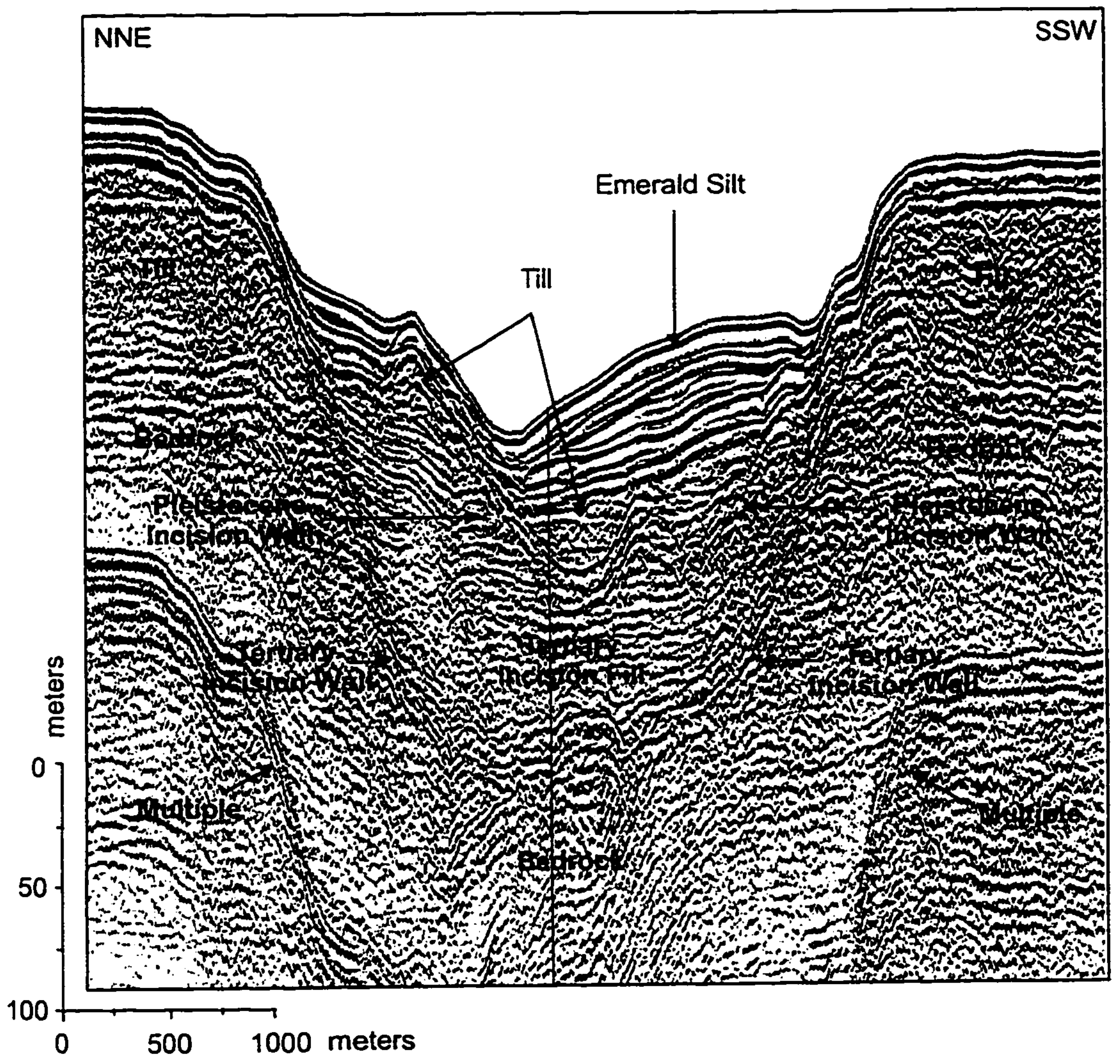



Figure 4.16 Sleeve gun profile of Subzone C' icision. This profile depicts a Tertiary incision (valley) with a superimposed Pleistocene incision. This is a good indication of the reworking of Tertiary incisions by Pleistocene processes, and therefore Pleistocene modification of a Tertiary landscape.

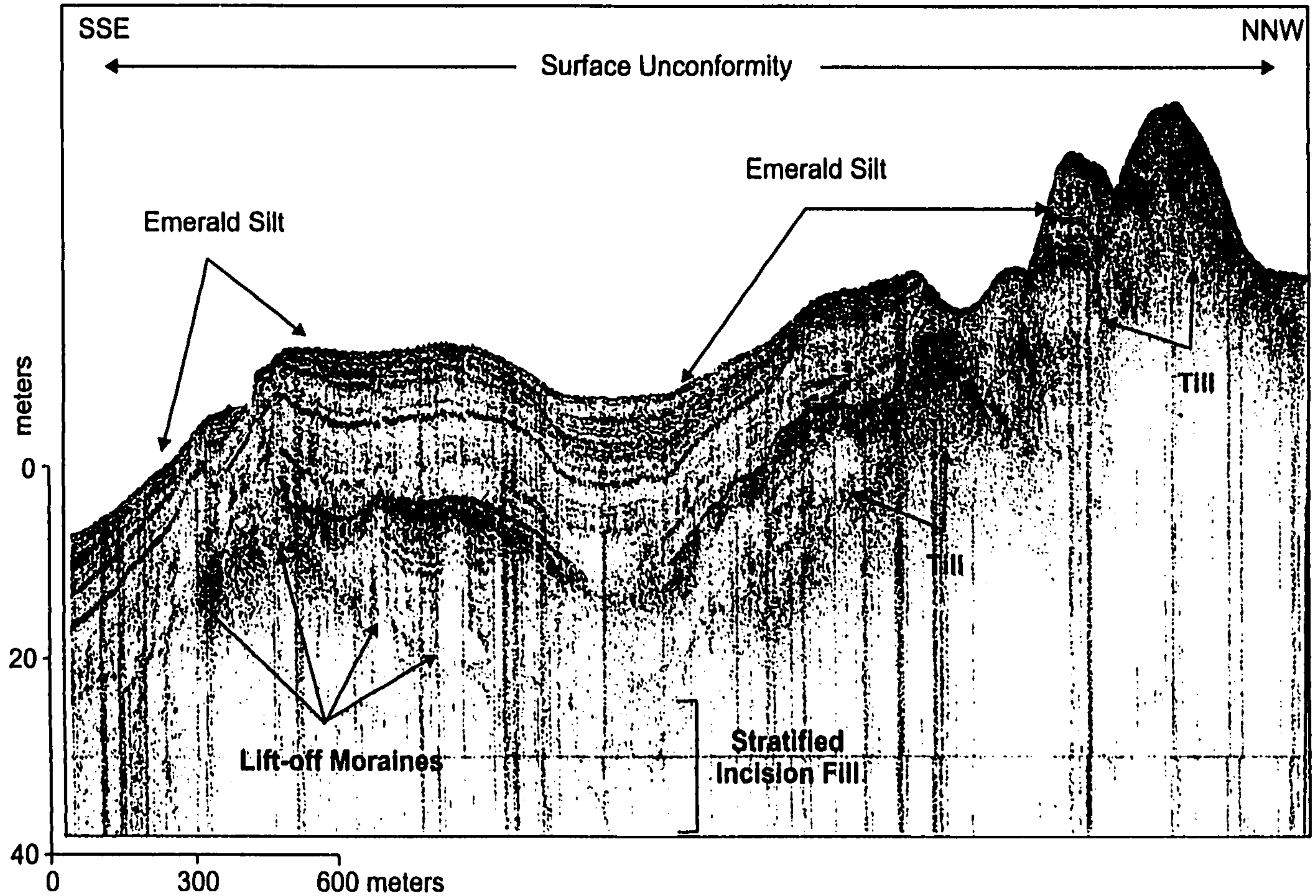



Figure 4.17 Huntec DTS profile in St. Anns Basin. This profile shows a small V-shaped incision which broadens out towards the surface. The V-shaped portion of the incision is filled with horizontally stratified sediments, while the broad area is covered with a typical basin sedimentary sequence (till and Emerald Silt). Does such a profile result from modification of a pre-glacial fluvial valley, or is the feature completely glacial in origin?

capped with thick glaciomarine sequences (Emerald Silt). This profile is from St. Anns Basin which is also filled with thick deposits of acoustically incoherent sediments (till) associated with the Scatarie Moraine of the Scotian Shelf End Moraine Complex, and thick glaciomarine deposits (Emerald Silt). In many areas multiple stacked sequences of stratified glaciomarine sediments (Emerald Silt) and incoherent acoustic facies (till/glacial diamict) were observed on seismic records from within the basin. The V-shaped incision base of figure 4.17 may be representative of a Tertiary valley and may be genetically related to the Tertiary valleys of figure 4.15 and 4.16. The interpretations for figures 4.15, 4.16, and 4.17 support the process of subglacial modification of pre-existing fluvial valleys. Evidence of till overlying the incision flanks and within the base of the Pleistocene incisions suggests ice was in contact with these incisions.

### 4.3.4: Zone D

Zone D (Brandal Basin), north of Sable Island Bank, and south of Middle Bank and the northwest end of Banquereau, exhibits an extensive system exposed incisions and a system of buried incisions in its northernmost area (directly south of Middle Bank; Fig. 4.1). Zone D incisions are similar in width and depth to those of Zone C. The exposed incisions are oriented north to south, and are linear and isolated. Isolated rectangular residuals (positive features) separate the incisions. The southern boundary of the zone is against the northern edge of Sable Island Bank and is characterized by triangular shaped residuals (Fig. 4.1). Data from the eastern portion of Zone D, south of the area between Middle Bank and Banquereau, are of poor quality and difficult to interpret. Bathymetrically, the incisions appear to resemble those in the western portion of the zone. South of the western extension of Banquereau, incisions change orientation from predominantly north-south to east-west trending, and appear as tributaries to The Gully (a submarine canyon). Many incisions in this Zone are likely Pleistocene in age and may be genetically related to Zone C incisions if they continue under the banks and join continuously with Zone D incisions, however not enough data exist to confirm this. (As incisions near The Gully area may have been strongly influenced by processes different from incisions elsewhere on the eastern Scotian Shelf—i.e. submarine canyon formation processes—these incisions are not included in Zone D.)

An acoustically incoherent unit which blankets large areas, including the near surface of the buried incisions and residuals occupies much of Zone D (Fig. 4.18). Incisions in Zone D are filled with similar depositional sequences to those in Zone C; an incoherent acoustic unit at their base (till?, glacial diamict?), subsequently covered by conformably draping glaciomarine deposits (Emerald Silt). A group of incisions with irregular and undulating bases appear to be filled with horizontally stratified sediments towards their bases (Fig. 4.18). The stratified deposits are subsequently overlain by a thick acoustically incoherent unit and in some areas multiple units (till). The acoustically incoherent unit (till) conformably to unconformably overlies the area, dipping into incisions and covering adjacent high surfaces. Stratified glaciomarine deposits (Emerald Silt) display a conformably draping to ponded character over the acoustically incoherent unit (till) where it dips into the incisions. It remains unclear whether the basal stratified sediments are Tertiary or Pleistocene in age (subglacial or interglacial). The multiple till units may indicate multiple ice advances which may have occupied the incisions. Where the blanketing

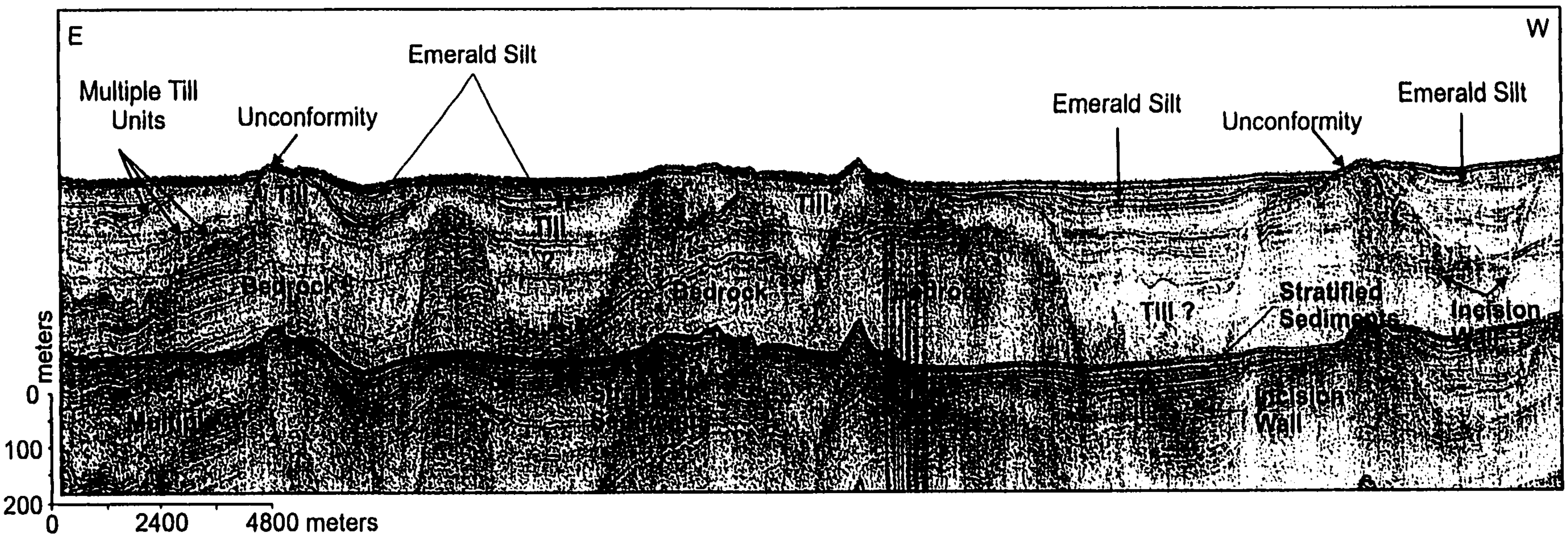



Figure 4.18 Airgun profile across Zone D (Brandal Basin) incisions. Note the extensive system of buried incisions in this area, in addition to the incoherent seismic reflector unit (till) which blankets the area. The buried incisions in this profile stand as an example of what buried incisions in the Bank areas may look like.

acoustically incoherent unit (till) occupies channels, it becomes evident that glacial ice also occupied the channels, at least to the base of these units.

Figure 4.19 displays an incision and a residual south of Middle Bank. The basal portion of the incision is filled with a unit generally displaying an acoustically incoherent signature, though there is some evidence of very weak stratification (diamict 1). This unit corresponds with a similar unit (diamict 1) capping the residual. This unit (diamict 1) is likely glacial in origin. A planar unconformity (uc1) separates diamict 1 from the underlying bedrock surface. In some areas diamict 1 is overlain by a thin acoustically incoherent deposit (diamict 2) which is separated by a second erosional unconformity (uc2). The incision and interfluve are subsequently overlain by conformably draping glaciomarine sediments (Emerald Silt). The thick acoustically incoherent unit (diamict 1) in Figure 4.19 may be genetically related to the acoustically incoherent capping unit described for Figure 4.18. This sequence indicates that subglacial erosion likely produced uc1 and subsequently, diamict 1 was deposited. Diamict 2 represents an anomalous process (possibly a minor readvance) as glaciomarine sedimentation was taking place at its time of deposition. The prograding unit of figure 4.19 shows different sedimentary depositional processes which occurred on the flank of Middle Bank. This likely represents the reworking and redeposition of Emerald Silt deposits during Pleistocene low sea level stands.

### 4.3.5: Areas Not Included in the Zone Classification

The eastern Scotian Shelf banks were not included in the zone classification. However, where seismic sources were able to penetrate the hard sands and gravels of the banks, information on incision fill was possible. Some

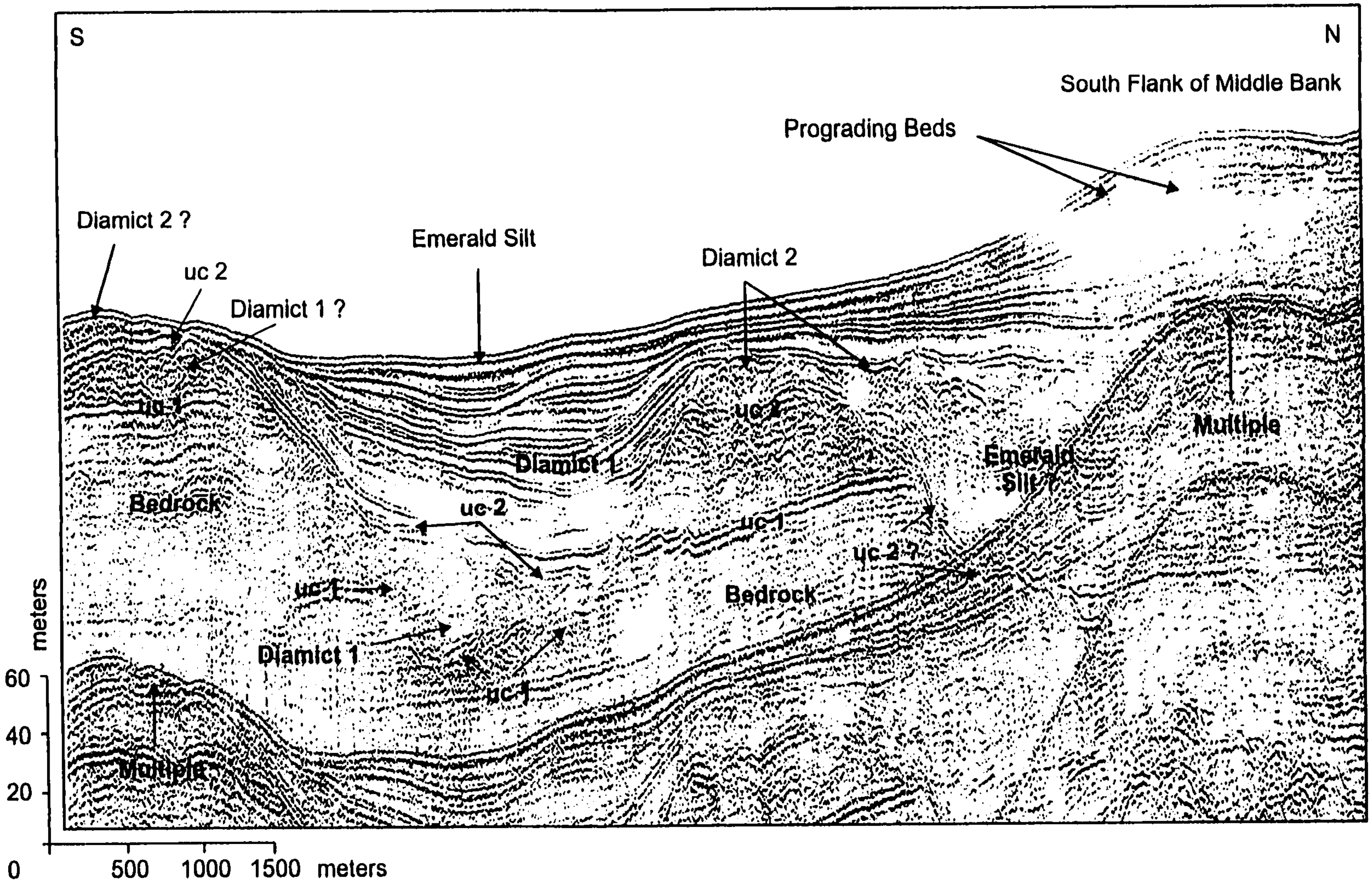



Figure 4.19 Sleeve gun profile of buried incisions and interfluve in Zone D. Two unconformities (uc 1, and uc 2) affect the southern incision and interfluve depicted in this profile. uc 1 and Diamict 1 may be related to Tertiary erosion and deposition, while uc 2 and Diamict 2 (till ?) is related to Pleistocene erosion and deposition. An alternate interpretation could be that uc 1 and Diamict 1 are related to an earlier glacial erosional and depositional phase, prior to uc 2. Based on the intensity of reflections, and the apparent weak stratification in Diamict 1 found within the buried incision and on the buried interfluve, Tertiary erosional and depositional processes are the favoured interpretation.

incisions were filled almost completely with acoustically incoherent sediments, whereas others showed some evidence of stratified deposits (Emerald Silt) which were truncated towards the surface by a planar unconformity above which lie sand and gravel deposits characteristic of the banks (Sable Island Sand and Gravel). Figure 4.20 shows evidence of both acoustically incoherent sediments, and stratified conformably draping sediments (Emerald Silt) truncated at the surface by an erosional unconformity. The maximum depth of most of these deeply incised buried incisions remains unknown, and in some cases the acoustical incoherence of the incision fill may be the result of poor quality data, and not real. The stratified sediments above (r), a reflector marking an unconformity, are likely related late glacial low-sea level stand, and represent the reworking of the uppermost incision fill deposits and deposition of Sable Island Sand and Gravel.

The area between Middle Bank and Banquereau was not assigned to one of the incision classification zones. Near the edge of Middle Bank, multiple stacked sediment sequences display a complex character (Fig 4.21). These sediments appear to be limited to a small area, as the closest seismic line, 8 km to the north, displays an acoustically incoherent unit (till/glacial diamict) covered by conformably draping glaciomarine sediments (Emerald Silt). Thick acoustically incoherent deposits (till/glacial diamict) are also found south of this area, directly between Middle Bank and the western end of Banquereau, and may control some of the positive relief in the bathymetry of this area. Glacial sedimentary sequences on Canso Bank are represented by complex successions of acoustically incoherent units (till), and glaciomarine sediments (Emerald Silt). Till tongues were also identified extending off Canso Bank (Fig. 4.13). These glacial deposits may represent a late glacial local ice cap over Canso Bank which fluctuated between grounded and floating conditions. Some

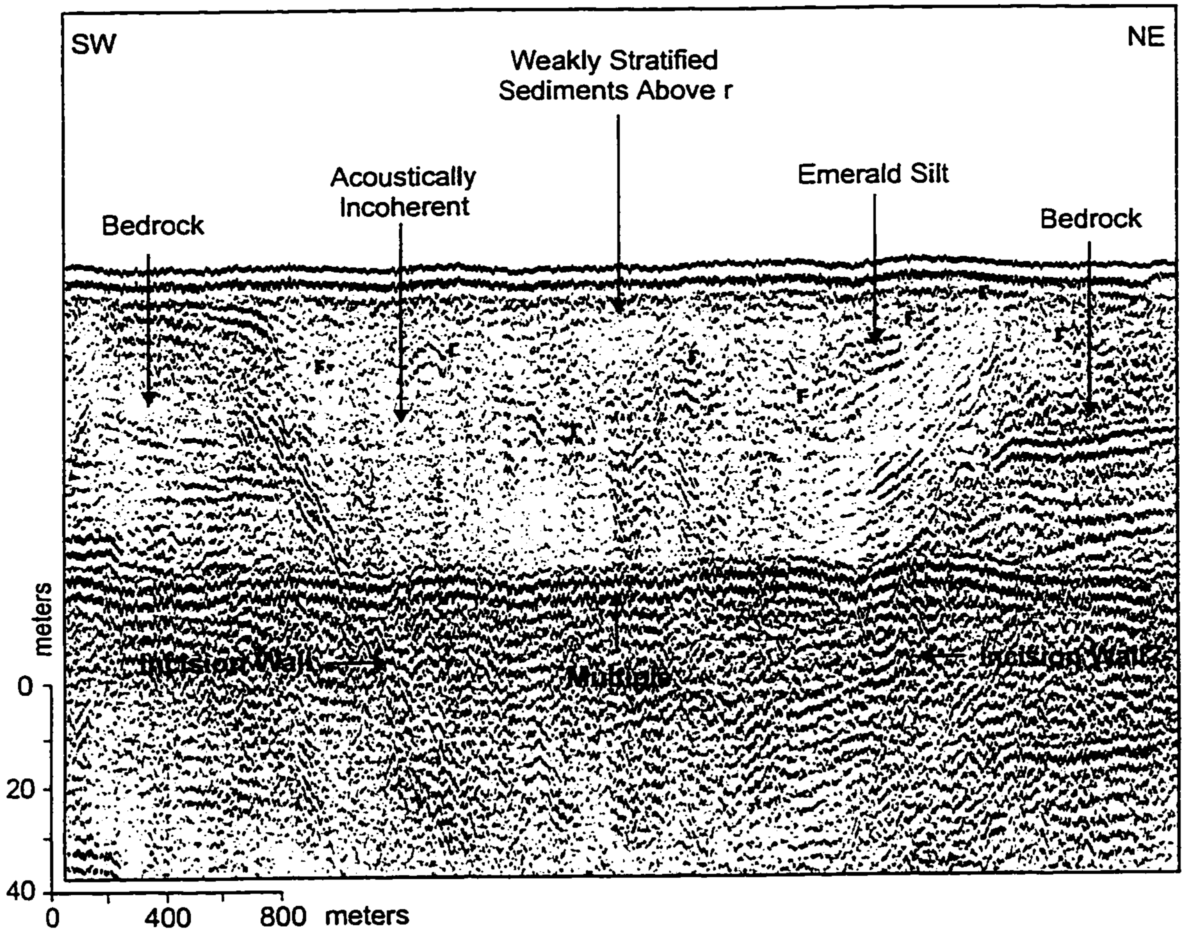



Figure 4.20 Sleeve gun profile of an incision under Banquereau. This incision appears to be filled with acoustically incoherent deposits, or deposits which cannot be resolved using this acoustic system, and Emerald Silt. The Emerald Silt unit has been truncated at the surface which is represented by an erosional unconformity. The Emerald Silt deposits where more extensive at one time, and likely conformably draped the bedrock walls bounding the incision. The reflector marked by 'r' represents an unconformity which may be related to erosion during low sea level stand when deposition of Sable Island Sand and Gravel took place. The sediments above 'r' appear to be weakly stratified.

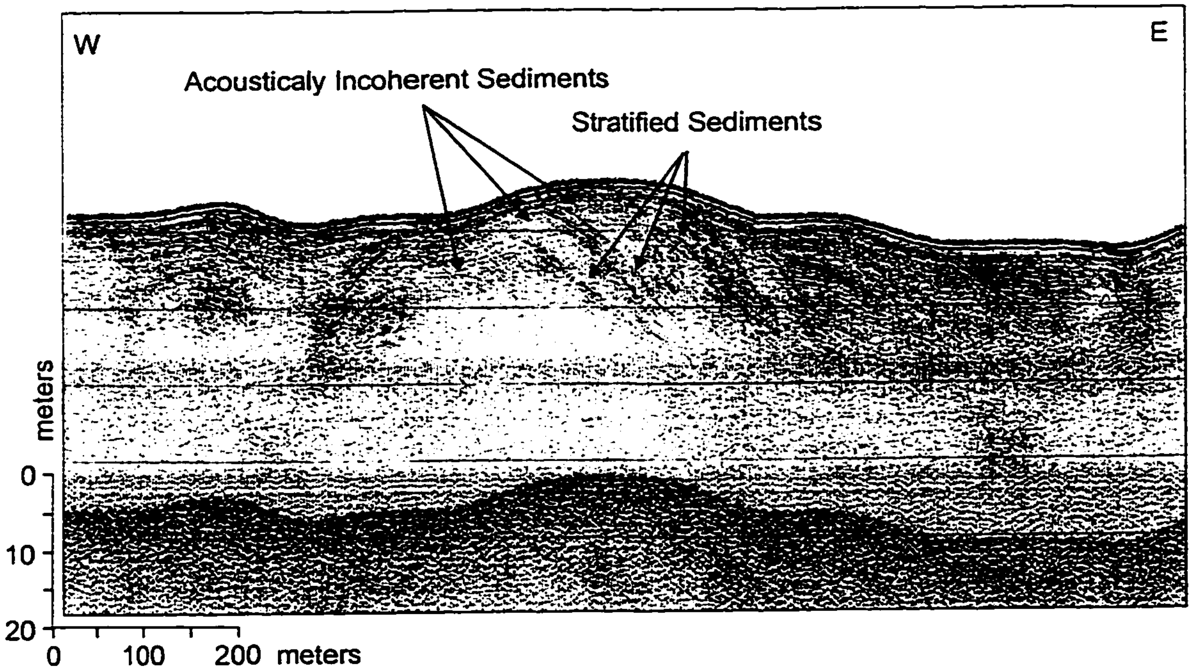



Figure 4.21 Huntec DTS profile of surface sediments in a Zone C[2] incision extending off Middle Bank. These sediments displaying a complex character of prograding (?) units displaying an incoherent seismic signature, and a stratified seismic signature, may be representative of ice contact deposits (proglacial delta?). Only two small locations around Middle Bank display this type of sediment character, on the eastern Scotian Shelf, and the origin remains questionable.

evidence of similar sequences were also observed in St. Anns Basin, and over Middle Bank, though resolution was not as high over Middle Bank as the surface sand and gravel deposits are thicker, lowering the quality of the seismic profiles. The complex stacked sequences in Figure 4.21 may represent ice marginal sedimentation deposited contemporaneous with Emerald Silt in other locations.

## 4.4: INCISION CROSS-PROFILE SHAPES

The eastern Scotian Shelf incisions are highly variable in cross-section shape and include V-shaped, U-shaped, parabola-shaped, and asymmetrical profiles, in addition to incisions displaying steep sides and flat bottoms, and irregular undulating bases.

Incisions of differing cross-profile shapes have been observed directly adjacent to one other, therefore, a zonal classification cannot be made on the basis of incision shape alone. However, while the spatial distribution of incision shapes is generally random, some small groups do display consistency. The deeply incised incisions in Subzone $C^1$ display an overall U-shape (Fig. 4.7), some incisions successive to one another in Subzone $C^2$ possess a parabola-shape (Fig. 4.9), and some incisions successive to one another in Zone D possess undulating irregular bases (Fig. 4.18). Seismic coverage is too sparse to determine if, and/or how much, incision cross-profiles vary along the length of a given incision.

### 4.4.1: Discussion

Is incision shape diagnostic of incision origin? This is a critical question generated by observations of differing incision shapes on the eastern Scotian Shelf. (Channels are defined as bank-full features, while valleys which are not bank-full features may possess channels towards their base.) It is a common belief that glaciated valleys and channels possess a U-shape, whereas river valleys possess a V-shape. However, it is not uncommon for fluvial valleys to possess irregular, undulating basal profiles or parabola-shapes, nor is it uncommon for a fluvial channel to possess a U-shape, parabola-shape, or an asymmetrical profile (Schumm, 1977; Knighton, 1984). Syvitski et al. (1987), discussed the character of fjords around the world, and identified V-shaped cross-profiles along narrow sections of some fjords, in addition to the classic U-shaped valley. J. Shaw (E; pers. comm., 1998) observed the same along Newfoundland fjords. Debris from slope failure infilling fjords/glaciated valleys can give V-shaped or parabola-shaped valley an apparent U-shape (different from a true U-shaped valley eroded into bedrock; Syvitski, 1987). Finally, the cross-profiles of incisions with irregular undulating bases strongly resemble the cross profiles of braided river valleys with multiple channels. Sections 4.4.1a and 4.4.1b consider fluvial and glacial channel and valley cross-profiles shapes in detail. (Note: J. Shaw E. is a scientist at BIO, different from J. Shaw W., of the University of Alberta.)

#### *4.4.1a: Fluvial Channels and Valleys*

Schumm (1977) presented a detailed study on fluvial geomorphology and the evolution of fluvial channels and valleys. If the morphology of the eastern Scotian Shelf is a product of subglacial glaciofluvial erosion then it stands to reason that the channel forms would resemble fluvial erosional channels.

Schumm (1977) showed through laboratory experiments that as water incises into bedrock, a channel would progressively develop into an overall U-shape. The laboratory experiment showed that as incision progressed it produced erosional forms such as potholes, lineations, and erosional ripples. The lineations were eroded into large grooves, and finally a single inner channel (U-shaped) developed which contained the entire flow. Knighton (1985) presented a study on local variations in cross-profile forms of river channels. He showed how asymmetrical forms develop by enhanced bank erosion along meanders, and his conclusions are in agreement with studies by Schumm (1977) concerning asymmetrical alluvial channel development. The origin of the U-shaped and asymmetrical shaped channels observed on the eastern Scotian Shelf may be explained this way.

It is also important to discuss the shape of fluvial valleys and paleovalleys with regards to the eastern Scotian Shelf, as evidence exists for extensive Tertiary fluvial activity (King et al., 1974). Many fluvial valleys tend to possess an overall V-shape as a result of valley side-wall slope failure. However, Schumm (1977) also discussed paleovalleys with U-shaped, and asymmetrical cross-profile profiles. Such valleys are filled with sediments, into which smaller modern fluvial channels are superimposed. Schumm (1977) also suggested that paleovalleys are not likely the result of erosion by water completely infilling the valley, rather the valley may be the result of differential erosion as a river channel shifts laterally during downward erosion. He further commented on the difficulty in distinguishing between channels and valleys based on shape alone. Channel and valley shapes often tend to be similar, and size may be one criterion for distinguishing between the two. Based on this discussion, some 'channels' on the eastern Scotian Shelf may in fact be largely inherited from a Tertiary fluvial valley morphology.

#### *4.4.1b: Glacial Channels and Valleys*

The fact that the eastern Scotian Shelf was subjected to glaciation requires that consideration be given to the cross-profiles of glaciated channels and valleys. Many of the studies considering glaciated valley cross-profiles were done in mountainous environments (Graf, 1970; Hirano and Aniya, 1988; Harbor, 1990; Harbor, 1992; Augustinus, 1995) different form the coastal plain environment of the Scotian Shelf. However, ice streams can be equated to mountain glaciers, with the exception that the fast moving ice is bounded by slower moving ice rather than rock (Hughes, 1992).

Graf (1970) applied the concepts of fluvial geomorphology to glacial valley and network evolution, and acknowledged that most glaciated valleys are modified pre-glacial fluvial valleys. He proposed that the width and depth of the glaciated channel are related to the discharge moving through the channel cross-section which also reflects the ordering of the channel, much like fluvial channels are ordered (greater discharge equals higher order–a primary channel; lower discharge equals lower order–a tributary channel). There is however, a key difference in that discharge increases to the river mouth of fluvial systems, whereas glacier discharge increases only to the ELA, from where is decreases to zero at the terminus. Graf (1970) proposed that glacial valley cross-profiles are generally parabolas, and greater discharge would result in more intense glacial erosion resulting in deeper and narrower valley cross-sections. Asymmetrical profiles were observed and explained by enhanced bedrock erosion along the outside of the curve as an ice stream corners (Graf, 1970), much like enhanced erosion along the outside of a fluvial channel meander.

Johnson (1970), Hirano and Aniya (1988), and Harbor (1990; 1992), and Augustinus (1995) discuss the evolution of a V-shaped fluvial valley to a U-shaped glacial valley. They all agree that the glaciated valley is generally parabola shaped (U-shaped) and results from plucking and abrasion of the bedrock by glacial ice.

Johnson (1970) proposed that as ice invaded a V-shaped fluvial valley, areas of dead ice (areas of negligible erosion) existed at the ice surface and along the valley sides and bottom. The concept of inactive ice along the valley bottom contrasts with other authors (Hirano and Aniya, 1988; Harbor, 1990; Harbor, 1992; Augustinus, 1995) who suggest the most active and erosive ice is along the valley bottom. Johnson (1970) further proposed that as differential erosion progresses along the valley, the valley evolves to a parabola shape and areas of dead ice disappear. Erosion thus affects the valley sides resulting in the U-shaped valley. In this explanation, where basal ice is less erosive, a remnant of the V-shape fluvial valley is left at the base of the glaciated valley.

Graf (1970) and Hirano and Aniya (1988) suggested that V-shaped glaciated valleys are related to lower valley order and hence, the extent of glacial erosion, Whereas, Harbor (1992) suggested that glaciated valleys retaining some form of a V-shape have not been subjected to as much glacial erosion as U-shaped valleys (over time). Augustinus (1995) stresses the importance of rock mass strength as a major control of slope stability, and that weakened rock would result in selective glacial erosion and strongly influence the final form of the glaciated valley. None of these authors consider that the original shape of a paleovalley may have been U-shaped or parabola-shaped, and that in some cases, glacial ice may have only acted to remove paleovalley fill with little modification to the original paleovalley shape.

Hirano and Aniya (1988) propose two types of glacial valley development: (1) the Rocky Mountain model; and (2) the Patagonia-Antarctica model. The Rocky Mountain model results in development of a wide V-shaped valley into a deep U-shaped valley, and the Patagonia-Antarctica model results in development of a steep and narrow V-shaped valley to a broad U-shaped valley. The Rocky Mountain model profile is a result of excavation by alpine valley glaciers, whereas the Patagonia-Antarctica model is a result of excavation under continental ice. Hirano and Aniya (1988) suggest the Patagonia-Antarctica model requires convergent flow of ice into the glacially excavated valleys from higher surfaces adjacent to the valleys. The Patagonia-Antarctica model may best be applied to the Zone C and D incisions of the eastern Scotian Shelf if they are a product of glacial erosion.

#### *4.4.1c: Conclusion*

From the above discussion, pertaining to the genesis of channel and valley shapes, it can clearly be seen that interpretations of channel and valley origins based strictly upon the shape of the channels and valleys can be extremely unreliable. All of the incision cross-profiles observed on the eastern Scotian Shelf can be attributed to varying degrees of fluvial and/or glacial erosion and processes. It remains unclear as to whether or not a particular channel or valley shape can be strictly associated with a particular channel or valley forming process, and here is where the fill and plan-form pattern (as revealed in the bathymetry) of the eastern Scotian Shelf incisions provides important clues (e.g. Tertiary incision fill indicates an incision may be a product of Tertiary fluvial valley evolution, tributary/distributary incision networks may indicate a fluvial origin rather than an origin by ice erosion).

## 4.5: SUMMARY AND CONCLUSIONS

In summary, Sable Island Bank, Banquereau, Middle Bank, Canso Bank and Artimon Bank are all underlain by networks of buried incisions. The relationship of the buried incisions to the exposed incisions adjacent to the banks is uncertain, as data are sparse and seismic penetration and resolution through the hard surface sands and gravels of the banks is poor. It remains unknown whether Zone A, Zone C, and Zone D incisions extend beneath the adjacent banks, though limited seismic coverage indicates that some may do so.

Incision fill towards the surface of the majority of exposed eastern Scotian Shelf incisions appears to be composed of glacial diamict overlain by thick glaciomarine deposits (Emerald Silt), and patchy deposits of marine clay (LaHave Clay) (Figs. 4.4, 4.11, 4.12, 4.13, 4.14, 4.17, 4.18). Where identified, fill towards the base of the incisions and incision shapes are highly variable (Figs. 4.7, 4.8, 4.9, 4.10, 4.18, 4.19). Some acoustically incoherent incision fill facies may be till deposits as they resemble the seismic signature of well sampled subglacial deposits interpreted as till overlying incision flanks (Fig. 4.7, 4.16, 4.17, 4.18, 4.19).

Residuals (surfaces between incisions) between the exposed incision networks appear to be composed predominantly of bedrock (Tertiary and Cretaceous), though in some areas glacial diamict forms some of the residuals, and salt intrusions (diapers) may provide some structural control. Variable amounts of glacial diamict and glaciomarine deposits overlie the bedrock controlled residuals, and in most cases glaciomarine deposits conformably drape the incisions and residuals, with thicker glaciomarine deposits found in the incisions (4.4, 4.15, 4.16, 4.18, 4.19).

Evidence exists for Pleistocene (subglacial) modification of Tertiary fluvial valleys based on evidence of Tertiary sediments infilling incisions (4.6, 4.9, 4.10, 4.15, 4.16). The southern incisions shown in figure 4.7 and 4.17 display small V-shapes towards their bases, though their overall form is U-shaped. This corresponds with the final form of a glaciated valley proposed by Johnson (1970) which contains a remnant of the original V-shaped fluvial valley at the base. Alternatively, it may represent a final stage of under-fit stream development.

A primary observation pertaining to the surficial succession of the eastern Scotian Shelf, including the exposed and buried incisions of Zones A, B, (where observable in airgun data), C, and D, is that the upper sediments are similar. The majority of the eastern Scotian Shelf incisions appear to contain an acoustically incoherent facies interpreted to be till and glacial diamict, with frequent occurrences of lift-off moraines (see section 2.3.2b) on the surface of this unit. This facies is subsequently overlain conformably by variable thicknesses of Emerald Silt and patches of uppermost LaHave Clay. This sequence likely represents Late Wisconsinan glacial deposits as the Late Wisconsinan was the last extensive glacial event to cross the Scotian Shelf, and indicates that ice occupied at least the upper sections of the majority of the exposed incisions.

Extensive erosional unconformities exist at the surface of Emerald Silt deposits over much of the eastern Scotian Shelf (Figs. 4.4, 4.11, 4.12, 4.13, 4.17). It is not unreasonable to postulate that these unconformities are the result of erosion by strong ocean currents toward the end of glaciation as sea level was rising and the ocean was adjusting to modern circulation (J. Shaw E, pers. comm., 1998). This is further supported by an extensive unconformity (R1) mapped on Banquereau and Sable Island Bank (Amos and Knoll, 1987; McLaren, 1988; Amos and Miller, 1990) which has been attributed to erosion by marine transgression following the Late Wisconsinan low sea level stand.

# CHAPTER 5: DISCUSSION

## 5.1: INTRODUCTION

The main focus of this chapter is to discuss the origin of the morphology of the eastern Scotian Shelf (section 5.4). It begins by presenting background information on subglacial hydrology which forms the basis for theories favouring incision of subglacial terrains by subglacial meltwater (section 5.2.2). This is followed by presenting the most popular models proposed for the formation of subglacial incisions (section 5.2.3). Studies concentrating on modern glacier processes are then presented as modern analogues (section 5.3). This background information is incorporated into and referred to in the discussions concerning the origin of the morphology of the eastern Scotian Shelf and provides greater detail than what is presented in the discussions.

## 5.2: THEORIES ON THE FORMATION OF INCISIONS IN SUBGLACIAL ENVIRONMENTS

### 5.2.1: Introduction

Section 5.1 gives a brief description of the major hypotheses proposed for the origin of incised subglacial morphologies. These hypotheses have been grouped into: (1) catastrophic subglacial meltwater release (section 5.1.3a); (2) non-catastrophic meltwater release (section 5.1.3b); (3) excavation by glacial ice (section 5.1.3c); and (4) multiple processes and generations (section 5.1.3d). Before presenting these hypotheses, a basic understanding of subglacial hydrology is necessary as it forms the basis of many of the hypotheses.

This study is concerned with the origin of the exposed incisions of the Scotian Shelf, the bases of which are all below maximum estimates of Late Wisconsinan low sea-level stand (110-120 m bsl; King and Fader, 1986; Fader, 1989). Therefore, it is assumed that incision which took place during the Pleistocene on the eastern Scotian Shelf occurred in a subglacial environment, though the existence of previous fluvial channel and valley systems may have influenced subsequent subglacial incision patterns.

### 5.2.2: Basic Subglacial Hydrology

Subglacial hydrology is pertinent to the discussion of the origin of the eastern Scotian Shelf morphology as meltwater is important to subglacial processes. Some of these processes are: (1) sliding; (2) till dilation and deformation processes; (3) evacuation through various types of subglacial conduit and channel systems; and (4) eroding and removing subglacial debris.

The following review of subglacial hydrology is based on Paterson (1994) which summarizes work by others (refer to Paterson, 1994, Chapter 6 for more details, and references therein). Subglacial meltwater is derived from two sources: (1) basal melting; and (2) surface meltwater and rain flowing through crevasses and moulins to reach the glacier bed. Some water storage can take place at the base of the glacier in subglacial lakes and cavities. When water reaching the bed is in excess of what is drained away, water pressure builds counteracting the weight of the glacier. The piezometric surface, that is the surface to which water rises under hydrostatic pressure, is an important consideration for understanding glacier hydrology and erosion by subglacial meltwater. This is because the piezometric surface controls the direction in which water flows. Depending on the ice thickness distribution, this property can allow water to flow up hill. Hydraulic conditions at the glacier bed are also very important when considering glacier movement such as sliding, surging, and bed deformation.

General considerations for the direction of water flow include: (1) the hydraulic gradient as determined by the piezometric surface; (2) the concept that larger conduits grow at the expense of smaller conduits; (3) the fact that subglacial water will tend to follow valley floors and cross divides at the lowest point as a result of dependence of flow on bed slope; and (4) the concept that the build up of water pressure can force water out of a channel into a sheet.

Steady-state water flow at the glacier bed may occur as: (1) thin sheet flow between the glacier and bed; (2) through channels cut in bedrock (N-channels or Nye channels); (3) channels cut upwards into the glacier ice (R-channels or Röthlisberger channels); (4) in subglacial sediments acting as aquifers; (5) through interconnected canals at the ice-substrata interface; and (6) in linked cavity systems.

In the linked cavity system, water flow takes place along a hard impermeable bed such as bedrock, where it collects downstream of topographic highs in cavities which are linked by a network of N-channels and/or R-channels. It is possible this type of flow may not be restricted to hard beds. Water flow along a soft bed (i.e. till or weakly lithified sediments), may be evacuated via sheet or channel flow along the ice-till interface, and/or the till and hard bed interface. This would occur once water input exceeds the capacity for the soft bed material to evacuate it via porous flow. Walder and Fowler (1994) proposed that sheet flow with a variable thickness is unstable, due to viscous dissipation in the water, can lead to the initiation of R channels.

Likewise, a system of canals, sometimes bifurcating, may develop in the surface of soft subglacial sediments. The canal system would be kept open by fluvial removal of inwardly creeping till, similar to the model for tunnel valley development proposed by Boulton and Hindmarsh (1987). Walder and Fowler (1994) hypothesized that a drainage network of wide, shallow, braided canals would develop along the ice-sediment interface of a deforming bed under which subglacial water pressure would be close to ice-overburden pressure, and the hydraulic gradient is low (Walder and Fowler, 1994; Clark and Walder, 1994). Under rigid bed conditions, water would drain through an aborescent network of a few large tunnels (R-channels), and at higher velocities than for the soft bed condition (Walder and Fowler, 1994; Clark and Walder, 1994).

Water may also escape catastrophically from an ice marginal environment or from beneath a glacier. Jökulhlaups, catastrophic releases of ice marginally stored or subglacially stored meltwater, are well documented events occurring in presently glaciated terrains (i.e. Nye, 1976). Shaw et al. (1996; and references therein) describe landform evidence in Saskatchewan, Ontario, and Alberta, for catastrophic subglacial meltwater release (referred to as the outburst-flood hypothesis from here on) evolving from unstable sheet flow to channelized flow. These events, however are at a significantly larger scale than Jökulhlaups and have not been observed in modern glaciated environments. In addition, the sheet flow stage of this model is much thicker than the sheet flow stage discussed by Walder and Fowler (1994), and Clark and Walder (1994).

## 5.2.3: Proposed Models for Subglacial Incision Formation

The following sections summarize the most popular models proposed to explain the origin of incised subglacial morphologies. They have been divided into four groups: (1) catastrophic meltwater release (section 5.1.3a); (2) non-catastrophic meltwater release (section 5.1.3b); (3) excavation by glacial ice (section 5.1.3c); and (4) multiple processes and generations (section 5.1.3d). Praeg (1996) presented an excellent review of the history of thinking on origin of tunnel-valleys (referred to as incisions in this work), and Ó Cofaigh (1996) provided a critical review of many of these models to which the reader is referred to for more information.

### *5.2.3a: Catastrophic Subglacial Meltwater Release*

Models attributing channel incision to catastrophic subglacial meltwater release assume flow was determined by the piezometric surface which is commonly related to the slope of the glacier surface, and not the bed. This helps explain how some channels can slope up hill. In other words, water would flow from thick ice to thin ice, rather than with the topographic gradient.

The catastrophic release of subglacial meltwater (outburst-flood hypothesis) is a relatively new hypothesis that has been proposed as an origin for channelized morphologies. The outburst-flood hypothesis favours comtemporaneous channelization following sheet floods to explain anabranching channel systems and enclosed basins in Alberta and Ontario (Shaw, 1983; Shaw and Kvill, 1984; Boyd et al., 1988; Shaw, 1989; Shaw et al., 1989; Shaw and Gilbert, 1990; Gilbert and Shaw, 1994; Sjogren and Rains, 1994; Brennand and Shaw, 1994; Shaw et al., 1996 ). Under this hypothesis, large portions of the glacier would have been decoupled from the bed, flooding the entire land surface during unstable sheet flow. As meltwater discharge declines, channelization would take place with possible local overbank flow as channels pinch off during waning flow (Brennand and Shaw, 1994). The resulting channel pattern is described as anabranching. It is attributed to the inability of a single channel to accommodate high discharges, resulting in widespread overbank flooding and the production of new channels by avulsion (Shaw et al., 1989). The observations presented in Chapter 4, describing the zonation of the eastern Scotian Shelf do not support an anabranching channel pattern. The outburst flood hypothesis would interpret incisions as channels and scour depressions rather than valleys, and presumes channels (i.e. zone B) to have been bank full during formation while scours were formed under sheet flow conditions (i.e. Subzone $C^2$ and Zone D).

Gilbert and Shaw (1994) explain the origin of enclosed basins in southern Ontario and Northern New York, which resemble the plan-form pattern of Zone C and Zone D incisions. In their model, they stressed the importance of escarpments which would act to confine water flow, locally increase velocity, and hence erosive power. The erosive vortices thus generated would then stream into channels creating enclosed basins. The origin of these channels was based on assuming surrounding streamlined terrain was formed by outburst-floods. Eyles and Boyce (1998) presented a theory of subglacial erosion for the formation of streamlined subglacial features which does not necessitate vortices derived from meltwater, and Pair (1997) presented a non-catastrophic model to explain the same features (see section 5.1.3d). However, Kor et al. (1991) presented strong evidence for meltwater erosional forms northwest of the study area. A fundamental question is are floods required to form these features?

Booth and Hallet (1993) favoured a location near the ice margin for channels (anabranching and isolated, with side walls ranging from 10 to >100 m high, and up to 500 m wide) attributed to a subglacial flood origin. They observe that channels occur within 20 km of the Puget lobe ice margin, and are generally absent in the interior. Boyd et al. (1988) use the concept of sheet flow evolving to channelization to explain the channels buried beneath Sable Island Bank (reviewed in Chapter 1, section 1.3). These channels would be located proximal (within a few kilometers) to the ice margin due to their location on the eastern Scotian Shelf (Figure 1.3).

Wright (1973) described a system of channels in east-central Minnesota, some of which are isolated, whereas others anabranch. They average widths of 300 m, and depths of 10 m, with a maximum observed depth of 70 m (much smaller than Zone C and D incisions of the eastern Scotian Shelf, but similar to that of Zone B). Wright (1973) proposed that accumulated subglacial meltwater was released catastrophically from a subglacial lake, eroding channels at relatively high velocities and under relatively high hydrostatic pressure. Wright (1973) suggested the possibility that all the channels were not cut simultaneously, and that they could have been cut progressively in a headward direction.

Wingfield (1990) also advocated subglacial meltwater flood events and attributed the origin of 'incisions' in the mid-North Sea to a singular mechanism, the outburst of intra-ice-sheet lakes. Major incisions are oval and slightly pear shaped depressions with widths of 2-5 km and depths of 100-360 m. Some depressions have a preferred orientation, while others radiate, and some are linked by cols. Smaller-scale depressions identified within the major incisions were attributed to the melting of buried ice, though they may be erosional marks if the channel was formed by meltwater. No channels observed on the eastern Scotian Shelf possess a distinct oval or pair shape.

Wingfield (1990) believes the incisions were cut during three flood episodes. Wingfield (1990) hypothesized that major incisions were formed time transgressively along the frozen toe of an ice sheet by jõkulhlaup plunge pools where a breach occurs in the frozen ice sheet toe. From the plunge pools water would spread forming a system of anastomosing channels. As drainage from the water source increased, the breach would widen and the plunge pool at the head of the incision would cut back upstream through the breach. Chaotic channel fill was interpreted to be the result of dropped sediment load and slumped material from the side walls.

Piotrowski (1997a) proposed a subglacial drainage cycle in which insufficient meltwater drainage through the substratum resulted in the ponding of water at the ice-bed interface. The system would change directly from aquifer-drainage to open-channel drainage by catastrophic releases of ponded meltwater through tunnel valleys. He does not give a description of how this occurs, only that no evidence of sheet floods as per Shaw (1989) exist in the area, though under the outburst-flood hypothesis, the channels would likely be interpreted as the evidence. Piotrowski (1997a) concluded that tunnel valleys in Northern Germany were formed by periodic subglacial meltwater erosion under high meltwater flow rates near the periphery of the ice sheet at its maximum extent and shortly thereafter. Piotrowski (1997b) further concluded that the tunnel valleys represent a stable, mature system lacking in remnants of transitional or short-lived features. This conclusion is based on the observation that there are few major tunnel valleys as opposed to complex anastomosing systems of smaller ones, and that the tunnel valleys appear to be inherited from earlier glaciations. This model does not consider drainage through a subglacial drainage system (section 5.1.2), or explain why the transition from meltwater drainage through the substratum to tunnel valley drainage has to be catastrophic and not non-catastrophic.

### *5.2.3b: Non-Catastrophic Meltwater Release*

Boulton and Hindmarsh (1987) hypothesized that subglacial tunnel valleys form near glacier termini through a process they call piping. In this process, saturated and dilated subglacial sediments would flow laterally towards a subglacial tunnel and would be subsequently flushed by water flowing along the axial tunnel/s. Ice would subsequently invade the valleys as piping progressed. This hypothesis suggested that large tunnel valleys can form under non-catastrophic conditions with limited meltwater input and that large valleys form from erosion by small channels. Boulton and Hindmarsh (1987) state that subglacial channels exist to drain excess water to allow stable deformation to occur. This is under the assumption that deformation occurs. The formation of tunnel valleys by the piping method is proposed to be a soft bed equivalent to eskers found in hard bed areas (Boulton and Hindmarsh, 1987). However, eskers have been identified in the prairies of Alberta (Shetsen, 1987; Shetsen, 1990; Munro and Shaw, 1997).

Praeg (1996), who supports the deformable bed theory of Boulton and Hindmarsh (1997), described a system of buried tunnel valleys in the North Sea Basin displaying a converging plan-form, with local divergent or anastomosing patterns. The tunnel valleys are up to 500 m deep, and 0.5-6 km wide. The plan-form of these valleys resembles Zone B incisions on the Scotian Shelf, but does not resemble Zone C or D incisions which do not display a convergent character, and display distinct cross-cutting relationships. Praeg (1996) described and interpreted tunnel valley fill stratigraphy as: (1) a basal unit of clinoforms interpreted as glaciofluvial sands; (2) subhorizontal onlapping deposits interpreted as sand and mud deposition in lake basins; (3) complex surfaces interpreted as deposition during marine transgression. These sequences do not correlate with Zone C or Zone D incision fill which is highly variable (see Chapter 4). Praeg (1996) proposed that the North Sea tunnel valleys were formed time transgressively in ice marginal areas through contemporaneous excavation and backfilling by a distributed system of streams in larger basins. This is assuming the tunnel valley fill is genetically related to the origin of the tunnel valleys which may not be the case.

Mooers (1989) describes a tunnel valley system in eastern Minnesota consisting of two main valleys 500-1000 m wide with average depths of 10 m, and local overdeepening up to 70 m. They are subparallel, occasionally branch and infrequently cross-cut, thus providing for determination of relative ages. Mooers (1989) favoured an ice marginal progressive origin for subglacial tunnel valleys, and attributed them to erosion by seasonally derived subglacial and englacial meltwater. He explained apparent anastomosing and continuity of tunnel valleys as a result of headward development of the drainage system, and the subsequent merging of valleys.

### *5.2.3c: Excavation By Glacial Ice*

Hughes (1987), in discussing ice streams stated:

> ...Pulling forces from ice buoyancy and surface slope combine to produce a maximum pulling force at the head of the ice stream, so that ice is dawn into the ice stream, creating a zone of converging flow... Converging flow maximizes the surface slope, which maximizes basal traction and allows rapid erosion of the bed, so that ice stream channels are characteristically foredeepened. Moreover, maximum erosion rates at the head of the ice stream

> extend the channel into the ice sheet. In this way the pulling power of the ice stream is able to reach far into the ice sheet...

Recent studies of the origin of the Norwegian Channel and its shelf edge deposits (North Sea Fan), North Sea, favour a process of fast-flowing ice stream conditions (Sejrup et al., 1996; King et al., 1997). Borehole and seismic data revealed multiple till units separated by unconformities and conformably overlying glaciomarine/marine sediments along the Norwegian Channel (Sejrup et al., 1996). King et al. (1997) did a very detailed analysis of core material including lithologies, grain size distribution, detailed analysis or rock fragments. They concluded there were three distinct phases of till deposition within the Norwegian Channel. Sejrup et al. (1996), and King et al. (1997) proposed that debris was derived from subglacial erosion through progressive glacial comminution and was subsequently transported via a deformable bed to the shelf edge, where it has deposited as glacigenic debris flows on the North Sea Fan (this model follows the concepts of ice stream flow proposed by Alley et al., 1989). Eittreim et al. (1995) proposed a similar concept for the Wilkes Land continental Shelf, Antarctica. King and Fader (1990; in prep.) also identified multiple distinct tills along the base and side walls of the Laurentian Channel . The Laurentian Channel is poorly sampled compared to the Norwegian Channel, however, similarities in form and sediment fill (as determined by acoustic character for the Laurentian Channel) suggest the Norwegian Channel is a good comparison. Zone C and Zone D incisions may have been subjected to similar erosional processes.

Carlson et al. (1982), and Herzer and Bornhold (1982) stress direct glacial erosion as the major process responsible for the formation of troughs identified in the Northeastern Gulf of Alaska, and the continental shelf off southwestern Vancouver Island respectively. White (1972) discussed the importance of glacial erosion in forming overdeepened ellipsoidal basins. He explains the Finger Lakes of New York, as former fluvial features which were overdeepened by accentuated preferential glacial erosion due to thick vigorous ice in the valleys, and thin weak ice on the uplands.

### *5.2.3d: Multiple Processes and Generations*

Pair (1997) suggested an alternative theory to the outburst-flood hypothesis for the evolution of enclosed basins (Gilbert and Shaw, 1994; section 5.1.3a). He studied isolated incisions 5-10 km long, 0.4-1 km wide, displaying meltwater erosional forms (s-forms) in southern Ontario and northern New York (the same incisions studied by Shaw and Gilbert, 1990) and Gilbert and Shaw, 1994). In his model, Pair (1997) acknowledged the ability of ice to deform and follow topographic variations. Through this recognition, he proposed that ice deforms into pre-existing subglacially-modified bedrock valleys resulting in a close ice-bed contact where meltwater would concentrate and flow. Pair (1997) concluded that the incisions were structurally controlled, and subsequently eroded by glacial action including plucking and dissolution, in addition to erosion by meltwater as evident by the s-forms.

Mullins and Hinchey (1989) and Mullins et al. (1996) proposed the process of ice streaming followed by, or in conjunction with, rapid outpourings of meltwater and sediment under high pressure as an origin for the New York Finger lakes, isolated depressions (comparable in dimensions to Zone C and Zone D incisions of the eastern Scotian Shelf). Unlike the theories for catastrophic subglacial meltwater release (section 5.1.3a), this process is more uniformitarian, and ice is an important component to the process. Mullins and Hinchey (1989), and Mullins et al. (1996) argue that the Finger Lakes are features formed near the ice margin during deglaciation, and recognize that the incision fill is Late Wisconsinan in age suggesting Late Wisconsinan ice eroded at least to the base of the incisions, though they also recognize that the incisions may be partially inherited from previous glacial events. Mullins et al. (1996) further concluded that the incisions may not have been formed synchronously.

Long and Stoker (1986), Cameron et al. (1987), and Piotrowski (1994) support the concept of multigeneration and multiprocess incision/valley formation based on incision fill characteristics. Ehlers et al. (1984) and Balson and Jeffery (1991) observed the fill of systems of incisions in northwest Germany, and the North Sea, respectively, and proposed an origin based on combined glacial erosion and meltwater processes.

Sættem et al. (1992), discussed incisions in the southwestern Barents Sea and proposed subglacial glacial/fluvial erosion and basinward submarine canyon development a possible origin, and Eyles (1987) acknowledged slumping and headward erosion along the shelf edge as a possibility for the formation of an incision in the Gulf of Alaska. The latter two processes may be important for incision evolution around The Gully area, and possibly parts of Subzone $C^1$ of the Scotian Shelf, bordering the Laurentian Channel.

### *5.2.3e: Summary*

Based on previous studies (discussions above), incisions observed in various glaciated terrains vary in scale, dimension, and plan-form. Some are oval and isolated in shape (basins), some are linear and isolated. Others form complex anabranching networks, or display a distributary/tributary character. Dimensions vary from a few tens of meters to kilometers in width, and a few tens of meters to hundreds of meters in depth. The eastern Scotian Shelf displays a variety of incision characteristics, and thus may be products of numerous formative processes, rather than of a single mechanism.

Except for the group which recognizes the possibility of multiple processes, most of the hypotheses for subglacial incision formation presented, are quite rigid in their conclusions. Meltwater hypotheses concerned with channel formation (catastrophic or non-catastrophic) ignore the ability of ice to erode. Theories which support ice erosion alone, don't give enough consideration to the fact that glaciers produce meltwater which must have a significant influence in glacial processes, be it a contribution to the erosion of a channel, or for merely evacuating glacially eroded debris.

Most of the models proposed for the origin of subglacial incised morphologies are based on theory and field observation in previously glaciated terrain, lacking in modern ice cover. They need to be complemented by observation of modern glaciated environments (section 5.2), thus providing greater insight into subglacial processes responsible for incision formation.

## 5.3: MODERN ANALOGUES

Few studies concerned with the origin of subglacial morphologies formed under former ice sheets directly address modern subglacial morphologies. Present day continental ice sheets (i.e. Antarctica and Greenland) are difficult to observe directly making it difficult to determine the processes occurring in the subglacial environment and details of morphology. Most observations of glacial processes are made near the margins of continental ice sheets, and at more accessible valley glaciers. Modern day continental ice sheets often terminate in marine environments on continental shelves. This is similar to the setting of the eastern Scotian Shelf.

Marine margins of continental ice sheets are characterized by tidewater glaciers, ice streams, and outlet glaciers which may feed ice shelves. Ice streams are part of inland ice sheets which flow rapidly through slower moving adjacent ice (Bentley, 1987). They are bordered by ice and not rock, and as they are part of the inland ice sheet, they are grounded and not floating (Bentley, 1987). Outlet glaciers are bordered by rock. Many glaciers show characteristics of both ice streams and outlet glaciers (i.e. Jakobshavns Isbræ, Greenland; Bentley, 1987).

### 5.3.1: Subglacial Morphologies

The West Antarctic ice streams flow over a broad coastal zone which is below sea level, lacks mountains, and is underlain by weak Tertiary sediments (Alley et al., 1989; Alley and Whillans, 1991), much like the Scotian Shelf. This is unlike the majority of present day marine terminating glaciers which originate in more mountainous regions (i.e. Alaska, Greenland, Trans Antarctic Mountains). The West Antarctic ice streams which flow to form the Ross Ice Shelf provide a present day analogue for past subglacial morphologies in coastal regions. It remains unclear if ice on the Scotian Shelf was always temperate. It is quite possible that cold based and warm based ice affected the Scotian Shelf at various times, during various stages of ice cover. Regardless, the West Antarctic ice streams, though located in a polar region, are warm based.

Extensive radar sounding studies revealed that the West Antarctic ice streams (all of which are pure ice streams except for ice stream A which borders the Trans Antarctic mountains) occupy subglacial troughs (Shabtaie and Bentley, 1987; Shabtaie and Bentley 1988). Shabtaie and Bentley (1988) mapped ten deep troughs which converge into and along ice streams resulting in four major troughs occupied by ice streams A, B1, B2, and C. The bases of the major troughs display irregular morphologies (Shabtaie and Bentley, 1988). The dimensions of the West Antarctic ice streams (A, B, C, D, E, and F) vary from 30-80 km wide and 300-500 km long, and their depths average about 500 m (Engelhardt, et al., 1990). The mapped troughs vary in width along their length, and display a down ice tributary system.

Ice streams which terminate as outlet glaciers and flow along steeper topographic gradients also occupy subglacial channels. They are, however, much narrower and deeper than the West Antarctic ice streams (Bentley, 1987). Jakobshavns Isbræ, West Greenland, is an ice stream terminating as an outlet glacier, which has a steep surface slope, and two up-ice coalescing branches. The entire glacier, including the portion considered to be a true ice stream, is underlain by a deep U-shaped subglacial trough which varies in width (>10 km to <5 km) and depth along its length, and reaches maximum depths of about 1500 m bsl (Clarke and Echelmeyer, 1996).

### 5.3.2: Subglacial Processes

Studies of the subglacial processes of modern glaciers provide important insight for the subglacial processes which may have occurred on the eastern Scotian Shelf.

Observations of modern ice sheet margins, including Jakobshavns Isbræ, West Greenland, and the West Antarctic ice streams, indicate an intimate interaction between glacial ice processes and glaciofluvial processes. This supports hypotheses for subglacial landform evolution which incorporate multiple processes. How ice moves and the evolution of subglacial morphology is largely dependent upon: (1) the substrate (till and/or bedrock, and bedrock type); (2) topography (flat or mountainous); (3) meltwater input and drainage systems; and (4) time. The substrate, hard or soft, will have a strong influence on the ability and the extent to which ice and meltwater can erode. Variations in topographic gradient can influence the direction, speed, and pattern of ice and meltwater flow. The amount of meltwater input in combination with the nature of the substrate and topography will influence the type of drainage system which develops (section 5.1.2). Time is important for progressive systems. More time can allow for progressive development of landforms, making them larger, wider and/or deeper. Time may not be important for landforms produced by a catastrophic process. All of these factors are important in understanding the development of glacial landforms.

The West Antarctic ice streams provide good evidence that subglacial processes are not constant through time and space. Studies show that the West Antarctic ice streams flow at different rates. Ice stream B, which is the most studied, has a significant negative net balance, similar to ice streams A and D, though to a greater degree than ice streams A and D (Shabtaie and Bentley, 1987). Ice stream B has shown a 50% decrease in velocity in less than three decades (Bindschadler, and Vornberger; 1998). Ice stream C shows a positive net balance (Shabtaie and Bentley, 1987). The lower portion of ice stream C appears to have stopped flowing about 130 years ago while the upper portion appears still to be moving, though at an unknown and reduced rate (Shabtaie and Bentley, 1987; Walker, 1993). Ice streams E and F do not show a significant negative net balance. In addition, ice stream B is widening (4 km in 20.1 years; Bindschadler and Vornberger, 1998) and developing headward toward the interior of the ice sheet, and ice stream C occupies a shallower trough (Bentley, 1987; Shabtaie and Bentley, 1987). Diversion of meltwater from ice stream C to ice stream B and/or ponding of meltwater under ice stream C (large bodies of water are believed to exist beneath ice stream C; Shabtaie and Bentley, 1987) have been used to explain the differences in velocity, and the apparent halt of ice stream C (Engelhardt, et al., 1990; Alley and Whillans, 1991; Walker, 1993). Shabtaie and Bentley (1988) suggested that complex bedrock morphologies beneath the ice streams and adjacent ridges separating them, reflect bed molding by ice flow with concentrated erosion in the basal troughs.

Ice flow may be controlled by different glacial processes in different environmental settings. Clarke and Echelmeyer (1996) proposed that internal deformation plays a dominant role in the dynamics of Jakobshavns Isbræ due to large shear stresses related to its steep gradient, and lack of a deformable bed, even though meltwater channels and till were observed at the base of the glacier. Joughin, et al. (1996) observed a mini-surge of Ryder Glacier, Greenland, and suggested that flow instability may be due to changes in the basal water system alone. Studies regarding the movement of the West Antarctic ice streams concentrate on deformable bed theories which include both erosion of subglacial bedrock and sediments, and deposition (i.e. Alley, 1989; Engelhardt, et al., 1990; Alley and Whillans, 1991). More recently Engelhardt and Kamb (1997) hypothesized that high water pressures causing the ice to lift from the bed are responsible for the fast movement of ice Stream B. Studies of Worthington Glacier, Alaska and Haut Glacier d-Arolla, Switzerland (Harper and Humphrey, 1995; Harbor et al., 1997), suggest the importance of nonuniform basal hydraulic and bed conditions for glacier flow, including differences in subglacial water pressures and systems, and the presence or absence of a deformable bed. It has been suggested that some ice within the West Antarctic ice streams may be frozen to the bed, and that bedrock outcrop acting as sticky spots could also result in, and/or emphasize, irregular basal morphologies and add to the complexity of ice flow (Alley and Whillans, 1991; Walker, 1993).

Examples from other glaciers also provide insight into the processes occurring at the base of glaciers. Humphrey and Raymond (1994) observed a glacial surge of Variegated Glacier, Alaska, and proposed it was influenced by two types of subglacial drainage systems: (1) the surging zone was underlain by a basal hydraulic zone of low water velocity and high water storage in the form of a distributed-flow system; and (2) the region down glacier of the surge front was underlain by a high velocity, low storage zone in the form of a conduit system (Humphrey and Raymond, 1994). They interpret extensive bedrock erosion at the base of the glacier and propose that over the course of a 20 year surge cycle, 0.3 m of bedrock was eroded, two thirds during two years of peak surge activity, with the majority of erosion during this time occurring over two months. It was concluded that basal erosion under Variegated Glacier occurred, and the basal hydraulic system acted to flush the eroded sediments (Humphrey and Raymond, 1994). This concept is further supported by observations of Sharp et al., (1994), who observed the basal ice layer of Variegated Glacier, and concluded that bedrock fracture was significant during surge conditions, as well as meltwater flushing of comminution products. They further stated that meltwater flushing of the ice-bed interface could maintain a clean ice contact, keeping the process of abrasion effective. Gustavson and Boothroyd (1987), studied Malaspina Glacier, Alaska, and suggest that subglacial regolith is entrained and transported by meltwater via subglacial and englacial tunnel systems. They point out the importance of englacial tunnel systems for transporting and evacuating debris which can be derived from the glacier bed as well as the surface.

Catastrophic releases of subglacial meltwater as per the outburst-flood hypothesis, have not been directly observed as a process responsible for subglacial landform evolution, and circumstances are such that they probably never will be. However, Jökulhlaups–sudden, rapid drainage of glacier-dammed lakes and/or water impounded within a glacier (Paterson, 1994)–have been observed in modern glaciated areas, and some floods have been extremely large (Paterson, 1994). Grímsvötn, is a large subglacial lake beneath the Vatnajökull ice cap, Iceland, which drains catastrophically approximately every 5 years (Nye, 1976). Approximately 25% of the lake water is derived from surface runoff, and 75% from geothermal melting. Lake Vostok, is large lake freshwater lake which lies under 4 km of ice in central Antarctica. This lake is 200 km long, 14 000 $km^2$ in area, and up to 500 m deep, and occupies a bedrock basin resembling a rift valley (Kapitsa, et al., 1996; Ellis-Evans and Wynn-Williams, 1996). Lake Vostok was predicted to have developed over the course of 1 million years. It represents the ability of large quantities of meltwater to be stored beneath an ice sheet.

The preceding discussions are referred to and provide background information to the following discussions on the origin of the morphology of the Eastern Scotian Shelf.

## 5.4: ORIGIN OF THE OF THE MORPHOLOGY OF THE EASTERN SCOTIAN SHELF

### 5.4.1: Introduction

Section 5.4 presents proposed origins for the morphology of the eastern Scotian Shelf. It is divided into two sections. First the outburst-flood hypothesis is presented (section 5.4.2). Next, the origin of the morphology of the eastern Scotian Shelf will be considered according to the zone classification presented in Chapter 4 (section 5.4.3). Each zone will be discussed individually except for Zone C and Zone D which, due to their similarities will be discussed together. Finally a regional synthesis, including a proposed model for deglaciation on the eastern Scotian Shelf, timing of incision formation, and an attempt to explain why the eastern Scotian Shelf if morphologically different from the western Scotian Shelf.

### 5.4.2: The Outburst-Flood Hypothesis

Anabranching subglacial channels (channels which bifurcate and rejoin) have been attributed to meltwater origins based on the difficulty of explaining such patterns by ice erosion alone (i.e. Brennand and Shaw, 1994), though this is only problematic if the channels are assumed to have formed synchronously. It is now widely accepted that the Channeled Scablands of Washington, which display an anabranching channel network, were formed by catastrophic releases of meltwater from a glacially dammed lake which flooded and eroded the pro-glacial landscape (Bretz, 1969; Baker, 1973).

Upon first glance, the morphology of the eastern Scotian Shelf (Fig. 1.2) appears to bear a strong resemblance to the anabranching system of the Channeled Scablands. However, upon close examination of the eastern Scotian Shelf morphology, it appears that there is not one system of channels or incisions, but several, and that the incisions do not anabranch, but converge or diverge, occur as isolated linear depressions, and display cross-cutting relationships (discussed in detail in Chapter 4).

The outburst-flood hypothesis invokes catastrophic releases of large quantities of subglacially stored meltwater flowing at great velocities beneath ice sheets (see Shaw et al., 1996 and references therein). This highly controversial theory has been proposed as a primary mode of subglacial landform evolution by both erosional and depositional processes, and challenges widely accepted theories on subglacial landform evolution by ice erosion, deposition, sediment molding/deformation, combined with non-catastrophic meltwater release.

The outburst-flood hypothesis originated through analogy, upon observing the similarity between erosional marks formed by turbulent water flow and streamlined subglacial landforms on glaciated terrain (i.e. Shaw, 1983; Shaw and Kvill, 1984; Shaw and Sharpe, 1987; Shaw et al., 1989; Kor et al., 1991). An outburst flood originates as turbulent sheet flow forming erosional and depositional features such as drumlins, Rogen moraine, flutes, scours, hummocky terrain, and bedrock erosional marks (s-forms) (Shaw et al., 1996 and references therein; Munro and Shaw, 1998). Vertically imposed, progressive, channelization by erosion of conduits into the substrate occurs upon the failure of unstable broad sheet flow as ice settled towards the bed (Brennand and Shaw, 1994; Gilbert and Shaw, 1994; Sjogren and Rains, 1994). Streamlined residuals between anabranching channels have been interpreted as drumlins (Shaw et al., 1996 and references therein). This hypothesis is based on field observations of morphology and sedimentology in the Livingstone Lake area (northern Saskatchewan), Georgian Bay area (southern Ontario and northern New York), Alberta, and Quebec (Shaw et al., 1996 and references therein; Brennand et al., 1996; Munro and Shaw, 1997).

Recently, Eyles and Boyce (1998) challenged the outburst-flood hypothesis by observing bedrock erosional forms such as flutes, crescentic grooves, and rat-tails along bedrock fault slip planes in unglaciated environments. They observed the similarity of these kinematic indicators to subglacial landforms, and showed through analogy, that turbulent meltwater is not necessary to form such features. Based on this observation, Eyles and Boyce (1998) concluded that streamlined subglacial landforms can result from the generation and transport of deformation till beneath an ice sheet, much like the generation and movement of a layer of diamict (gouge debris) between fault planes.

Eyles and Boyce (1998) do not address the origin of channelized morphologies, however, Gorrell and Shaw (1991) noted striations and meltwater features, also interpreted as abrasion features by Eyles and Boyce (1998), along the wall of a channel in which an esker lies. How much this channel was modified by meltwater, and how much this channel was modified by actively eroding ice needs to be addressed in more detail. Polished surfaces have been attributed to meltwater activity (i.e. Gorrell and Shaw, 1991) however, river pebbles and cobbles are smoothed and rounded, but not polished, while glacially striated surfaces are commonly polished. More obvious meltwater erosional features such as potholes are found on bedrock outcrops strongly affected by glaciofluvial activity. Many of these potholes are not polished or striated, while surrounding bedrock often is, indicating that polishing may be the result of ice and diamict abrasion against bedrock rather than meltwater activity (personal observation, NWT, Canada).

Pair (1997) noted erosional features (i.e. flutes and rat tails) along incisions, and attributed them to thin sheets of subglacial meltwater modifying pre-existing subglacially-modified (glacial erosion) bedrock valleys. This interpretation also discounts the necessity of outburst-floods for forming channels, though it does not acknowledge that some of the observed erosional features could have been formed by glacial abrasion rather than meltwater.

The amount of inheritance the landscape has from pre-glacial landform evolution (i.e. Jens-ove-Nöslund, 1997, and references therein; Lidmar-Bergstrom, 1997), pre-glacial fluvial activity, or earlier glacial episodes, including pro-glacial processes also needs to be addressed in more detail.

Isolated channels or scours have been explained by the outburst-flood hypothesis, and the possibility of these features being formed by preferential erosion by fast-flowing ice was acknowledged (Gilbert and Shaw, 1994). However, fast flowing ice was discounted as an agent of formation based on regional application of the flood hypothesis to the origin of surrounding streamlined landforms (Gilbert and Shaw, 1994). This argument loses its strength upon application of alternate theories for forming streamlined subglacial landforms (i.e.: Boyce and Eyles, 1991; Hart, 1997; Eyles and Boyce, 1998), and direct observation of streamlined forms forming and emerging from beneath modern glaciers (i.e. van der Meer, 1997; Hart and Smith, 1997; Bindschadler, and Vornberger, 1998), independent of significant meltwater activity.

Is the spectacular morphology of the eastern Scotian Shelf a product of outburst-flood events? The following application of the outburst-flood hypothesis is based on assuming certain conditions existed and these assumptions are presented only to provide parameters necessary to test this hypothesis.

First, for the eastern Scotian Shelf to be a product of outburst-floods, a source of meltwater is required. The most likely source of meltwater would be from overbanking of the Laurentian Channel assuming the Laurentian Channel was at maximum carrying capacity. Second, evidence for this overbanking must be found. The channels covering the eastern Scotian Shelf can account for this. A similar morphology is not documented on the Grand Banks, peripheral to the Laurentian Channel, though buried incisions have been identified, and incisions in an arcuate pattern along the Grand Banks, distal from the mainland of Newfoundland are evident in the bathymetry (Fig. 1.1). The incisions in an arcuate pattern on the Grand Banks do not appear to be related to discharge overbanking from the Laurentian Channel. It is unknown if buried incisions adjacent to the Laurentian Channel are related to overbanking as no pattern has been determined. Third, the incisions on the eastern Scotian Shelf must be explained by fluvial processes. Theories of how channelized morphologies, both anabranching and isolated, develop subsequent to sheet flow have been presented (Chapter 1, section 1.2; Chapter 5, section 5.1.3a). Discussions in Chapter 4, (section 4.4.1a), make it clear that fluvial channels can possess a variety of shapes, though the most common shapes for fluvial erosional channels are U-shaped and asymmetrical. Seismic profiles from the eastern Scotian Shelf show evidence of both of these shapes. Sediment fill does not have to be related to an erosional form, therefore, if incision fill does not correlate with a fluvial process, (i.e. till infills a channel), the origin of the fill can be disregarded as significant to channel formation. If, however, the incision fill does support the theory, then it can be considered as evidence. Such assumptions presume the origin of the incisions is known.

It appears that all the parameters exist to explain the origin of the eastern Scotian Shelf incisions by the outburst-flood hypothesis. However, the following points raise some critical questions which need to be answered to make this conclusion valid. Answering them is beyond the scope of this thesis, however, they do provide some guidelines for future research.

1. Meltwater Source

The location of a subglacial meltwater reservoir large enough to produce and source the discharges necessary to form landforms described as products of outburst-floods has yet to be resolved. Sources of meltwater for pro-glacial outburst floods are well documented by extensive lacustrine sediments marking the extent of former pro-glacial lakes (Bretz, 1969; Baker, 1973; Christiansen, 1979; Kehew, 1982; Kehew and Lord, 1986). It would follow that thick extensive lacustrine sediments should exist marking the extent of a large former subglacial lake. Such sediments have not been identified.

Shoemaker (1991; 1992) presented theories on how large subglacial lakes might form. He proposed that Hudson Bay would be a likely setting for the development of a large subglacial lake which would supply meltwater for outburst-floods. If the Hudson Bay acted as a reservoir for large volumes of subglacial meltwater, than it would be expected that thick lacustrine sediments would exist in the Hudson Bay. This is not the case. Till, sometimes multiple tills, overlie a smooth bedrock unconformity (based on seismic interpretations and core samples), and this is subsequently overlain by thin (>5 m) glaciomarine deposits (Josenhans and Zevenhuizen, 1990). An assumption that meltwater was subglacially stored beneath Hudson Bay and flooded towards the southwest and southeast ignores the existence of the Hudson Strait which would logically be the path of least resistance. The model assumes ice would dam this outlet. Shoemaker (1991; 1992) does not consider the possibility that ice may have been cold based over Hudson Bay and presents his theories assuming ice was warm based in order to provide a necessary parameter for theorizing the development of an extensive subglacial lake.

Josenhans and Zevenhuizen (1990) mapped fluted tills over large areas of the Hudson Bay floor, and sidescan sonograms reveal they are identical in form to streamlined features attributed to outburst-floods. These forms are supported by evidence from airphotos of streamlined landforms on land surfaces within and adjacent to the Hudson Bay. Fluted surfaces on the floor of Hudson Bay converge towards the centre of Hudson Bay which does not support catastrophic flow out of the bay. Subparallel flutes were mapped in the northern Half of Hudson Bay. These flutes orient themselves towards Hudson Strait. If these flutes are attributed to outburst-floods, then a meltwater source would have to be located southwest of Hudson Bay. If the flutes are reinterpreted to be oriented southwesterly, towards the Livingstone Lake flood path, then a meltwater source would have to be located northeast of Hudson Bay, within or beyond Hudson Strait. This is highly unlikely as it would require meltwater to flow towards land rather than away from it, and it ignores all evidence for ice flow out of Hudson Bay into Hudson Strait (Josenhans and Zevenhuizen, 1990; Shilts, 1982; Shilts, 1986). Anastamosing meltwater channels, interpreted to flow into Hudson Bay, were also identified on the floor of Hudson Bay on seismic records and sidescan sonogram records (Josenhans and Zevenhuizen, 1990).

All of the features identified on the floor of Hudson Bay have been explained as products of outburst-flood erosion by Shaw et al. (1996) and references therein. If this is the case, then finding a source of meltwater to explain outburst-flood meltwater forms across Canada becomes more controversial. The geomorphic and sedimentological evidence on the floor of Hudson Bay, a proposed meltwater reservoir, does not support a quiet environment of meltwater ponding. Instead, it supports an environment of active subglacial erosion. If the features on the floor of Hudson Bay are genetically related to any of the proposed outburst-flood events, increased amounts of meltwater are required, as is a reservoir outside Hudson Bay.

A model proposed by Fisher et al. (1988) suggests that thick ice frozen to the bed of Hudson Bay existed during the last glacial maximum, about 18 ka. If this model is correct, the time-frame for the buildup of a large subglacial lake within Hudson Bay is very short to nonexistent (i.e. outburst-floods across Alberta are expected to have occurred between 18-14 ka BP; Rains, 1993; Shaw 1994; Shaw, 1996). Lake Vostok, a large lake beneath Antarctica is located in a structural basin and had over 1 million years to develop (Kapitsa et al., 1996). Lake Grímsvötn, Iceland, forms near a source of great geothermal heat (Nye, 1976). None of these conditions (time and structure in the case of Lake Vostok; intense geothermal heat in the case of Lake Grímsvötn) existed in North America during the last glaciation.

The evidence presented here does not support the hypothesis that Hudson Bay was a potential source of large quantities of subglacially ponded meltwater. As yet, Hudson Bay is the only area where it has been proposed a large subglacial meltwater reservoir may have existed. The problem of a meltwater source for catastrophic subglacial meltwater flood events remains.

## 2. Landform Evidence

Convincing evidence of landforms created by catastrophic releases of subglacial meltwater needs to be identified in a path from a source to the Laurentian Channel. As it stands, the model of subglacial meltwater overbanking from the Laurentian Channel onto the eastern Scotian Shelf and eroding the eastern Scotian Shelf morphology is based on assuming this occurred. No evidence exists for meltwater erosion as the primary mode of origin for the Laurentian Channel except that it is an erosional feature. Multiple till units have been mapped at the base and along the flanks of the Laurentian Channel (King and Fader, 1990; King and Fader, in prep.), however, the outburst flood theory would disregard this evidence as not relevant to the creation of an erosional form (i.e. tills are produced and deposited after the major erosional event by meltwater occurred).

3. Incision Fill

Is incision fill relevant to an erosional form? Attempting to answer this question is a circular argument. Since the outburst-flood hypothesis requires meltwater to erode a form, all evidence of subglacial deposits (tills, glacial diamicts) which do not support the theory would be regarded as not relevant to the origin of the incision form. This is because deposition within the incision occurs subsequent to the erosional phase. If, however, incision fill does support the hypothesis (i.e. glaciofluvial deposits, hyperconcentrated flow deposits) then it is considered relevant to the origin of the incision and representative of depositional phases of waning flow. It is possible that glaciofluvial deposits are also not related to the erosional form and were deposited subsequent to glacial erosion (i.e. Zone C and Zone D).

Some fill towards the surface of the eastern Scotian Shelf incisions can confidently be identified as part of a till blanket subsequently overlain by glaciomarine deposits (Figs. 4.4, 4.18). Some incisions contain a basal acoustically incoherent unit (Figs. 4.5, 4.6, 4.7, 4.8, 4.16) overlain by glaciomarine deposits. It is possible this sedimentary sequence is the same as that mentioned above. However, without samples, the issue regarding the origin of the acoustically incoherent incision fill still remains. Proponents of subglacial erosion will favour a till interpretation (i.e. Amos and Knoll, 1987), and proponents of glaciofluvial erosion will favour a hyperconcentrated flow deposit interpretation (Boyd et al., 1988). Boulders identified within the fill of a sampled incision (Amos and Knoll, 1987), and point source reflectors within many of the acoustically incoherent incision fill deposits (Figs. 4.5, 4.6, 4.7,.8) indicate a bouldery deposit, and not massive sands.

With the exception of a few incisions near the inner shelf in Zone A which show clinoform fill (Fig. 4.3), basal stratified fill is generally horizontally stratified, and ponded to onlapping (Figs. 4.8, 4.9, 4.10, 4.15, 4.16, 4.18). This seismic signature indicates relatively passive incision filling in a low energy environment, and does not support deposition by glaciofluvial processes. Some incision fill displaying this seismic signature was interpreted as Tertiary valley-fill sediments (Figs. 4.9, 4.15, 4.16; Chapter 4). Figures 4.10 and 4.18 show stacked sequences within the incisions, which may be interpreted as Pleistocene deposits alternating between glacial diamict, glaciomarine, and normal marine sedimentation.

4. Plan-Form Patterns of Incisions

The plan-form patterns of the incisions identified on the eastern Scotian Shelf may provide some clues to incision genesis. The Bathymetry (Fig. 4.1) reveals that there are two dominant incision directions: (1) Zone B and Subzone $C^1$ incisions are oriented in a northeast-southwest direction; (2) Zone A, Subzone $C^2$, and Zone D incisions are oriented in a north-south direction, slightly skewed to the southeast or southwest. Two dominant incision patterns are evident: (1) Zone B displays a distributary/tributary pattern; (2) Zone C and

Zone D display isolated linear incisions, though Subzone $C^1$ incisions appear to have distributary/tributary extensions (Fig. 4.2, area in box).

If the incisions are a product of catastrophic subglacial meltwater erosion, questions such as these must be answered: Are all of the incisions evident on the eastern Scotian Shelf the product of one flood event or numerous flood events? Is the source of meltwater always the same (from the Laurentian Channel)? Where can alternate sources of meltwater be derived from? If the incision pattern is the product of simultaneous erosion, how are the cross-cutting relationships between Zone B and Subzone $C^2$ incisions (Area BC; Fig 4.1) explained, and how are the cross-cutting relationships between the linear incisions of Subzone $C^2$ explained? These questions cannot be answered with the limited data available, though they may be answered theoretically.

The plan-form pattern of Zone B incisions cannot be explained by ice erosion, therefore, water must have been the primary erosive agent. There is insufficient evidence to prove whether or not these incisions were formed by catastrophic meltwater erosion, or by a non-catastrophic subglacial drainage system. Seismic records reveal that the morphology may be partly inherited from pre-glacial fluvial erosion (Figs. 4.6, 4.15, 4.17). The origin of Zone B incisions is discussed in more detail in section 5.4.3b

If all of the incisions were modified and occupied during a single flood event, then complex meltwater flow patterns and changing hydraulic dynamics would have to be invoked to explain the cross-cutting relationships observed in area BC, where Subzone $C^1$ cross-cuts Zone B. In addition, changes in flow dynamics would have to be invoked to explain the difference in plan-form patterns, and the transitional area between Subzone $C^2$ and Zone B. A convincing argument has to be presented to explain why southwestward flow switched to southward flow (indicated by cross-cutting in area BC), and meltwater began eroding isolated basins (Subzone $C^2$ and Zone D), rather than anabranching channel networks (Zone B).

Gilbert and Shaw (1994) presented a theory for forming isolated basins, similar to Zone C and Zone D incisions. This model proposes that meltwater flowing over a scarp generates erosive vortices which erode linear depressions (see sections 1.2 and 5.3.1a). Such a scarp does not exist to the north of Subzone $C^2$ incisions or Zone D incisions, though a scarp does exist for Subzone $C^1$ incisions along the Laurentian Channel. Could a switch in flow regime due to switches in topographic gradients between Subzone $C^1$ and Zone B result in the transition from eroding deep linear channels or scours (Subzone $C^1$) to eroding relatively small radiating channels (Zone B)? The residuals in Subzone $C^1$ are at the same bathymetric contour as the residuals in Zone B, therefore, there is no evidence to suggest there was a difference in the morphology prior to incision erosion, to cause a switch form confined to less confined flow over Misaine Bank in Zone B.

Explaining the variable directions, and types of incisions identified on the eastern Scotian Shelf by the outburst-flood hypothesis will require a great deal of modeling and theorizing.

5. Ice Sheet/Ice Shelf Stability

The stability of the ice covering the Scotian Shelf at the time of an outburst-flood event needs to be considered. Under this theory, it would be assumed the ice sheet was grounded, and supporting evidence for grounded ice on the Scotian Shelf is abundant. Assuming zones of incisions on the eastern Scotian Shelf formed simultaneously, once a flood event is initiated, the ice sheet would be lifted from the bed during sheet flow, and portions of the ice sheet would become a floating ice shelf with turbulent meltwater flowing beneath. Rains et al. (1993), argue that little landscape modification would take place during deglaciation following a catastrophic meltwater flood event due to low ice sheet gradients and the remaining presence of thin clean ice. Would the ice shelf have been able to survive such turbulent subglacial meltwater flows in a marine environment, or would it simply have disintegrated? Even if rapid ice surging occurred during a flood event, it is still questionable if a thin ice shelf could have been maintained for regrounding in an open marine environment. In calm environments, ice shelves are at delicate equilibria as indicated by observation of a highly sensitive 'lightly grounded area' (sensitive to ice thickness causing rapid advance or retreat) near the terminus of West Antarctic ice streams A, B, and C, (Shabtaie and Bentley, 1987), and the sensitivity of other marine terminating glaciers and ice shelves (i.e. oscillations in flow rate of the Larsen Ice Shelf; Bindschadler et al., 1995; Skvarca, et al., 1995).

The subglacial meltwater flood hypothesis proposes that channelization takes place upon regrounding of the ice sheet. If the ice sheet no longer exists, this process is not possible. Can extensive channels be simultaneously eroded >300 m bsl without the confinement of an overlying ice sheet in the marine environment? without an overlying ice sheet, would meltwater have retained high enough velocities to continue eroding and carrying sediments, or would the meltwater have lost much of its velocity upon entering dense saline marine water, and simply dumped it's load before it could erode channels? These factors also need to be considered for the Laurentian Channel. Grounded ice at the mouth of the Laurentian Channel may have been extremely unstable and prone to rapid thinning and retreat via calving as a result of ungrounding (section 2.3.3). If this were the case, then it would not have been necessary for meltwater to find an alternate route over the eastern Scotian Shelf.

### *5.4.2a: Conclusions*

Many questions remain to be answered to conclude that the outburst-flood hypothesis is a valid explanation for the origin of the morphology of the eastern Scotian Shelf. In particular, the source of meltwater creates a problem. The floor of the Hudson Bay does not contain ponded lake sediments, it contains tills with superimposed erosional forms. These forms have been interpreted and explained as a product of turbulent subglacial meltwater under the outburst-flood hypothesis, and as products of subglacial erosion by others (i.e. Shilts, 1982; Shilts, 1986; Eyles and Boyce, 1998). It is difficult to explain how a potential meltwater reservoir could contain 'turbulent meltwater' erosional forms. Meltwater erosional forms oriented towards the centre of Hudson Bay require large subglacial lakes outside of the Hudson Bay. If the forms are interpreted to flow southwestward, towards the Livingstone Lake flood path, (disregarding flow directions interpreted on surrounding land), a source of meltwater northeast of the Hudson Bay would be required; an unlikely setting.

The origin of landforms 'up-flow' from the Laurentian Channel need to be convincingly explained by the subglacial meltwater flood hypothesis, and they need to show an obvious flow path which emanates from a potential source of meltwater. Hypothesized meltwater flow patterns need to be presented in a form which will explain the cross-cutting relationships, and various orientations and plan-forms of incisions on the eastern Scotian Shelf. The ability of an ice shelf to survive turbulent meltwater sheet flow also needs to be addressed in detail. If the subglacial channels were eroded under bank-full closed conduit conditions, the preferential erosion of the substrate over the overlying ice must be addressed. If erosion of the overlying ice roof also occurred, than additional quantities of meltwater are also required.

It is concluded that the subglacial meltwater flood hypothesis does not sufficiently explain or answer some critical questions which are necessary for assuming the morphology of the eastern Scotian Shelf is a product of this mechanism.

### 5.4.3: Origin of the Morphology of the Eastern Scotian Shelf According to the Zone Classification

The origin of the morphology of the eastern Scotian Shelf is now considered according to the zone classification presented in Chapter 4. The classification recognizes areas of incisions as morphologically different, indicating they may have been subjected to different formational processes. The approach taken for this analysis includes consideration of processes and morphologies observed in modern glaciated environments. It is believed modern glaciated environments provide important clues to processes which occurred during glaciation on presently deglaciated terrains such as the eastern Scotian Shelf.

#### *5.4.3a: Zone A*

Zone A, which is adjacent to a major highland area of Nova Scotia, Cape Breton Island, falls within the landward confines of the Scotian Shelf end moraine complex (Fig. 4.1). This is within the limits of the minimum model of Late Wisconsinan ice cover on the Scotian Shelf (Dyke and Prest, 1987). The

Scatarie moraine, which is a part of the Scotian Shelf end moraine complex is in close proximity, and occupies portions of St. Anns Basin (in Subzone $C^1$). Due to it's inner shelf location, Zone A may have been affected by local glacial advance and retreat phases, and associated glaciofluvial activity, which did not affect the middle and outer shelf areas.

Incisions towards the north of the Zone A are eroded into crystalline bedrock and this may account for their smaller scale. They may be largely inherited from fluvial drainage systems developed over the Appalachian Region since Late Jurassic and Early Cretaceous times (King and MacLean, 1976). Towards the south, Zone A occupies an area transitional to Subzone $C^2$, and may be genetically related (the origin of Zone C incisions is discussed in section 5.3.2c). In this transitional area, Zone A incision characteristics, including incision fill, are similar to that of Subzone $C^2$, though they are of a smaller scale (fill is generally an acoustically incoherent basal unit overlain by glaciomarine sediments). The area where Zone A joins Subzone $C^2$ between Canso and Misaine Bank (where it cross-cuts Zone B incisions) may represent the path of a large inland ice stream extending further off shore and cutting between Middle Bank and Banquereau (the ice stream hypothesis is presented in section 5.3.3c).

Meltwater discharge from land, and marine transgression may have had a stronger influence on the morphological evolution of this zone than on the middle and outer shelf zones during deglaciation, particularly in the areas closest to land. Zone A incisions may be comparable to the Valley Subzone of the inner shelf terrain zone classification (Stea, 1995). The Valley Subzone channels are of a similar scale to Zone A incisions to the north. Stea (1995) suggested some may be meltwater conduits, and he interpreted surfical channel fill (sampled) as a product of wave-current redistribution of shoreface deposits related to sea-level lowstand (Clinoforms in Fig. 4.3 possibly represent this).

### *5.4.3b: Zone B*

Zone B incisions are interpreted as a system of distributary/tributary R-channels subglacially eroded into surficial sediments and bedrock (Figs. 4.1, 4.5; section 5.1.2). It is difficult to follow the track of a single channel branching and rejoining, so it s not appropriate to describe the system as anabranching. These incisions modify a pre-glacial drainage system, though it remains unclear how much of this surficial morphology is inherited from subaerial Tertiary erosion. Ice is unlikely to erode a distributary/tributary system of small sinuous channels, therefore it is not considered as the primary formative agent.

There appears to be clear evidence of a pre-glacial fluvial system crossing Zone B of the eastern Scotian Shelf (Figs. 4.6, 4.15), however, this system would have been modified by subglacial processes. More data, particularly high resolution seismic profiles are required to reach conclusions regarding the formation of these channels. Questions remain such as: How much of the morphology is inherited from a pre-glacial landscape? Is the system tributary or distributary? Is it genetically related to the surrounding Zone C incisions? Is the system a result of subglacial outburst floods, progressive erosion, and/or non-catastrophic subglacial drainage? The lack of data makes it difficult to answer these questions, and relying on morphology alone can be extremely misleading.

Conclusions as to whether or not this system originated via progressive channelization following sheet flow from an outburst-flood event, or is representative of non-catastrophic drainage processes are not possible.

Apparent anastomosing channels have been explained by both catastrophic processes (Brennand and Shaw, 1994), and by non-catastrophic processes (Moores, 1989).

Zone B incisions are interpreted as older than Subzone $C^2$ incisions as indicated by cross-cutting relationships in the area marked BC (Fig. 4.1). Subzone $C^1$ incisions appear to be transitional to, or from, Zone B incisions (Fig. 4.2, area in box). Cross-cutting relationships do not exist here, therefore, relative ages cannot be determined. It is possible that incisions in this area are genetically related, or Subzone $C^1$ incisions are younger, and modify and (over)deepen Zone B incisions.

Zone B incisions may represent an exposed version of buried incisions above R2 unconformity within the Sable Island Sand Body described by McLaren (1988; section 1.3), beneath the surface of Banquereau (Amos and Knoll, 1987; section 1.3), and western Sable Island Bank (Amos and Miller, 1990; section 1.3). McLaren (1988) interpreted incisions above R2 (facies 1) as channels produced by subglacial and subaerial glaciofluvial erosion. Amos and Knoll (1987) interpreted incisions as channels which are glaciofluvial in origin, and Amos and Miller (1990) interpreted incisions as tidal channels. If the Misaine Bank, Zone B, incisions represent an exposed version of the buried incisions discussed above, it is clear that they are not the result of tidal scour. However, accentuated scour as a result of storm driven tidal currents resulting from water-mass acceleration over shallow banks (Amos and Knoll, 1987), may have played a role in the development of surface expression of these incisions during low sea level stands. This does not mean that the channels interpreted as products tidal scour on western Sable Island Bank (Amos and Miller, 1990) were not produced in this way. It does, however, point out the limitations of 2-D seismic surveys.

An alternative interpretation is that an extensive Tertiary delta system may have evolved in Zone B and the morphology represents a preserved landscape. A major continental drainage system fed such a delta system during the Late Jurassic-Early Cretaceous and formed the Sable Delta complex (Wade and MacLean, 1990).

### *5.4.3c: Zone C And Zone D*

Zone C and D are considered together in this section. This is because their morphology and incision fill sequences are similar indicating they may be a product of similar formative processes.

The formation of linear isolated incisions has been explained by Gilbert and Shaw (1994), and Wingfield (1990) who propose catastrophic meltwater release. Mullins and Hinchey (1989), and Mullins et al. (1996) propose high pressure non-catastrophic meltwater flow under ice streaming conditions, whereas others propose direct glacial abrasion (i.e. Carlson et al., 1982). These theories were presented in section 5.1.3

Many of the catastrophic meltwater release theories (section 5.1.3) used to explain linear isolated incisions suffer from the same problems outlined in section 5.3.1, in addition to those discussed by Ó Cofaigh (1996).

The process of ice streaming (the movement of faster flowing ice through slower moving ice which does not necessitate confinement by bedrock walls) is considered as a primary mode of origin for incisions in Zone C and Zone D. This discussion will take the same approach as the discussion for the outburst-flood hypothesis. A series of criteria need to be met and questions need to be answered if this is to be a valid hypothesis for the formation of Zone C and Zone D incisions.

1. Do Ice Streams Occupy Channels and/or valleys?

Section 5.2.1 describes in detail the existence of a series of troughs beneath the West Antarctic ice streams, and the existence of a deep U-shaped valley beneath Jakobshavns Isbræ, Greenland. This includes the portion which is a pure ice stream, in addition to the portion which is an outlet glacier. Section 4.4.1b, in Chapter 4 discusses in detail the evolution of glaciated valleys and their resulting form. It is generally agreed that the final form of a glaciated valley is U-shaped or parabola shaped, however, fjord cross-profiles have also revealed that glaciated valleys can also be V-shaped (Syvitski, 1987; J. Shaw E, pers. comm., 1998). Zone C and Zone D incisions possess a variety of shapes which can be described as typical of glaciated valleys (section 4.4.1b). Figure 4.7 provides a good example of U-shaped valleys. It is concluded that ice streams do sometimes occupy channels and/or valleys

2. Size of an Ice Stream

The modern analogues presented above (the West Antarctic ice streams and Jakobshavns Isbræ) are of a much larger scale than Zone C and Zone D incisions. Can ice streams at the scale of the eastern Scotian Shelf Zone C and Zone D incisions exist (>300 m deep and >4 km wide)?

Satellite imagery over James Ross Island (Skvarca, et al., 1995), and the Antarctic peninsula (Bindschadler, et al., 1995) displays an abundance of ice streams and outlet glaciers along their coasts which are comparable in plan view to the eastern Scotian Shelf Zone C and Zone D incisions. Figure 5.1, of the Antarctic peninsula, shows ice streams and outlet glaciers flowing into Larsen Ice Shelf which are almost identical in width, spacing, and length (i.e. 1-2 km wide; 1-2 km between the ice streams and outlet glaciers; 3-4 km long) to

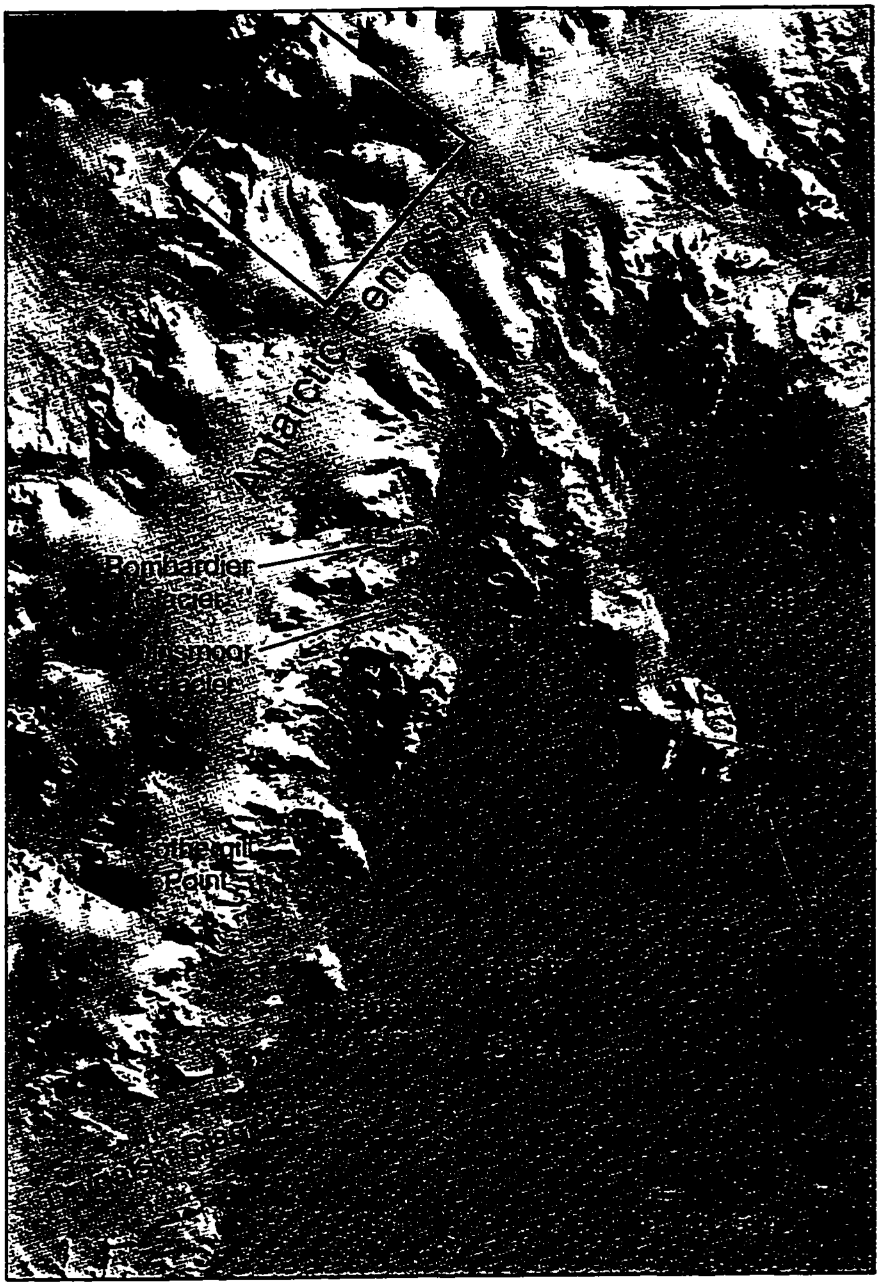


Figure 5.1 Landsat-MSS image of the northern Larsen Ice Shelf off the Antarctic Peninsula. This image depicts numerous outlet glaciers and ice streams feeding the Larsen Ice Shelf which are similar in scale to Zone C and Zone D incisions of the eastern Scotian Shelf. This image may be used as a modern analogue representing what the eastern Scotian Shelf may have looked like when it was covered by ice (compare with Figure 1.2). The area in the box marks an area of converging ice streams (compare with area in box in figure 4.2) Modified from Bindschadler et al. (1994).

Zone C and Zone D incisions. This may provide insight into the configuration of the eastern Scotian Shelf when it was covered by ice.

The terrain of the Antarctic Peninsula is mountainous. How accurate an analogue is this image to the eastern Scotian Shelf, considering the eastern Scotian Shelf is not mountainous, but a flat coastal plain. This question is difficult to answer. Ice streams have been described as similar to valley glaciers with the exception that they are bounded by ice and not rock (Hughes, 1992). The Antarctic Peninsula ice cap is much smaller than the West Antarctic Ice Sheet which feeds the West Antarctic ice streams. It would stand to reason that ice streams and outlet glaciers around the Antarctic Peninsula would therefore be much smaller. Local ice centres on Nova Scotia (Stea, 1995) would be more comparable to the scale of the Antarctic Peninsula than the West Antarctic Ice Sheet or the Laurentide Ice sheet.

Why are the West Antarctic ice streams, which lie on a substrate similar to the Scotian Shelf, much larger than Zone C and Zone D incisions? This is likely because the West Antarctic ice streams are much longer-lived features than similar features on the eastern Scotian Shelf would have been. Therefore, they have had a much longer period of time to develop. The West Antarctic ice streams would not have originated at their present scale, they would have started as much smaller features which grew with time. The length of ice streams on the Scotian Shelf may have simply been constrained by the shelf width and the size of local ice caps feeding the ice streams. Dispersal centres over the Appalachians would have been much smaller and closer to the coast than the Antarctic ice streams which are fed from dispersal centres very far from the coast.

3. Can Ice Streams Erode?

This question is the most critical one to answer. If an ice stream cannot erode, then the hypothesis that Zone C and Zone D incisions were formed by ice streams is invalid. What evidence from modern environments suggest that ice streams can erode? The fact that they are lie in troughs and are underlain by till is good evidence. Recently, Bindschadler and Vornberger (1998) provided evidence that West Antarctic ice stream B eroded 14 km of Ridge BC (a ridge between ice stream B and ice stream C). They predicted an erosion rate of 447±34 m/yr over a time span of 29.1 years. This statistic represents erosion of the ice sheet ice, however, it likely would result in subsequent erosion of the substrate. Sharp et al. (1994) studied Variegated Glacier, a surge type glacier, and pointed out the importance of both bedrock fracture and meltwater flushing during surge conditions. They suggested that meltwater flushing of the ice bed interface during surge conditions could act to keep the basal ice clean and allow for effective abrasion of the glacier bed to occur. Humphrey and Raymond, (1994) concluded Variegated Glacier eroded 0.3 m of bedrock during a 20 year surge cycle, and most of this debris was eroded during two months. These studies show that abrasion and erosion can occur in temperate glaciers where melting occurs at the glacier bed. In addition, modern marine terminating glaciers have been observed expelling large quantities of debris laden meltwater at their terminus (i.e. Gustavson and Boothroyd, 1987). It is concluded that ice streams erode (see section 5.3.2).

4. Incision Fill

Discussing the incision fill in this section has the same problem as it did for discussing the outburst-hypothesis. Incision fill may not be relevant to the origin of the erosional form.

Evidence exists for subglacial modification of Tertiary fluvial valleys (Figs. 4.9, 4.16). Incision fill can be variable in character (Fig. 4.8) indicating that unique depositional processes occur in incisions adjacent to one another. Acoustically incoherent deposits occur towards the base of many incisions (Figs. 4.7, 4.8) which can be interpreted as a product of deposition by hyperconcentrated flows, debris flows, slope failure, or subglacial processes. Stacked sequences occur in a series of incisions in Zone D (Fig. 4.18) which may represent multiple till sequences (multiple till sequences are recognized on adjacent surfaces). Where units show variable thicknesses and acoustic signatures (Figs. 4.10, 4.18), they may represent alternating deposits of glacial diamict, glaciomarine, and normal marine deposits such as were identified by Sejrup et al. (1996) in the Norwegian Channel.

The surficial sediments in the majority of incisions consists of till overlain by glaciomarine deposits (Figs. 4.11, 4.12, 4.13, 4.17, 4.18, 4.19). Figure 4.7 depicts the cross-profile of Subzone $C^1$ incisions adjacent to the Laurentian Channel. They contain simple sequence similar to the surficial deposits; an acoustically incoherent deposit overlain by Emerald Silt. If this unit is interpreted as till, than it correlates to the surficial sediments confidently identified by high resolution Huntec DTS data and indicates ice occupied the base of these incisions.

The fact that West Antarctic ice streams adjacent to one another display unique characteristics indicates that ice stream activity is not uniform through time or space (section 5.2.2). This may explain the differences in incision fill, and stacked sequences observed in Zone C and Zone D incisions. The differences in fill of the eastern Scotian Shelf incisions may represent different phases of morphological evolution beneath ice streams.

4. Plan-form Pattern of Incisions

Linear incisions can be explained theoretically by catastrophic meltwater erosion (section 5.1.3a). However, they can also be explained by ice streams. This is evident from the form of the West Antarctic ice streams, the Antarctic Peninsula ice streams, and other outlet glaciers and ice streams (Fig. 5.1). Area BC (Fig. 4.1) can simply be explained as a product of different processes occurring at different times, rather than of changing meltwater flow regimes and directions during a single event. Area BC can be interpreted as a product of ice streams developing headward at an ice margin, cross-cutting an earlier Tertiary and/or glaciofluvially influenced landscape. The area which is transitional between Zone B and Subzone $C^1$ (Fig. 4.2, area in box; interpreted as flow into the Laurentian Channel) strongly resembles the area marked in a box on Figure 5.1 which shows converging ice streams flowing out of the ice sheet covering the Antarctic Peninsula. This area also resembles the larger system of troughs mapped beneath the West Antarctic ice streams (Shabtaie and Bentley, 1988; section 5.3.1). These incisions may be interpreted as overdeepenings of the earlier meltwater influenced landscape. Because they have the same orientation as Zone B incisions, no cross-cutting relationships can be identified.

Seismic profiles suggest that Zone C and Zone D incisions are not completely glacial in origin. Some incisions are inherited from fluvial valleys (Figs. 4.9, 4.15, 4.17). Some of the morphology of the eastern Scotian shelf is clearly the product of a pre-glacial fluvial activity. Patterns toward the south of Subzone $C^1$ (Fig. 4.1) display a meandering form which has not been explained by the catastrophic outburst flood mechanism or by ice erosion. King et al. (1974) mapped the proposed Tertiary drainage pattern which flowed across the eastern Scotian Shelf (Fig. 2.4).

The cross-cutting patterns observed within Subzone $C^2$ may be attributed to different ice margin configurations. Stea (1995) mapped a complex series of glacial ice flow patterns for the Wisconsinan glaciation on the mainland of Nova Scotia. The earliest of the most extensive ice-flow phases (70-30 ka; Phase 1a) flowed eastward and may have influenced Subzone $C^1$, adjacent to the Laurentian Channel. This area may also have been influenced by flow radiating off of an ice centre situated over Misaine bank (ice-flow Phase 3), in addition to influencing incisions to the south in Subzone $C^2$. An ice centre situated over Misaine bank would also explain why Zone B, an older landform assemblage, was preserved. Four ice flow phases were mapped for the Wisconsinan. Deep water and terrestrial evidence (reviewed in section 2.3.4a) indicates that the Early Wisconsinan and Illinoian were times of more extensive glaciation over Nova Scotia than the Late Wisconsinan. Ice flow directions, and the effects of each of these glaciations on the eastern Scotian Shelf cannot be determined. Could it be that the effects of all these glaciations are represented in this morphology?

The boundary between the Laurentian Channel and the eastern Scotian Shelf is characterized by till overlying bedrock (MacLean and King, 1970). This deposit may be viewed as a medial moraine composed of tills from both an outlet glacier/ice stream flowing through the Laurentian Channel and tributary glaciers from Subzone $C^1$ joining the Laurentian Channel ice. The northern edge of Sable Island Bank and Banquereau display a cuspate, irregular morphology. This is comparable to the cuspate form of the Browns Bank Moraine (Scotian Shelf), and to what Boulton et al. (1996) identify as a lobed frontal margin which they describe as reflecting glacier lobes resulting from a recent ice surge of glacier Sefstrombreen, Spitsbergen.

## 5. Stratigraphic Evidence

On the outer banks of the eastern Scotian Shelf (Sable Island Bank and Banquereau), large buried incisions are located below regional unconformity R2 (dated at approximately 28 000 yr BP; Amos and Miller, 1990). Dates from sediments below and above this unconformity range from 32 000 yr BP to 28 000-3000 yr BP respectively (Amos and Miller, 1990). Foraminiferal evidence indicates fluctuations between open marine and ice shelf conditions during this time (Amos and Miller, 1990). The lithostratigraphy does not provide any evidence of massive sands from catastrophic debris dumping. The sediments covering Sable Island Bank and Banquereau are interpreted to be representative of the Emerald Silt formation (Amos and Knoll, 1987; Amos and Miller, 1990). Foraminifera within the sediment sequences are marine and indicate fluctuations between ice shelf and open marine conditions across the outer banks of the eastern Scotian Shelf. There is no evidence in the biostratigraphic record of sudden large quantities of freshwater input. Based on these observations, there is no evidence for outburst-flood sedimentation on the eastern Scotian Shelf outer banks during the Mid to Late Wisconsinan. The same formations exist on Banquereau. The lithostratigraphy and biostratigraphy of western Sable Island bank and Banquereau provide strong evidence that during the Mid to Late Wisconsinan, the outer eastern Scotian Shelf Banks were affected by open marine and ice shelf conditions.

Alley et al. (1989), and Fahnestock (1996) suggested rapid discharges from outlet glaciers and ice streams would likely have been necessary for the maintenance of a temperate ice shelf. This is exemplified by the high input of ice from Jakobshavns Isbræ, necessary to maintain its ice shelf. Ice streaming may have been necessary to maintain ice shelves on the outer eastern Scotian Shelf banks.

## 6. Additional Considerations

King and Fader (1992) observed buried and open incisions on the southeastern part of the northeastern Newfoundland Shelf. These incisions are comparable in dimension to Zone C and Zone D incisions. They display a radiating pattern, and independent moraines with complex internal structure at their terminus. King and Fader (1990) propose that these features represent the tidewater terminus of an isolated ice lobe emanating from north-central Grand Bank. The complex internal structure and coarseness of these moraines indicates the importance of meltwater input for their formation.

These incisions are part of an arcuate pattern of incisions visible on the bathymetry of the Grand Banks of Newfoundland (Fig. 1.1). This type of pattern likely marks an ice marginal position. It is unclear if the terminus was tidewater or not. The concept of tidewater termini is based on associating meltwater channels with this type of terminus. If the incisions are not completely glaciofluvial in origin, but features formed by ice and meltwater erosion under ice streaming conditions, than the terminus need not be a tidewater terminus.

Approximately 90% of present day ice sheets covering Antarctica and Greenland is discharged by marine ice streams (Hughes, 1992). Modern ice streams occupy subglacial channels and/or valleys, and have the ability to erode. Adjacent ice streams do not display consistent behavour through space or time (section 5.3) which may explain variable fill among Zone C and Zone D incisions. Interpretations of seismic profiles suggests some of the eastern Scotian Shelf morphology is inherited from the Tertiary and suggests some of the incisions are representative of Tertiary fluvial valleys. The ice flow history documented on land reveals a complex pattern of ice flow directions during the Wisconsinan, this does not include the glaciations prior to this. Cross-cutting relationships of incisions observed within Subzone $C^1$ and area BC may likewise represent shifting ice terminus configurations. As well, a complex ice flow history may explain the difference in the two dominant incision orientations (north-south and northeast-southwest).

It is concluded that Zone C and Zone D incisions were most likely developed via a combination of glacial erosion and meltwater erosion under ice streaming conditions.

### *5.4.3d: Conclusions*

The conclusions reached concerning the origin of the morphology of the eastern Scotian Shelf within each zone are derived from a more local approach than the outburst-flood hypothesis. This negates the necessity to explain morphologies, both regionally and from far distances, by a singular mechanism. Many answers to the questions posed above can be supported by evidence from modern glaciated environments. The analogues used in this analysis are not perfect as they include polar glaciers and mountain glaciers, however, they do provide pertinent information and evidence to validate processes which may have occurred at the margins of the Laurentide ice sheet, in particular the eastern Scotian Shelf region.

It is concluded that Zone C and Zone D incisions of the eastern Scotian Shelf were most likely formed by erosion by both ice and meltwater, near the ice margin, under ice streaming conditions. The cross-cutting relationships and differences in orientations of Zone C and Zone D incisions are likely the result of different ice sheet configurations (Loncarevic et al., 1992) during different Pleistocene glaciations, reconfigurations during ice retreat, and ice flow phases (Stea, 1995). Zone B incisions likely represent a pre-glacial landscape modified by a subglacial drainage system which is older than Zone C incisions, and Zone A represents a combination of these processes, including influence from a greater number of glacial phases due to its inner shelf location.

## 5.5: REGIONAL SYNTHESIS

### 5.5.1: A model of Deglaciation for the Eastern Scotian Shelf

The outer banks of the eastern Scotian Shelf are largely devoid of bouldery deposits except for some buried incisions (Amos and Knoll, 1997; Amos and Miller, 1990). However, sand and gravel deposits are abundant. It is suggested that the large scale buried incisions (below R2) were pathways occupied by ice streams during early extensive ice advance which carried debris further off shore towards the margin of the continental shelf. This could explain why some of the incisions contain bouldery debris, different from the local bedrock. Slower moving ice on the banks may just have eroded the underlying Tertiary and Cretaceous sediments and reworked glaciomarine deposits, thus explaining the lack of bouldery deposits on the outer banks. Faster flowing ice in ice streams may have derived a large portion of glacial debris from greater distances, while slower moving surrounding ice may have derived glacial debris from local sources.

As ice retreated, it may have grounded in positions of stability from which ice streams could feed unstable temperate ice shelves. Non-catastrophic debris laden meltwater would simultaneously discharge, producing sediment plumes which deposited the Emerald Silt facies. This may have been the situation along the north end of Sable Island Bank and Banquereau. The grounding line would be the cuspate northern end of the banks relative to where the thick bank sediments begin. Upon ungrounding (indicated by crevasse intrusion features–lift-off moraines on the surface of tills within the incisions; crevasses were only observed at the base of floating ice in the Antarctic ice streams, section 5.3) ice may rapidly retreat, which would explain why Zone C and D incisions are incompletely filled. Regrounding in a stable up-ice position would result in deposition of thick glaciomarine sediments at a new location.

The lack of positive features representing moraines at ice marginal positions would be due to a result of reworking and erosion during low sea-level stands and marine transgression. Such sediments would subsequently have been deposited as Sable Island Sand and Gravel deposits. This situation may be represented by the incisions observed on the Grand Banks (King and Fader, 1990). Moraines are preserved at the distal end of incisions in deeper water, where they would not have been subjected to erosion and reworking by low sea-level stand and marine transgression. Incisions in shallower waters lack similar moraines because they were probably eroded during low sea level and marine transgression. Dates provided by Amos and Miller (1990) for the Emerald Silt formation covering southwestern Sable Island suggest progressive deposition between 32 000 yr BP, and 3000 BP. Biostratigraphic analysis suggests conditions fluctuated between open marine and glaciomarine conditions. The time period between 26 000-13 000 yr BP was a time of warmer open marine conditions than prior to 26 000 yr BP. This in combination with sedimentary structure in the sand deposits (see Amos and Miller, 1990) does not support rapid sedimentation resulting in massive sands, and therefore does not support the concept of deposition by catastrophic floods after 30 000 yr BP. Whether or not this occurred prior to 30 000 yr BP cannot be determined.

This model suggests that the eastern Scotian shelf would have been deglaciated progressively, and that incisions would therefore be older towards the outer shelf than on the inner shelf. This model may apply to the Late Wisconsinan, and earlier glacial advances over the Scotian Shelf when the major excavation of the incisions likely took place. Local ice caps may have existed on the banks at various times, and local readvance off of the banks may also have reoccupied incisions (i.e. the till tongue in a channel off Canso Bank; Fig. 4.14a). These concepts need to be tested with additional seismic data, dating control, and sampling.

### 5.5.2: Timing of the Formation of the Incisions of the Eastern Scotian Shelf

Based on evidence from on and offshore (section 2.3.3a), the pre-Wisconsinan and Early Wisconsinan were the times of maximum glaciation on the Scotian Shelf. This may have been the time of major incision excavation on the eastern Scotian Shelf. Piper et al. (1994) proposed OIS 14 to OIS 12 (Mid Pleistocene) was a time of extensive erosion and excavation of Cretaceous and Tertiary bedrock.

Boyd et al. (1988) obtained a date of 37 000 ± 400 yr BP (RIDDL #639) from upper sediments infilling one of the buried channels beneath Sable Island Bank, and Amos and Miller (1990) recovered dates from just above R2 giving an approximate time bracket of 28 000-26 000 yr BP on Sable Island Bank. Amos and Knoll (1989) Recovered dates just above R2 on Banquereau providing an approximate time bracket of 18 000-26 000 yr BP for its formation. All of the large scale incisions comparable to those of Zone C and Zone D are located below R2, which indicates they are older than at least 18 000 yr BP. King (1993) suggested that some of the wide range of dates may be due to reworking during a later glacial advance. Dates reported by Gipp and Piper

(1989) and Piper et al. (1990b) for Emerald Basin and LaHave Basin respectively indicate lift-off at approximately 18 yr BP. These dates, in addition to dates from Emerald Basin in ice-distal Emerald Silt facies B of 14.5-10.0 yr BP (King and Fader, 1988) suggest ice had retreated to the inner shelf before 14 yr BP, at least in the west.

Estimates of ice thickness, based on sediment geotechnical properties, indicate that ice on the Scotian Shelf was relatively thin during the Late Wisconsinan (minimum 175 m, maximum 250 m along the outer shelf; Mulder and Moran, 1995) and the existence of till tongues and lift-off moraines within eastern Scotian Shelf channels lends evidence to support the concept of a lightly grounded ice sheet at the time of deglaciation during the Late Wisconsinan. The original hypothesis of lift-off moraine formation, in which moraines are formed by the squeezing of subglacial material into bottom crevasses (King and Fader, 1986) is supported by the identification of basal crevasses limited to floating areas of West Antarctic ice stream C which are (Shabtaie and Bentley, 1987). This evidence suggests that the Scotian Shelf ice was partially buoyant and in a very sensitive state during the time of Late Wisconsinan deglaciation.

It remains unknown whether the eastern Scotian Shelf incisions were formed simultaneously or time transgressively. The outburst-flood hypothesis suggests they were not formed time transgressively, the ice streaming hypothesis suggests they were and that Zone B represents a preserved landscape assemblage. This would mean D incisions would be older than Subzone $C^2$ incisions. If Subzone $C^2$ incisions are connected under the banks to Zone D incisions, then it would mean the incisions formed progressively from Zone D to Subzone $C^2$.

## 5.5.3: Comparison of the Eastern Scotian Shelf Morphology With the Remainder of the Scotian Shelf

The morphology of the eastern Scotian shelf is distinct from the remainder of the Scotian Shelf which is characterized by broad open basins and troughs between banks. Why do these morphologies differ? This is a perplexing and difficult question to answer. How much of the morphology is influenced by a pre-glacial/Tertiary drainage system? King et al. (1974) mapped this morphology (Fig. 2.4), which shows that the eastern Scotian Shelf was dissected more extensively than the remainder of the Scotian Shelf. An assessment of the bedrock geology of the Scotian Shelf shows the development of the Sable Island area during the Late Jurassic and Early Cretaceous (Wade and MacLean, 1990). Here, a sediment transport path was documented flowing out of the pre-glacial Laurentian Channel, with a portion of the flow path branching across the eastern Scotian Shelf producing the Sable Delta complex. It is possible a similar flow pattern occurred during Tertiary fluvial development on the eastern Scotian Shelf which initiated the template for Pleistocene processes to follow. This could in part explain the northeast-southwest trend of incisions in zone B and Subzone $C^1$ and for the greater intensity of dissection on the eastern Scotian Shelf compared to the western Scotian Shelf. It appears that this possible flow path across the eastern Scotian shelf may be against topographic gradient, however, the present morphology is a product of glacial modification and overdeepening. The Laurentian Channel is interpreted as a glacial feature (King and Fader, 1986; King and Fader, 1990; King and Fader, in prep.), therefore, such a flow path may have been possible prior to glaciation and excavation of the Laurentian Channel.

It is possible that the differences in the Scotian Shelf morphology represent different types of glacial erosion. Van dijke and Veldkamp (1996) support the concept of high pressure subglacial meltwater eroding channels, recognizing the importance of ice in the process. They propose a reason for channelized versus basin type morphologies. In their model, ice initially advances and creates a subglacial depression. Once deglaciation begins, greater quantities of meltwater are produced, and channels develop evacuating meltwater during retreat. Basins are proposed to form by erosion during readvance over the channelized morphology.

The fact that many incisions on relatively flat lying continental shelves (including the West Antarctic ice streams) reach maximum depths of approximately 500 m may be an indication that some type of threshold depth may be involved, after which lateral erosion becomes more dominant. This may in part be due to increases in the size of ice streams through lateral accretion and/or increased input of ice from the head of the ice stream as it matures and develops headward. Hirano and Aniya (1988) propose that continental ice sheets may produce broad U-shaped valleys in low lying areas as opposed to mountainous areas. The area of St. Anns Basin on the eastern Scotian Shelf may be a product of adjacent ice streams widening and coalescing resulting in an apparent basin. The basins to the west of the eastern Scotian Shelf may be the result of this process and represent more advanced ice stream development. In addition, the joining of incisions as a result of lateral migration may also be a possibility, as exemplified by the West Antarctic ice streams which display evidence of lateral migration (Shabtaie and Bentley, 1988). St. Anns Basin is close to the inner shelf and contains a portion of the Scateri Moraine of the Scotian Shelf End Moraine Complex. Readvance into this area may have resulted in the basin. The eastern and western Scotian shelf may show different degrees of glacial modification resulting in different morphologies. One morphology may be more mature than the other. Consequently, what is observed may represent different degrees of morphological development.

The above are only suggestions for the different morphologies of the eastern and western Scotian Shelf. It remains unknown why they differ so considerably.

### 5.5.4: Conclusion

According to ten Brink and Schneider (1995), the eastern Scotian Shelf morphology represents a 'young' glaciated continental shelf as opposed to a 'mature' glaciated continental shelf such as the Antarctic continental shelf. They envisage mature glaciated continental shelves as products of gradual and incremental development over many glacial advance and retreat cycles. Mature glaciated continental shelves are characterized by a stratal geometry and morphology dipping landward, as opposed to the morphology of the Scotian Shelf with its seaward-dipping, irregular morphology, and discontinuous seismic reflections, representative of a young glaciated continental shelf.

The eastern Scotian Shelf morphology likely reflects a number of processes including influence from pre-glacial fluvial systems which were not entirely removed by glacial erosion and deposition, subglacial meltwater erosion and deposition, and ice streaming. Evidence of multiple glaciations on the eastern Scotian Shelf suggests the morphology may also reflect the products of multiple ice advance and retreat phases.

# CHAPTER 6: CONCLUSIONS

## 6.1: CONCLUSIONS

The conclusions regarding the origin of the morphology of the eastern Scotian Shelf are based on observation (seismic stratigraphy and bathymetry), a comparison to observed modern glacial processes, and theory. They were derived from a more local approach to analyzing the morphology than the outburst-flood hypothesis requires. This negates the necessity to explain morphologies, both regionally and from far distances, by a singular mechanism.

Ice streaming is the model which best fits the origin of incisions in Zone C and Zone D. Ice streaming does not explain Zone B incisions which may be the result of a strong pre-glacial fluvial influence, a non-catastrophic subglacial channelized distributary system, and/or catastrophic meltwater release. Zone A incisions, close to the inner shelf, likely represent a combination of glaciofluvial processes and ice streaming processes, in addition to being affected by more glacial phases than the middle and outer shelf. The boundaries of each zone blend into one another, making it obvious that transitional areas exist, and that there must be some genetic relationship between the incisions, in addition to unique processes as displayed by their differing characteristics. These models are suggested to best fit the morphology of the eastern Scotian Shelf, and may be applied to incisions in other areas with similar characteristics. As clearly represented by the zone classification of the eastern Scotian Shelf morphology and in the varying morphologies of other channelized landscapes and continental shelves, incisions in different zones may have been influenced and formed by differing processes.

The lithostratigraphy and biostratigraphy of western Sable Island Bank and Banquereau do not support deposition from outburst-flood events. The foraminifers within the Emerald Silt formation covering these banks extend back to >30 000 yr BP, and are all marine. There is no evidence of unusual amounts of fresh water input on the outer banks of the Eastern Scotian Shelf from the Mid Wisconsinan to the Holocene.

From this regional analysis of the eastern Scotian Shelf, it can clearly be seen that the incision networks are probably not the result of one formational event, or of synchronous formation. Rather, they are more likely the result of multiple glaciations and differing ice sheet configurations which modified an ancient fluvial landscape by both ice and meltwater. The dominant incision fill composing at least the upper sediments appears to be Late Wisconsinan recessional deposits. This indicates that Late Wisconsinan ice was both extensive across the Scotian Shelf, and likely occupied at least the upper portions of all of the eastern Scotian Shelf incisions during ice advance and retreat. One important observation, independent from the question of the mode of incision origin, is the presence of tills in the upper regions of almost every incision surveyed. This lends strong support for the ability of glacial ice to mold and deform into very complex patterns, while still preserving older morphologies, even in the weakest of bedrock types, like those representative of the Tertiary and Cretaceous sediments of the Scotian Shelf.

Further study of the eastern Scotian Shelf will help clarify some of the debate concerning the mode of incision formation and contribute to a better understanding of the unusual and spectacular morphology of the eastern Scotian Shelf. In particular, closer spaced survey lines, using both Huntec and airgun systems with accurate navigation, and sampling of the basal incision sediments are necessary to gain a better understanding of the relationships of the various incision systems.

### 6.1.1: A Final Comment

In explaining the origin of incisions in glaciated regions, some validity can be found in all of the theories outlined in section 5.1. However, there are often trends which presume to explain the origin of apparently similar morphologies by a single mechanism. Based on variable terrain and bedrock types, basal ice conditions (temperate or cold based), hydrology, proximity to the ice margin, direct observation of glaciers and stratigraphic sequences, it becomes evident that the use of one mechanism to explain all features of similar morphology may be misguided. All factors relevant for a given area need to be considered. Incision may be the result of numerous processes, and multigenerational, or in some cases they may be the result of a singular process, and of a single generation. Detailed observation of morphology, sedimentology, and regional distributions suggest that processes which may explain incisions in one region may not explain them in another. Occam's Razor may be argued, however there may also be a problem with oversimplification. As so eloquently stated in a quote for all geomorphologists to heed, passed on by Fader (pers. comm., 1998) from an unknown originator, "beware of the lethal lure of morphological similarities."

### 6.1.2: Recommendations For Future Work

The following are recommendations for future research concerned with the origin of the morphology of the eastern Scotian Shelf.

1. 3-D seismic surveys to determine the degree to which the exposed network of channels extend under the adjacent banks. This will help determine if the buried channels are connected to the exposed channels.
2. Additional detailed surveys over Misaine bank to determine the extent of buried channels on Misaine Bank.
3. Additional high resolution seismic surveys to determine the nature of channel fill of the exposed channels.
4. Additional airgun surveys over Zone C and Zone D to determine the channel fill to the base.
5. Additional profiles down the length of the channels to compliment the cross-section surveys in order to determine if channel fill changes along the channel length, and to determine the nature of the long profiles of the channels
6. More work with the bathymetric data, including fleeter mouse fly-through, to understand the morphology better by viewing it from different perspectives, and to obtain long profiles of the exposed channels.
7. Sampling of the sediments at depth which at present are interpreted solely on the basis of seismic data. This includes both the stratified sediments and the acoustically incoherent sediments. This will help determine if the stratified sediments are glacial, interglacial, or pre-glacial in origin, and the nature of the acoustically incoherent sediments.
8. Study of industry seismic reflection data will help, particularly in the Bank areas covered with hard surface sands and gravels.

8. Lithological composition and sediment budget analysis of sediments found in deeper water off the shelf to determine if meltwater eroded the eastern Scotian Shelf channels in a single event.

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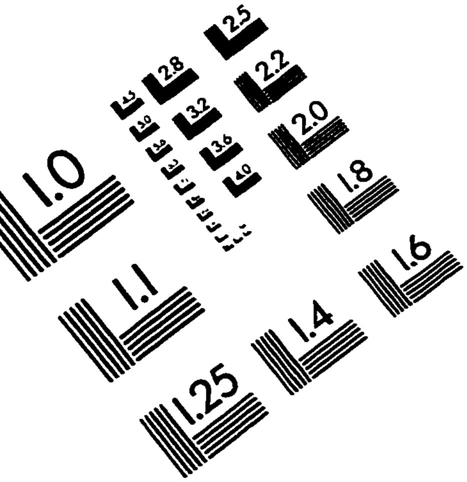
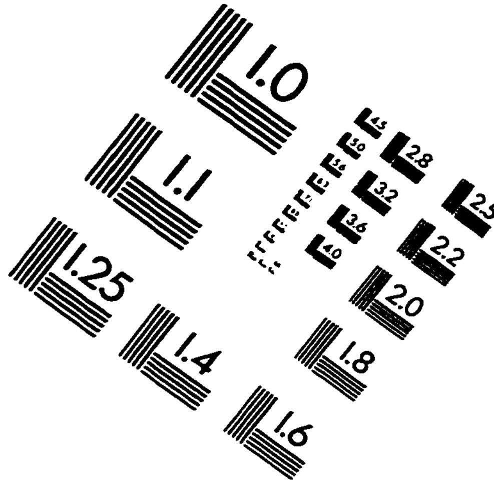
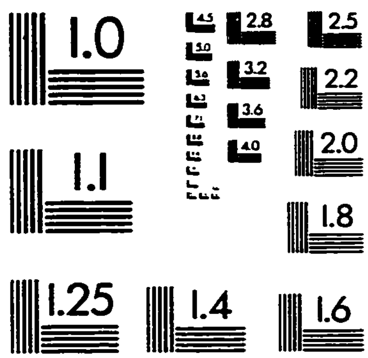
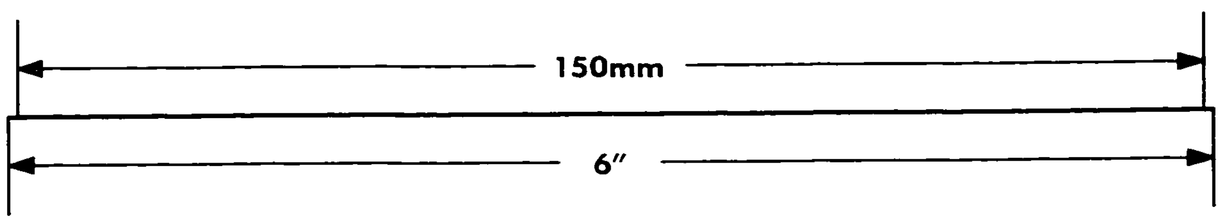
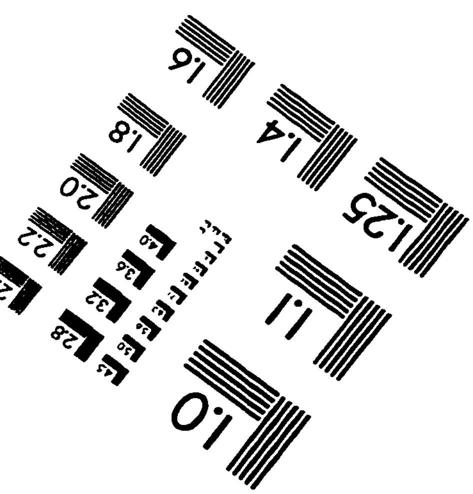
# IMAGE EVALUATION TEST TARGET (QA-3)

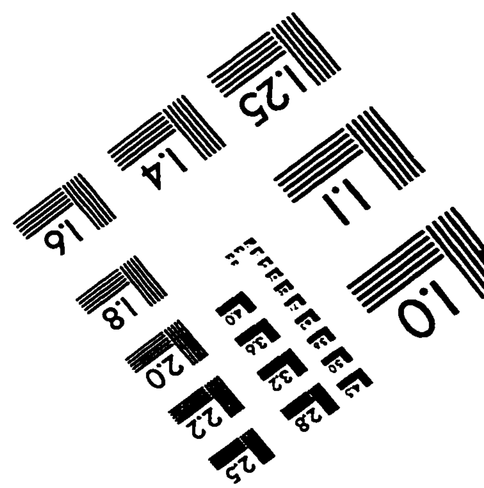








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