Plio-Pleistocene slope construction off western Nova Scotia, Canada

E. UCHUPI & S. A. SWIFT

Woods Hole Oceanographic Institution, Woods Hole, Mass. 02543, U.S.A.

ABSTRACT

Atlantic continental slopes have an average gradient of 7° and tend to be irregularly incised by submarine canyons with reliefs of about 1000 m. In contrast the slope off western Nova Scotia (from 61°30'W to 65°W), Canada, is smooth (relief <300 m) and has declivities ranging from 2.6° off Emerald Bank to 1.6° off LaHave Bank. This lack of deep relief is due to a combination of low subsidence and sedimentation rates during Late Cretaceous and Cenozoic. As a result of this low gradient, canyons that formed on the slope during periods of low sea level and high sediment supply rarely survived, being filled during subsequent regressions when sediment supply was less. High sedimentation rates by subglacial sediment plumes during late Wisconsin did lead, however, to extensive sediment failure that produced low relief slump and debris flow structures.

Key words: Continental slope morphology/construction, Nova Scotian Margin, Late Wisconsin glaciation, sediment plumes, ice shelf model, grounded ice model.

RESUMEN

Los taludes atlánticos tienen una pendiente media de 7° y presentan cañones submarinos con relieves de unos 1000 m que los cortan irregularmente. En contraste el talud frente a Nueva Escocia occidental (de 61° 30'W a 65°W),

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Canada, es suave (el relieve no supera los 300 m) con inclinaciones que oscilan entre 2.6° frente al Emerald Bank y 1.6° frente al LaHave Bank. La ausencia de relieves acusados se debe a la combinación de baja subsidencia y baja velocidad de sedimentación durante el Cretácico Superior y el Cenozoico. A causa de ese gradiente tan pequeño la supervivencia de los cañones que se formaron en periodos de nivel del mar bajo y elevado aporte de sedimento era muy escasa, pues se rellenaban durante las transgresiones subsiguientes cuando el aporte sedimentario era menor. Las elevadas tasas de sedimentación debidas a plumas (chorros) de sedimento subglaciares durante el Winsconsin Superior condujo, sin embargo, a desplomes sedimentarios generalizados que produjeron estructuras de desplome (*slump*) de poco relieve y debritas.

Palabras clave: Morfologia/construcción talud continental, Margen de Nueva Escocia, glaciación Wisconsin Superior, plumas de sedimento, modelo de plataforma de hielo, modelo de hielo encallado.

INTRODUCTION

The continental margin off Nova Scotia, Canada, is constructed of Mesozoic-Cenozoic strata deposited during the formation of the present North Atlantic. Emplacement of these sediments was controlled by changes in the configuration of the oceanic basin with time, variations in sea level (tectonic and eustatically induced), carbonate compensation depth, oceanographic circulation and climate. Plastic flow of Late Triassic-Early Jurassic evaporites from near the base of the section and intrusion into the overlying sediments created a zone of subsurface diapirs and ridges beneath the lower continental slope and upper continental rise between 2,000 and 3,000 m water depth (Fig. 1; Jansa & Wade, 1975; Holser *et al.*, 1988). Intrusion and deformation was probably episodic with events extending from the Mesozoic to the Pleistocene. With the exception of a few minor topographic irregularities, this flow has had little effect on the margin's surface morphology. Pronounced changes in sea level induced by the waxing and waning of continental glaciers since the Eocene led to the development of irregular morphology on the continental shelf and slope off Nova Scotia.

During late Pleistocene the continental margins off northeastern North America were dominated by fluvioglacial and glaciomarine processes that led to extensive erosion/deposition on the continental margin. Geologists working on deep-sea fans emplaced in non-glaciated areas rarely appreciate the extent of these glacial/fluvial processes in the construction of present margins. On shelves, direct deposition by ice was limited. Most of the detritus was emplaced by fluvio-glacial processes, an observation needed to be emphasized as there is tendency among some geologists to give the greater role to direct

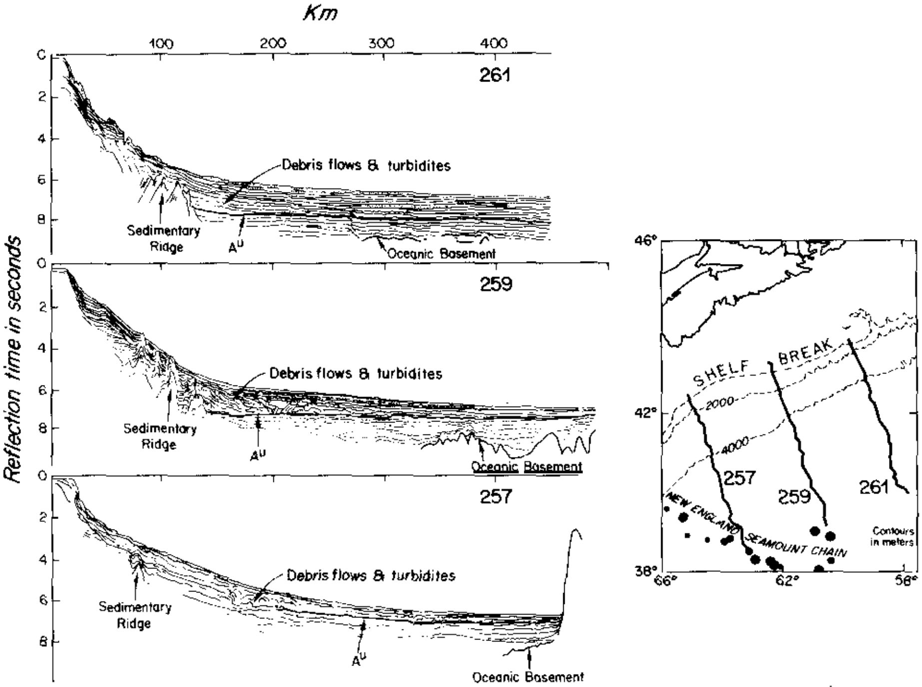


Fig. 1.—Single-channel seismic reflection profiles of the continental margin off Nova Scotia, Canada. Compiled from Emery *et al.* (1970). Note sedimentary ridge (diapirs) near the base of the continental slope formed by the plastic flow of Mesozoic evaporites and the lack of relief on the continental slope along profiles 257 and 259. The chaotic unit above Horizon Au (a Paleogene hiatus eroded by bottom currents) seaward of the sedimentary ridge is a deep-sea fan. See Fig. 4

Fig. 1.—Perfiles sísmicos de reflexión monocanal del talud continental frente a Nueva Escocia, Canada. Elaborado d partir de datos de Emery *et al.* (1970). Obsérvense la cresta sedimentaria (diapiros) cerca de la base del talud continental formadas por flujos plásticos de evaporitas Mesozoicas y la falta de relieve en el talud continental en los perfiles 257 y 259. La unidad caótica sobre el horizonte Au (una discontinuidad paleógena erosionada por corrientes de fondo) hacia el mar de la cresta sedimentaria es un abanico submarino profundo. Véase la Fig. 4.

glacial action. Below we briefly describe the late Pleistocene glacial history of the Canadian Nova Scotian margin to illustrate the influence this glacial event had on the construction of the continental slope off Nova Scotia, Canada.

TOPOGRAPHY

Like all formerly glaciated shelves, the Nova Scotian margin from the Laurentian Channel on the northeast to Northeast Channel on the southwest displays irregular morphology (Fig. 2). Non-glaciated Atlantic shelves are

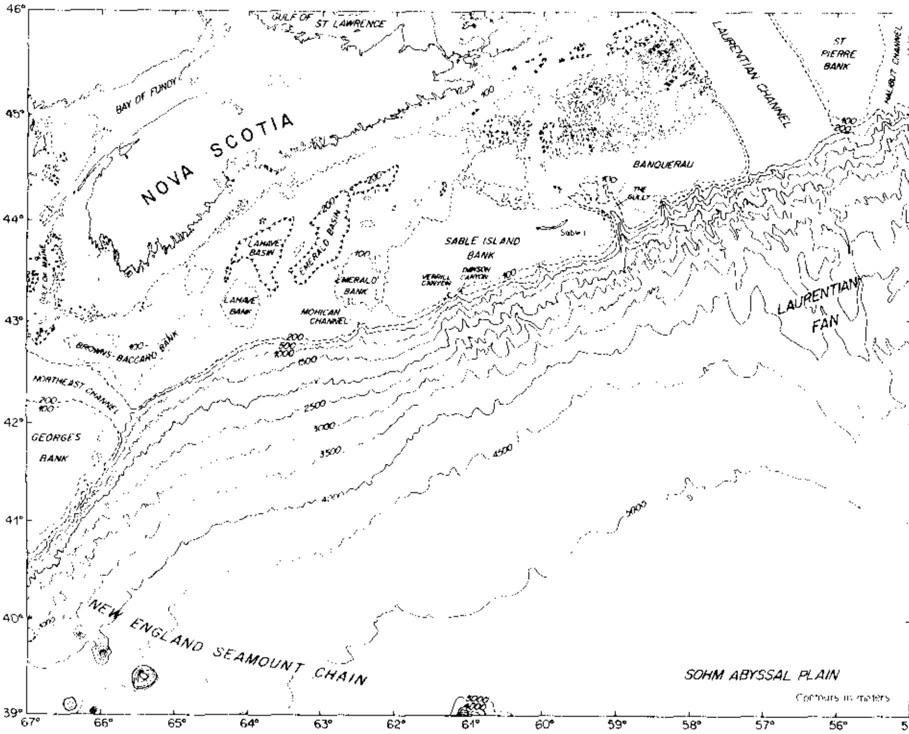


Fig. 2.—Bathymetry of the Nova Scotian continental margin. Compiled from Shor (1984a, 1984b). Note lack of relief on the continental slope from 61° 30'W to 65° W.

Fig. 2.— Batimetría del margen continental de Nueva Escocia. Recopilado de Shor (1984a, 1984b). Obsérvese la falta de relieve en el talud continental desde 61° 30'W a 65° W.

relatively smooth and dip gently seaward to a depth of about 120 m where a sharp increase in gradient leads to the continental slope (Emery & Uchupi, 1984, p. 46). In contrast, the Nova Scotia shelf is irregular and deeper (Uchupi, 1969; King, 1970; King & MacLean, 1976). Immediately offshore is a narrow zone less than 120 m deep where pre-Mesozoic rocks are exposed. On the mid-shelf is a series of depressions with water depths of 150-290 m. On the eastern end of the shelf the mid-shelf lows consist of a mosaic of narrow north-south and east-west trending highs and depressions terminating abruptly against the outer banks and the western edge of Laurentian Channel. The head of the Gully extends across the outer shelf to this irregular mid-shelf zone via a 25-50 km wide trough separating Banquereau and Sable Island banks (Fig. 2). The mid-shelf depressions on the western half of the shelf (Emerald and LaHave basins) trend southwesterly to the shelf's edge via a saddle between Emerald and LaHave banks. Along the outer shelf is a chain of banks less than 100 m deep separated from one another by U-shaped troughs and oriented either parallel or perpendicular to the shelf edge. Shelf edge depths beyond the outer banks range from about 100 m off Banquereau and Sable Island banks to 180 m south of Emerald and LaHave basins and to 140 m at the southwest end of the shelf (Stow, 1977).

The continental slope beyond the shelf edge displays marked changes in topography along strike. East of 61°30'W and west of 65° W the slope descends with a gradient of 5°-7°from a water depth of ~100 to 180 m at the shelf's edge to the upper rise at 2,500 to 3,000 m. The slope east of 61°30'W and west of 65° W is deeply entrenched by submarine canyons, with relief exceeding 1,000 m, that lead to deep-sea channels on the continental rise. This topography is typical of Atlantic Ocean continental margins. The morphology of the continental slope segment extending from 61°30'W to about 65°W is quite distinct from other Atlantic slopes in that it lacks prominent irregularities. In a basinward direction, the slope is made up of two segments. For less than 3 km beyond the shelf edge, the upper portion extends to a depth of 800 m with a gradient of 5° and is relatively smooth. The lower portion, which lacks appreciable irregularities, descends over 60-80 km with a gradient of 2° to a depth of 2,500 to 3,000 m where it merges with the continental rise.

Characteristically, both portions of the smooth slope from 64°W to 61° 30'W have maximum relief of only 150-200 m (Swift, 1985; Piper & Sparkes, 1987). Irregularities on the smooth segment from 65° W to 64° W have greater relief averaging about 200 m (Uchupi, 1969). Small scale relief includes eroded valleys (1-4 km wide and up to 100 m deep; Piper, Farre & Shor, 1985), acoustically stratified sediment blocks (35-150 m in relief and typically elongate downslope; Hill 1983; Swift, 1985; Piper & Sparkes, 1987), erosional scarps oriented downslope (relief typically of 30-150 but ranging up to 225 m; Swift, 1985, 1987; Hill, 1983; Mosher *et al.*, 1989), and sediment slides (Swift, 1985; Piper, Farre & Shor, 1985; Piper, Farre & Shor, 1989; Mosher *et al.*, 1989). These slides appear to converge down slope terminating at a depth of 4,800 to 5,000 m (near contact between the continental rise and the Sohm Abyssal Plain) in a series of discrete lobate deposits (Hughes-Clarke, O'Leary & Piper, *in press*). Sea MARC I deep-towed sidescan sonar recordings from the eastern edge of the smooth slope segment display slides and slide scars, that give the seafloor a step like morphology, and debris flows at the distal ends of the slides. Piper, Farre & Shor (1985) ascribed these sediment failures to a large earthquake that occurred between 12 and 5 ka. A survey using the GLORIA sidescan sonar imaging system revealed channels (200 to 1,000 m wide, generally less than 20 m deep and 5 to 100 km long) intricately connected over wide areas in the smooth segment. They appear to converge in water depths greater than 4,500 m on a few larger channels with reliefs of up to 50 m and widths of 3 to 5 km (Hughes-Clarke, O'Leary & Piper, *in press*). Other positive topographic features revealed by the GLORIA sidescan sonar on the slope include contour parallel and transverse ridges associated with slides. On the continental rise relief rarely exceeds 30-40 m.

Low-relief, braided channels and debris flows are ubiquitous west of about 61° W, whereas sediment waves and deep-sea channels are common to the east (Shor & Lonsdale, 1981; Swift, 1985; Hughes-Clarke, O'Leary & Piper, *in press*).

PRE-QUATERNARY DEVELOPMENT

Off Nova Scotia, sedimentation patterns during glacial epochs were influenced by pre-existing shelf and slope morphology. This is well-established on the shelf, where King, McLean & Fader (1974) showed that Tertiary fluvial erosion patterns of shelf basins and banks were inherited. These patterns directed ice flow basinward during the Quaternary. On the continental slope, Cenozoic morphological changes are less well known. We suggest, however, that they were no less influential.

Structurally, the outer Scotian shelf is divided near 61° W into the subsiding Sable and Abenaki Basins to the east and the relatively stable La Have Platform to the west (Jansa & Wade, 1975). The location of Mesozoic carbonate platform development followed the crustal hinge and, thus, is farther seaward off the western shelf than in the east. These carbonate buildups were best developed on stable basement highs, such as the outer horsts of the La Have Platform, where low rates of subsidence and siliclastic sedimentation provided the most suitable environment (Eliuk, 1978; Jansa, 1981). In the Abenaki and Sable Basins, carbonate deposition during the Mid-Late Jurassic was intermittent and diluted by deltaic siliclastics. This Sable Island Delta prograded southward covering the carbonate platform and side escarp-ments beneath a thick regressive wedge by the Early Cretaceous (Given, 1977). On the western shelf, the carbonate buildup on the La Have Platform was «drowned» during the Neocomian by rising relative sea level (Eliuk, 1977). Early Cretaceous deposition was much slower on the LaHave Platform than in the basins to the east, so parts of the escarpment fronting the carbonate platform that existed seaward of the shelf's edge remained exposed into the Late Cretaceous long after the demise of the platform (Fig. 8 in Jansa, 1981; Fig. 7 in Swift, 1987). By the Late Cretaceous, crustal down-warping of the outer shelf basinward (seaward) had increased water depth above the paleo-carbonate platform such that an upper slope region with gradient intermediate between the prograding shelf wedge and the carbonate escarpment developed. By mid-late Tertiary, upper rise deposits lapped onto early Tertiary shelf deposits landward of the carbonate platform burying the escarpment and creating a smooth, low-gradient slope (Swift, 1987). Despite rapid Pliocene-Pleistocene progradation of the shelf, the present shelf edge west of 61°31W still remains landward or above the carbonate escarpment, and the slope deepens at only 1.5°-3° rather than at 5°-7° to the east.

Since the Late Cretaceous, canyons formed episodically on the slope and incised the shelf edge (Swift, 1986, 1987). The history of these canyons west of 61°-30° W differs from those elsewhere in the Atlantic. All except one, the Mohican Channel, were filled by subsequent shelf progradation. Reflection profiles and borehole data indicate that Mohican channel, part of the Schubenacadie drainage system, is partially filled and has a vestigial morphology only because of its enormous original size (Given, 1977; Swift, 1987). On the shelf this canyon fill is mainly Neogene, whereas on the shelf's edge and slope it is Pleistocene. D.J.W. Piper (personal communication) also believes that the canyon fill on the smooth slope is Pleistocene.

GLACIAL CHRONOLOGY

Despite over a century of detailed study, the Quaternary glacial chronology for the Canadian Maritime provinces is incomplete and still lacks agreement with the marine record in essential ways. On Anticosti Island, in the Gulf of St. Lawrence, three tills have been recognized, an early Wisconsin (85,000 B.P. based on acid racemization of shell fragments in overlying marine sediments), a middle and a late Wisconsin (St-Pierre, Gwyn & Dubois, 1987). According to St-Pierre, Gwyn & Dubois, after 36,000 B.P. the ice front in the Gulf of St. Lawrence was located along the south coast of Anticosti Island. There is poorly dated evidence from Nova Scotia (Grant, 1976, 1977, 1989) and southeast Massachusetts (Oldale & Eskenasy, 1983) for a pre-Wisconsin glaciation (Illinoian; ca. 150 ka or older). Although exposures are too sparse to infer any details about the nature of this event, Grant (1989) used these deposits to infer that Wisconsin glaciations that followed were weaker or more limited in extent. In Nova Scotia, a sequence of organic rich beds overlying the Illinoian deposits record an interglacial epoch correlated to the Sangamonian (oxygen isotope stage 5; 125- 75 ka). Palynological evidence suggests environmental conditions were similar to the present interglacial. Grant (1989) infers two major glacial pulses for the Wisconsin. The first, more extensive, Laurentide ice event penetrated southward across Nova Scotia and, synchronously, out the Laurentian Channel in early Wisconsin (ca. 75-60 ka; oxygen isotope stage 4). The lack of evidence for glaciation in some highlands and absence of post-glacial marine deposits indicate thinner or more restricted glacial ice than in regions closer to Laurentia (eg. New England). Based on the interpretation and chronology of King & Fader (1986), Grant (1989) suggests that grounded ice extended across the Scotian shelf basins and outer shelf banks depositing the Scotian Shelf Drift, a till underlying most of the shelf. During the middle Wisconsin, younger tills on land indicate that the ice sheet reorganized into smaller, isolated domes on highlands and, possibly, offshore. The last ice advance phase occurred in late Wisconsin (ca. 25 ka). However, the extent, source, and direction of this last ice advance are disputed between a «minimal» ice event with local domes on highlands and ice margins near present shorelines and a « maximal» model with a single southward ice advance across Nova Scotia to the shelf edge (Grant, 1989). We tend to favor the «maximal» model and our discussion below is based on this premise.

King & Fader (1986) present a glacial scenario for the Scotian shelf based on Quaternary marine stratigraphy. Their model indicates that grounded ice extended across the Scotian shelf basins and outer banks depositing the Scotian Shelf Drift, an overconsolidated till. Based on total-carbon C^{14} age dates, they assign an early Wisconsin age (70-45 ka) to this advance. Between 50-32 ka, grounded ice retreated to the inner shelf and the outer banks while an ice shelf covered the middle shelf basins. Final ice retreat began around 32 ka opening the inner shelf. However, the chronology by King & Fader based on C^{14} dates on total organic matter is systematically too old due to the inclusion of reworked material. Gipp & Piper (1989) dated un-reworked mollusc shells from the Emerald Silt above the till in Emerald Basin with accelerator mass spectrometry (AMS) and estimated that the top of the Scotian Shelf Drift as 18 ka rather than 35 ka. Similar measurements from LaHave Basin yielded a date of < 18.5 ka for the top of the drift and 17.5 ka for the Emerald Silt immediately above the drift (Piper, Gipp & Moran, 1990). Such ages are consistent with a late Wisconsin glaciation, which deposited the Nova Scotian Shelf Drift, and an ice retreat from the outer shelf which began about 18 ka. Amos & Knoll (1987) found evidence on the subsurface of Banquereau for only one Wisconsin ice advance at 20-26 ka, and the chronology of Mosher *et al.* (1989) for the last glacial advance in Verrill Canyon is consistent with the Banquereau and Emerald Silt dates. Amos & Miller (1990) have reported the presence of pro-glacial, marine sediments beneath Sable Island Bank that they believed were deposited from an ice shelf prior to 35 ka, and ice proximal deposits dated as 28-15 ka. The older dates, however, are based on C^{14} determinations on molluscs that show evidence of reworking casting doubts on their accuracy.

Glacial episodes on the continental slope have been inferred primarily on the basis of seismic stratigraphy. Between 61° W and 64° W, Swift (1986, 1987) correlated two seismic unconformities formed during minor canyon-cutting events to early Pliocene and late Wisconsin ice advances. In the same region, Piper, Normark & Sparkes (1987) correlated a seismic unconformity, marking an increase in sediment accumulation rate, the beginning of a cut-and-fill sequence, and the first erosion of the Verrill-Dawson canyon system, to a low sea level stand in the late Pliocene. They also dated, by correlation to industry boreholes, two overlying erosional episodes associated with low sea level to the mid-Late Pliocene and the Plio-Pleistocene boundary. Piper, Normark & Sparkes (1987) suggest that the deepest of a sequence of «till tongues», acoustically-incoherent wedges thinning basinward from the shelf edge, was formed during the mid-Quaternary (~800 ka). They interpret these units as evidence for advance of grounded ice to the shelf edge. Piper (1988) reported that sequences of up to five «till tongues» occur along the Scotian margin, whereas Mosher *et al.* (1989) found three (dated to early and late Wisconsin and the Sangamonian) near 61° W. Farther east on the continental slope off St. Pierre Bank (Fig. 2), a poorly-sorted, silty diamict, extending from the shelf edge to ~600 m water depth, is overlain by 2 to 5 m of stratified sediments containing a benthic foraminiferal fauna dominated by *Elphidium excavatum* and *Cassidulina reniforme* which Bonifay & Piper (1988) interpret as indicating ice-margin deposition. Radiocarbon determinations by AMS yielded ages of 3.3 and 11.8 ka for the stratified unit which implies an age of more than 11.8 ka for the diamict unit. Bonifay & Piper (1988) proposed that the diamict is a slumped morainal facies and proglacial sediments resulting from a late Wisconsin ice surge through Halibut Channel.

QUATERNARY LITHOFACIES

On the Scotian shelf, a surficial Pleistocene-Holocene succession, comprising five formations 10 m thick on the inner shelf and up to 200 m on the outer shelf, rests unconformably on pre-Pleistocene and possible older Pleistocene strata. At the base of the succession is the Scotian Shelf Drift, an over-consolidated, poorly sorted deposit containing appreciable amounts of gravel regarded as a till deposit (King & MacLean, 1976; King, 1980). Where exposed its surface is irregular in part due to relict iceberg furrowing. Interbedded with and overlaying the till is the Emerald Silt, a fine-grained, well-bedded silt to sand, locally gravelly, proglacial formation deposited under brackish to normal marine conditions. The *Elphidium* fauna associated with the silt probably reflects turbid glacial meltwater conditions from 20,000 to 10,000 years B.P. (Scott *et al.*, 1984).

Using the Antarctic ice shelf as a present-day analog, King & Fader (1986) proposed that the Emerald Silt was deposited beneath a floating ice shelf. Oldale, Williams &Coleman (1990) rejected the ice-shelf model because of the formation's wide extent, its well bedded characteristics, scarcity of coarse clasts, and incompatibility of an Antarctic-type ice shelf with the temperate climate that existed at the end of the Wisconsin at the latitudes of Nova Scotia. They proposed instead that the silt was the product of glacial meltwater that was discharged directly into the sea via subglacial streams or indirectly via subaerial streams.

In the center of the central shelf depressions the Emerald Silt is capped by the LaHave Clay, a Holocene deposit winnowed out of the sediments from adjacent banks and land areas. Coeval to the clay are the Sambro Sand, a sub-littoral deposit now found at a depth of 115 m reworked from the Scotian Shelf Drift and Emerald Silt. The Sable Sand and Gravel forms a basal transgressive unit atop eroded remnants of the Scotian Shelf Drift and Emerald Silt, and confined to the outer banks and inshore areas in water depths of less than 115 m (King, 1980).

The outer margins of the banks along the shelf's edge represent local accumulations of proglacial sediment. On Banquereau, at the northeast end of the Nova Scotian Shelf, they consist of nearly 100 m of a gravel-rich unit barren of fauna occurring within stratified proglacial Emerald Silt deposited during the late Wisconsin 26,000 to 18,000 B.P. (Amos & Knoll, 1987). At the top of the section is a sequence of channel deposits that are coeval with a eustatic low sea level stand 110 m below the present level ca. 14,000 B.P. The marine transgression that took place 8,000 B.P. is marked by a widespread erosional surface above which is as much as 40 m of the Sable Island Sand and Gravel. Data from three boreholes, ten vibrocores, and seismic reflection profiles indicate that the stratigraphy of the topmost 50 m of Sable Island Bank is comparable to the 100 m thick late Pleistocene section on Banquereau to the east: the Emerald Silt at the base consists of four members, a lower stratified sand, barren gravelly sand, an upper channel gravel, and stratified sands (Amos & Miller, 1990). The Sable Island Sand and Gravel above the Emerald Silt is made up of a lower bedded sand emplaced under heavy sea ice, cross-bedded sand, and a trough-bedded sand. High resolution seismic reflection profiles from the northern edge of Sable Island Bank reveal the presence of a sub-surface channel network extending over 450 m below the present sea level. Boyd, Scott & Douma (1988) proposed that the network represents sub-ice tunnel valleys.

During the Plio-Quaternary, thick, prograding sediment wedges were deposited at the shelf edge along the entire Scotian margin (Jansa & Wade, 1975). While the outer shelf banks and gullies were extended seaward, erosion incised canyons into the slope to the east of 61°30'W and to the west of 65° W. In Verrill and Dawson canyons along the eastern edge of the smooth slope segment, canyon cutting can be documented in the earliest late Pliocene and late Pliocene (Piper, Normark & Sparkes, 1987). Canyon cutting by turbidity currents continued during the Pleistocene causing sediment to bypass the slope and sedimentation rates to diminish to less than 0.1 m/1,000 years (Piper, Normark & Sparkes, 1987). Most coarse sediments reaching the shelf edge were transported to the continental rise.

On the smooth slope segment, post mid-Pliocene sediments are over 1,000 m thick near the shelf edge (Jansa & Wade, 1975; Austin *et al.*, 1982; Swift, 1987). At the base are Pliocene and older Pleistocene deposits and above them is a late Pleistocene sequence of sand and mud turbidites, gravelly sandy mud debris flows, ice rafted debris, and hemipelagic sediments (Hill, Aksu & Piper, 1982; Hill, 1984) which accumulated at rates of about 1 m/1,000 years (Mosher *et al.*, 1989). The topmost part of the section consists of 1 to 2 m of Holocene hemipelagic sediment deposited during the last 13,000 years at a rate 0.05 m/ 1,000 years (Mosher *et al.*, 1989). The late Pleistocene sediments display slump structures and closely spaced gullies, instability features related to the high rates of glacial deposition (Piper & Sparkes, 1987). West of the Mohican channel near 62° W (Fig. 2), these failures produced a 150 m high scarp on the upper continental slope at the base of which is a massive debris flow (Hill, 1983) surrounding autochthonous sediment blocks up to 50 m thick (Piper & Sparkes, 1987). The deposits rest unconformably on and preferentially filled the canyons eroded prior and during the last glaciation producing local sediment accumulations in excess of 1,000 m with Mohican Canyon containing 1,410 m of Plio-Pleistocene fill (Hill 1983; Swift 1986, 1987; Piper, Normark & Sparkes, 1987). Further landward, mid-Tertiary reflectors, dated at shelf boreholes, can be traced across the buried expressions of the canyon cut-and-fill (Swift, 1986) indicating that the canyon extensions on the shelf were filled earlier than those on the slope. On the slope, canyons cut into Plio-Pleistocene deposits are being filled now or were buried during the Pleistocene (Hill, 1983; Swift, 1987; Piper, Normark & Sparkes, 1987). This illustrates the origins of the differing slope morphologies. In the late Neogene, the supply of sediment to the shelf edge episodically increased enough to bury the canyons on the shelf and maintain the canyons on the slope, much as elsewhere in the Atlantic. Unlike elsewhere, however, the canyons between 61°30'W and 65° W were subsequently filled during the Pleistocene.

Seaward of the diapiric sedimentary ridge province, the seismic unit immediately above Horizon Au, a Paleogene unconformity, displays hummocky reflectors (Fig. 1). This unit has a fan-like distribution and maximum thickness of 1.2 km (Fig. 3; Swift, 1987). We believe the unit was emplaced by turbidity currents, whereas Ebinger & Tucholke (1988) concluded that these deposits underwent reworking by bottom currents after deposition. The source of these sediments was the offshore diapirs which were uplifted during the Miocene (Swift, 1987). Resting on this unit is a Pliocene-Pleistocene sequence 300-400 m thick displaying a seismic signature analogous to olistostromes. Sediment failure of the youngest unit, which was reworked by bottom currents during the Pleistocene, also resulted from the rapid uplift of the sedimentary diapiric ridge (Swift, 1987). Growth of the diapirs in the sedimentary ridge pro-vince at both times resulted from sediment loading that caused the Mesozoic evaporites to flow. This flow in turn led to subsidence landward of the ridge enhancing deposition on the slope.

GEOLOGIC SCENARIO FOR SLOPE DEVELOPMENT

The present morphology of the Nova Scotia margin is the result of glacial-marine processes superposed on a topography molded by preglacial fluvial and marine processes. From the Eocene to the Pleistocene, the shelf was subjected to subaerial fluvial erosion during the regressions and marine deposition during the transgressions. The continental slope was a site of submarine erosion during the regressions and marine deposition and possibly

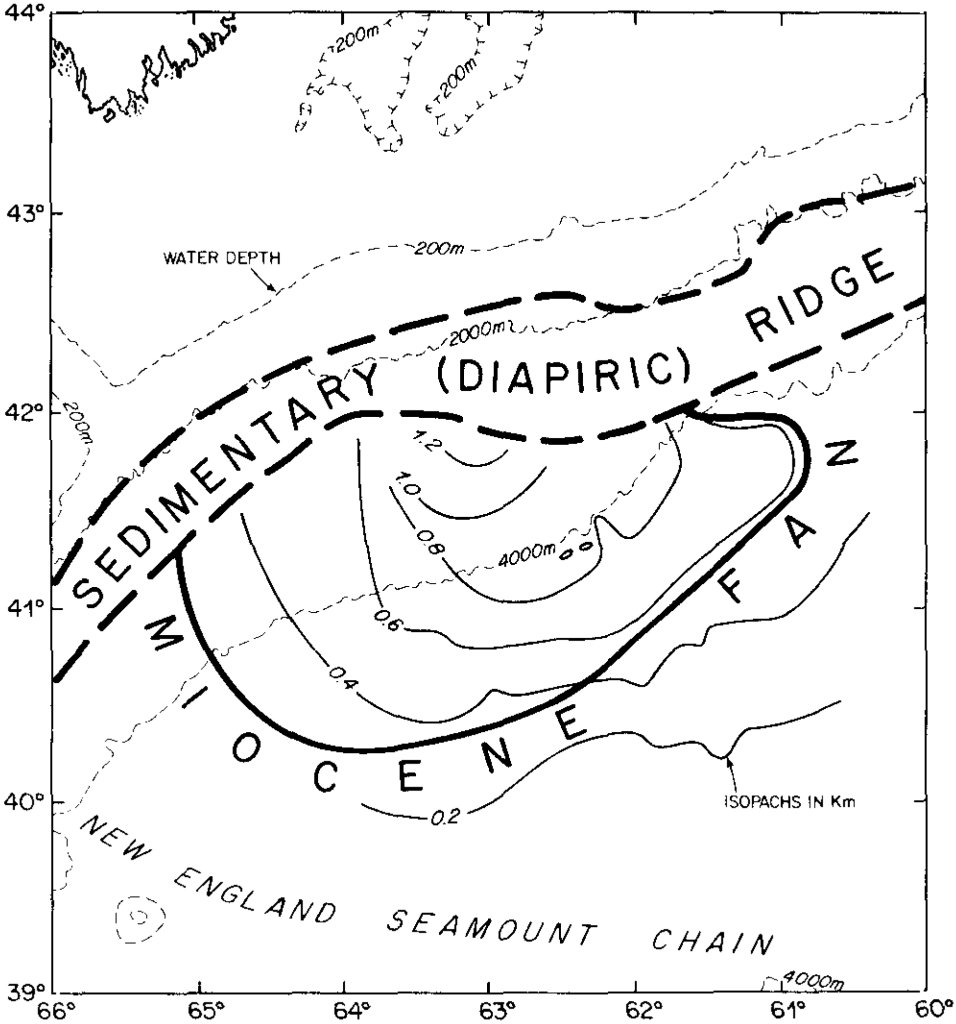


Fig. 3.—Isopach map of a Miocene submarine fan south of diapiric ridge. This fan was constructed from sediments eroded from diapirs uplifted at that time. Modified from Swift (1987).

Fig. 3.—Mapa de isopacas de un abanico submarino Mioceno al sur de la cresta diapírica. El abanico se construyó con sedimentos erosionados de diapiros que estaban elevados en esa época. Modificado de Swift (1987).

non-deposition during the transgressions. These regressive/transgressive (erosional/depositional) cycles were the result of waxing and wanning of continental ice sheets. During the late Pleistocene, the cut and fill terrane on the shelf formed during the Cenozoic by fluvial and marine processes was modified by glacial action. Although there is evidence that the northeastern North American shelf was glaciated at least once prior to 50 ka, the effect of pre-Wisconsin glaciations on the morphology of the shelf appears to have been minor or effectively masked and modified by the late Wisconsin glacial event.

The late Wisconsin glaciation probably reached its maximum extent and volume ~26,000-20,000 years ago. As a result of this ice advance, eustatic sea level dropped to 121-126 m below its present level (Fairbanks, 1989). We postulate that the southern limit of the late Wisconsin Laurentian Ice Sheet was along the shelf's edge adjacent to the smooth slope segment from 63°30'W to 62°30'W, at the mouths of Northeast and Laurential channels, the Gully and the smaller channels between the shelf edge banks, and along the northern edges of Georges, Browns-Baccaro Sable Island, and Banquereau banks. These banks were above sea-level at that time (Fig. 4). Retreat of the ice front probably began soon after the Laurentian Ice Sheet reached its southern limit, a retreat marked by extreme melting events and glacial surges of short duration. As the base of the grounded ice began to melt the drift found on much of the Scotian shelf was deposited subaerially. Extensive glacial melting and runoff lead to deposition of glaciofluvial sediments on the banks, while glaciomarine sediments were deposited on the outer edges of the banks and on the smooth slope segment. The volume of freshwater released in meltwater and calving of icebergs must have been quite large. Jones & Ruddiman (1982), for example, have calculated that the volume of such freshwater during the Pleistocene may have been 1.5-2.0 times the discharge of the Amazon River. No ice shelf will develop in an environment characterized by such melting rates. For a present-day analog to the late Wisconsin glacial front on the Nova Scotia margin, the grounded ice margin in the northern Barents Sea (Pfirman, 1985) is more realistic than the ice shelves off much drier Antarctica (Oldale, Williams & Colman, 1990). As the ice retreated northward away from the outer shelf, glaciomarine sediments were deposited on the mid-shelf depressions. Glacial ice retreat patterns, determined from AMS dates, indicate that the ice edge was near the Nova Scotian coast 14,500 B.P. (Stea & Mott, 1989), the Bay of Fundy was virtually ice free 14,000 B.P (Stea & Wightman, 1987), and deglaciation of Nova Scotia may have been nearly complete by 11,000 B.P. Freshwater discharges during the ice retreat must have affected not only the local sediment regime but also climatic conditions. Lehman *et al.* (*in press*) believe that such a freshwater discharge about 15,000 B.P. from the Fennoscandian Ice Sheet reached the northern North Atlantic and may have contributed to extreme oceanic cooling similar to that suggested for the Younger Dryas. This general retreat was interrupted about 11,000 B.P. According to Mott *et al.* (1986) and Stea & Mott (1989) there is evidence in eastern Canada for a local cool period and a local glacial readvance between 11 and 10 ka, coeval with the Young Dryas cooling in northern Europe. Based on these suppositions we reconstruct the Wisconsin glacial history of the region as follows.

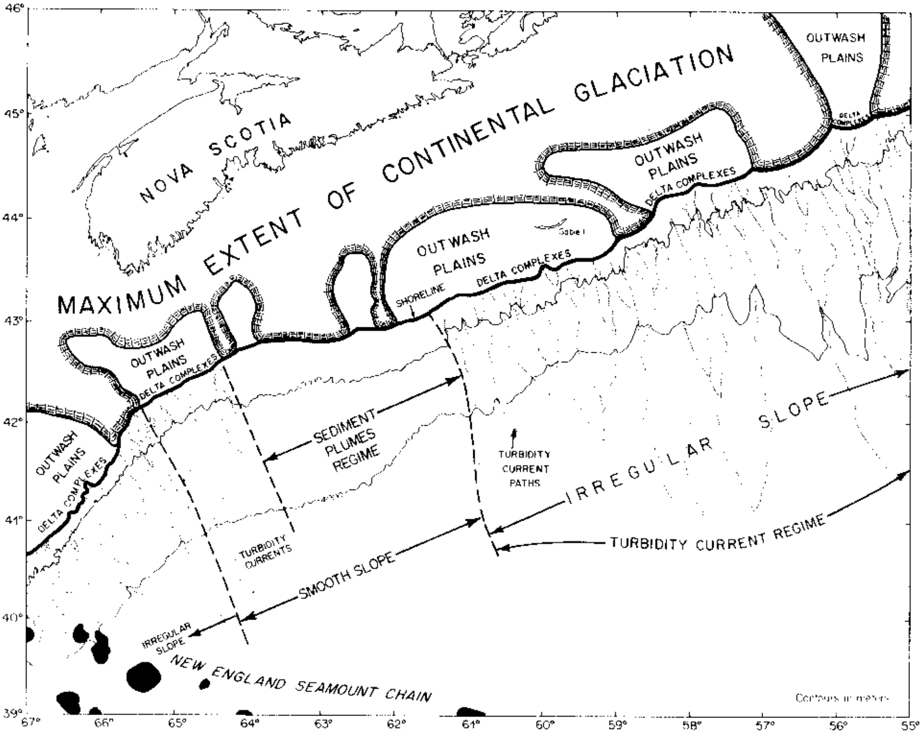


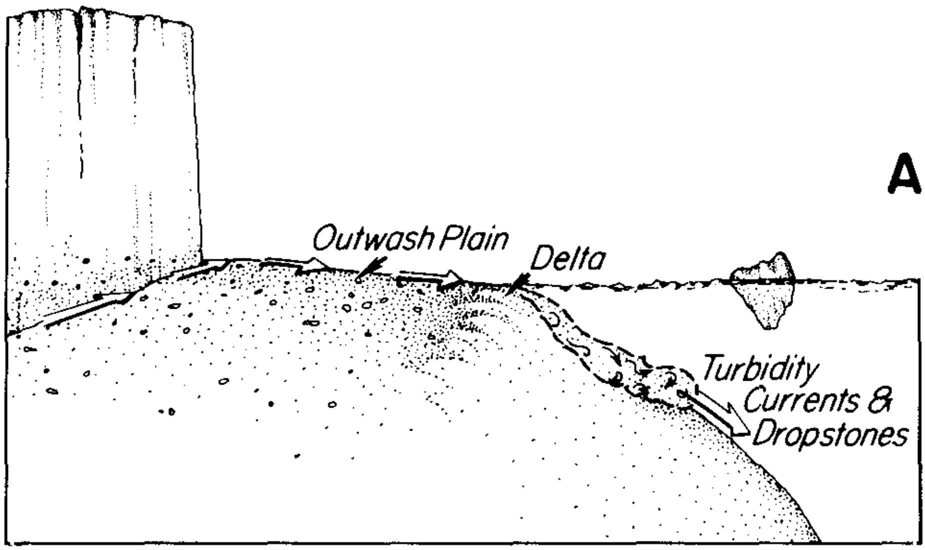
Fig. 4.—Paleogeography of the Nova Scotia margin during late Pleistocene (late Wisconsin) approximately 23,000 years ago. Note that along the eastern and western ends of the Nova Scotian shelf the glacier is fronted by an outwash plain/delta system, whereas in the center the ice extends to the marine realm.

Fig. 4.—Paleogeografía del margen de Nueva Escocia durante el Pleistoceno Superior (Wisconsin Superior) hace unos 23.000 años. Nótese que el glaciar terminaba en sistemas de llanura aluvial/ delta en los extremos este y oeste de la plataforma de Nueva Escocia mientras que en el centro el hielo se extendía hasta el dominio marino.

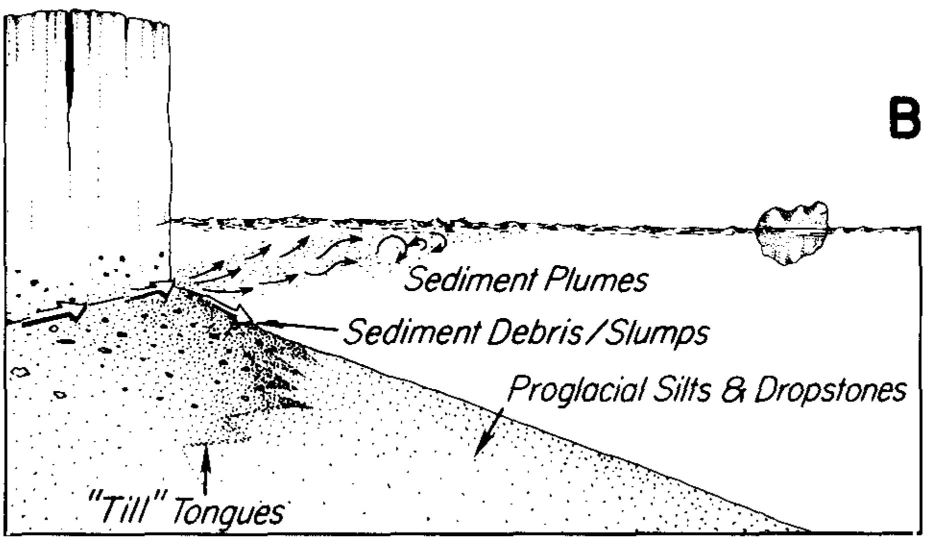
As a result of pre-Wisconsin regressions and possibly pre-Wisconsin glaciation (Moshner, *et al.*, 1989) the Nova Scotian shelf surface was somewhat irregular and the continental slope was indented by a system of submarine canyons that served as passageways for sediment to the deep-sea. Prior to about 23,000 B.P. the Laurentian Ice Sheet reached the general vicinity of the Scotian shelf edge (Fig. 3). Laurentian and Northeast channels served as drainage paths for ice filling the gulfs of St. Lawrence and Maine, respectively. As the glacier extended across the Scotian Shelf, erosion by ice streams, the most energetic elements of large ice sheets (Drewry, 1983), modified the pre-existing topography. The seaward limit of the ice sheet is reflected by the morphology of the continental slope. At the eastern end of the shelf, where canyon erosion continued into the late Pleistocene, the ice margin was along the northern edges of Banquereau and Sable Island banks. The streams extended their courses across the banks which were above sea level at that time, and deposited their load along the southern edge of the banks along the contemporary shoreline located on the uppermost continental slope 121-126 m below the present sea level to form pro-glacial delta systems. This sediment input must have been massive leading to unstable conditions and downslope transport by turbidity currents that maintained the canyons inherited from pre-glacial times and cut new ones (Fig. 5A). These currents by-passed the slope and deposited their load on the continental rise and adjacent Sohm Abyssal Plain.

At the western end of the smooth slope segment from 64° W to 65° W the ice terminated along the northern edge of Browns-Baccaro Bank. Along the southern edge of this high were a series of pro-glacial deltas constructed by glacial streams that extended their courses across the bank to the open sea. Episodic sediment failure led to formation of slumps and slides and down slope sediment transport. In contrast to the slope seaward of Sable Island Bank and Banquereau whose steepness enhanced turbidity current activity, the gentler slope seaward of Browns-Baccaro Bank inhibits such high currents preserving the general smoothness of the slope.

In the smooth slope segment from 62°30'W and 63°30'W, the ice sheet advanced to the edge of the shelf (the contemporary shore) through the wide divide between Emerald and LaHave banks, and the narrower troughs between LaHave and Browns-Baccaro banks and between Emerald and Sable Island banks. The north-south trending Emerald and LaHave banks probably were formed in part by lateral deposition from these glacial lobes. Sediment reached the continental slope via sediment-laden meltwater plumes generated by subglacial streams debouching out of a grounded ice wall (Fig. 5B). This sediment-laden meltwater, which reached the open sea through tunnels at the base of the ice front, deposited their coarse load near the grounding line on the uppermost continental slope forming «till tongues» identified by Piper, Mormark & Sparkes (1987) and Mosher *et al.* (1989). Finer sediments were carried beyond the shelf edge into deeper water by near surface sediment plumes active along the ice front. According to Molnia (1983) such plumes may transport glaciomarine muds as much as 50 km offshore. Deposition on the smooth slope off Nova Scotia reached thicknesses of over 1000 m at rates of 1 m/1,000 years. This subaqueous sedimentation, however, was not responsible for the lack of relief on the slope segment between Sable Island and Browns Banks. If such processes were responsible, then the slope off the Laurentian and Northeast Channels, where the ice front also extended to the top of the slope, would have low reliefs. Yet both of these segments display well developed submarine canyons with the one off the Laurentian Channel having a massive submarine fan (Laurentian Fan, Fig. 2) at its base. Construction on the smooth segment continued because of its inherited low declivity, a gradient resulting from low subsidence rate and sediment supply that prolonged the exposure of the carbonate front during early



Delta-Turbidity Current Regime



Sediment Plume Regime

Cenozoic. Seismic reflection profiles oriented parallel to the contours on the outer shelf and upper slope show numerous examples of buried canyons (Austin *et al.*, 1982; Piper, Farre & Shor, 1985; Piper, Normark & Sparkes, 1987; Swift, 1985,1987). We suggest that sediments were able to fill these canyons because the overall inclination of the continental slope was too low to sustain canyon excavation. Although valleys were eroded locally at times of low sea level and high glacial sediment supply, extensive turbidity current activity could not be generated on the smooth slope over scales of 100,000 years to remove the fill from the canyons. What role the offshore diapiric structures played on the geohistory of the smooth slope is yet to be resolved. Some of the structures opposite the smooth slope were uplifted during the Miocene and Plio-Pleistocene becoming the major sediment contributor to the region south of the high. Possibly at the same time, as a result of these uplifts, the diapirs may have partially dammed the sediments behind them. Such an entrapment will tend to enhance deposition on the slope.

With time the glacial front retreated across the shelf and the continental slope became a site of pelagic deposition. As the ice retreated subglacial runoff deposited the Emerald Silt on the mid-shelf depressions atop the basal drift. As the ice retreated from the region and the Holocene sea transgressed across the Nova Scotian shelf, proglacial sediments were reworked from the offshore banks to form the Sable Island Sand and Gravel, the Sambro Sand, and the LaHave Clay.

Fig. 5—Schematic diagram showing nature of processes along an ice front near the outer continental shelf. (A) On the eastern and western ends of the shelf outer banks blocked the ice front and subglacial streams deposited their load on an outwash plain and delta. Massive sediment deposition along delta fronts on the crest of the steep continental slope on the astern shelf led to considerable instability, downslope failures, formation of turbidity currents and canyon cutting on the slope. On the western end of the shelf where the declivity of the slope was much gentler erosion by gravitational processes was much less. B. Along the ice front segment in contact with marine waters sediment emplacement on the slope was via direct deposition along the ice front on the upper slope and pelagic deposition from sediment plumes farther seaward. Massive sediment buildup on the upper slope led to extensive sediment failure producing low relief gravitational structures.

Fig. 5.—Diagrama esquemático que muestra la naturaleza de los procesos en un frente de hielo cerca de la plataforma continental externa. (A) En los extremos oriental y occidental de la plataforma los bancos exteriores bloqueaban el frente de hielo y las corrientes subglaciales depositaban su carga bloqueaban el frente de hielo y las corrientes subglaciales depositaban su carga en una llanura aluvial (outwash) y delta. En el flanco este el depósito masivo en los frentes deltaicos situados en la cresta de un talud continental abrupto condujo a una inestabilidad muy considerable, desplomes, formación de corrientes de turbidez y excavaciones de cañones en el talud. En el extremo oeste de la plataforma, donde la inclinación de la pendiente era mucho menor, los procesos gravitacioales produjeron menos erosión. (B) En el segmento de frente glaciar que estaba en contacto con las aguas marinas el emplazamiento del sedimento se realizó más hacia el mar, las plumas de sedimento generaron sedimentos pelágicos. La acumulación exagerada de sedimentos en el talud superior dio lugar a desplomes generalizados que produjeron estructuras gravitorias de bajo relieve.

CONCLUSION

The continental slope off western Nova Scotia is unusual in that it lacks the rough morphology typical of Atlantic continental slopes. Such turbidity current terrane, however, is found on the subsurface, eroded, presumably, by turbidity currents during low sea level. The overall low gradient inherited from early Neogene as a result of low subsidence and deposition rates, however, inhibited turbidity currents and the maintenance of these canyons. This allowed smoothing of slope valley walls by local erosion and onlapping of continental rise sedimentation to persist subduing and even the burying the rough terrane formed by the turbidity currents.

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