CENOZOIC GEOLOGY OF THE CONTINENTAL SLOPE AND RISE OFF WESTERN NOVA SCOTIA

by

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Submitted to the Department of Earth, Atmospheric,and Planetary Sciences in October, 1985, in partial fulfillment of the requirements for the degree of Doctor of Philosophy

ABSTRACT

The outer continental margin of Nova Scotia is divided by a diapir province, 40-110 km wide and ~1000km long, that trends subparallel to the shelf edge along the upper continental rise and slope. The growth pattern for a small region of this margin (61°-64°W) during the Late Cretaceous and Cenozoic was studied using seismic stratigraphy and well data. Structure maps show that a steep continental slope existed landward of the diapir province (~2200-3800 m water depth) from Early Cretaceous until Miocene time when onlapping upper rise sediments reduced the gradient. Shelf edge canyons were cut during the late Maestrichtian-early Paleocene, Eocene-Oligocene, and Pleistocene. Extensions of Tertiary canyons onto the slope are poorly defined, but small Paleocene fans of interbedded chalk and mudstone on the upper rise indicate that slope canyons existed at that time. Abyssal currents eroded the upper rise and smoothed relief on the continental slope in the Oligocene and middle(?) Miocene. In the Miocene, turbidites may have ponded on the upper rise landward of seafloor highs uplifted by salt ridges or pillows. Pliocene-Pleistocene sediments drape over pre-existing topography. At the beginning and end of the Pleistocene, turbidity currents, caused by delivery of large sediment loads to the shelf edge by glaciers, eroded the present canyon morphology.

The late Cenozoic section of the lower continental rise thins seaward from ~2 km near the diapir province and rests on Horizon A^u, a prominent unconformity eroded during the Oligocene by abyssal currents. The morphology of the lower rise is largely due to construction by down-slope deposits shed in the Miocene-Pliocene from uplift of the diapir province. Abyssal currents episodically eroded sediment, but current controlled deposition formed only a thin (<300 m) deposit in the Pliocene(?). Uplift in the diapir province accelerated during the Pleistocene and olistostromes up to 300 m thick were shed onto the lower rise. In the latest Pleistocene, sediments transported down-slope by near-bottom processes accumulated west of a sharp boundary running near 62°30'W from 500 m seaward to the abyssal plain. To the east, hemipelagic sediments accumulated above 4300 m, while turbidity currents, originating in deep canyons to the east, and abyssal currents reworked sediments below 4300 m. A glacial sediment source and relict shelf morphology controlled sedimentation processes and, thus, the location of depocenters on the slope and rise.

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INTRODUCTION

Continental rises dominate passive continental margins in terms of area and sediment volume (Emery and Uchupi, 1972; Drake and Burke, 1974). Most rise sediments lie on oceanic crust; a small portion, variable along strike, may lie over rifted continental crust (eg. Emery et al., 1970; Keen and Barrett, 1981). Sediments thin from greater than 10 km beneath the lower slope and upper rise to less than 250 m at the edge of the abyssal plain (Tucholke et al., 1982). Rise sediments are generally flat lying and undeformed except where diapirs have intruded near the rise-slope boundary.

Although many models describing the development of continental rises have been proposed, a clear consensus on continental rise origin does not exist in the literature. The debate can be summarized as a conflict over the relative importance of three principal sedimentary constructional mechanisms: (1) growth and coalescence of deep-seas fans (Emery, 1960); mass movements of slope sediments (Emery et al., 1970), and (3) deposition from contour-following abyssal currents (Heezen et al., 1966).

Morphology, structure, development, sedimentary processes, and deposits of deep-sea fans are known from modern and ancient analogs (eg. Normark, 1974; Nelson, 1976). Recognition of buried fans in ocean basins relys on identification of fan facies in deep drilling cores (eg. Lancelot, 1977), on seismic facies identified by internal reflection parameters and by external form in 2-dimensional seismic reflection profiles (Mitchum et al., 1977), or on isopach mapping of acoustic intervals (Mountain and Tucholke, in press). The acoustic character of seismic intervals is sometimes an ambiguous indicator of sedimentary facies, but three-dimensional seismic mapping reveals internal shape, continuity, and asymmetries which assist in identification of emplacement processes and allow volume calculations (Mitchum et al., 1977).

Slump scars and deposits are well known along passive continental margins (eg. Embley, 1980). "Chaotic" acoustic patterns within a seismic interval have been the principal criteria used for identification of buried slump deposits (Emery et al., 1970). Unfortunately, such an acoustic signal also might be identified as channel cut-and-fill associated with a buried fan deposit. Fan and slump deposits can be distinguished from one another on the basis of three-dimensional shape and orientation of deposits.

Surface abyssal current deposits have been mapped based on the presence of current-produced bedforms in bottom photographs or side scan sonar images, and based on echocharacter in 3.5 kHz echograms. (Heezen et al., 1966; Flood and Hollister, 1974; Damuth, 1980). Buried current deposits have been identified by deep-sea drilling (Ewing, Hollister, et al., 1972), by acoustic geometry of deposits (eg. Mountain and Tucholke, in press), by the presence of buried mud waves (eg. Mountain and Tucholke, in press), and by lack of internal reflectors (Laine and Hollister, 1981).

A small region of the continental margin off western Nova Scotia was chosen for a detailed seismic reflection survey to investigate the relative significance that deposits from different sediment transport mechanisms have had in constructing a rise prism. The Nova Scotian rise was selected for study because the transport mechanism responsible for the post-Eocene construction of the rise has been debated in the literature (Emery et al., 1970; Hollister and Heezen, 1972; Uchupi and Austin, 1979), the geologic history of the shelf is well known from boreholes and seismic studies (Jansa and Wade, 1975; Given, 1977), and the surface sedimentary processes are the subject of study by the HEBBLE program (Hollister and McCave, 1984). The continental margin off Nova Scotia differs from that off the U.S. in that the upper rise is intruded by piercement structures caused by movement of salt and shale in a diapir province 40-60 km wide. Landward of the diapir province is a narrow basin containing at least 6-8 km of sediment fill. The northern edge of this basin near the present shelf break is cut by slope canyons east of 61°W and has a relatively smooth topography to the west. Seaward of the diapirs, a Neogene sediment sequence forms a seaward thinning wedge atop a prominent unconformity eroded in the mid-Tertiary.

The first paper in this thesis describes the bathymetry, physiography, and the 3.5 kHz echocharacter of the rise off Nova Scotia. These features are interpreted in terms of the sedimentary processes active in the late Pleistocene.

The second paper describes the correlation of regional deep-water seismic stratigraphy from drill holes in the western North Atlantic Basin across the New England Seamounts onto the Nova Scotian continental rise. The diapir province effectively blocks direct correlation of shelf and slope facies into the basin. Drill holes in the basin northeast of the seamounts were located on basement highs and could not be used to constrain the stratigraphy along the continental rise.

The third paper in the thesis describes the geologic development during the Cretaceous and Cenozoic of a portion of the outer Scotian margin (61°-64°W). Diapirs intrude upper rise sediments between the 2200-3800 m isobaths. Landward seismic profiles and wells were used to study the development of the continental slope, history of canyons, and the history of sediment deposition and erosion. Seaward, the geometry and seismic character of intervals between seismic unconformities are used to infer sedimentary processes.

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CHAPTER 2

LATE PLEISTOCENE SEDIMENTATION ON THE CONTINENTAL SLOPE AND RISE OFF WESTERN NOVA SCOTIA

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ABSTRACT

Reflectivity patterns in echograms recorded from relatively smooth continental slope and rise off Nova Scotia can be grouped into four depositional associations. The shallowest asssociation corresponds to gravelly sand on the outer shelf that spills over the shelf edge and mixes with mud down to 500 m water depth. A sharp boundary near 62°30'W, extending from 500 m down to the abyssal plain, separates associations further seaward. Sediments transported downslope by near-bottom gravity processes accumulated west of the boundary; hemipelagic sediments accumulated east of the boundary from 500 to 4300 m. Below 4300 m, sediment waves are common in a contourite-fan association.

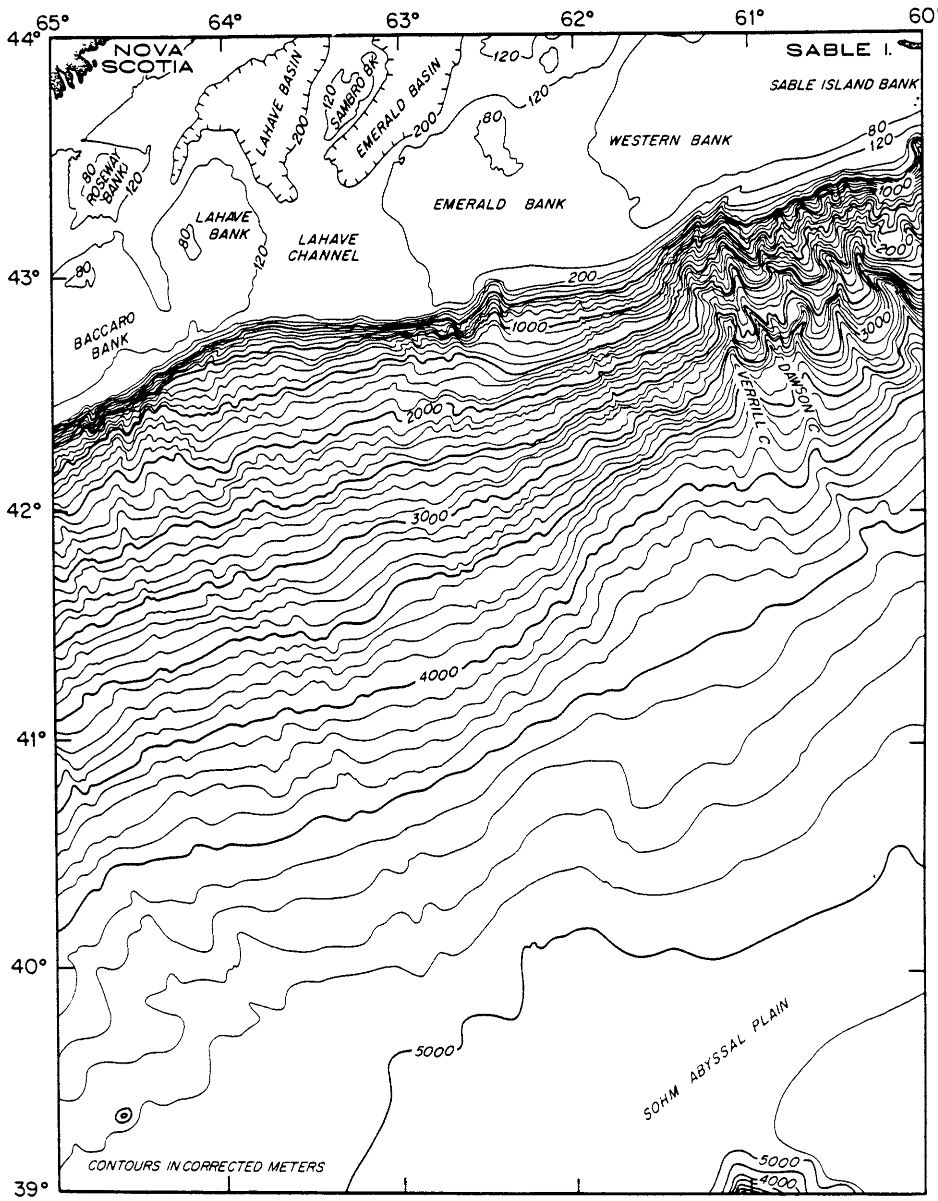
These associations indicate that during the late Pleistocene radically different sedimentation processes were juxtaposed across a sharp boundary down the slope and rise. West of 62°30'W small-scale slope failure occurred over a wide area due to relatively rapid accumulation of fine-grain glacial debris. The slope prograded seaward; the lower rise received distal turbidity current and debris flows deposits. East of 62°30'W relatively less failure occurred in hemipelagic sediment. Lower on the rise seaward of the hemipelagics, turbidity currents and contour currents carried sediment, originating in deep canyons to the east of 61°30'W, along the lower rise towards the south and west. In the west the primary depocenter was the slope and upper rise; in the east the depocenter was on the lower rise. A glacial sediment source on land and relict shelf morphology probably controlled sedimentation processes and, thus, the location of depocenters seaward of the shelf edge.

INTRODUCTION

The Nova Scotian continental slope between 65° and 61°30'W is among the smoothest in the North Atlantic. The steepest relief (maximum of 300 m over 2 km, 1:7) occurs in a narrow zone about 3 km wide between the shelf edge and the 800 m isobath (Fig. 1). Seaward, seafloor relief is typically much less than 50 m over 2 km (1:40). In contrast, canyon relief east of 61°30'W is of the order of 200-1000 m over 6 km (1:30-1:6) and extends about 90 km from the shelf edge to the 3500 m isobath.

Emery and Uchupi (1967) found evidence of slope progradation off LaHave Channel (informal name proposed here for the shallow depression between LaHave Bank and Emerald Bank), and Uchupi (1969) noted that the slope to the east was eroded by canyons. More recent seismic and well data support their observations and suggest that Tertiary shelf edge topography between 65° and 61°30'W was filled and smoothed during the Plio-Pleistocene, probably by slumps and turbidites, whereas canyons formed on the slope to the east (Jansa and Wade, 1975; Parsons, 1975; King and Young, 1977; Uchupi and others, 1982). East of 61°30'W, Stanley and Silverberg (1969) and Stanley and others (1972) concluded that erosion of glacial outwash debris exposed on Sable and Emerald Banks during the Pleistocene led to instability and slumping along the heads and flanks of canyons. Piper (1975) suggested that during low sea level canyon heads received sediment from longshore drift. Piper and Slatt (1977) and Stow (1978) proposed that ice-rafting from Laurentian Channel was also a significant sediment source.

Detailed studies within the smooth slope region west of 61°30'W found a striking change in sedimentation between 63° and 62°W. To the west, Hill



<u>Figure 1</u> Bathymetry of Nova Scotian continental slope and rise from Shor (1984) modified using RC2408 data.

(1983, 1984) mapped small turbidity current, channel/fan, and slump/slide features of late Pleistocene age between 180 and 1500 m water depth (Box I in Fig. 2). Between 500 and 2500 m just east of 62°30'W (Box II in Fig. 2), shallow seismic structure and sedimentology of piston cores suggest sedimentation was predominantly hemipelagic in the late Pleistocene (Hill, 1981; Piper and others, in press)

On the lower Nova Scotian continental rise (4600-5000 m) studies show that at the present time southwest-flowing bottom currents and "abyssal storms" erode and deposit sediment (High Energy Benthic Boundary Layer Experiment, HEBBLE; Box III in Fig. 2; Hollister and McCave, 1984, and references therein). Mean flow at these depths is probably due to recirculation of the Gulf Stream within this basin and not to shallower thermohaline circulation of the Western Boundary Undercurrent (Hogg, 1983). Hollister and McCave (1984) attributed the intense abyssal storms to mesoscale variability of the Gulf Stream; Hogg (1983) suggested the storms are related to instability of Gulf Stream recirculation. Pleistocene paleoceanography for this region is uncertain (Hill, 1981; Alam and others, 1983), and HEBBLE results can not be confidently extrapolated to the late Pleistocene.

Sedimentology studies on the rise have produced conflicting conclusions. Hollister (1967) and Hollister and Heezen (1972) suggested that reddish lutite interbedded with well-sorted silt laminations predominate in Pleistocene sediments of piston cores and that reflectivity patterns of 12 kHz echograms parallel isobaths. They inferred that contour currents deposited both surface and subsurface sediment. Using core data,

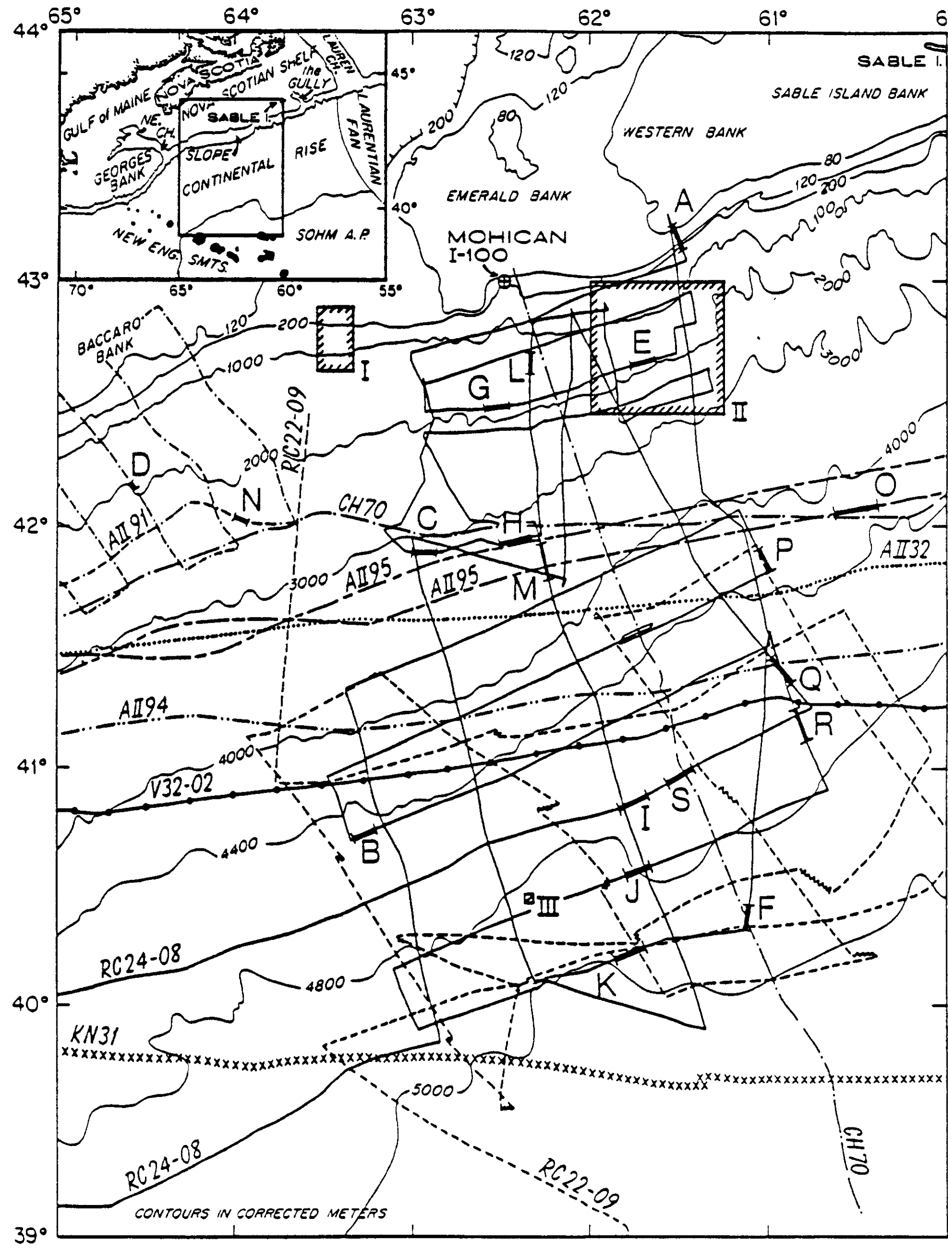


Figure 2 Echosounding tracklines. Thickened lines and letters show locations of profiles illustrated in Figs. 4-7. Box I is detailed study area of Hill (1983, 1984), Box II is study area of Piper and others (in press), and Box III is HEBBLE study site (Hollister and McCave, 1984). Tracklines are labeled with vessel abbreviation (AII=ATLANTIS II, CH=CHAIN, KN=KNORR, RC=ROBERT CONRAD, V=VEMA) and cruise number. Cruises AII 32, KN 31, and CH 70 used 12 kHz echosounding; all others used 3.5 kHz. Isobaths from Fig. 1.

Zimmerman (1972) also concluded that southwest sediment transport occurred during the Pleistocene.

In contrast, Emery and Uchupi (1972, p. 397-399) concluded that downslope processes (turbidity currents, slumps, and slides) emplaced most of the sediment on this rise. Others argue that Pleistocene sediments on the rise west of the Laurentian Fan (insert in Fig. 2) are turbidites: Stanley and others (1972) on the basis of sediment texture and mineralogy, Piper (1975) using sedimentary structures, and Stow (1978, 1979) using clay mineralogy, heavy mineral analyses, and sediment texture and structure. Piper and Slatt (1977) and Stow (1979, 1981) suggested that south of Nova Scotia the reddish lutite and montmorillonite interpreted by Hollister and Heezen (1972) as contourite was derived from ice transport across Nova Scotia and from coastal plain erosion followed by transport onto the rise by downslope processes.

Recent studies suggest that <u>both</u> cross-isobath and parallel-to-isobath processes may have been active. Shor and others (1984) inferred both types of processes from magnetic grain alignments in lower rise cores. Damuth and others (1979, 1981) observed turbidite channels, mass transport deposits, and migrating sediment waves in 3.5 kHz echograms from the lower rise.

DATA

In order to constrain interpretation of late Pleistocene sedimentation, the bathymetry, physiography, and reflectivity patterns (echo character) along continuous echosounding lines have been compiled principally between 64°-60°W and between the shelf break and abyssal plain (Fig. 2). About 4000 km of 3.5 kHz and single channel seismic data were collected in June-July, 1983, aboard the R/V ROBERT CONRAD (cruise RC 2408) and supplemented with about 5000 km of 12 and 3.5 kHz data (pulse length less than 5 msec) archived by Woods Hole Oceanographic Institution (WHOI) and Lamont-Doherty Geological Observatory (LDGO). Using RC 2408 records, seafloor physiography (seafloor channels, scarps, and sediment waves) was mapped. Bathymetry was contoured at 50 m interval after correction for speed of sound in water using Mathews (1939) and surface water mass locations (determined by 15 expendable bathythermographs to 500 m, by hourly sea-surface temperature measurements, and by weekly sea-surface oceanographic analyses from the National Earth Satellite Service, NOAA). To enlarge coverage, bathymetric contours (100 m interval) of Shor (1984) were modified to agree with the new data (Fig. 1). Published and unpublished descriptions of WHOI and LDGO cores were examined but not used extensively in this study (except as noted below), because the cores did not adequately sample all reflectivity patterns in the region mapped.

Damuth and others (1979, 1981) compiled an echo character map showing the distribution of reflectivity patterns for the continental rise off western Nova Scotia. Their study does not include the RC 2408 data and does not extend north of 42°N (about 3000 m depth). Their analysis differs from mine in that they used HEBBLE results in their interpretation and that they did not group reflectivity patterns into associations.

ECHOCHARACTER ASSOCIATIONS

In continental slope and rise environments elsewhere, echo character surveys have been successfully used to interpret sedimentation processes active both at present and during the late Quaternary (Damuth, 1980, and references therein). Although echo character maps are often useful presentations of these data, recognition of major depositional systems in such maps is sometimes difficult because (1) interpretation based on echo character alone is often ambiguous, (2) the nature of contacts are usually not included, and (3) echo character produced by related sedimentary processes (eg. slump, slide, and turbidites of a mass sediment failure) are often mapped separately. Another difficulty, in this study at least, is that tracklines may be too widely spaced to adequately map spatial variation in echo character.

For these reasons, echo character data in this study are organized into four "echo character associations". Within an association, a characteristic process or closely related processes produce sediment types and accumulation patterns -and thus, presumably, echo character- that are distinct from those in other associations. Associations may include regions of erosion (eg. the source region of a slide or slump) as well as deposition. Echocharacter associations are analagous to "facies associations" of Reading (1978, p. 5) and to "facies tracts" of Teichert (1958).

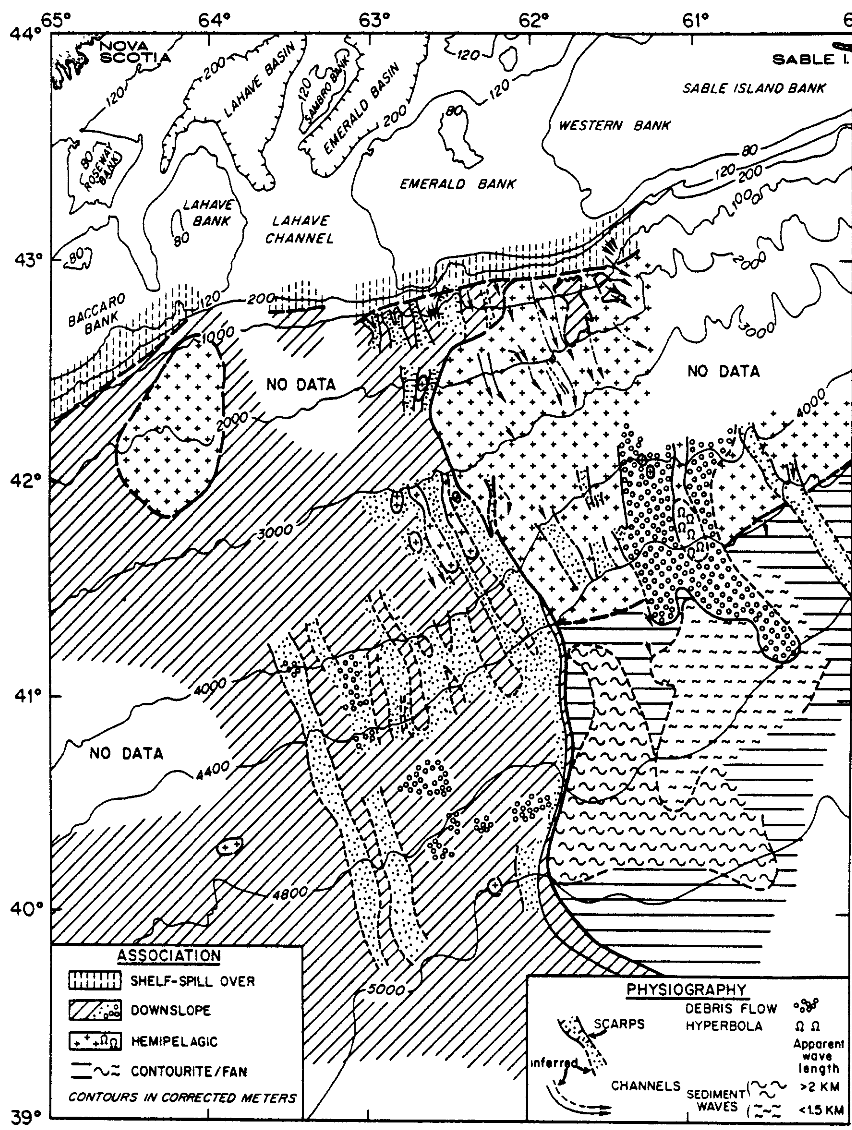
Interpretation of echo character associations is based on (1) reflectivity pattern, (2) water depth, (3) physiography, and (4) detailed morphologic, seismic, and sediment studies, (5) contacts between associations. Each association includes one or more echotypes defined by Damuth (1975, 1978, 1980); no new echotypes were recognized. Since many echotypes and structures seen in these data are described elsewhere (Jacobi, 1976; Flood and others, 1979; Damuth, 1980; Embley 1980, 1982; and references therein), not all echotypes are illustrated here.

The associations probably reflect sedimentation conditions during the late Pleistocene. No outcrops of significantly older rocks, as found elsewhere off Nova Scotia (Stanley and others, 1972; King and Young, 1977), were noted except for scarps and canyons indicated in Fig. 3. Holocene sediment on the Nova Scotian slope and rise is commonly less than 2 m thick, and average late Pleistocene accumulation rates are typically 100-500 $m/10^6$ yrs (Stanley and others, 1972; Piper, 1975; Stow, 1979; Hill, 1981). Thus, the deepest reflectors at about 100 msec vertical incidence two-way travel time (90 m at 1800 m/sec) are probably no older than $9x10^5$ yrs but may be much younger.

Shelf-Spillover Association

The flat outer shelf is characterized by a short (less than 10 msec) echo with no subbottom returns (profile A, Fig. 4; echotype IA of Damuth, 1978). The echo lengthens basinward of the shelf edge, and a single subbottom reflector similar to echotype IC of Damuth (1978) typically appears at about 400 m. In deeper water east of 62°15'W, several distinct, coherent subbottom reflectors (hemipelagic association; profile E, Fig. 4) appear with a transition zone about 10 km wide. West of 62°15'W distinct subbottom reflections are rare.

Surface sediments on the outer Scotian shelf are glacial tills deposited during late Pleistocene low sea levels and coarse sands and gravels formed by submarine reworking of tills and outwash during the Holocene



<u>Figure 3</u> Distribution of echo character associations. Heavy line separating associations is broken where boundary is uncertain. Channels (dots and arrows) and debris flows (circles) are downslope deposits which could be distinguished by echo character and physiography. Channels indicated by arrows have apparent widths less than about 1 km. Channel boundaries and arrows are dashed where uncertain; scarp boundaries are hatched. "Horseshoe" pattern indicates hemipelagic deposits with hyperbolic echo character (Profile P, Fig. 7). Wavy patterns mark contourite-fan association with sediment waves divided into two wavelength classes; limits are dashed where known and unmarked where questionable. Isobaths from Fig. 1.

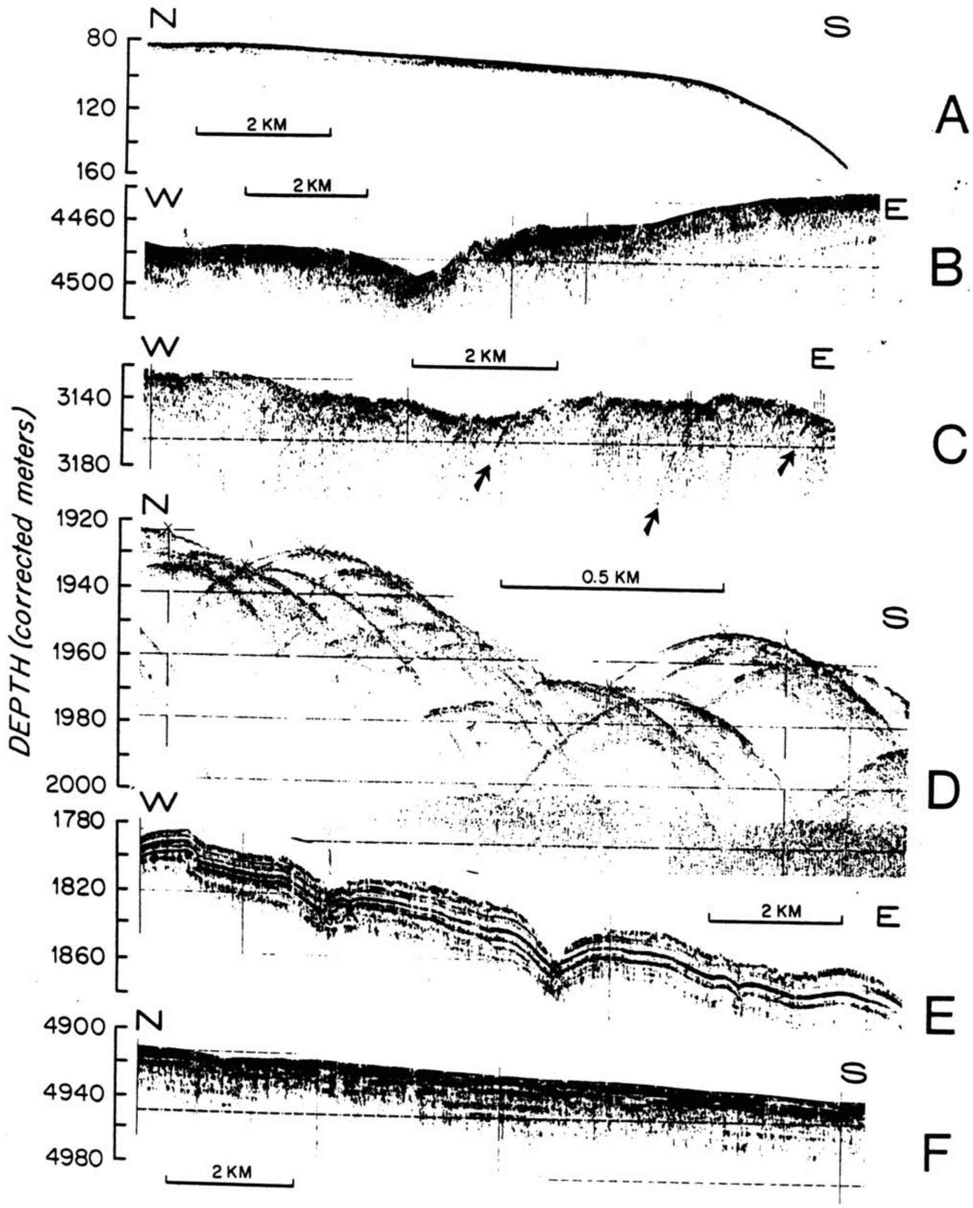


Figure 4 Profiles (3.5 kHz) from each echo character associations. A: shelf-spillover, B-D: downslope, E: hemipelagic, F: countourite-fan. Locations in Fig. 2. Profiles B and C cross channel deposits on the rise. Arrows in profile C indicate "tails" of some hyperbolae. In profile E, note continuity of some reflectors beneath two topographic channels.

transgression (King, 1970; Drapeau and King, 1972). Beyond the shelf edge silt and clay content of surface sediment increases with depth of water until mud becomes the most common facies at about 500 m (Stanley and others, 1972; Hill and Bowen, 1983; Hill and others, 1983). As the mud content of coarse shelf sediment increases seaward, the acoustic velocity and bulk density decreases (Hamilton and Bachman, 1982). As a result, the seafloor acoustic impedance contrast decreases and penetration increases. The seaward transition to subbottom "fuzziness" (prolonged, laterally incoherent echo) and subbottom reflectors may be due to deeper penetration of the acoustic signal, to actual changes in the number and thickness of beds (see for example, Tucholke, 1980; Damuth, 1980), or to scattering from increasing seafloor roughness (Bryan and Markl, 1966). In any case, because of the masking effects of the seafloor, little can be inferred about late Pleistocene sedimentation along the shelf edge from the echo character and distribution of shelf-spillover association.

Downslope Association

In general, downslope association includes areas where subbottom reflectors are scarce or absent and where the seafloor return is prolonged compared with the shelf seafloor echo. This association includes reflectivity patterns which elsewhere have been mapped separately (Damuth, 1980).

The predominant echotype from 1000 m depth to the abyssal plain is an echo 50-100 msec long with low coherence, only vague and fuzzy subbottom reflectors, and no hyperbolae (profile B, Fig. 4; prolonged echotype (IIB)

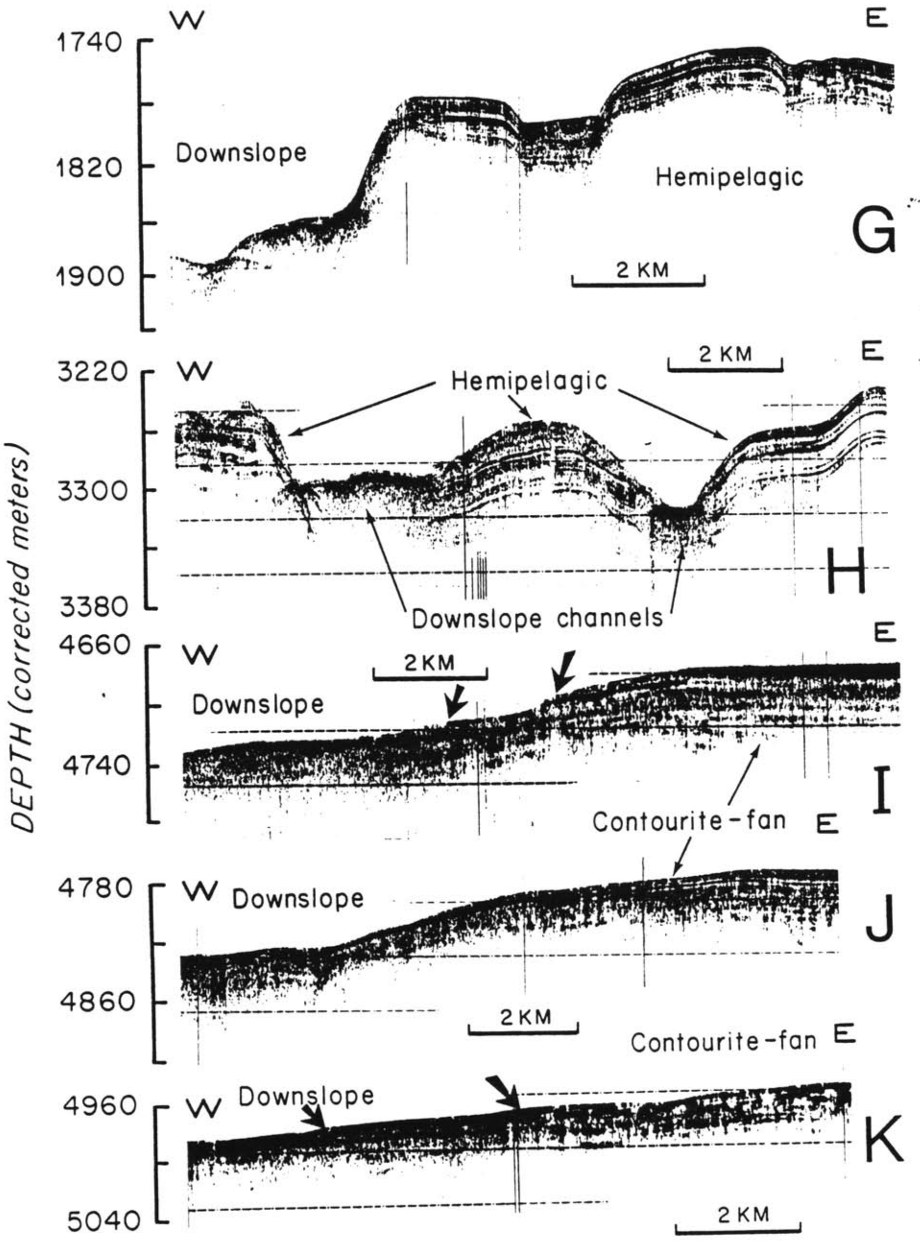


Figure 5 3.5 kHz echosounding records across eastern boundary of downslope association. Locations in Fig. 2. Seismic data indicate that the west-facing scarp forming the association boundary in profile G is the wall of a partially filled channel cut in mid-Tertiary (?). In profile H, reflectors in only the left-most channel wall are hyperbolic; reflectors in other boundaries indicate drape over pre-existing topography (note that the arrow marking the right channel points to a reflector that is continuous beneath the channel). Short arrows in profiles I and K indicate reflector outcrops. In profile J, units thin westward but are not truncated.

of Damuth, 1980). The Sohm Abyssal Plain also returns echoes of this type; distinct subbottom reflectors, commonly found on other abyssal plains and ascribed to turbidites (Damuth, 1975; Tucholke, 1980), were not found.

Prolonged echo may be due in part to oblique return from seafloor roughness features and, in part, due to the nature of interference and backscattering returns from many thick, coarse-grained beds (Bryan and Markl, 1966; Tucholke, 1980). Damuth's (1980 and references therein) echo character studies show that piston cores from areas of prolonged echoes (IIB) contain much thicker and more abundant silt/sand beds than do cores from other echotypes. Thick sand units and gravel are present in piston cores from areas identified as downslope association but are rare elsewhere (Hollister, 1967; Stanley and others, 1972; Stow, 1977; Tucholke, unpublished data).

On the levees of rise channels and on the slope between 400-1500 m, the prolonged echotype often includes one or two coherent subbottom reflectors (north end of profile L, Fig 6). The disappearance of the reflector along a trackline into prolonged echoes (not shown in Fig. 6) is not accompanied by any noticeable change in fuzziness. This suggests that sufficient energy penetrates subbottom in some regions of prolonged echo to return a reflector if a sufficient subbottom impedance contrast exists.

Large, irregular hyperbolae (profile D, Fig. 4; echotype IIIA of Damuth, 1978) were found only on the slope south of Baccaro Bank. Side echoes and diffraction off rough seafloor probably produce such hyperbolae (Damuth, 1975, 1980; Damuth and Hayes, 1977).

Regular, overlapping hyperbolae with no subbottom reflectors and with

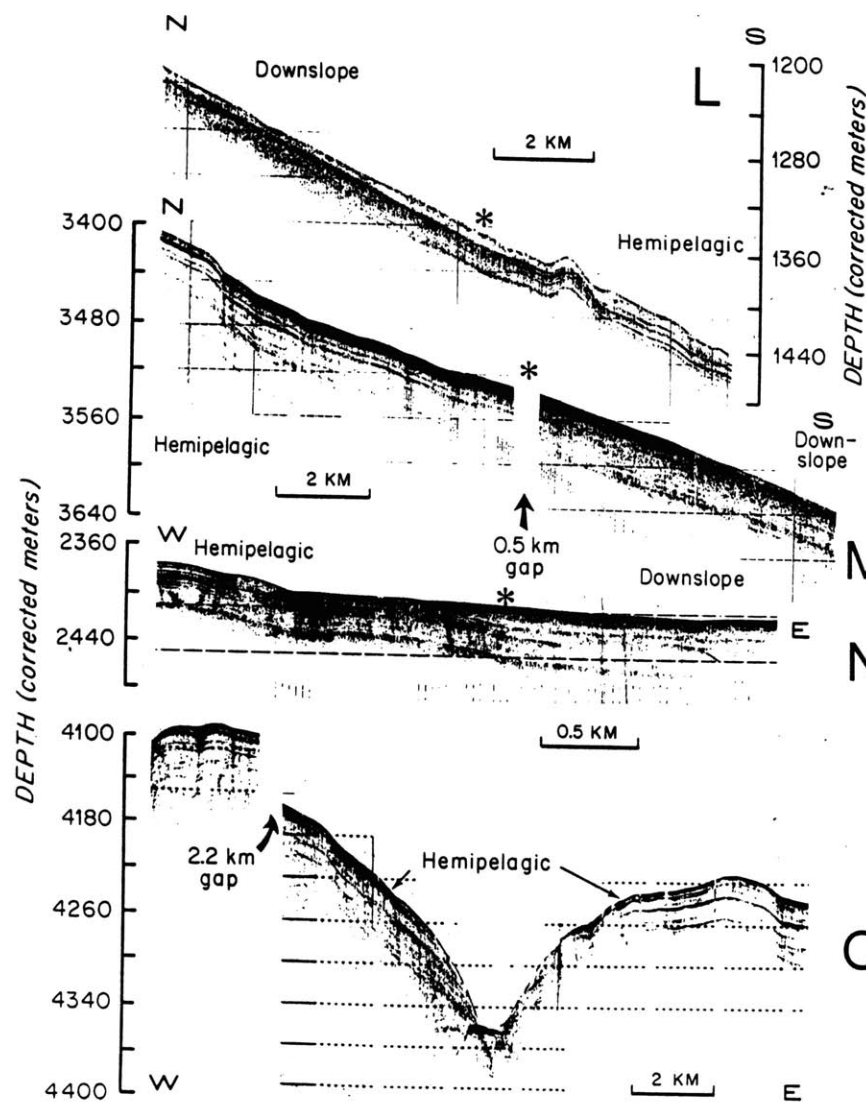


Figure 6 3.5 kHz echosounding records across hemipelagic facies contacts. Profiles L and M are gradational contacts. Stars indicate approximate center of transition. Locations in Fig. 2. In profile L, the uppermost unit may be a debris flow; deeper units show a southward change in echo character. Echograms from tracklines parallel to isobaths indicate that the south end of profile M is in a channel (FIg. 3) suggesting that the downdip loss of coherent subbottom reflectors may be due to erosion or to masking by channel axis deposits. In profile N, a debris flow (east) laps over dipping hemipelagic deposits (west) masking returns from subbottom units. Subbottom returns from the east channel wall in profile O can be traced eastward 50-60 km to the next channel (discussed in text).

vertices either at the seafloor or at varying elevations above the seafloor (profile C, Fig. 4; echotypes IIID and IIIC of Damuth, 1980, and Damuth and Hayes, 1977) were uncommon on the rise. However, Damuth and others (1979, 1981) map large fields between 65-64°W below 3000 m where my data are sparse and of poor quality. East of 64°W, such hyperbolae were recorded from the floors of channels (eg. profile C) and from a large debris flow deposit which fills a topographic channel near 42°N, 61°W. Such hyperbolic echoes are characteristic of the upslope portions of slump-debris flow deposits (Embley, 1980; Damuth and Embley, 1981).

On the slope and upper rise to the west of 62°15'W no evidence was found in the physiography, reflection pattern, or seismics for large-scale late Pleistocene slumps. In particular, no evidence such as documented by Embley (1980, 1982, and references therein) elsewhere in the Atlantic could be found on the slope in this data set or that of Hill (1983, 1984) for a slump greater than 5 km wide. On the rise semi-transparent layers resembling debris flow deposits could not be mapped over distances greater than about 20 km with the exception of a large deposit near 61°W, 42°N (Fig. 3). Damuth and others (1979, 1981) map a large area of debris flow deposits between 40°30'-41°30'N and 62°-63°30'W. However, the RC 2408 data in this region show channeled physiography and widespread areas of prolonged echo with minor hyperbolae. This evidence suggests turbidite deposits, although the occurrence of debris flow fill in channels may be underestimated due to the difficulty of distinguishing thick debris flows in areas of prolonged echo.

Late Pleistocene sediments mapped here as <u>down-slope</u> <u>association</u> are interpreted as deposits emplaced predominantly by near-bottom gravity flows: turbidity currents and mass-transport mechanisms. The most compelling reason is the distinct, continuous eastern boundary near 62°W. This feature stretches nearly 370 km from about a depth of 500 m down to the abyssal plain (Fig. 3). Echocharacter changes in 5 km or less across the boundary. Neither hemipelagic nor contour-following sedimentation can account for these characteristics.

The association boundary on the slope sometimes coincides with a seafloor scarp and outcropping reflectors clearly indicating channel cutting to the west (Profiles G and H, Fig. 5). Occasionally, reflectors in channeled walls indicate that sediment was draped over pre-existing topography or went through several cycles of drape and erosion (for example, the eastern half of profile H, Fig. 5). Rarely, the change in echo character across this boundary is gradational (Profiles L and M, Fig. 6) probably due to a gradual change in seafloor or subbottom sedimentology.

On the lower rise the seafloor deepens about 100 m from east to west across the boundary. On two crossings outcropping reflectors indicate erosion (Profiles I and K, Fig. 5). Elsewhere, continuous units as well as eroded units overlie the boundary (Profiles J, Fig. 5). The thicknesses of continuous units are typically unchanged across the boundary or are thicker to the east than on the sloping boundary suggesting that sediment was draped over pre-existing topography.

Three additional lines of evidence support the interpretation of downslope processes to the west of the boundary. First, Hill (1983, 1984) found that slumps, slides, and channel-fan systems, with widths less than 2 km along isobaths, originate at the base of a 150 m escarpment which parallels the 500 m isobath. In cores, sand and silt layers with chaotic stratigraphy are common within reddish-brown to brown mud indicating that much slope sediment was emplaced by gravitational processes in the late Wisconsin (Hill, 1983, 1984). Although none of the echosounding lines used in this study cross Hill's area, the physiography is similar to that where my coverage shows downslope association.

The second indication of downslope transport is that well-developed channels are more common west of the boundary than east (Fig. 3). Channels with steep scarps (3-6°; 1:19-1:7; 80-150 m over 1-1.5 km) occur between 1700-2300 m (profile G, Fig. 5). Shallow (30-40 m), steep-sided channels 4-15 km wide can be traced between 3000-3800 m through outliers of well-stratified sediment (profile H, Fig. 5). Between 3500 m and the abyssal plain, channels, mapped primarily by seafloor morphology, are broader (about 15 km) and less well-defined features (Fig. 3).

Third, detailed topographic mapping near 62°30'N, 40°W indicates that lower rise morphology is controlled by erosional channels (Shor and Lonsdale, 1981). Magnetic anisotropy measurements on gravity cores at the HEBBLE site (Box III in Fig. 2) show grains aligned downslope in the pre-Holocene section (Flood and others, in press; Shor and others, 1984).

Hemipelagic Association

Hemipelagic association occurs in areas where the upper 50-100 msec of echograms show several distinct, coherent subbottom reflectors that are laterally coherent over 5 km or more (profile E, Fig. 4; distinct with parallel subbottoms (IB) of Damuth, 1980; Fig. 5 in Damuth and Hayes, 1977). In general, this association occurs between depths of 500 and 4300 m east of the contact with downslope association at about 62°30'W (Fig. 3). The association could not be mapped on the slope east of 61°30'W because of lack of data. However, cores from intercanyon ridges to the east suggest similar sedimentation patterns (Stanley and Silverberg, 1969), and the association can been identified on the rise south of the canyons at least as far as 60°W (Fig. 3). The southern boundary is poorly constrained. Isolated blocks of well-stratified sediments with seafloor relief of 5-30 m occur within the region of downslope processes (Fig. 3; for example, 40°15'N, 63°50'W). Some block edges show truncated reflectors suggesting an erosional origin (profile H, Fig. 5).

V-shaped channels up to a few hundred meters wide (for example, profile E in Fig. 4) can be traced on the slope for up to 50 km (Fig. 3; Piper and others, in press), but most are considerably shorter. Round bottom channels 5-10 km wide could be identified at 4000-4200 m by changes in physiography and by fading out of subbottom reflectors, but these channels could not be confidently traced to upslope sources (Fig. 3).

Sediment in this association is interpreted to be predominantly hemipelagic (ie., composed primarily of fine-grained terrigenous mud transported seaward well above the seafloor, sensu Pierce, 1976) because reflector units are continuous over a wide range of topography and slope both in strike and dip lines. Such an origin is also suggested by the gradational contact with spillover association near 500 m and continuity of the association from the slope onto the upper rise (Fig. 3).

Piper and others (in press) report that slope seismic units thin downslope not along slope. Locally on the slope and upper rise subbottom units drape over relic channels and can be correlated over 10-20 km along isobaths (Profile E, Fig. 4; Piper and others, in press). Subbottom units can also be traced with uniform thickness for 20-30 km onto the west wall of Verrill Canyon at 1000-1500 m depth and for 30-50 km across the channel at 4000-4300 m in Profile O (Fig. 6). These thickness changes suggest that canyon/channel wall units may not have a turbidity current origin and, if so, that the canyon/channels east of 61°W may not have been conduits for significant transport of fine-grained sediment during the late Pleistocene.

Thin slumps and slides between 500 and 2000 m redistributed 10-20 m of surface sediment downslope in post-glacial time (Piper and others, in press, Box II in Fig. 2; Fig 3). Mass-transport deposits may be more common in this region than is indicated by Fig. 3. However, core descriptions (Piper and others, in press; Hill, 1981) indicate that late Pleistocene sediment has a hemipelagic source and that both deposition and reworking processes are significantly different from those west of 62°30'W.

A large debris flow deposit fills channels near 42°N, 61°W and laps over the uppermost layers of well-stratified hemipelagic deposits which form the channel walls. Although the deposit could not be traced shallower than 3900 m depth with the data available, seafloor topography suggests the flow may have originated in Verrill Canyon. For these reasons this deposit is interpreted to be the result of sediment failure within the steep-walled canyon region of the slope to the east of 61°W and not representative of late Pleistocene sedimentation conditions on this portion of the rise.

Hummocky seafloor (Profile P; Fig. 7) occurs on a ridge between two arms of the debris flow and returns hyperbolic echoes with seafloor relief of 10-20 m. Subbottom reflectors are conformable and do not show migration. The only pair of tracklines which intersect on the structures shows no difference in wavelength suggesting no elongation. These features appear similar to structures ascribed elsewhere to creep (Hill and others, 1982).

Contourite-fan Association

Contourite-fan association is mapped by occurrence of subbottom reflectors which are laterally discontinuous over distances greater than 1-2 km (profile F, Fig. 4; Fig. 5 in Damuth, 1975; echotype IIA, Damuth (1980, and references therein). The association is found south of the 4300 m isobath and east of 62°30'W. Although the northern boundary is poorly sampled and not well constrained, it appears gradational over 2-12 km. Some reflectors in the hemipelagic unit crop out south of the boundary (Profile Q, Fig. 7). The west boundary is sharp; the transition to no subbottom reflectors occurs over 2-5 km and is accompanied by the depth changes described above (Profiles I, J, and K; Fig. 5). South and east boundaries are not well surveyed.

Seafloor sediment waves occur only within this association (Profile Q, R, and S; Fig. 7). Subbottom reflectors are either offset with depth indicating upslope migration or conformable. Wavelengths at any trackline azimuth are shortest near Profile Q and increases southward and westward (Figure 3 and 7). Wavelength measurements at locations where 3 or more

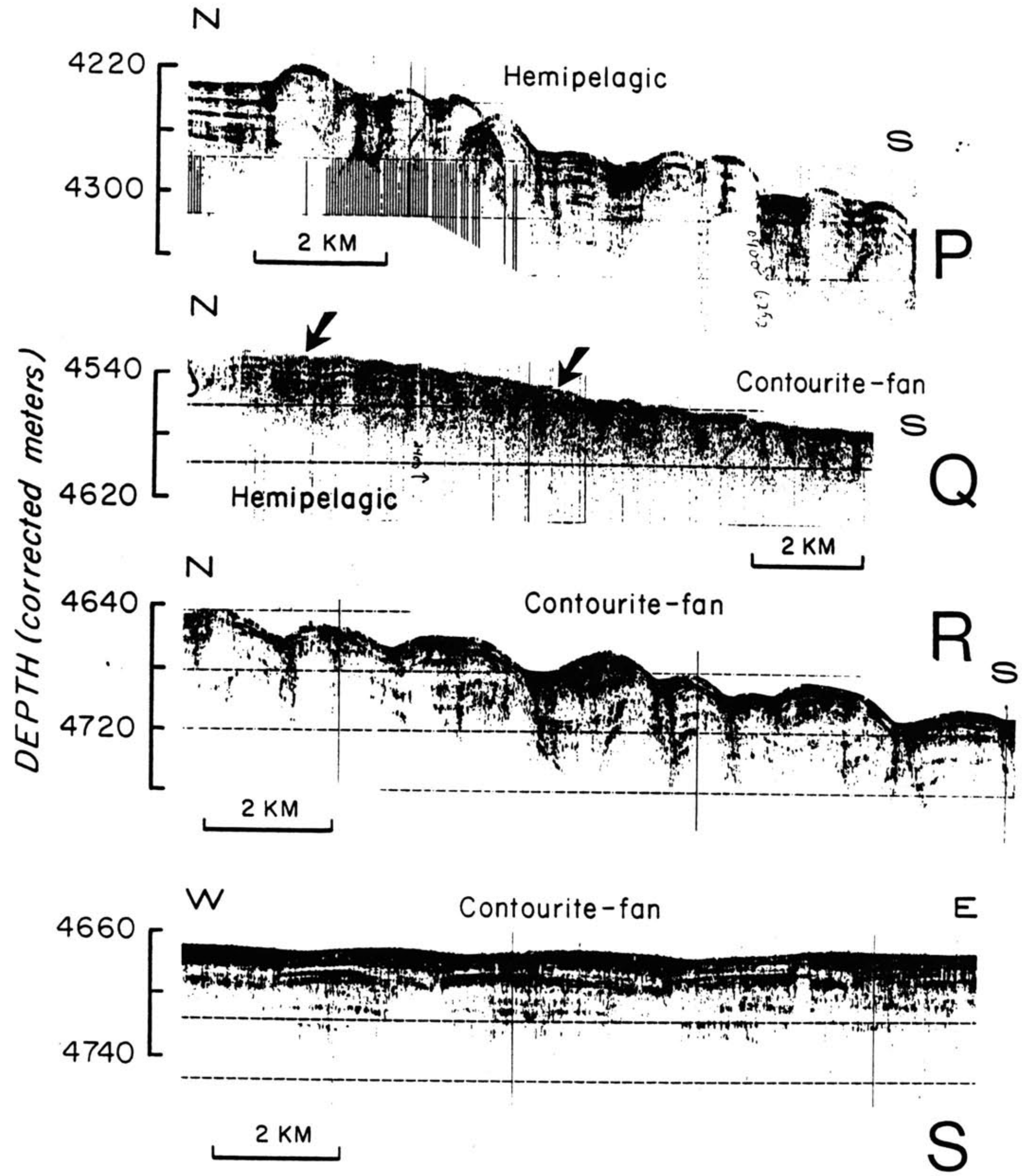


Figure 7 3.5 kHz echosounding records across hyperbolae in hemipelagic deposits (profile P) and sediment waves (profiles Q-S). Locations in Fig. 2. Short arrows in profile Q indicate reflector outcrops. Note increase in apparent wavelength southward from Q to R and westward from R to S.

tracklines cross are not consistent enough to confidently determine wave orientations.

Contourite-fan association shares features with turbidite and contourite deposits which have been clearly identified on other continental rises, so probably both gravity-driven flows and abyssal currents influenced sedimentation here in the late Pleistocene. On one hand, a turbidite origin is suggested by echo character (see discussion in Damuth, 1980), proximity to canyons east of 61°W, occurrence of well-developed rise channels, and similarities of echo character to that of Sohm Abyssal Plain deposits. In addition, sediment waves are known from abyssal fan environments (Damuth, 1979, Normark and others, 1980, and references therein). On the other hand, reflectors cropping out at the upslope boundary suggest erosion by contour-following currents. Echograms recorded from channel walls appear hemipelagic in character (Profile O, Fig. 6; see discussion above) suggesting that late Pleistocene turbidity currents may not have been large enough to transport significant sediment southwest along the lower rise. Lastly, sediment waves are a well-known characteristic of contour current deposits (Damuth, 1975; Embley and Langseth, 1977; Damuth, 1980, and references therein). Waves here are best developed between 4500-4900 m where current erosion and deposition are most active at present (Hollister and McCave, 1984). In many of these aspects, contourite-fan association appears similar to contemporaneous turbidite channels and contour-current deposits on the continental rise off northeast South America (Embley and Langseth, 1977). Here, further distinction between turbidite and contourite was not possible with this data set.

DISCUSSION

The principle finding of this study is the region of down-slope association centered seaward of a gap in the outer shelf banks with a sharp eastern boundary which extends from the shelf edge to the abyssal plain. To account for this pattern and its relationship to shelf topography, a model is proposed below presuming that the upper 50-100 m of sediment, which produce reflectivity patterns, are roughly late Pleistocene everywhere and that temporal changes in sedimentation, whether episodic or fluctuating (periods less than 100,000 yrs), are less important than spatial changes. These assumptions are consistent with Hill's (1983, 1984) studies of Wisconsin stratigraphy on the slope, lack of evidence for wide-spread episodic mass failure events west of 62°W, and occurrence of drape, as well as erosional, relationships along the eastern boundary of the down-slope asssociation.

Sedimentation Model

During the late Pleistocene, grounded ice probably reached no further than the inner Nova Scotian shelf, and sea level on the outer shelf was 100-120 m lower relative to present (King, 1970, 1983). Under these conditions LaHave Channel was the only substantial waterway (about 60 km wide, 30-50 m deep) between grounded ice and the continental slope; the Gully (inset Fig. 2) is significantly narrower and more sinuous. Midshelf basins probably trapped sediment transported from the glacier near the seafloor, while ice-rafting and meltwater plumes carried sediment southward through LaHave Channel. Wisconsin ice may have abutted the landward edge of outer banks off western Nova Scotia (Stanley and Cok, 1968; Emery and Uchupi, 1972; Schlee and Pratt, 1970; Schlee, 1973). If so, ice streams probably funneled sediment through LaHave Channel as near-bottom transport as well as surface plumes.

Seaward of the outer banks, in contrast to conditions seaward of LaHave Channel, erosion of the outer banks, ice-rafting from Northeast or Laurentian Channels, and erosion of canyon walls were probably the principal sediment sources for the slope. Most supply was localized at canyon heads (Piper, 1975; Stow, 1978).

Based on lithology of slope sediments (Hill, 1983, 1984) and distribution of echo character associations, the dominant seaward sediment flux through LaHave Channel is inferred to be a sea surface plume analogous to plumes off modern glaciers (Edwards, 1978; Pfirman, 1984). Near-bottom transport probably carried to the upper slope lesser amounts of coarse debris released from icebergs grounded along the northern channel edge.

Most fine-grained sediment in the LaHave Channel plume probably settled out upslope of the 1200 m isobath. Because failure of fine-grain sediment on a slope is more likely in faster accumulating sediments (Morgenstern, 1967; Coleman and others, 1983), the boundary between downslope and hemipelagic associations at 500-1000 m and 62°15'W probably corresponds to the southeast limit of significant sediment transport in surface and intermediate waters. Other explanations are not consistent with existing geologic and oceanographic data: (1) Multi-channel seismic lines (obtained from Canadian Oil and Gas Lands Administration) show no faults and no changes in sediment structure of the upper 2 km associated with the boundary. (2) Surficial pockmarks, which on the shelf indicate weakening of surface sediment by release of pre-Pleistocene gas (Josenhans and others, 1978), have not been found on the slope or near the shelf edge (Hill, 1983; Hill and others, 1983, King, 1983). (3) Hill and Bowen (1983) suggest that spatial changes in present shelf edge sediment transport may be due to longitudinal changes in spawning rates of warm-core rings off the Gulf Stream. However, recent data show no sharp change in ring spawning (Joyce, 1984), and the appearance during glacials of discontinuities in rate and location of ring genesis seems unlikely.

As a result of even sediment dispersal within the LaHave Channel plume, deltas did not develop on the slope, failure was small in scale, and canyons and steep slopes similar to those east of 61°W did not develop. Small slides, debris flows, and turbidites transported sediment from the slope to the upper rise forming a depocenter termed the "slope-rise apron" (Stow and others, 1984). The lower rise aggraded as a distal province of the slope-rise apron in a manner significantly different than the lower rise elsewhere off Nova Scotia, which developed as a result of slope by-passing and deep-sea fan construction (eg. Piper, 1975). Downslope transport of sediment occurred in many channel systems; no dominant channel-fan system developed on the rise.

To the east of 62°15'W hemipelagic mud accumulated on smooth slope. Downslope transport in narrow channels and thin, broad slump-slide features (Piper and others, in press) was less common here than to the west. Further east, hemipelagic accumulation was restricted to canyon divides (Stanley and Silverberg, 1969). Turbidity currents transported sediment down canyons to fans on the lower rise. Below 4500 m channel overbank flow may have deposited load as sediment waves (Normark and others, 1980). Alternatively, abyssal currents may have redistributed sediment to the southwest either by deflecting overbank spillage or by resuspension and transport. Winnowing may have occurred, but probably was insignificant because cores from this region contain substantial amounts of mud (Piper, 1975; Stow, 1977, 1978).

The southwestward bulge of contours near 62°30'W appears to mark the southwest limit of significant lower fan progradation. However, reflector relations in the termination of the bulge show little evidence of progradation in the section resolvable with 3.5 kHz echosounding (Profiles I, J, and K; Fig. 5). Thus, the topography was likely relict during the late Pleistocene. Reflectors cropping out to the west at the association boundary indicate that erosion by turbidity currents helped maintain the relief.

Geologic Significance

Sharply defined variations in late Pleistocene sedimentation occurred along strike in the slope and rise off Nova Scotia. Aggradation of the slope-rise apron west of 62°15'W was controlled by small-scale downslope movement without a central canyon-channel system. This is unlike most fan models (for example, Stow, 1981) and unlike the slope by-passing/ coalescing rise-fan model (Piper, 1975) which applies better to the east.

The reason for lateral variability was variation in rate, process, and grain-size of sediment supplied from the shelf. During low sea levels in the late Pleistocene, LaHave Channel influenced slope and rise accumulation patterns by funneling fine-grain sediment (glacial) to the slope. Thus, outer shelf morphology played a prominent role in slope-rise sedimentation. In the past, under similar conditions of sediment source and sea level, laterally variable shelf morphology might also have influenced deepwater sedimentation patterns. Temporal change in shelf morphology (the present Nova Scotian shelf morphology was largely determined by fluvial processes during a mid-Tertiary low sea level, King and others, 1974) might also have produced stratigraphic change in patterns of accumulation, lithology, and seismic facies on the slope and rise.

CONCLUSIONS

Late Pleistocene sedimentation varied along the slope and rise off western Nova Scotia as a result of a glacial source on the shelf and spatial variation in outer shelf morphology. Seaward sediment transport in deep water differed across a sharp boundary near 62°15'W extending 370 km from the upper slope to the abyssal plain. Fine-grained glacial debris accumulated relatively rapidly on the slope seaward of LaHave Channel. Small-scale sediment failure created a slope-rise apron that filled rather than carved topographic relief. Turbidity currents, slides, and debris flows distributed sediment onto the lower rise through many small channels which lacked a major slope or shelf source. On smooth slope east of the boundary, failure rate of hemipelagic sediment was significantly less. Well-stratified sediment accumulated down to 4300 m depth with minor sliding and channeling. Further east, outer banks reduced the rate of supply to the slope of fine glacial outwash and ice-rafted debris. Turbidity currents transported sediment eroded from outer banks down canyon-channel systems to the lower rise where both overbank spillage and contour-following currents deposited sediment in a lower rise fan.

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CHAPTER 3

SEISMIC STRATIGRAPHIC CORRELATION ACROSS THE NEW ENGLAND SEAMOUNTS, WESTERN NORTH ATLANTIC

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ABSTRACT

Correlation of deep-water seismic stratigraphy from the western North Atlantic Basin across a basement high within the New England Seamounts shows that stratigraphy on the Nova Scotian rise is broadly similar to that off the United States. Age of volcanism inferred from stratigraphy above the basement high is consistent with a hot spot origin, but other stratigraphic data suggest eruptions occurred significantly before and after passage of the hot spot. Episodic volcanism along a crustal fracture must have occurred in addition to hot spot volcanism. The New England Seamounts were not a barrier to stratigraphic development of the continental rise. However, patterns of erosion by abyssal currents in the late Tertiary (A^u) and subsequent deposition were centered symmetrically about the seamounts. We attribute this symmetry to interaction between abyssal currents and the Gulf Stream.

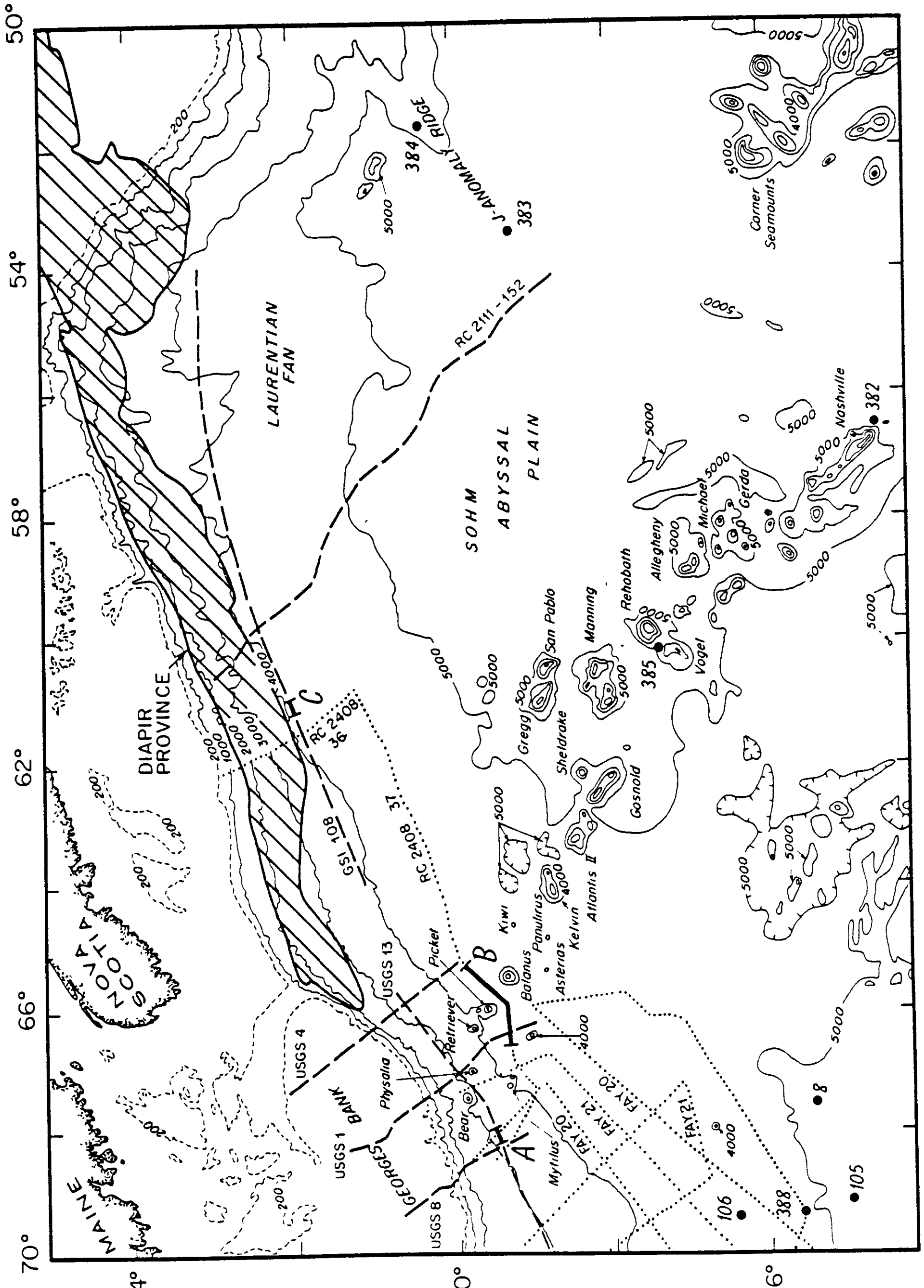
INTRODUCTION

The Sohm Abyssal Plain and the Nova Scotian continental rise (hereafter, Sohm basin) are separated from the Western North Atlantic basin by the New England Seamounts, variously interpreted as eruption centers along a linear crustal fracture (Uchupi and others, 1970) or as the trace of a hot spot on the North American plate (Vogt and Tucholke, 1979). Duncan (1984) reported radiometric age dates supporting southeast movement of hotspot volcanism between 110 and 70 Ma (Albian- Maestrichtian). Prior to our studies (this paper; Ebinger and Tucholke, in prep.) it was not known whether seismic horizons in the Western North Atlantic basin (Tucholke and Mountain, 1979;

Klitgord and Grow, 1980) continue north of the seamount chain or whether a different seismic stratigraphy developed in the Sohm basin as a result of either circulation changes caused by the seamounts or variations in continental sediment source unrelated to the seamounts.

Several factors hampered earlier attempts to establish a stratigraphic framework within the Sohm basin. The sedimentary section is intruded by the seamounts themselves and, between the seamounts, by irregular, poorly surveyed basement highs (Uchupi and others, 1970; Tucholke and others, 1982). Jurassic-Lower Cretaceous stratigraphy cannot be correlated around the southeast end of the seamount chain because these sequences pinch out on oceanic crust northwest of Nashville Seamount (Fig. 1; Mountain and Tucholke, in press). Reflector identifications based on tracing horizons upslope from deeper-water sequences and around the northwest end of the seamounts off Georges Bank are hampered by changes in sedimentary facies, poor seismic resolution due to deep burial and seafloor canyons, an Upper Jurassic/Lower Cretaceous carbonate reef beneath the continental shelf edge and slope, and numerous unconformities beneath the slope and upper rise (eg. Schlee and others, 1979). Shelf stratigraphy of Nova Scotia and Grand Banks, determined by extensive drilling and seismic surveys, cannot be carried seaward confidently through a diapir province developed beneath the continental slope and upper rise (Jansa and Wade, 1975; Given, 1977; Fig. 1).

In the absence of direct ties across the New England Seamounts, prior identification of western North Atlantic horizons in the Sohm basin was based on relative stratigraphic position and seismic character (Emery and others, 1970; Jansa and Wade, 1975; Uchupi and Austin, 1979; Wade, 1981;



<u>Figure 1</u>

Location of primary seismic data. Seismic profiles (A,B, and C) along thickened tracklines appear in Figs. 2 and 3. Location of diapir province (hatchured) from Uchupi (1984a and 1984b) and Emery and Uchupi (1984). Dotted tracklines are single-channel data; dashed tracklines are multi-channel data. Solid dots are Deep Sea Drilling Project sites. Bathymetry in corrected meters from NAVOCEANO chart NA-6.

Wade in Beaumont and others, 1982). Abbreviated sedimentary sections, drilled on the flanks of two New England Seamounts (Deep Sea Drilling Project [DSDP] Sites 382 and 385, Fig. 1) and on the J-Anomaly Ridge (DSDP Sites 383 and 384), provide little stratigraphic control (Tucholke, Vogt, and others, 1979). Seismic markers previously identified only within the Sohm basin include horizon L, a prominent reflector and unconformity at the base of the youngest sediment wedge in the Laurentian Fan sequence (Uchupi and Austin, 1979; Stow, 1981; Piper and Normark, 1982) and horizons R and S, reflectors within the youngest wedge (Normark and others, 1983). Uchupi and Austin (1979) inferred a Pliocene-Pleistocene age for L, Normark and Piper (1982) placed L at the PLiocene-Pleistocene boundary, and Ebinger and Tucholke (in prep.) suggest L may be as old as mid-Miocene.

Ebinger and Tucholke (in prep.) identified seismic sequence boundaries in a multi-channel dip line across the Nova Scotian rise (RC2111 line 152, Fig. 1; Tucholke and Ludwig, 1982). Along the line, age of oceanic basement was determined by identifications of M-series magnetic anomalies and by extrapolating an M22-M25 seafloor spreading rate of 11 mm/yr into the Jurassic Magnetic Quiet Zone (JMQZ). Minimum ages of deep sequence boundaries were then dated by basement age at the location of seaward pinchouts.

In this paper we correlate Upper Jurassic-middle Tertiary seismic stratigraphy south of Georges Bank across the New England Seamounts using new data. We also present a seismic stratigraphic reference section for the continental rise off Nova Scotia. Our purpose is to evaluate effects of the New England Seamounts on development of the sedimentary sequence and to present new data relevant to the origin of the New England Seamounts.

DATA

A single-channel seismic line (RC2408 line 37) was shot southwestward along the Nova Scotian continental rise and through the New England Seamounts, avoiding much of the known basement relief associated with the seamounts (Fig. 1). Southwest of the seamounts the line crosses several U.S. Geological Survey (USGS) multi- and single-channel lines whose stratigraphy has been carried northward from DSDP holes (Klitgord and Grow, 1980; Mountain and Tucholke, in press; Schlee and others, in press; K. Klitgord, personal communication, 1983). Mountain and Tucholke (in press) calibrated the seismic stratigraphy by borehole correlations and reflector pinchouts on basement as follows: J2 (middle Callovian-lower Kimmeridgian), J1 (upper Oxfordian-Tithonian), ß (Hauterivian-Barremian), A^* (upper Maestrichtian), and A^u (upper Eocene- lower Miocene). Northeast of the seamounts, seismic horizons can be traced confidently eastward along line 37 as far as 63°W (Fig. 1). Farther east, the post-A^u section thickens and deeper reflectors become less continuous. However, reflector identifications can be jump-correlated 90 km landward to GSI line 108, a multi-channel strikeline lying seaward of the diapir province (Fig. 1), and traced northeastward along the Nova Scotian rise. Reflector identifications in line 108 agree with those in lines 36, 37, and 152.

CROSSING THE NEW ENGLAND SEAMOUNTS

Along line 37 southwest of the New England Seamounts, oceanic basement, J1, ß, and A^u are prominent seismic markers that can be confidently traced from trackline crossings with USGS lines (Fig. 2 and 3). A^* can be

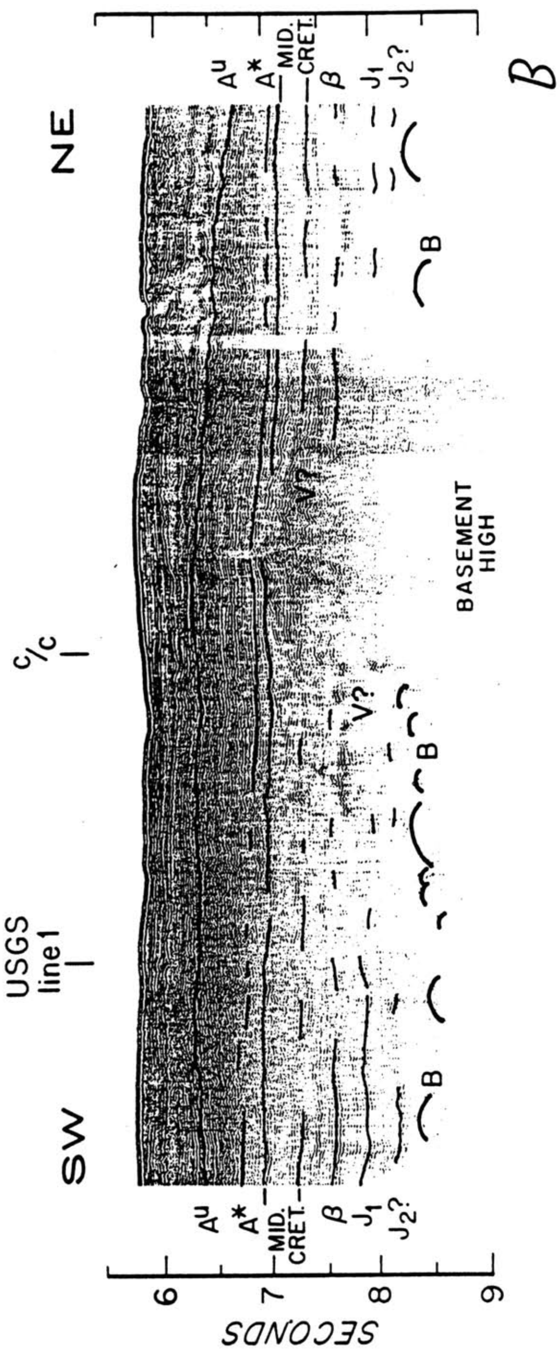


Figure 2
Basement high between Picket and Balanus Seamounts; Profile B. See Fig. 1 for location. Single-channel ship-board record shot using four 466 cu. in. air guns and bandpass filtered at 80-170 Hz. C/c indicates course change; V indicates possible volcanogenic horizons; B indicates hyperbolae characteristic of oceanic basement. Crossing with USGS line 1 is indicated. Vertical exaggeration of seafloor is ~17.

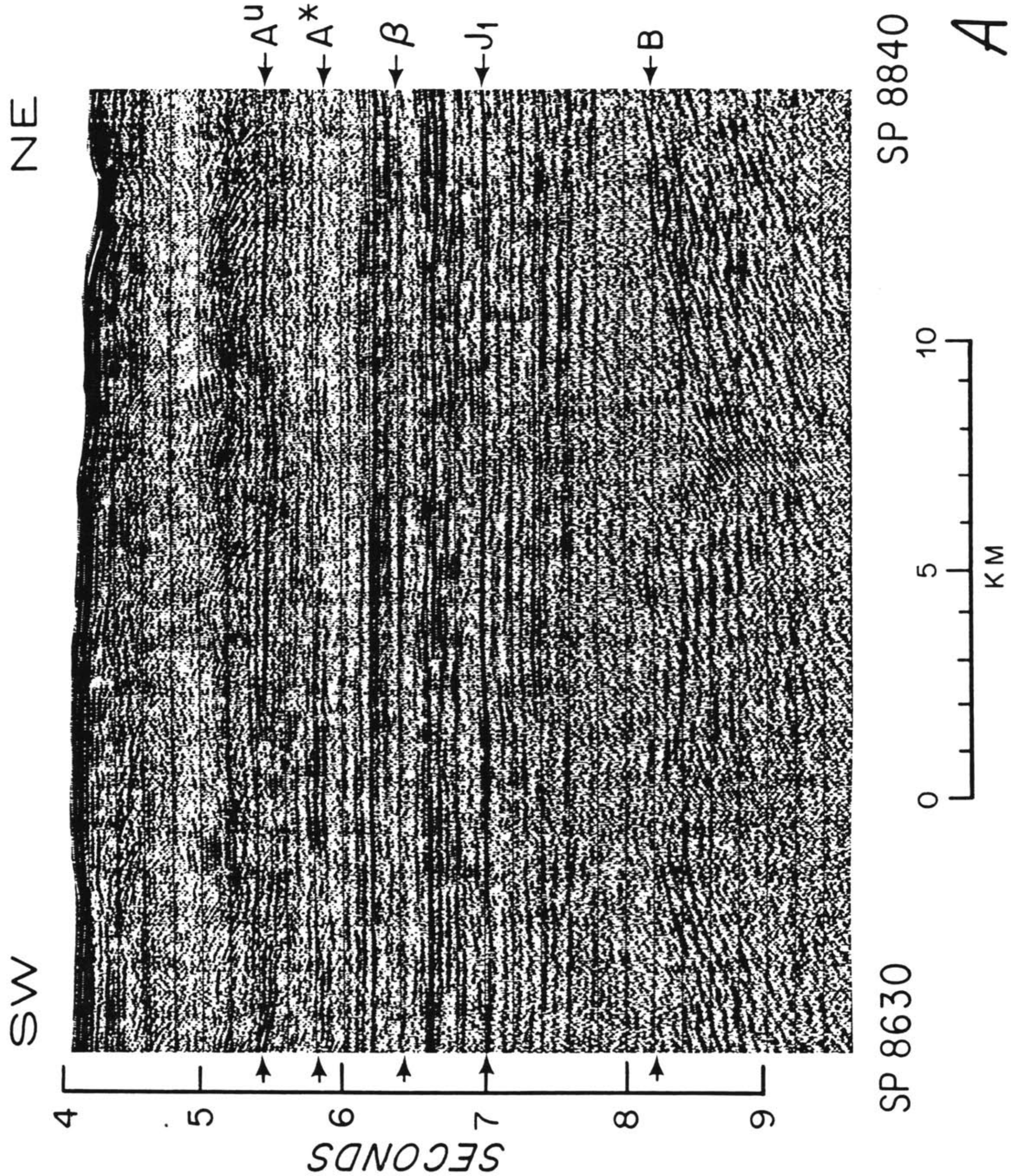


Figure 3

Processed multi-channel seismic sections illustrating similarity of stratigraphy across New England Seamounts. a) Profile A along USGS line 13 southwest of the New England Seamounts. b) Profile C along GSI line 108 to northeast. Profiles are ~210 km apart; locations in Fig. 1.

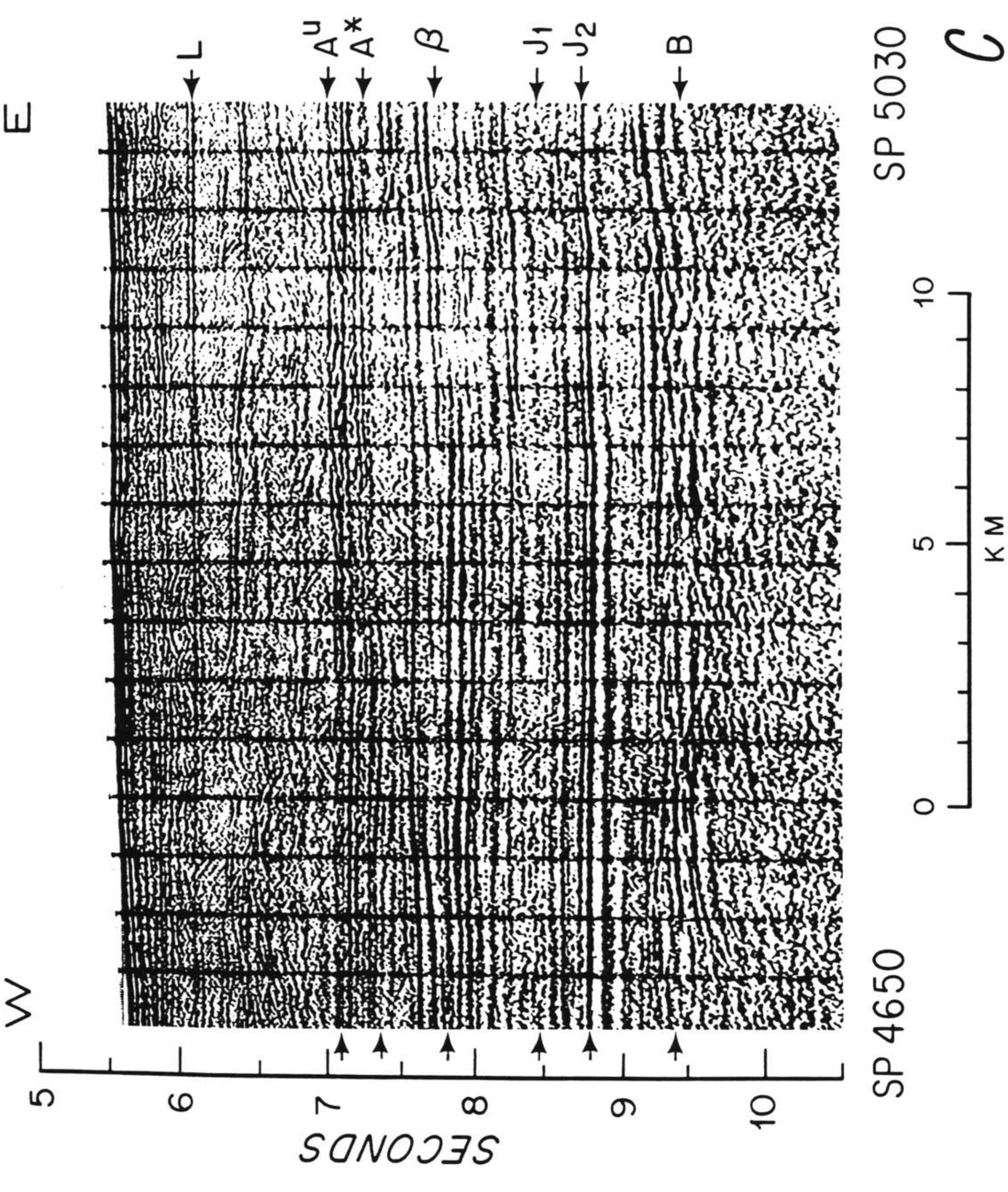


Figure 3b

traced as a horizon within a sequence of discontinuous, low-amplitude reflectors. Two unnamed seismic horizons occur between ß and A^* (Fig. 2 and 3) and are, therefore, middle Cretaceous in age. The deeper horizon is a discontinuous, sometimes high-amplitude, reflector at the base of a seismically incoherent layer. The shallower horizon is a prominent reflector. An intermittent reflector between basement and J1 may be J2.

Between Picket and Balanus Seamounts, reflectors deeper than the younger mid-Cretaceous reflector terminate against a basement high (Fig. 2). ß and the older mid-Cretaceous reflector can be extrapolated across the ridge; J1 is poorly constrained.

The basement high is a seismically incoherent zone capped by irregular, high-amplitude reflectors and hyperbolae (Fig. 2). The crest has a width of about 35 km and a minimum height above surrounding oceanic basement of 1.3-1.4 km (at 2 km/s). Stratigraphically, the crest reaches to the middle Cretaceous. A flat, high-amplitude reflector occurs at 7.7 s between J1 and ß on both sides, but most clearly to the southwest (Fig. 2). The reflector extends 5-6 km away from the high. High-amplitude hyperbolae (volcanics?) overlie ß on the northeast (Fig. 2). Above the high, A^* and the younger middle Cretaceous reflector occur 0.1-0.15 s shallower than they do to either side.

Seismic character of the middle Cretaceous to A^u section changes little across the high, while the post-A^u section changes significantly. Post-A^u reflectors southwest of the high are relatively coherent and flat; reflectors northeast of the high include irregular, dipping unconformities and cut-and-fill structures (Fig. 2).

SEISMIC STRATIGRAPHY OF THE SCOTIAN RISE

The basal seismic unit seaward of the diapirs consists of flat-lying, low-amplitude reflectors that fill basement lows. Conformably atop this unit is a high-amplitude, continuous couplet. Ebinger and Tucholke (in prep.) show that this couplet pinches out seaward within the JMQZ on crust of approximately Callovian-Oxfordian age and suggest a correlation to J2 (Fig. 3).

Overlying J2 is a sequence of flat, continuous reflectors of moderate to high-amplitude capped by J1, a high-amplitude, continuous reflector (Fig. 3). This sequence thins seaward and southwestward of Sable Island (~ 60°W) by downlap onto J2 and by internal pinchout. Ebinger and Tucholke (in prep.) confirm a late Kimmeridgian age for J1 by tracing it seaward to a basement pinchout near magnetic anomaly M25. This reflector was previously identified as ß by Wade (in Beaumont and others, 1982; picked assuming 2.5 km/s) and by Uchupi and Austin (1979). Overlying reflectors lap out onto J1 along line 108 for about 100 km northeastward and southwestward symmetrically about 58°W; elsewhere the overlying unit appears conformable.

In some areas, J1 and J2 are seismic unconformities as well as prominent reflectors. Klitgord and Grow (1980) noted that J1 and J2 appear conformable south of the seamounts. This discrepancy may be due to a difference in resolution of seismic data rather than to a real difference in stratigraphic relationship.

Above J1 lies a sequence of flat reflectors with few internal unconformities. Along line 108, amplitudes change over 10-15 km creating a discontinuous appearance. ß can be traced west of 58°30'W, in general, as a conformable upper sequence boundary and, rarely, as a distinct reflector. Beneath the Sohm Abyssal Plain, β becomes a flat, widespread reflector that pinches out on crust near magnetic anomaly M4, suggesting a Hauterivian age (Ebinger and Tucholke, in prep.). Between 62°W and 59°40'W, reflectors above β undulate with relief of 0.2-0.4 s over 10-20 km (Fig. 3) similar to channel cut-and-fill structure identified off New Jersey (Mountain and Tucholke 1979, their Fig. 14). Our correlations place β shallower than previous identifications. Horizon β identified by Jansa and Wade (1975, their Fig. 9) ties to a reflector at about 8.25 s in Fig. 3, mid-way between J1 and β.

West of 58°30'W, the seismic unit above β is characterized by an upward decrease in amplitude and continuity of reflectors. Reflectors are generally flat and parallel, but loss of definition makes it difficult to resolve terminations. Discontinuous reflectors grade upward into a thin, seismically incoherent unit 0.1-0.3 s thick (Fig. 3). Our correlation tentatively places A^* at the top of the incoherent layer, agreeing well with identification of A^* by Wade (in Beaumont et al., 1982; assuming 2.0 km/s).

West of 58°30'W, A^* and the incoherent layer are overlain by a thin (<0.3 s) sequence of flat, high-amplitude reflectors that are typically continuous for 10-20 km. Along line 37, the erosional unconformity at the top of this sequence correlates to A^u. From 66°W, at the median line of the New England Seamounts, to 63°W, A^u dips northeastward at the expense of the post-A^* section; the post-A^u section thickens in parallel, so there is no significant change in seafloor relief. A^* and the mid-Cretaceous reflectors dip northeastward more gently than A^u. Along line 108 east of 58°30'W, A^u truncates A^*, the incoherent layer, and ß, suggesting unconformable contact of Miocene(?) on Lower Cretaceous sediments beneath the Laurentian Cone; this relationship was unappreciated in earlier interpretations (eg. Uchupi and Austin, 1979). Our A^u agrees well with identifications by Emery et al. (1970), Jansa and Wade (1975), and Uchupi and Austin (1979), but lies slightly (0.3 s) below A^u identified by Wade (in Beaumont and others, 1982).

Two seismic sequences generally occur above A^u: a basal layer with abundant hyperbolae and short (1-5 km), discontinuous reflectors; and an upper sequence, commonly lacking hyperbolae, of incoherent layers and high-amplitude, semi-continuous (>10 km) reflectors. Along line 108 east of 61-62°W, these two sequences are separated by L, and the whole post-A^u section thickens from 0.5 to 2.5-3.0 s (Fig. 3b).

DISCUSSION

Age of the New England Seamounts

The seamount age-progression of Duncan (1984) predicts an age of about 100 Ma (Albian) for eruption of volcanics forming the basement high between Picket and Balanus Seamounts. This date coincides with the approximate age of sediments overlying the high as inferred from the seismic stratigraphy (ie., between the mid-Cretaceous reflectors). Thus, our data are consistent with a hot-spot origin for this feature.

Other evidence suggests, however, that eruptions occurred, here and elsewhere along the seamount trend, at times significantly different from those predicted by Duncan's age progression. On the continent, Foland and Faul (1977) noted episodic volcanism in the White Mountain magma series. Mountain and Tucholke (in press) found that ß and J1 onlap a basement high located along USGS line 13 (near Bear Seamount) about 125 km landward of our crossing (see Schlee and others, in press, their Fig. 13). They inferred that volcanism pre-dated the Kimmeridgian/Tithonian and, thus, the passage of the hot-spot by ~50 Ma. Schlee and others (in press) report that middle Cretaceous seismic units onlap a "volcaniclastic apron" surrounding nearby Bear Seamount and suggest that construction of the seamount terminated in the Early Cretaceous (Barremian-Hauterivian ?), ~20 Ma before Duncan's age date for Bear. Along line 37 southwest of the basement high (Fig. 2), a high-amplitude reflector lying below ß may have formed by extrusive processes at the end of the Jurassic or earliest Cretaceous: 30-40 Ma before passage of the hot-spot. Mid-Tertiary uplift or volcanism of Mytilus Seamount is required to place the summit in shallow water as suggested by Eocene coralline algae recovered from the seamount crest (Zeigler, 1955). This event is unrelated to hot-spot activity because the 40-50 Ma interval between hot-spot passage and uplift of Mytilus is much greater than characteristic periods of recurrent volcanism associated with most oceanic islands (ie. 5-20 myrs, Vogt and Tucholke, 1979). A reheating event is consistent with the uplift of middle and Upper Cretaceous reflectors above the basement high (Fig. 2), although the effect of compaction of pre-middle Cretaceous sediments in the basins to either side may be important. In conclusion, we suggest that episodic volcanism along a crustal "zone of weakness" must be considered a viable hypothesis <u>in addition to</u> hot spot volcanism for the origin of New England Seamounts.

Oceanographic Barriers

Seismic stratigraphy northeast of the New England Seamounts is broadly similar to that southwest of the chain. We infer that the seamounts posed no barrier along the continental rise to cause differentiation in development of stratigraphic sequences within the western North Atlantic during the Cretaceous and early Tertiary.

Our data indicate, however, that after the Eocene sedimentation near the chain differed from that in the Sohm basin. A^u cuts stratigraphically deeper northeastward to ~200 km from the crest of the chain, while above A^u over a similar distance, less sediment accumulated near the seamounts than in the basins (see also Mountain and Tucholke, in press, their Fig. 8-23). These seismic stratigraphic relationships suggest that near the seamounts erosion was less during the formation of A^u and that erosion (or non-deposition) increased after A^u, although we cannot rule out spatial variations in rate of terrigenous sediment supply. Since the current that cut A^u probably flowed in only one direction (southwestward, Miller and Tucholke, 1983), the symmetric sedimentation pattern is not due to downstream effects caused by the seamounts themselves.

Based on the broad region of the rise affected (~500 km along contours), we speculate that the velocity of southwest-flowing current eroding A^u was reduced by interaction with eastward flow driven by the Gulf Stream (Richardson, 1985), the northern edge of which crosses the continental rise south of Georges Bank (Fisher, 1977). We also suggest that, as abyssal current strength waned (Miller and Tucholke, 1983), a change in the interaction beneath the Gulf Stream reversed the net effect on bottom sediments. Erosion or non-deposition became more common on the continental rise beneath the unsteady northern boundary of the Gulf Stream (Hollister and McCave, 1984), and particularly beneath the region of high surface kinetic energy above the New England Seamounts (Richardson, 1983). In this view, post-Eocene sedimentation patterns can be attributed to the New England Seamounts only to the extent that they affect the stability of the Gulf Stream.

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Chapter 4

CENOZOIC DEVELOPMENT OF THE OUTER CONTINENTAL MARGIN OFF WESTERN NOVA SCOTIA

Stephen A. Swift

ABSTRACT

The outer continental margin of Nova Scotia is divided by a diapir province, 40-110 km wide and ~1000km long, that trends subparallel to the shelf edge along the upper continental rise and slope. The growth pattern for a small region of this margin during the Late Cretaceous and Cenozoic was studied using seismic stratigraphy and well data. Structure maps show that a steep continental slope existed landward of the diapir province (~2200-3800 mwater depth) from Early Cretaceous until Miocene time when onlapping upper rise sediments reduced the gradient. Shelf edge canyons were cut during the late Maestrichtian-early Paleocene, Eocene-Oligocene, and Pleistocene. Extensions of Tertiary canyons onto the slope are poorly defined, but small Paleocene fans of interbedded chalk and mudstone on the upper rise indicate that slope canyons existed at that time. Abyssal currents eroded the upper rise and smoothed relief on the continental slope in the Oligocene and middle(?) Miocene. In the Miocene, turbidites may have ponded on the upper rise landward of seafloor highs uplifted by salt ridges or pillows. Pliocene-Pleistocene deposits on the continental slope and upper rise are generally hemipelagic sediments or unchanneled turbidites. At the beginning and end of the Pleistocene, turbidity currents, caused by delivery of large sediment loads to the shelf edge by glaciers, eroded the present canyon morphology.

The late Cenozoic section of the lower continental rise thins seaward from ~2 km near the diapir province and rests on Horizon A^u, a prominent unconformity eroded during the Oligocene by abyssal currents. The morphology of the lower rise is largely due to construction by down-slope

deposits shed in the Miocene-Pliocene probably from uplift of the diapir province. Abyssal currents episodically eroded sediment, but current controlled deposition formed only a thin (<300 m) deposit in the Pliocene(?). Uplift in the diapir province accelerated during the Pleistocene and olistostromes up to 300 m thick were shed onto the lower rise. Late Pleistocene turbidity currents built leveed channels while contour currents reworked but did not significantly redistribute these deposits.

INTRODUCTION

Off western Nova Scotia, shelf sediments overlying the LaHave platform (Fig. 1) thicken seaward to ~6 km at the basement hinge-line near the present shelf edge (Wade, 1977; Schlee and Jansa, 1981). Seaward of the hinge-line, a 40 km wide sediment basin, informally termed the Schubenacadie basin, lies between an Upper Jurassic-Lower Cretaceous carbonate platform beneath the continental slope and a diapir province beneath the upper rise (Fig. 2). Rifted continental crust in the basin is overlain by 8-10 km of sediment (Keen et al., 1975; Keen and Cordsen, 1981; Schlee and Jansa, 1981; Beaumont et al., 1982). Deep salt(?) structures there uplift pre-Upper Cretaceous sediments (Jansa and Wade, 1975; Ellis et al., 1985).

Seaward on the upper rise, diapirs deform sediments in a 40-110 km wide province (Fig. 1; Emery et al., 1970; Jansa and Wade, 1975; Parsons, 1975; Uchupi et al., 1977; Wade, 1981). The seaward edge of the diapirs coincides with the East Coast Magnetic Anomaly, which may mark the landward edge of oceanic crust (Keen et al., 1975). In the study area between 61°-64°W,

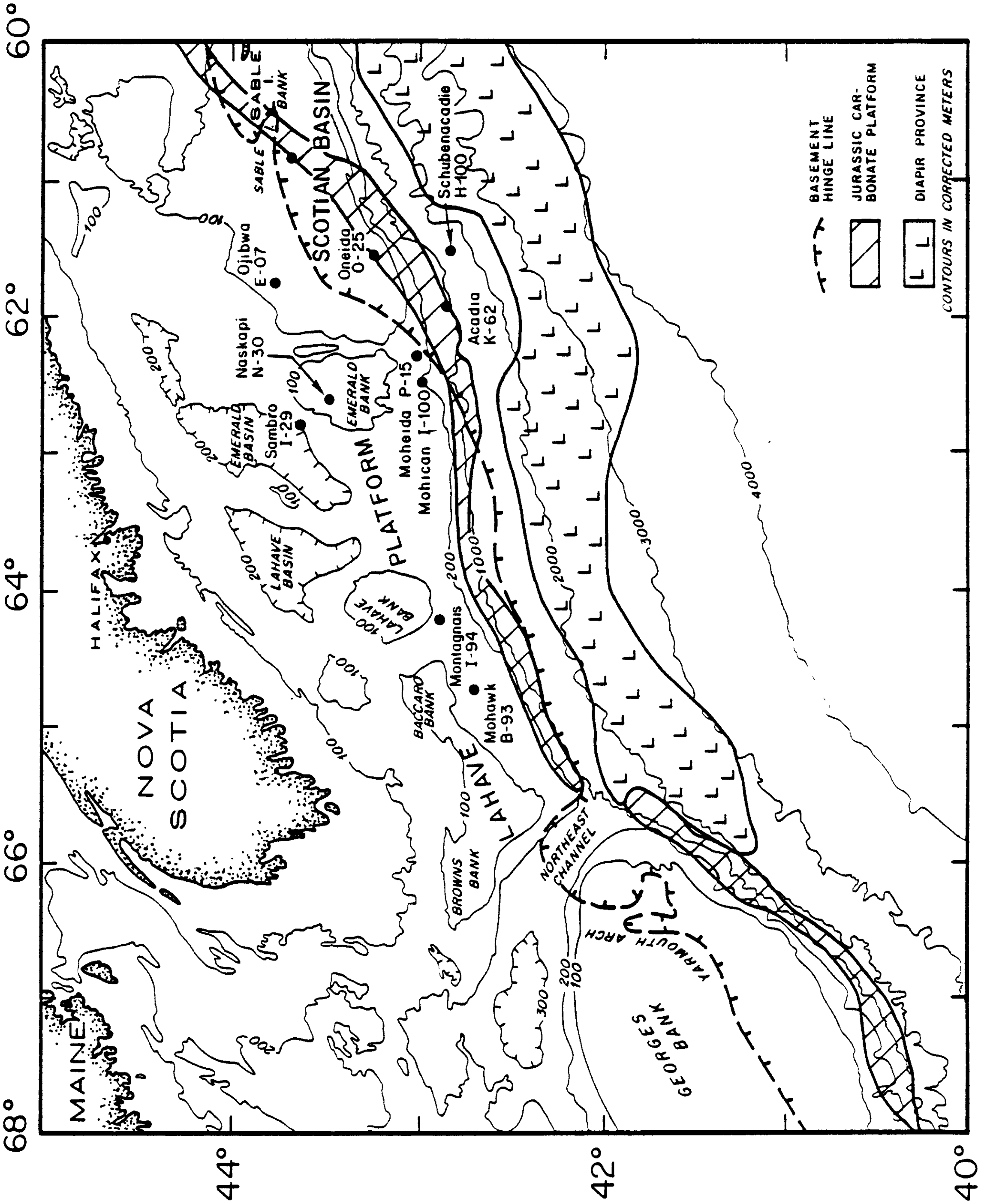


Figure 1 Physiographic and tectonic features of the western Nova Scotian continental margin. Bathymetry taken from Shor (1984a, b). Location of basement hinge-line, the edge of the Late Jurassic-Neocomian carbonate platform, and the diapir province taken from Uchupi (1984a, b). Dots mark well locations.

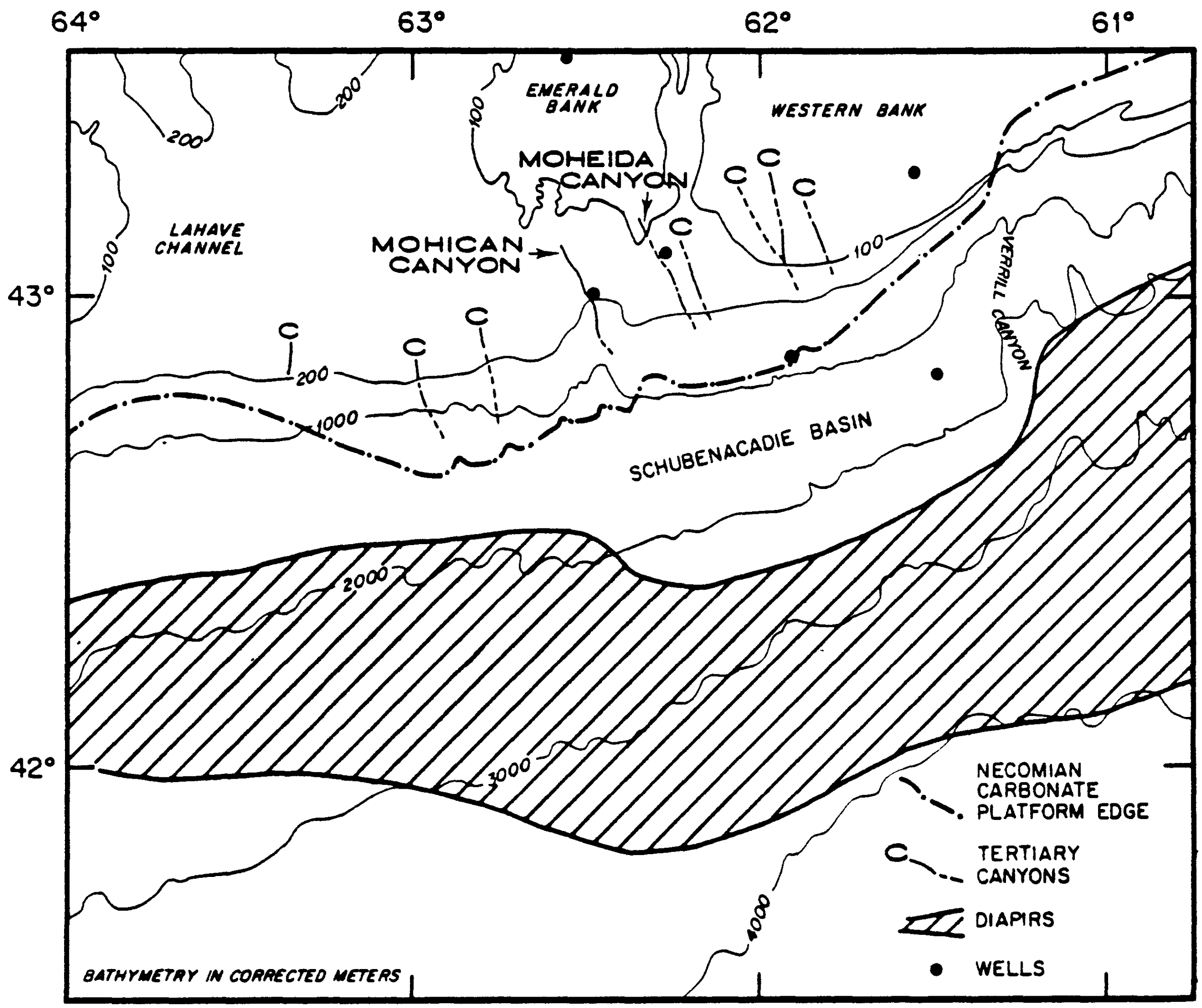


Figure 2 Tectonic and geologic structures in study area. Location of diapir province modified from Wade (unpublished data) and Uchupi (1984a, b). Seaward edge of province is sharp; landward edge is poorly defined. Deeply buried swells or pillows occur within Schubenacadie basin. Locations of Tertiary canyons and carbonate platform edge from this study. Dots indicate well locations. LaHave Channel is informal name for shallow depression west of Emerald Bank.

closely-spaced diapiric stocks 5-10 km in diameter may be rooted in deep, linear salt ridges 5-10 km wide and 70-120 km long (Wade, 1981; Emery and Uchupi, 1984). Fault deformation above these stocks extends to the seafloor (Fig. 3; Jansa and Wade, 1975; Parsons, 1975). Jansa and Wade (1975) and Parsons (1975) ascribe deformation to both shale flowage and halokinesis of the Upper Triassic-Lower Jurassic Argo Formation. Movement may have begun as early as Jurassic (Jansa and Wade, 1975; Smith, 1975), but deformation of the present seafloor indicates significant late Cenozoic activity (Webb, 1973; Jansa and Wade, 1975; Parsons, 1975; Ebinger and Tucholke, in prep.).

On the lower continental rise, Upper Cenozoic sediments form a wedge up to 2 km thick extending from the diapir province seaward about 300 km (Emery et al., 1970 and Jansa and Wade, 1975). The wedge overlies Horizon A^u (Swift et al., in press), a prominent western North Atlantic seismic reflector and unconformity eroded by abyssal currents between late Eocene and early Miocene (Tucholke and Mountain, 1979). The section above Horizon A^u accounts for about a third of the sediment overlying oceanic crust (Wade, 1981; Ebinger and Tucholke, in preparation).

General features of the geologic evolution of western Nova Scotia have been described (Jansa and Wade, 1975; King and MacClean, 1976; Given, 1977; Wade, 1977; Uchupi et al., 1977; Austin et al., 1980; and Emery and Uchupi, 1984), but detailed studies have not been published. The purpose of this paper is to describe the Cenozoic development of a relatively small, topographically smooth portion of the margin from the outer shelf to lower rise with the emphasis on interpreting sedimentary processes.

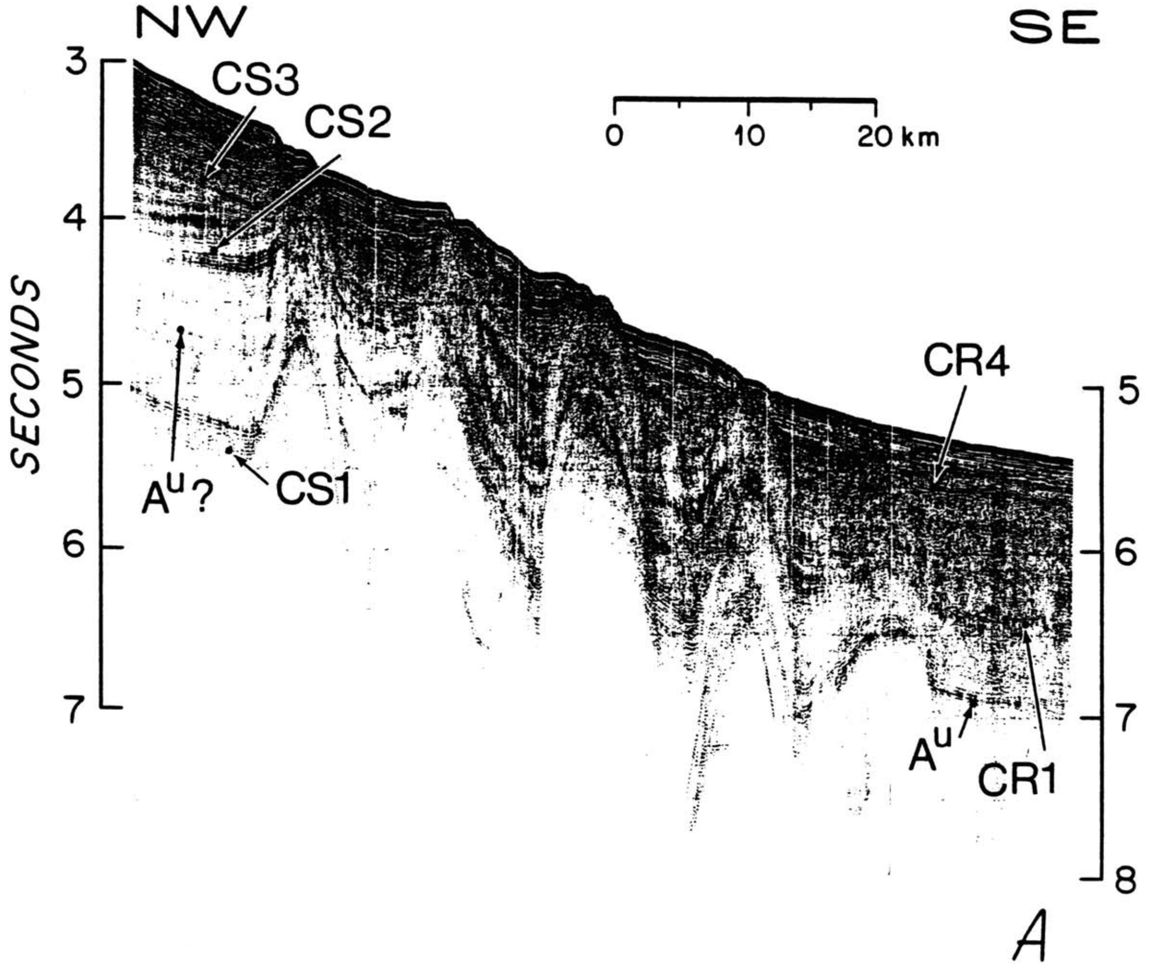


Figure 3 Single-channel airgun profile (unprocessed shipboard record at 80-170 Hz) across diapir province. Location in Fig. 5 (Profile A). Vertical exaggeration of seafloor about 16. Horizon A^u seaward of diapirs is from Swift et al. (in press), and landward of diapirs from correlation to the Schubenacadie H-100 well. To the northwest, numbers indicate slope horizons, while to the southeast, they indicate rise horizons. Note flat reflectors between Horizons CS2 and CS3.

This interpretation is based on seismic stratigraphy and, for the shelf and slope, also on well stratigraphy. Stratigraphic control on the lower rise is poor because there are no boreholes and because seismic horizons cannot be correlated unambiguously from shelf and slope wells through the diapir province (Swift et al., in press). Seismic character of sedimentary sequences, structure contours on seismic horizons, and isopach maps were used to interpret the geologic development of the margin. No corrections were applied in structure and isopach maps for sediment loading and compaction. Because sediment deformation within the diapir province prevents direct correlation of seismic stratigraphy, horizons choosen for mapping on either side of the diapir province were selected independently. In general, horizons selected are seismic disconformities where a significant change in sedimentation pattern occurred. These are indicated by numerous, angular reflector terminations (seismic unconformities) or by major changes in seismic character.

Three seismic horizons were identified in the Schubenacadie basin and mapped landward onto the shelf (Fig. 4). The deepest horizon (CS1) is a relatively continuous, high-amplitude reflector within the Upper Cretaceous. The upper two horizons (CS2 and CS3) are conspicuous, widespread angular seismic unconformities within the Tertiary section that, respectively, mark a significant change in deposition and an erosional episode related to uplift in the diapir province. Prominent Pleistocene erosional unconformities are not mapped because reflection profiles in this data set are not spaced closely enough to resolve adequately the relief on these surfaces.

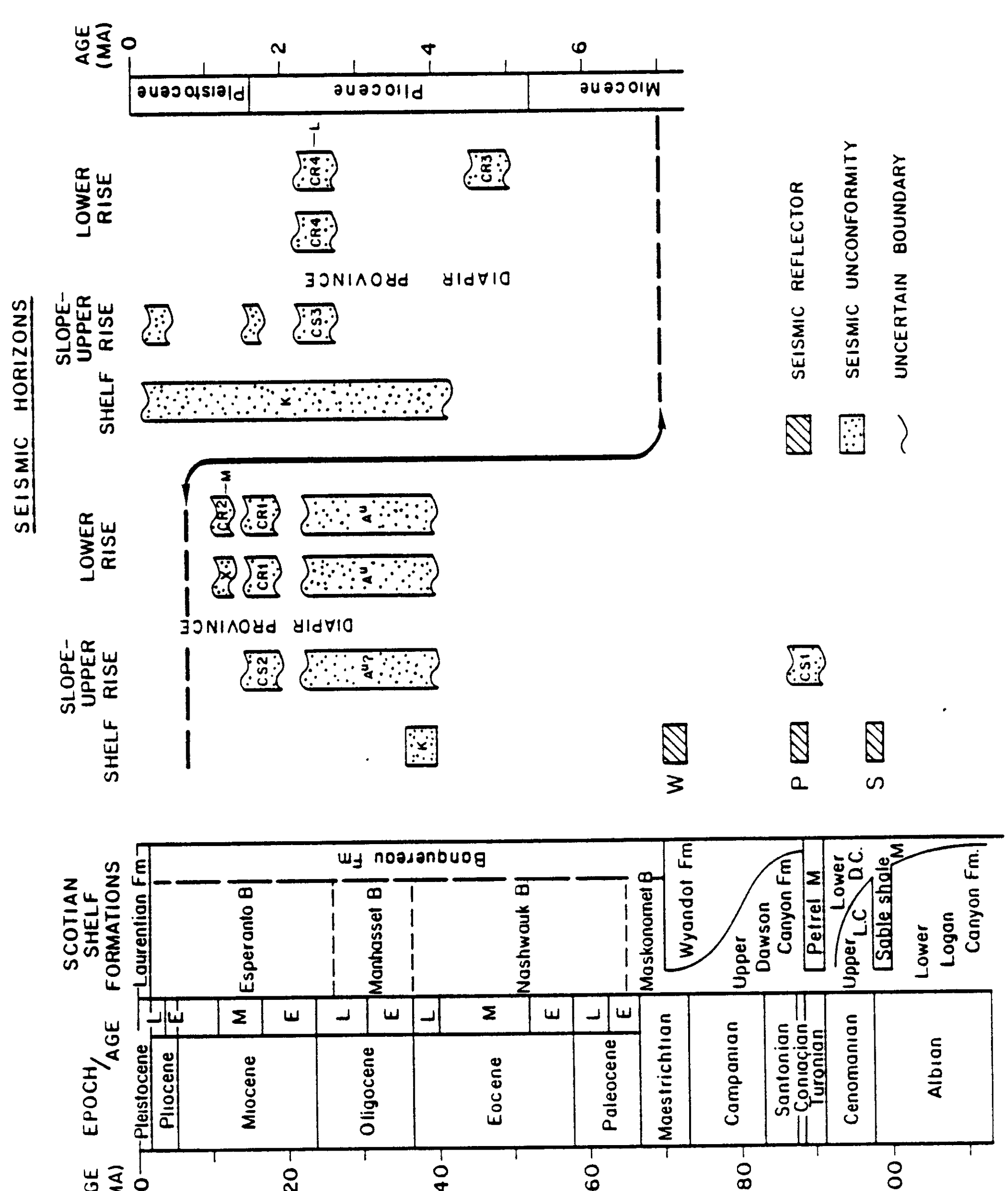


Figure 4 Correlation diagram of Late Cretaceous to Cenozoic seismic stratigraphy for western Nova Scotia continental margin. Expanded time scale for Pliocene-Pleistocene is shown on the right. Age of formations from Jansa and Wade (1975), Hardy (1975), Given (1977), and Barss et al. (1979). K indicates seismic horizons from King et al. (1974). Horizons W, P, and S taken from Jansa and Wade (1975). Horizon A^u is from Tucholke and Mountain (1979). M indicates age of Horizon Merlin (Mountain and Tucholke, in press); L indicates age of Horizon L (Uchupi and Austin, 1979; Normark et al., 1983). Length of bars indicates inferred temporal range of seismic horizons. Time scale from Berggren et al. (in press, Cenozoic) and Harland et al. (1982, Cretaceous).

Seaward of the diapirs, the seismic section above Horizon A^u was divided into four layers separated by three prominent seismic unconformities (CR1, CR3, and CR4; Fig. 4). Although additional less prominent seismic unconformities could be distinguished within these layers, they were not mapped. Two of these (Horizons X and CR2) are also discussed.

DATA

Landward of the diapir province, seismic data studied include industry multi-channel profiles shot and processed in the early 1970's and single-channel profiles shot in 1983 on the R/V ROBERT CONRAD (cruise RC2408; Fig. 5). Additional profiles and line drawings from Emery et al. (1970), Bhat et al. (1975), Jansa and Wade (1975), Given (1977), and Jansa (1981) were studied. Age and lithologic data for six wells drilled in the region (Fig. 5) were obtained from Ascoli (1976), Barss et al. (1979), Poag (1982), and unpublished well history reports. Jansa and Wade (1975), Hardy (1975), Given (1977), and Eliuk (1978) provide additional descriptions of these wells. Seismic horizons in this area were dated primarily by ties to published biostratigraphy of the Oneida O-25 well (Ascoli, 1976; Barss et al., 1979) and to preliminary biostratigraphy of the Schubenacadie H-100 well (unpublished well history report; Fig. 6 and 7). Available biostratigraphy is inadequate to correlate seismic stratigraphy with higher order features in global sea level curves (Vail and Hardenbol, 1979; Watts and Thorne, 1984; Miller et al., 1984). The time scales of Berggren et al. (in press; Cenozoic) and Harland et al. (1982; Cretaceous) were used. Seismic travel time was converted to depth using analytical functions

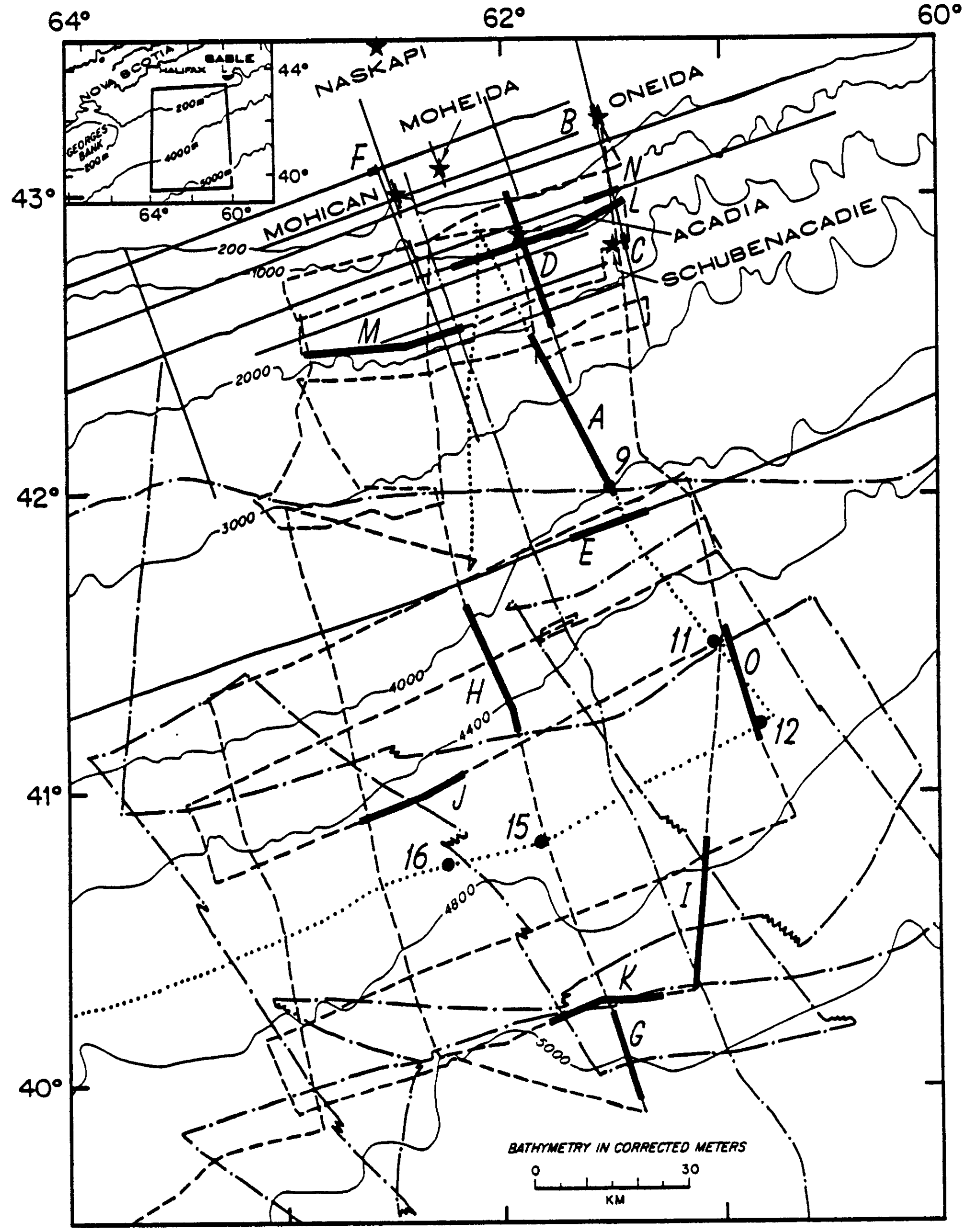


Figure 5 Location of seismic profiles and wells (stars) used in compiling structure and isopach maps. Heavy lines indicate illustrated profiles. Numbered heavy dots indicate launch positions of sonobuoys. Broken lines indicate data collected on RC2408 using either four 466 cu. in. airguns (dots) or an 80 cu. in. water gun (dashes). Single-channel profiles collected on CHAIN 70 and RC2209 are indicated by dot-dash lines. Solid lines are 24-fold stacked profiles shot by industry. Bathymetry from Swift (1985).

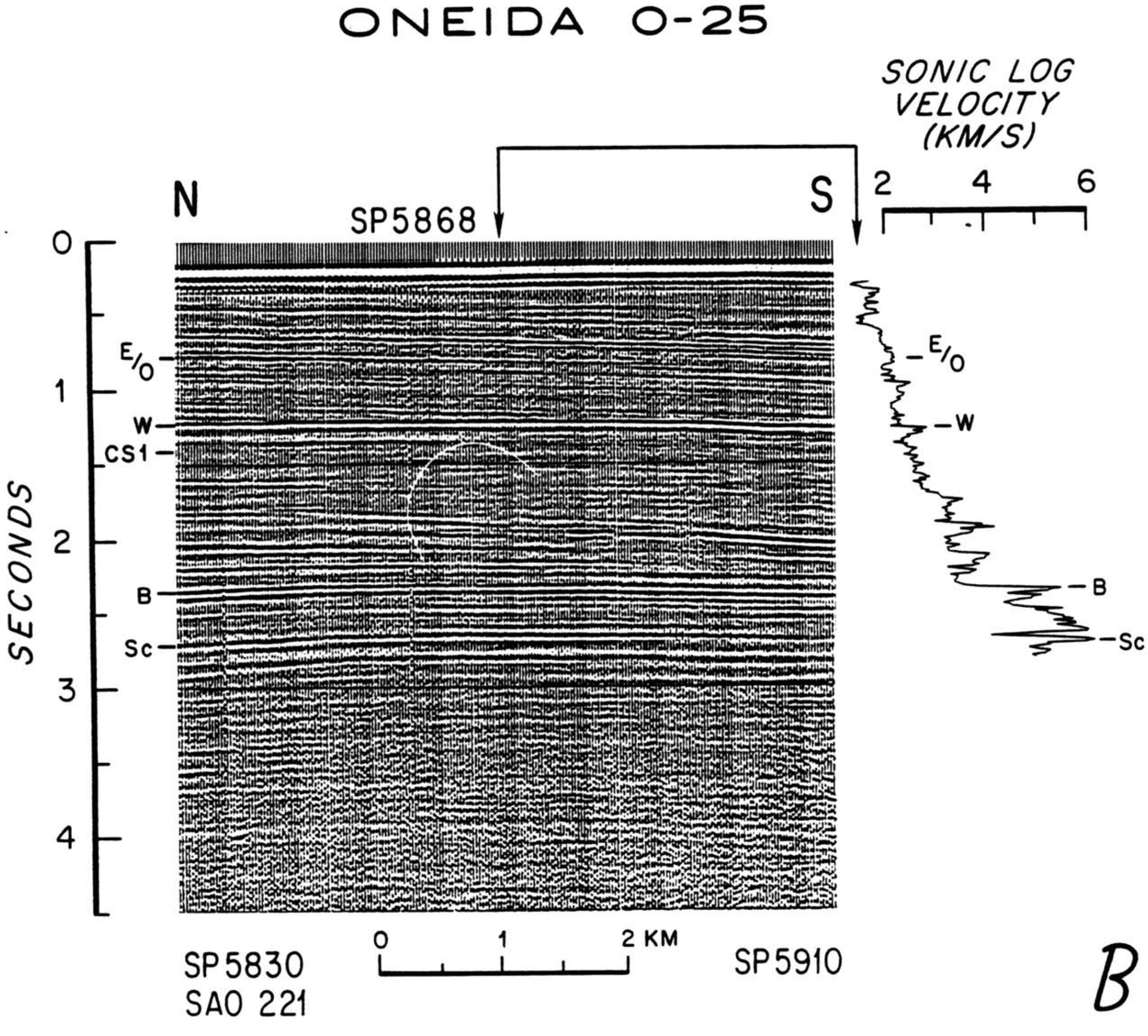


Figure 6 Profiles across wells used for stratigraphic ties. Locations in Fig. 5. Stratigraphic sections in Fig. 7. (a) Oneida O-25 (Profile B, water depth= 56 m, TD= 4084 m below sealevel). Formation tops are Wyandot chalk (W), Baccaro limestone (B), and Scatarie limestone (Sc). Sonic velocity calculated by digitizing the integrated time marks on analog sonic log record every 5 msec to the nearest foot. (b) Schubenacadie H-100 (Profile C, water depth= 1477 m, TD= 4176 m below sealevel). Sonic log velocity calculated by digitizing every one msec to the nearest 0.1 m.

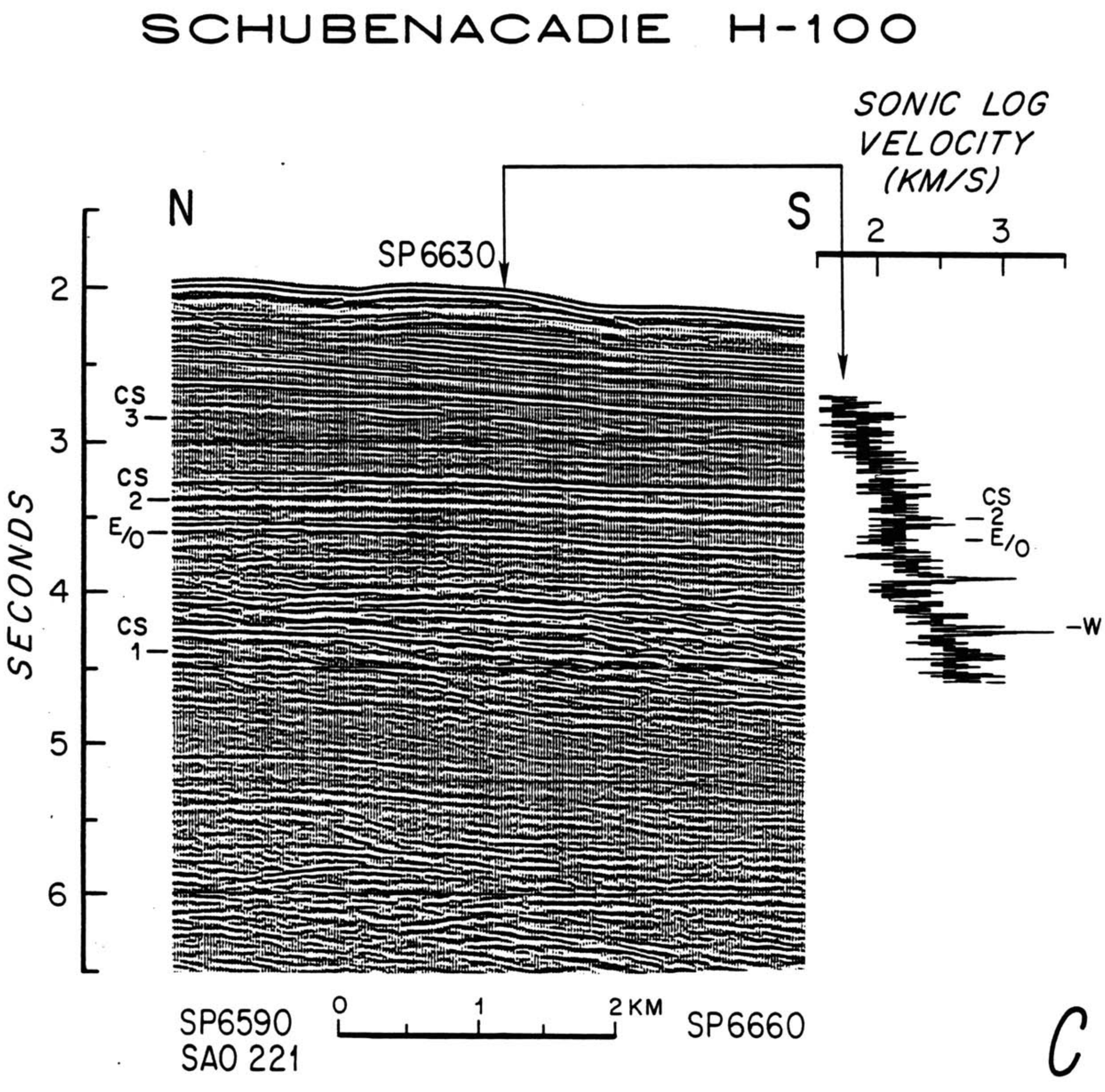

SCHUBENACADIE H-100
SONIC LOG
VELOCITY
(KM/S)
N
S
SP6630
2
3
SECONDS
2
3
4
5
6
CS
3
CS
2
E/O
CS
1
CS
2
E/O
W
SP6590
SAO 221
0
1
2 KM
SP6660
C

Figure 6b

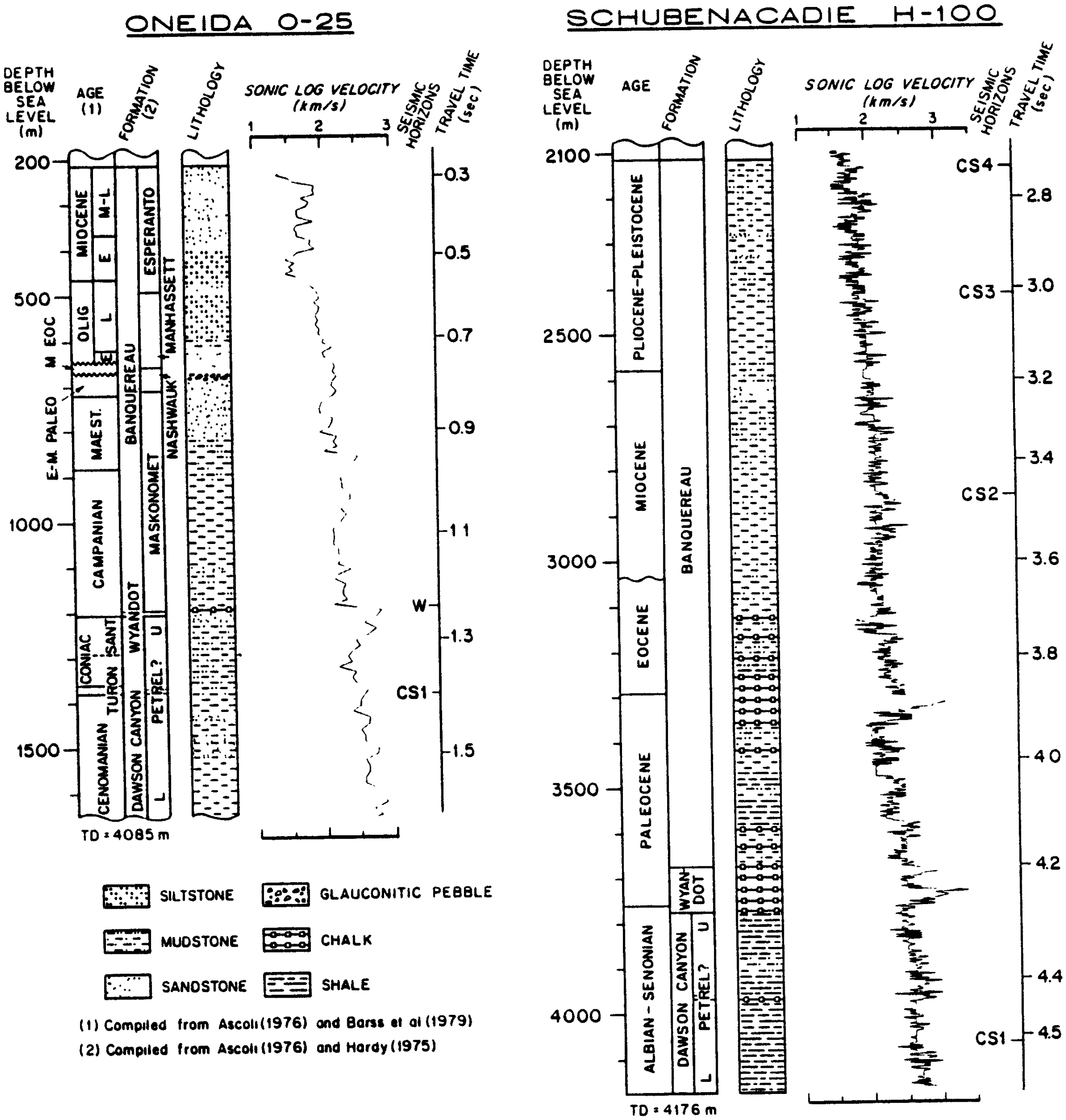


Figure 7 Stratigraphy of the Oneida O-25 and Schubenacadie H-100 wells. Depth to horizons calculated using functions in Table 1. Age in the Oneida O-25 well from Ascoli (1976) and Barss et al. (1979), lithology from well history report, and formations from Hardy (1975) and Ascoli (1976). Age and lithology in the Schubenacadie H-100 well from well history report; formations assigned on the basis of lithology.

derived from sonic logs made at the Oneida O-25 and Schubenacadie H-100 wells. Straight lines were fitted to sonic velocity vs. travel time plots, and a function for depth below seafloor was derived by integration (Table 1; Swift, 1986).

Seaward of the diapirs, interpretations are based primarily on unprocessed single-channel seismic profiles collected on cruise RC2408 (Fig. 5). These data were supplemented with an industry multi-channel profile (GSI 108) and single-channel profiles shot on CHAIN 70 (line 259, Emery et al., 1970) and on the R/V ROBERT CONRAD in 1978 (RC2209). Five expendable sonobuoys shot on RC2408 (Fig. 5) were analyzed for seismic velocity. Travel times of wide-angle reflections and refractions were simultaneously modeled using ray-tracing to give velocity-depth functions (Table 2; Swift, 1986). In three of five buoys, a velocity offset of 0.5-0.6 km/s occurred within 0.2 s of Horizon A^u. An offset of about 0.3 km/s occurred near a shallower reflector (CR1). The overall linear velocity gradient in the section above Horizon A^u was estimated from travel time plots and integrated to yield a function for depth (Table 1, Swift, 1986):

CRETACEOUS SEDIMENTARY FRAMEWORK

On the outer shelf off western Nova Scotia, a prominent seismic reflector ("C" of Given, 1977) marks the upward, Berriasian transition from carbonate platform (Baccaro Member of the Abenaki Formation) to clastic deposition (Missisauga and Verrill Canyon Formations) (Figs. 6, 8; Jansa and Wade, 1975; Given, 1977; Eliuk, 1978; Jansa, 1981; Ellis et al., 1985). Near the seaward edge of the platform, the reflector is generally flat but

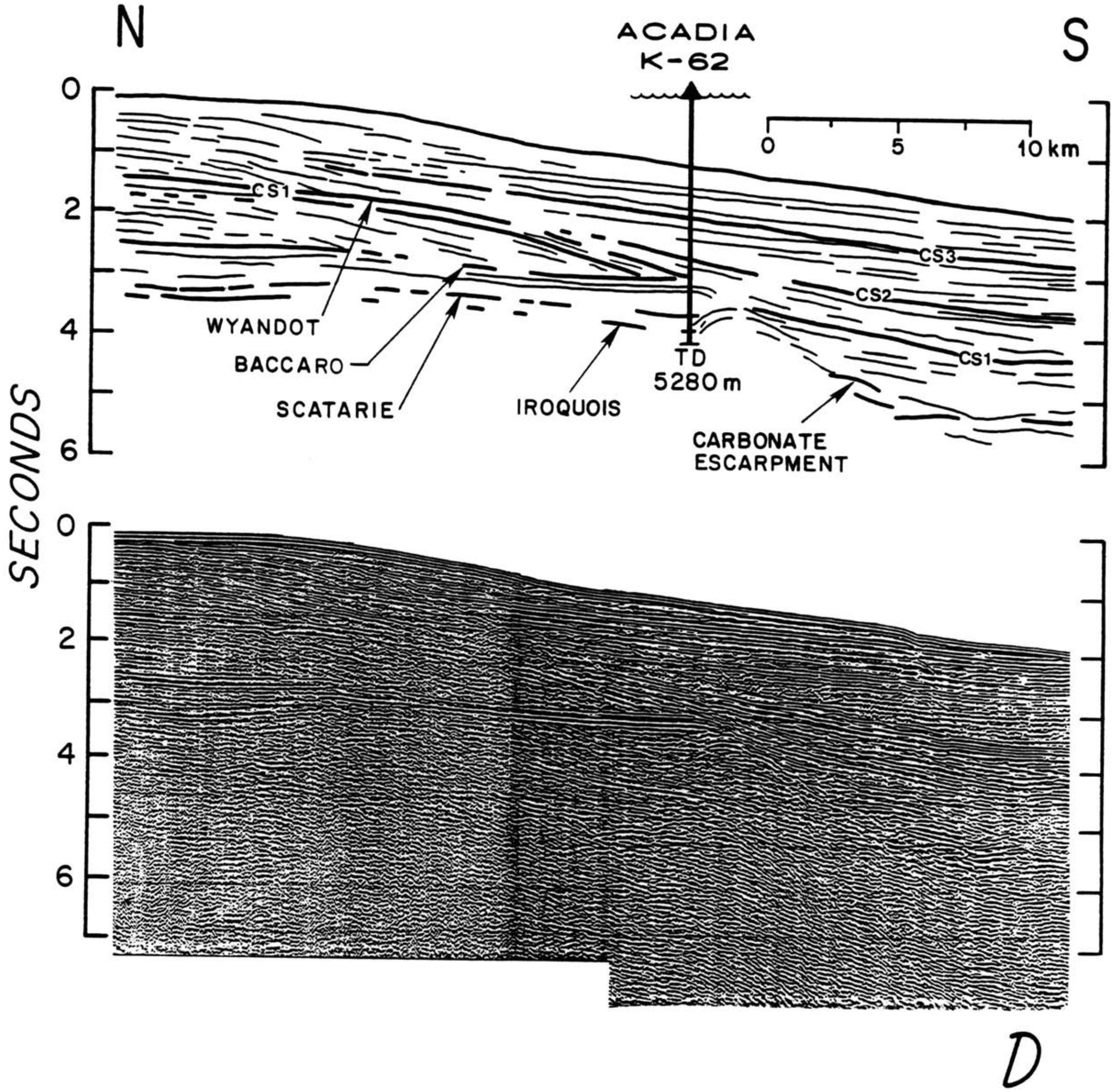


Figure 8 Multi-channel dip profile (line 212) across the Schubenacadie basin from shelf to diapir province. Location in Fig. 5 (Profile D). Vertical exaggeration of seafloor about 3. Nearby profiles are illustrated in Bhat et al. (1975, Fig. 8), Jansa and Wade (1975, Fig. 9), Given (1977, Fig. 2), Jansa (1981, Fig. 8), and Ellis et al. (1985, Fig. 5). Continental slope (CS) horizons, 1-3, discussed in text are shown. Tops of formations are traced from the Acadia K-62 well. (a) Continental slope (SAS 044). (b) Upper continental rise (SAW 054).

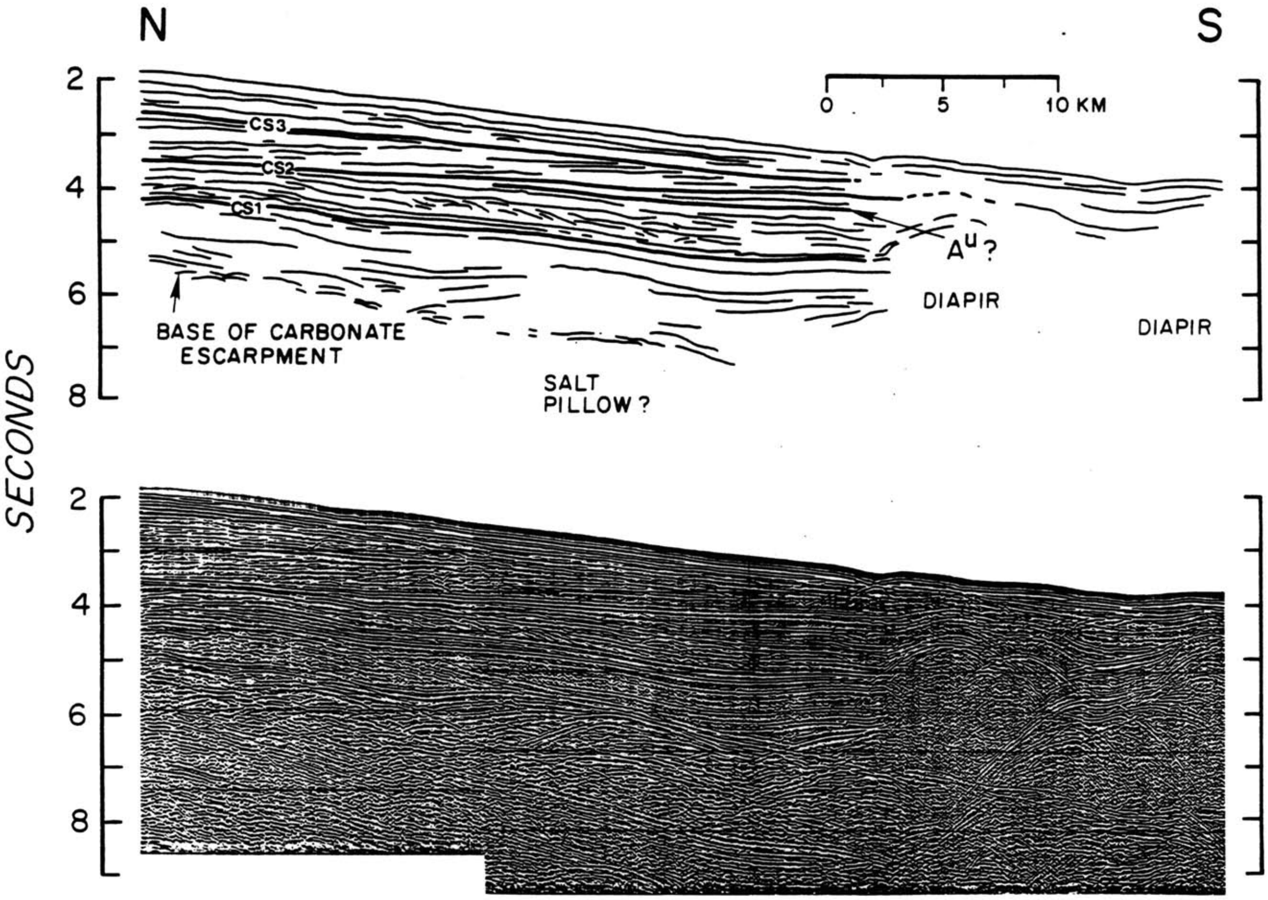

N
S
0
5
10 KM
CS3
CS2
CS1
A^u ?
DIAPIR
DIAPIR
BASE OF CARBONATE
ESCARPMENT
SALT
PILLOW ?
SECONDS
2
4
6
8
2
4
6
8

Figure 8b

is locally seaward dipping or faulted down-to-basin (Fig. 8). Eliuk (1978) interpreted local re-entrents in the platform edge (Fig. 2) as 'channels'. South of the carbonate platform, a series of shingled, high-amplitude reflectors plunge steeply seaward marking an escarpment with relief of 1.5-2.2 km (Fig. 8; Jansa, 1981; Ellis et al., 1985). These reflectors then flatten and continue seaward to the diapir province suggesting a carbonate talus slope.

Seaward of the diapir province (Fig. 9), Horizon ß, a reflector marking the upward transition from deep-sea carbonates to clastics in the Hauterivian-Barremian (Tucholke and Mountain, 1979; Mountain and Tucholke, in press), terminates against the seaward edge of the diapir province (Swift et al., in press). Sediment deformation by diapirs obscures the relationship between Horizon ß and the carbonate talus slope.

Isopachs for Lower Cretaceous sediments on the shelf (Given, 1977; Wade, 1981) show that the study area (~61°-64°W) is near the western limit of the Sable Island delta, that was deposited on the eastern Scotian shelf from the end of the Late Jurassic to the Albian (Jansa and Wade, 1975; Given, 1977). West of 61°W, reflectors of the Missisauga (Berriasian-Aptian) and Logan Canyon (Aptian-Cenomanian) Formations downlap seaward onto the Baccaro Member of the Abenaki Formation, while reflectors of equivalent age in the Schubenacadie basin (Verrill Canyon and Dawson Canyon Formations) lap onto the carbonate escarpment (Fig. 8). Thus, terrigenous clastics did not bury the carbonate platform edge during the Early Cretaceous regression as they did to the east beneath the Sable Island delta (Jansa and Wade, 1975; Given, 1977; Eliuk, 1978). Western shelf sources (Schubenacadie drainage system;

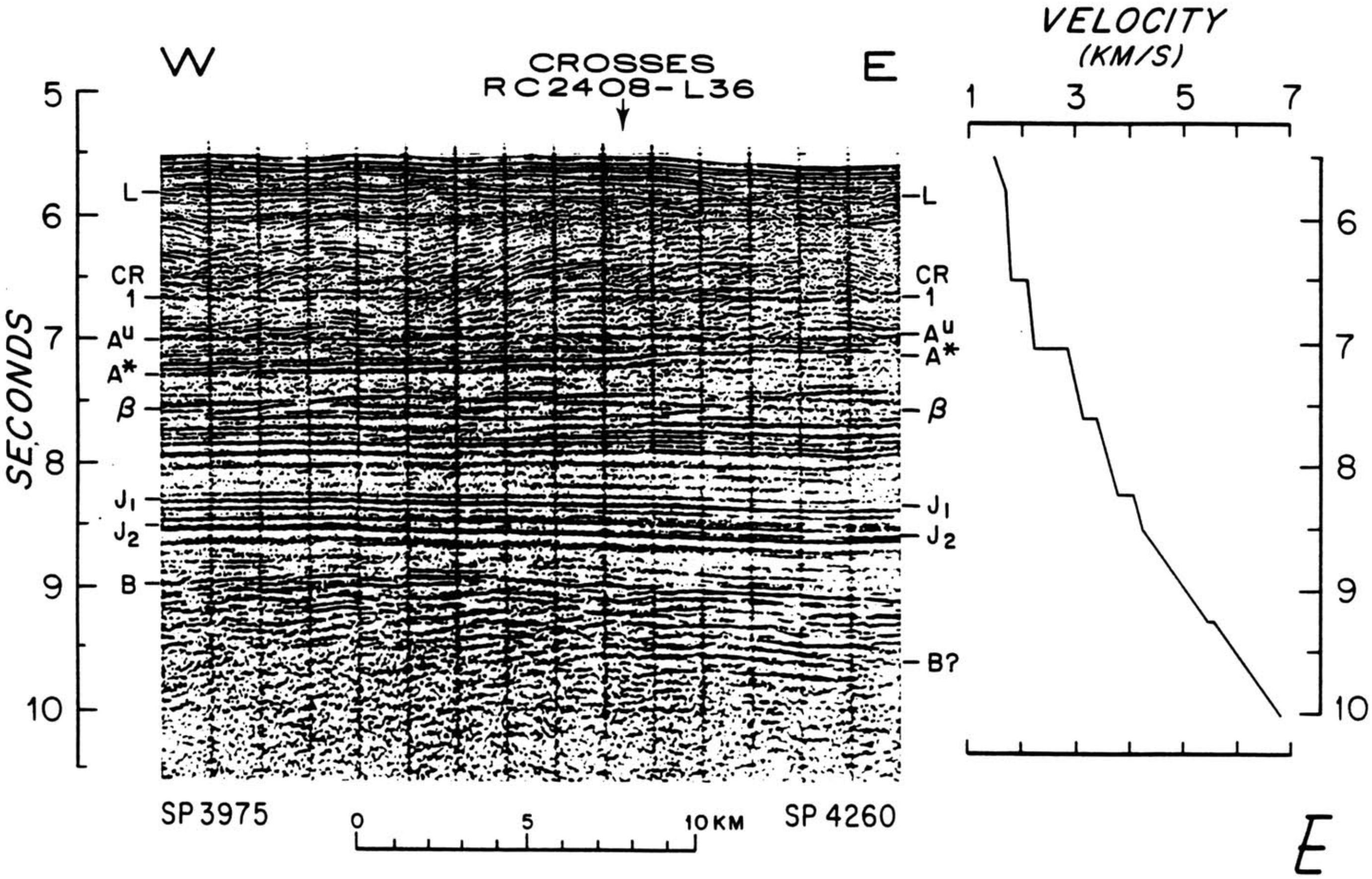


Figure 9 Multi-channel profile from GSI 108 just seaward of diapir province and velocity function from sonobuoy 9 shot on a crossing line (RC2408). Location in Fig. 5 (Profile E). Identification of seismic horizons is from Swift et al. (in press).

Given, 1977) were probably much less developed than those feeding the Sable Island delta. Down-slope processes probably carried terrigenous sediment to the upper rise through gaps in the carbonate platform (Eliuk, 1978). Erosion of the seaward edge of the platform (Eliuk, 1978) probably contributed shallow water carbonate detritus to the onlapping wedge.

Horizon CS1, correlated to the middle Upper Cretaceous (possibly the Turonian Petrel Member of Dawson Canyon Formation) at the Schubenacadie H-100 well (Fig. 7), occurs within a sedimentary sequence that fills a channel cut on the carbonate platform. Contours on depth to Horizon CS1 follow the trend of the platform edge (Fig. 10a). Thus, by early Late Cretaceous time the platform edge was not buried, and the steep slope of the escarpment controlled the location of the paleo-slope.

During the Late Cretaceous transgression on the Scotian shelf, deposition of inner neritic sandstones and shales of the Dawson Canyon Formation continued on the shelf (Jansa and Wade, 1975; Given, 1977), while mudstones and shales accumulated on the upper rise (Fig. 7). These sequences were interrupted by deposition of thin, neritic shale (Sable Shale Member of the Dawson Canyon Formation; Fig. 4) and outer neritic to upper bathyl carbonate units (Petrel Member and Wyandot Formation; Jansa and Wade, 1975; Given, 1977). The Wyandot Formation (Santonian-lower Maestrichtian), a nannofossil chalk interbedded with shale and mudstone, produces a high-amplitude, flat seismic marker on the shelf (Fig. 6), while Horizon A^*, the deep-sea stratigraphic equivalent in the western North Atlantic (Tucholke and Mountain, 1979; Tucholke, 1981), is a low-amplitude, often discontinuous reflector seaward of the diapirs (Fig. 9; Swift et al., in press).

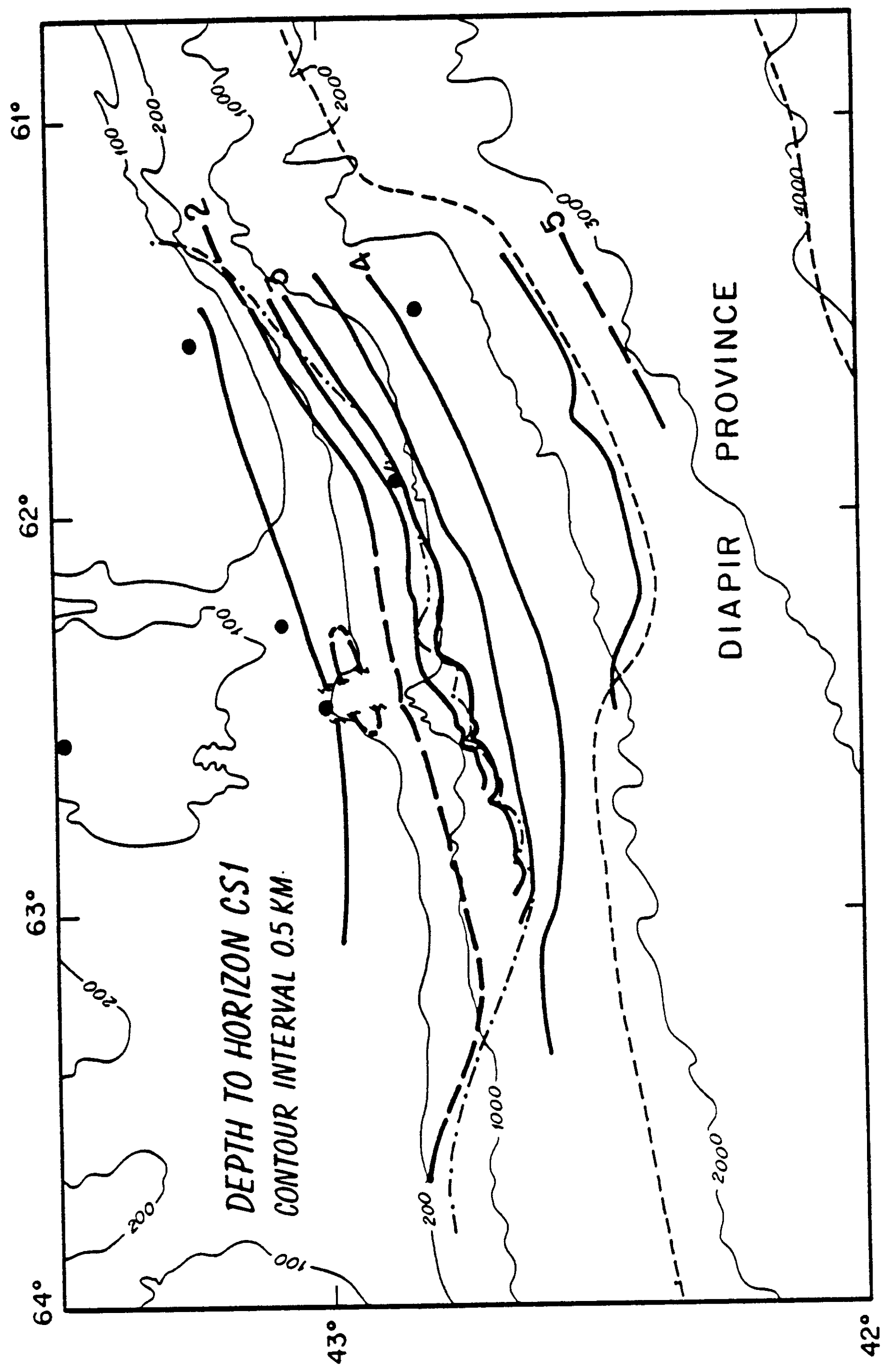


Figure 10 Depth (km) below sea level and isopachs (km) between seismic horizons on the continental slope and upper rise. (a) Horizon CS1 (Turonian?), (b) Horizon CS2 (lower Miocene?), (c) Horizon CS3 (Pliocene), (d) interval between Horizon CS1 and CS2, (e) interval between Horizons CS2 and CS3, (f) interval between Horizon CS3 and seafloor. Small dots in (f) are data points. Data coverage for other maps varies slightly. Dot-dashed line indicates edge of Neocomian carbonate platform. Dashed line indicates landward edge of diapirs. Heavy dots are well locations.

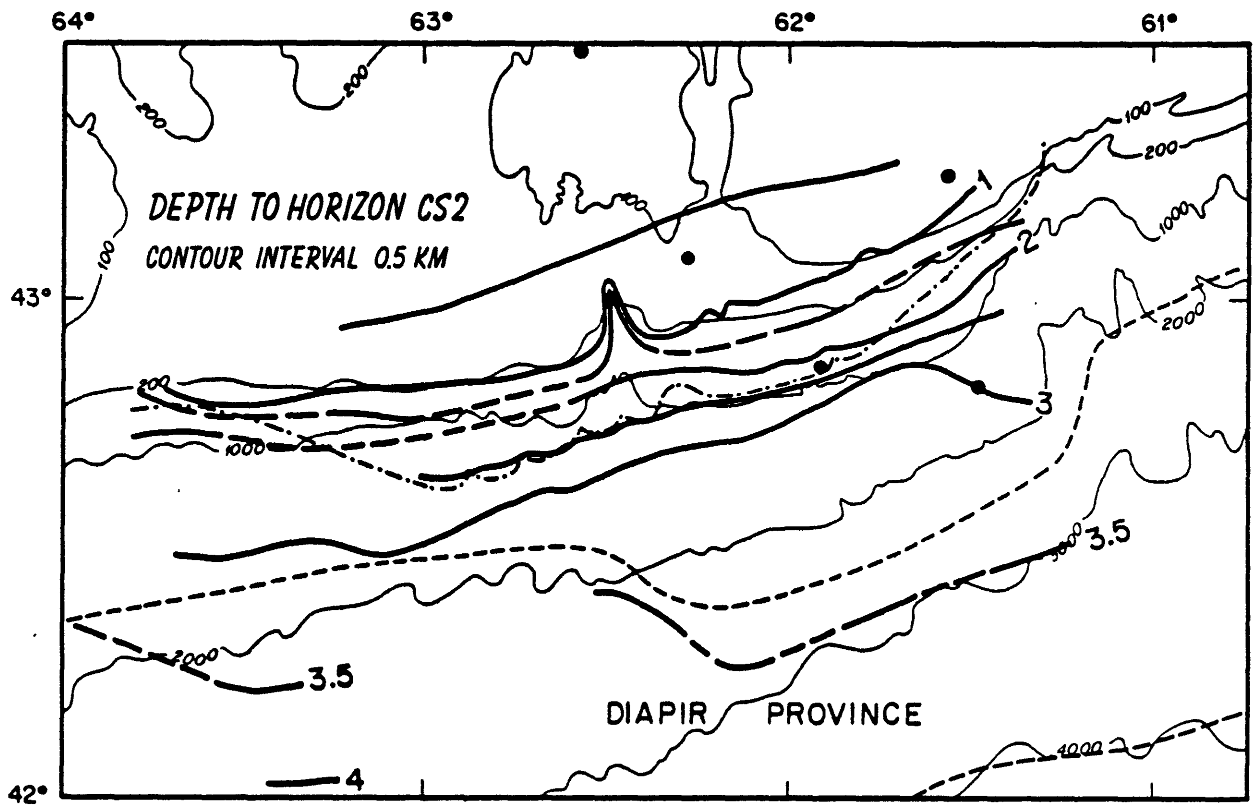

64°
63°
62°
61°
43°
42°
DEPTH TO HORIZON CS2
CONTOUR INTERVAL 0.5 KM
DIAPIR PROVINCE
1
2
3
3.5
3.5
4
100
200
1000
2000
4000

Figure 10b

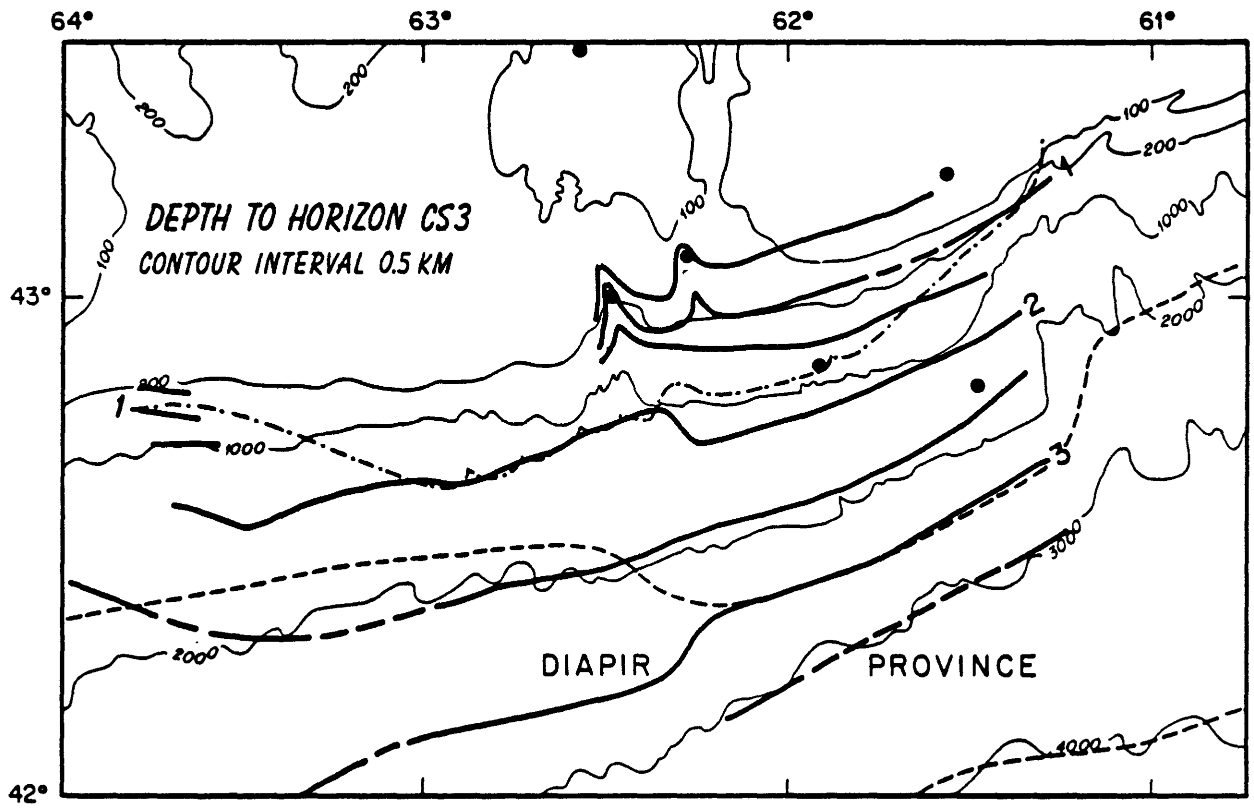

64°
63°
62°
61°
43°
42°
DEPTH TO HORIZON CS3
CONTOUR INTERVAL 0.5 KM
DIAPIR
PROVINCE
100
200
1000
2000
3000
4000
1
2
3

Figure 10c

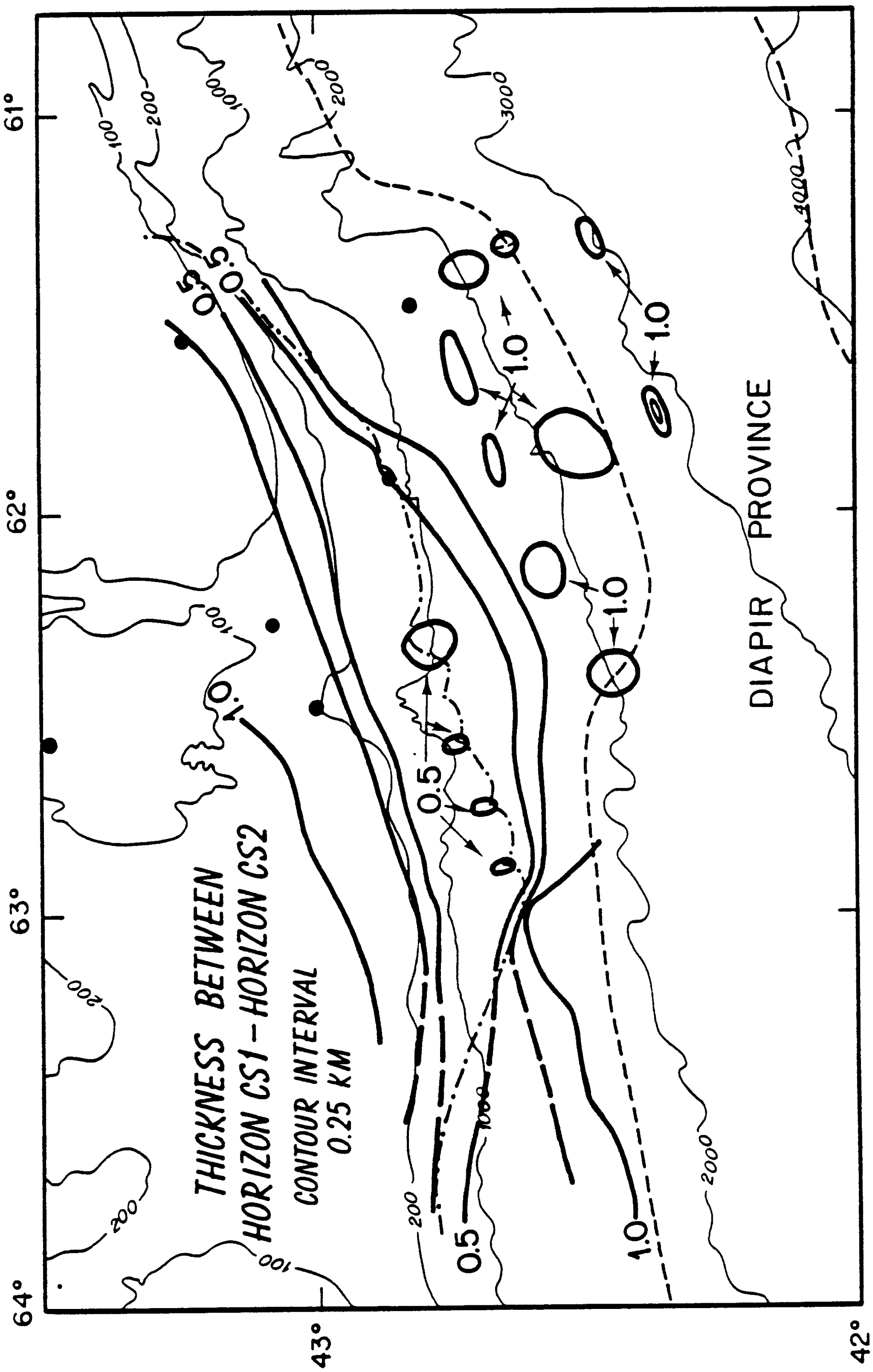

THICKNESS BETWEEN
HORIZON CS1 – HORIZON CS2
CONTOUR INTERVAL
0.25 KM
DIAPIR PROVINCE
61°
62°
63°
64°
43°
42°
0.5
1.0
100
200
1000
2000
3000
4000

Figure 10d

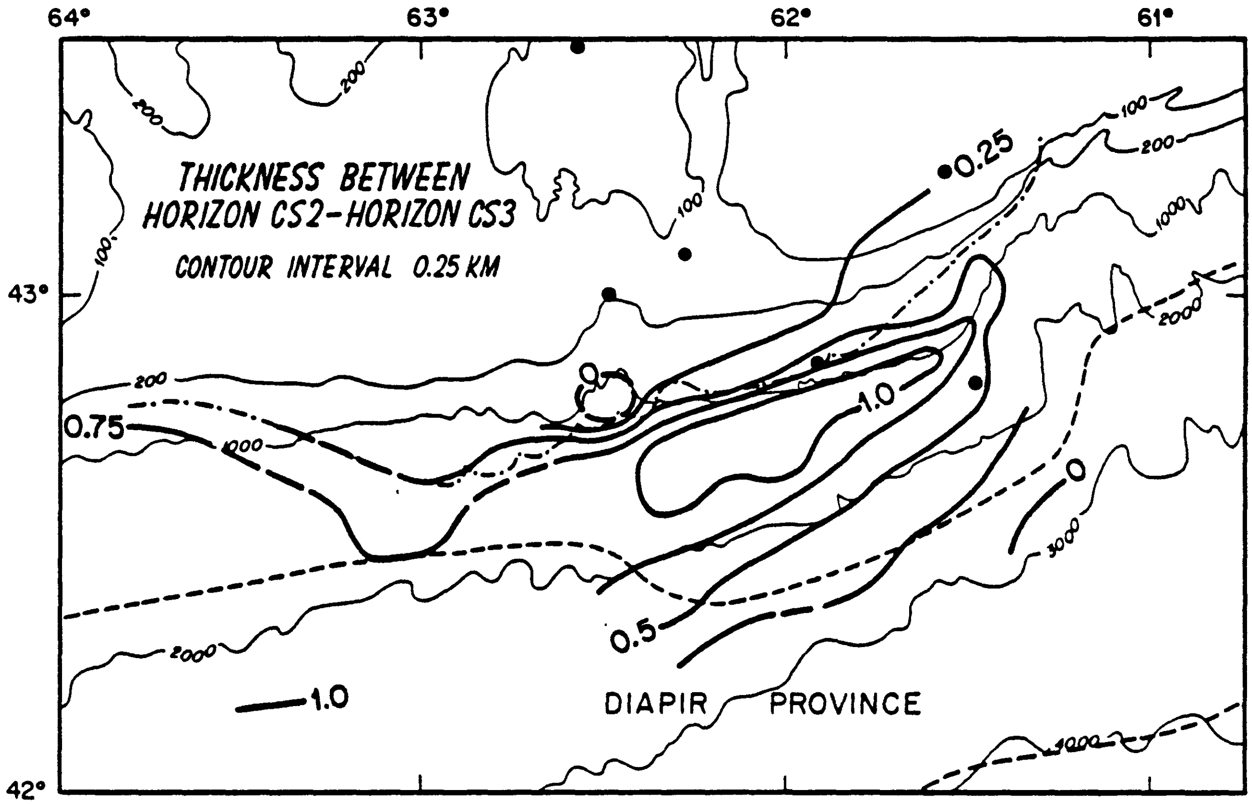
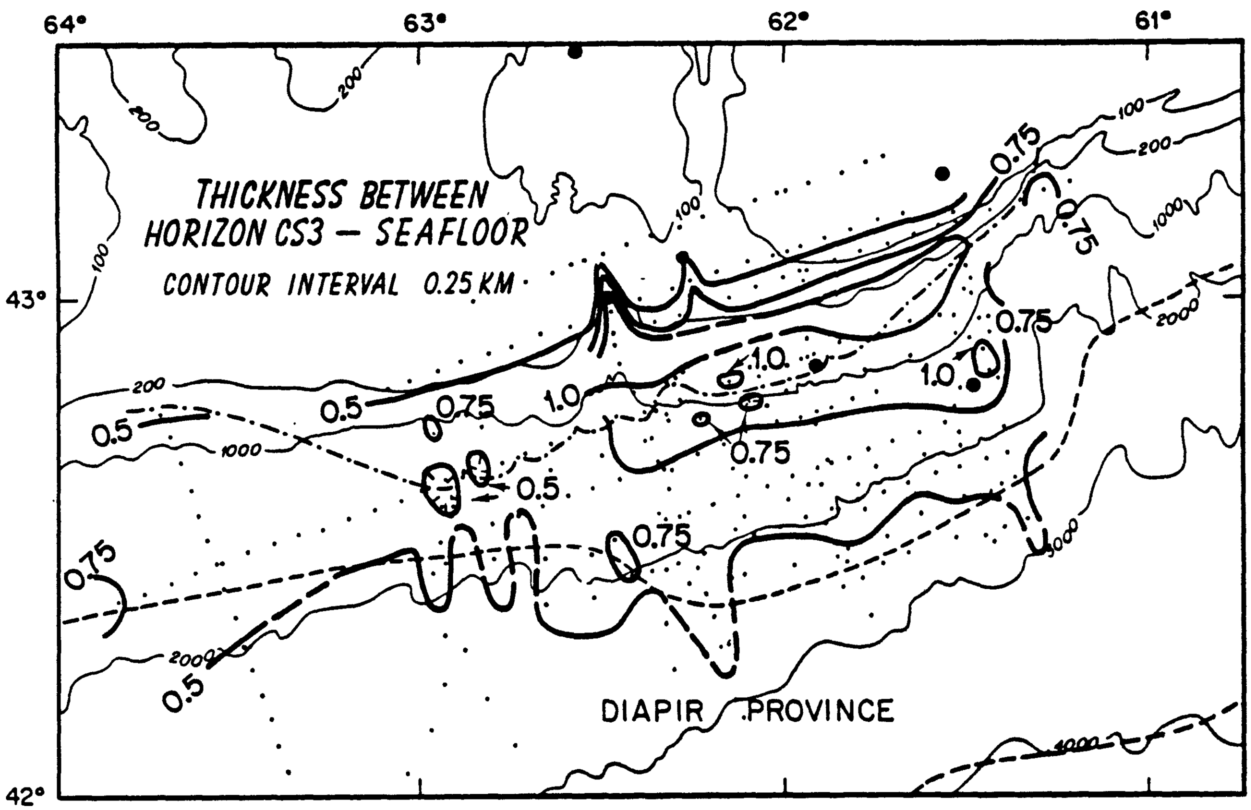

64°
63°
62°
61°
43°
42°
THICKNESS BETWEEN
HORIZON CS2-HORIZON CS3
CONTOUR INTERVAL 0.25 KM
0.25
0.75
1.0
0.5
0
1.0
DIAPIR PROVINCE
100
200
1000
2000
3000
4000

Figure 10e



64°
63°
62°
61°
43°
42°
THICKNESS BETWEEN
HORIZON CS3 — SEAFLOOR
CONTOUR INTERVAL 0.25 KM
DIAPIR PROVINCE

Figure 10f

The Wyandot shelf edge lies 10-20 km landward of the Berriasian platform edge (Fig. 8). It is unlikely that shelf edge retreat was caused by contemporaneous slope failure (eg. King and Young, 1977) because basinward deposition products typical of slope erosion (slumps, slides, and fans) are missing or too thin to be resolved seismically. Marine transgression had already begun on the western shelf by the Barremian when delta development to the east was at a maximum (Given, 1977; her Fig. 12). Thus, shelf edge retreat on the western Scotian shelf during the Cretaceous was more likely due to poorly developed sediment supply.

In summary, Upper Jurassic-Early Cretaceous carbonate sedimentation established a relatively flat shelf flanked seaward by an abrupt shelf edge and a steep carbonate escarpment. Seaward, pelagic carbonates probably intermixed near the base of the escarpment with carbonate talus. During the Neocomian regression, an outer delta environment briefly extended across the carbonate platform to the outer shelf. However, these deposits are thin, and the shelf edge remained unburied. From Barremian to Maestrichtian time, shelf water depths increased, and the shelf edge retreated. Transgressive marine shales and shallow marine sandstones on the outer shelf were interrupted briefly by pelagic chalk deposition. Downslope processes transported fine-grained sediment through channels in the exposed carbonate platform edge into the Schubenacadie basin. Accumulation there was probably slow due to a weak sediment source.

CENOZOIC DEVELOPMENT

Late Maestrichtian-early Paleocene

Clinoform reflectors of the Banquereau Formation (upper Maestrichtian-

Pliocene) downlap seaward onto the Wyandot Formation indicating seaward progradation of an outer shelf sediment wedge (Fig. 6; King et al., 1974; Jansa and Wade, 1975, Given, 1977). Although the contact is a seismic discontinuity especially on the inner shelf (King et al., 1974), the boundary is biostratigraphically conformable in outer shelf wells (Jansa and Wade, 1975; Hardy, 1975). Sediment accumulation rates in the Naskapi N-30 and Oneida O-25 wells (based on Ascoli, 1976) decreased from ~40 m/my for the Cretaceous to 10-20 m/my for the Cenozoic. This decrease may be the result of a decrease in the rate of terrigenous sediment supply. Alternately, erosion associated with either sea-level lowstands (eg. Vail et al., 1977) or fluctuating rates of sea level fall (eg. Pitman, 1978; Watts and Thorne, 1984) may have shifted the depocenter seaward. Regional isopach maps show that Cenozoic accumulation was still much greater on the eastern shelf and slope than on the west (Parsons, 1975; Given, 1977; Wade, 1981).

In general, sediments deposited in the late Maestrichtian to Eocene grade seaward from middle neritic-bathyl mudstones and argillaceous, glauconitic sandstone on the shelf to shale, chalk, and mudstones on the slope and upper rise (Fig. 7; Given, 1977; Hardy, 1975). In the Oneida O-25 well, a thick (~500 m), continuously-deposited sequence of silty, slightly calcareous mudstone (Campanian-lower Paleocene, Maskonomet bed) coarsens upwards to thin, glauconitic, calcaeous sandstone (upper Paleocene-middle Eocene, Nashwauk bed; Fig. 7; Hardy, 1975; Ascoli, 1976; Barss et al., 1979). These sediments are foreset beds in the prograding clinoform sequence (Fig. 6). Thin bottomset beds occur farther seaward on the shelf indicating that the shelf depocenter had not reached the shelf edge established at the end of the Cretaceous. A time correlative section lying above the Wyandot in the Schubenacadie H-100 well (Fig. 7) is thin (~150 m) and shaley suggesting a relatively low supply rate of fine-grained sediment to the slope and upper rise. Seaward of the diapirs, moderate amplitude reflectors above Horizon A* are continuous and flat-lying to gently seaward dipping (Fig. 9). These reflectors may represent fine-grained distal turbidite and hemipelagic equivalents of mudstones on the upper rise.

Several paleo-canyons with apparent relief of 0.5-1.5 km lie beneath the present outer shelf (Fig. 2). All were initially eroded in the late Maestrichtian or Tertiary. None can be traced seaward onto Tertiary paleo-slope surfaces because either current erosion there in the Miocene (see below) smoothed canyon relief or their relief is too low to resolve seismically with the data available. Most of the shelf canyons are filled; only Mohican Canyon still has bathymetric expression. Mohican Canyon is older and larger than the other shelf edge canyons. At the head of Mohican Canyon, Eocene/Oligocene and possibly late Paleocene reflectors can be traced continuously above the trace of the canyon (Fig. 11). Thus, headward erosion reached a maximum no later than early Paleocene and may correlate with sea level lowstands (Vail et al., 1977) or with fluctuating rates of sea level change (Watts and Thorne, 1984). A time correlative fan sequence does not occur in the Schubenacadie basin, but such a sequence may be too thin to identify with these data.

In summary, a clastic sediment wedge began to prograde across the relatively flat outer shelf in the Maestrichtian. Thin, bottomset beds

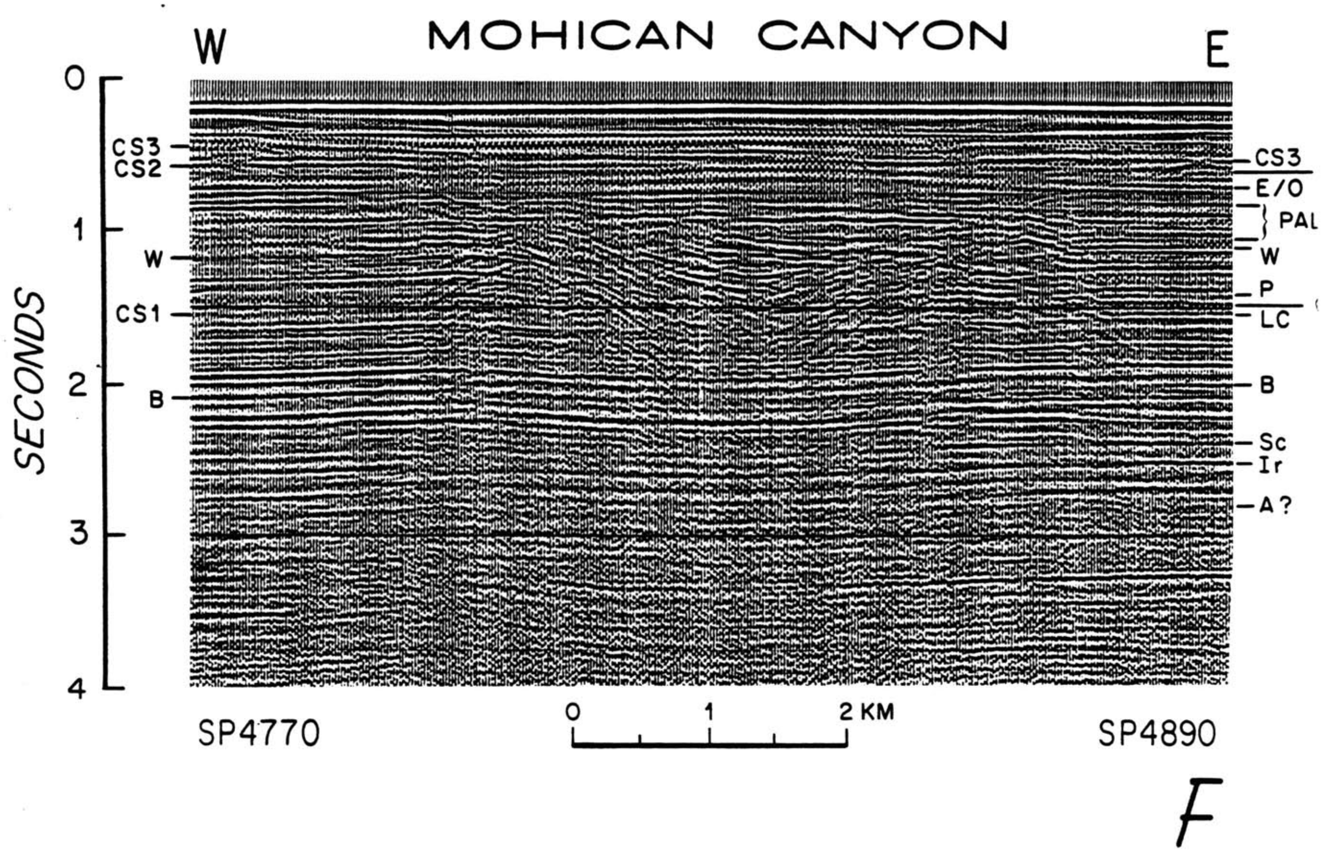


Figure 11 Profile from line SAO 170 illustrating buried Mohican Canyon (side echoes and diffractions). Location in Fig. 5 (Profile F). Note continuity of Paleocene(?) reflectors above the canyon and Petrel reflector below. Tops of formations correlated from the Moheida P-52 and Oneida O-25 wells: E/O- Eocene-Oligocene unconformity in the Oneida well, W- Wyandot chalk, P- Petrel chalk, LC- Logan Canyon, B-Baccaro limestone, Sc- Scatarie limestone, Ir- Iroquois dolomite, A?- Argo salt.

accumulated on the outermost shelf while fine-grained turbidites and hemipelagic sediments were deposited on the continental rise. Headward erosion of Mohican Canyon reached a maximum in the late Maestrichtian-early Paleocene with no time-correlative fan sequence on the upper rise.

Late Paleocene

A late Paleocene hiatus occurs in the Oneida O-25 well (Fig. 7; Ascoli, 1976; Nashwauk bed, Hardy, 1975). On the upper rise, hummocky reflector sequences similar to "compound fan complexes" of Mitchum et al. (1977) lie side-by-side with along-slope widths of ~5-10 km and thicknesses of up to 0.5 km. These seismic fan sequences correlate in the Schubenacadie H-100 well to Paleocene-Eocene chalks interbedded with mudstones but without significant sandstone (Fig. 7). Because the chalky "fans" are more numerous and closer spaced than paleo-canyons on the outer shelf (Fig. 2), they probably originated from erosion of slope canyons which did not reach the shelf edge. As suggested before, slope canyons either were smoothed by later current erosion or have relief too low to be seismically resolved. Likewise, the absence of Paleocene seismic fan sequences in the multi-channel profile just seaward of the diapirs either suggests that later erosion smoothed these features or very low relief sequences were deposited on the lower rise.

In summary, late Paleocene erosion formed a local hiatus on the outer shelf. This erosional episode differs from all other Cretaceous and Tertiary episodes in this region in that fans developed on the upper rise.

Eocene-Oligocene

During the Eocene well-sorted, glauconitic sandstone and sandy, calcareous mudstone accumulated on the shelf (upper Nashwauk bed, Hardy, 1975). In the late Eocene, widespread erosion cut up to 300 m of relief on the middle shelf leaving a prominent seismic unconformity (King et al., 1974). This erosion formed the present cuesta physiography of the outer shelf (King, 1967). The morphology of this unconformity suggests erosion by fluvial processes during subaerial exposure (King et al., 1974). Later, Oligocene and Miocene sedimentation filled topographic lows on this surface, but enough relief was preserved during these periods to control drainage patterns during a similar erosional event in the late Tertiary-early Pleistocene (King et al., 1974). At the Oneida O-25 well, upper Eocene foraminiferal and palynological zones are missing or very thin (Ascoli, 1976; Fig. 7). A correlative seismic unconformity can be traced as far seaward as the paleo-shelf edge. Oligocene and Miocene reflectors can be traced across small paleo-canyons east of Mohican canyon (Fig. 2) suggesting that headward erosion reached a maximum no later than the early Oligocene. Thus, late Eocene erosion affected the Scotian shelf seaward at least as far as the shelf edge.

On the paleo-slope, less than 150 m of sediment can be detected between Horizons CS1 (Turonian?) and CS2, an overlying unconformity eroded in the early(?) Miocene (Figs. 8, 10d). Eocene-Oligocene rocks are thin or missing. On the upper rise, hummocky reflectors of the Paleocene-Eocene fan sequence are overlain by sub-parallel, seaward dipping reflectors (Figs. 6, 8) that tie at the Schubenacadie H-100 well to Eocene and Miocene mudstones

(Fig. 7). Lack of prominent channel features in strike profiles suggests that these sediments were deposited by unchanneled down slope flows or from turbidity currents which spilled from low relief upper rise channels. Horizon CS2 truncates the updip end of these reflectors (Fig. 8). Thus, Miocene erosion prevents direct correlation of the late Eocene shelf unconformity southeastward onto the rise. Since no canyons occur on the upper rise below Horizon CS2, this region was probably a depocenter for sediment eroded on the shelf.

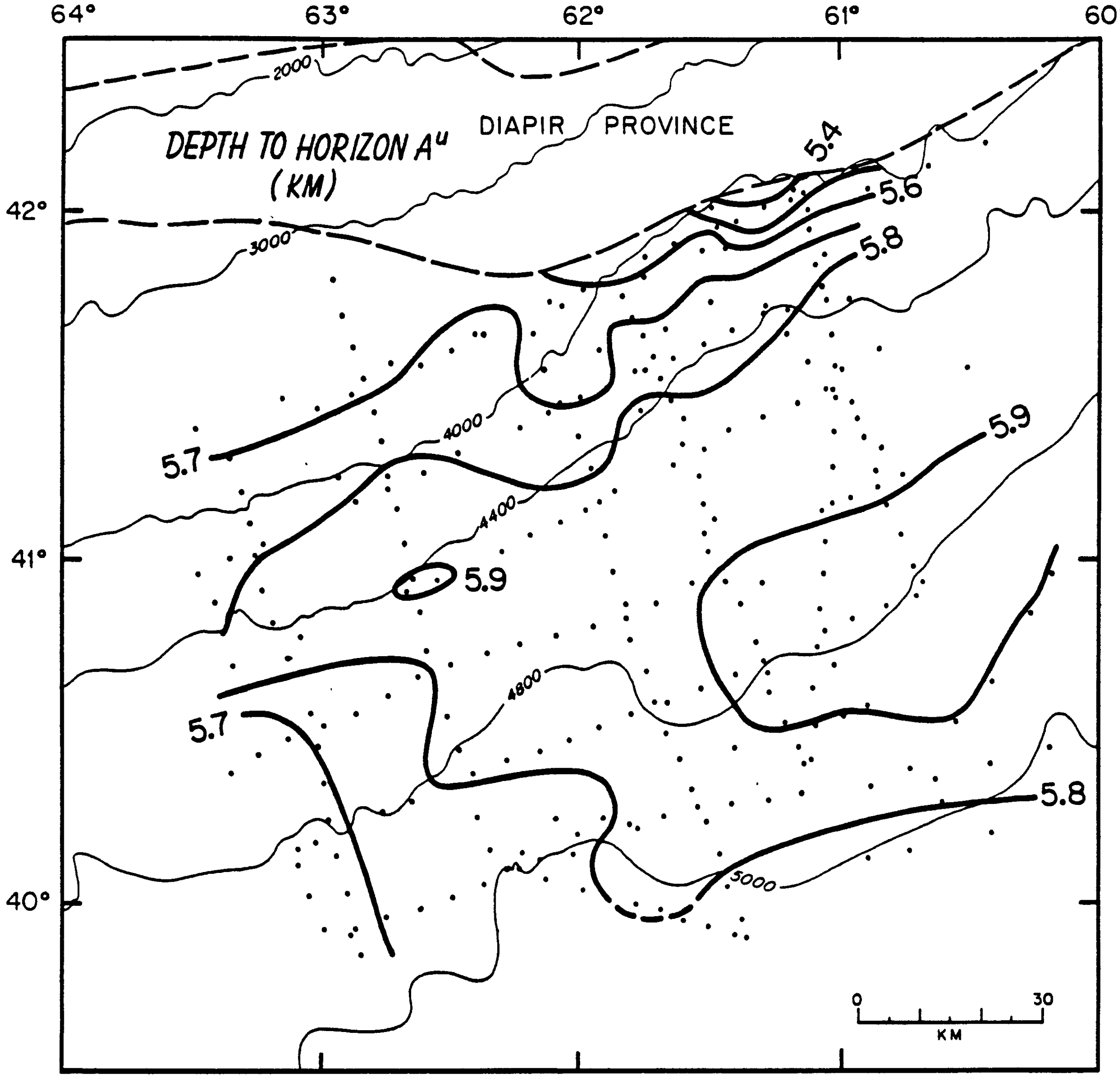
In contrast, between 57°-59°W, Parsons (1975) found paleo-canyons on the upper rise within the diapir province having relief of 200-300 m and correlated these features to the Eocene-early Oligocene. Presumably, upper rise canyons mapped by Parsons (1975) extend landward to paleo-slope canyons where they may correlate with the late Eocene unconformity on the shelf noted by King et al. (1974).

Oligocene sediments were continuously deposited on the Scotian shelf (Hardy, 1975; Ascoli, 1976). Oligocene rocks are typically glauconitic sandstone with fine-grained interbeds (Manhassett bed; Hardy, 1975). Paleontological indicators suggest that shelf water depths decreased from outer neritic-bathyl in the Eocene to inner-to-outer neritic in the Oligocene-Miocene (Ascoli, 1976; Given, 1977). Shelf well stratigraphy, in particular the Oneida O-25 well, does not support evidence from continental margins elsewhere in the North Atlantic for a middle Oligocene unconformity (Miller et al., 1985; Mountain and Tucholke, in press).

The Oligocene and possibly parts of the Eocene and Miocene are missing in the Schubenacadie H-100 well on the upper rise (Fig. 7). This hiatus correlates with a smooth, seaward dipping seismic reflector of moderate amplitude that appears conformable with surrounding reflectors (Fig. 6). There are no channels associated with this surface. The landward end of this reflector is truncated by Horizon CS2 (Fig. 8). The Oligocene hiatus is tentatively correlated across the diapir province with Horizon A^u based on similar dip and travel time below seafloor (Fig. 3).

On the lower rise seaward of the diapirs, Horizon A^u is a smooth surface dipping gently seaward and northeastward (Fig. 12a) that is marked by downlap of reflectors onto a sequence of relatively high-amplitude, continuous reflectors (Fig. 9). This horizon is an erosional unconformity based on coincident seismic velocity discontinuities and changes in velocity-depth gradient gradient (Table 2; Fig. 9). The unconformity ties to Horizon A^u in the North American basin (Swift et al., in press) and probably was cut by southwest flowing abyssal currents in the Oligocene (Tucholke and Mountain, 1979; Miller and Tucholke, 1983).

In summary, progradation of the shelf sediment wedge and accumulation of fine-grained sediments, probably turbidites, on the continental rise during the Eocene-Oligocene were interrupted by one or more widespread erosion episodes. The relationships among the upper Eocene-Oligocene unconformities on shelf, upper rise, and lower rise are uncertain because seismic ties cannot be confidently made across either the paleoslope or the diapir province and because the available biostratigraphy has poor resolution. If Horizon A^u on the lower rise and the Oligocene hiatus in the Schubenacadie basin are equivalent, major erosion on the shelf (late Eocene) appears to either preceed or coincide with the initiation of abyssal current erosion on



<u>Figure 12</u> Depth (km) below sealevel to lower continental rise horizons. (a) Depth (km) to Horizon A^u (Oligocene). Dots are data points. Data coverage for other maps varies slightly. (b) Horizon CR1 (lower? Miocene). Horizon CR2 excavated Horizon CR1 south of the hatchured line (see also Fig. 16). (c) Horizon CR3 (Pliocene?). Horizon CR3 merges with Horizon CR4 north of hatchured line. (d) Horizon CR4 (Pliocene/ Pleistocene).

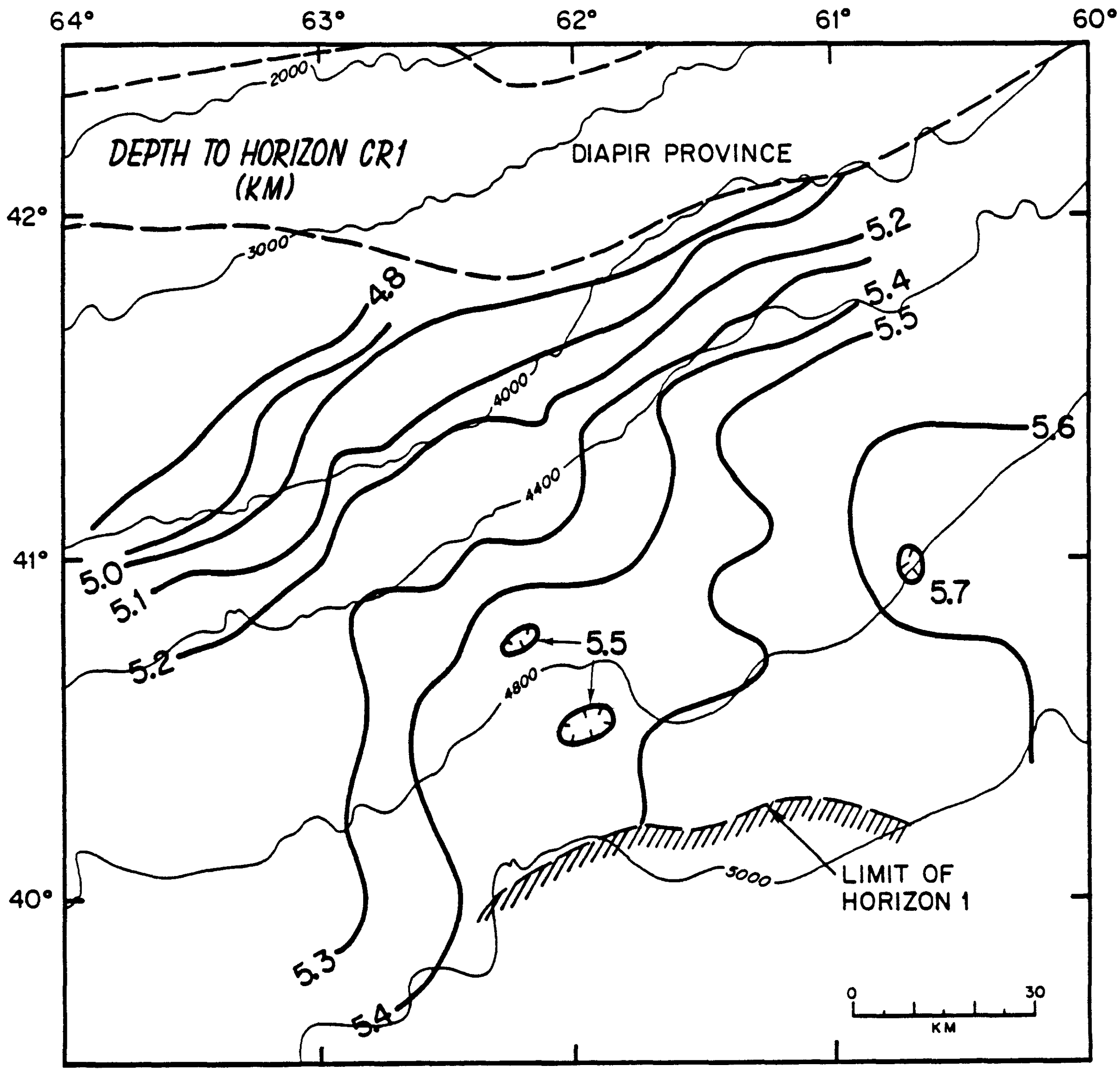
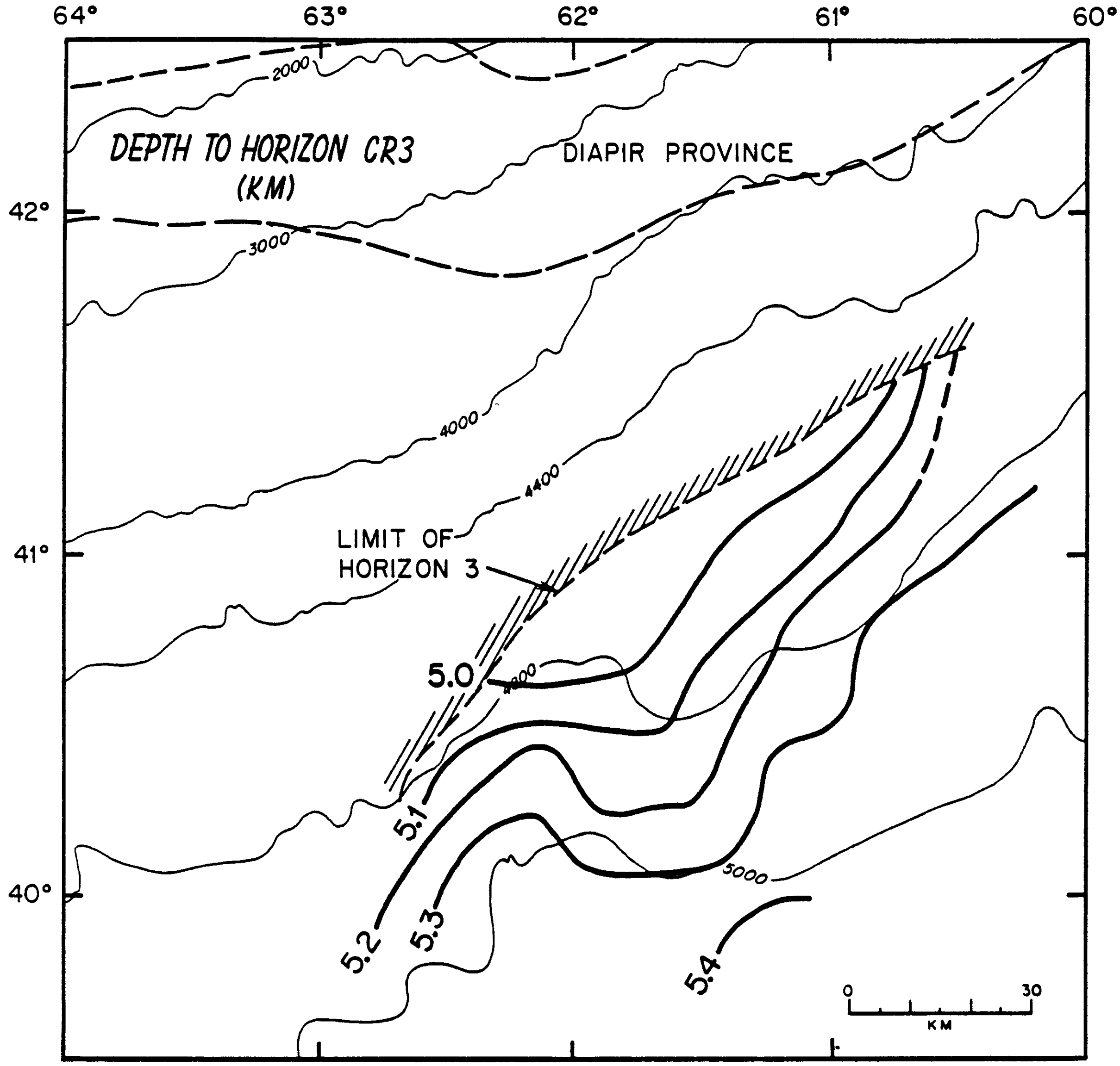

64°
63°
62°
61°
60°
42°
41°
40°
DEPTH TO HORIZON CR1
(KM)
DIAPIR PROVINCE
2000
3000
4000
4400
4800
5000
4.8
5.0
5.1
5.2
5.2
5.3
5.4
5.4
5.5
5.5
5.6
5.7
LIMIT OF
HORIZON 1
0
30
KM

Figure 12b



64°
63°
62°
61°
60°
42°
41°
40°
DEPTH TO HORIZON CR3
(KM)
DIAPIR PROVINCE
2000
3000
4000
4400
4000
5000
LIMIT OF
HORIZON 3
5.0
5.1
5.2
5.3
5.4
0
30
KM

Figure 12c

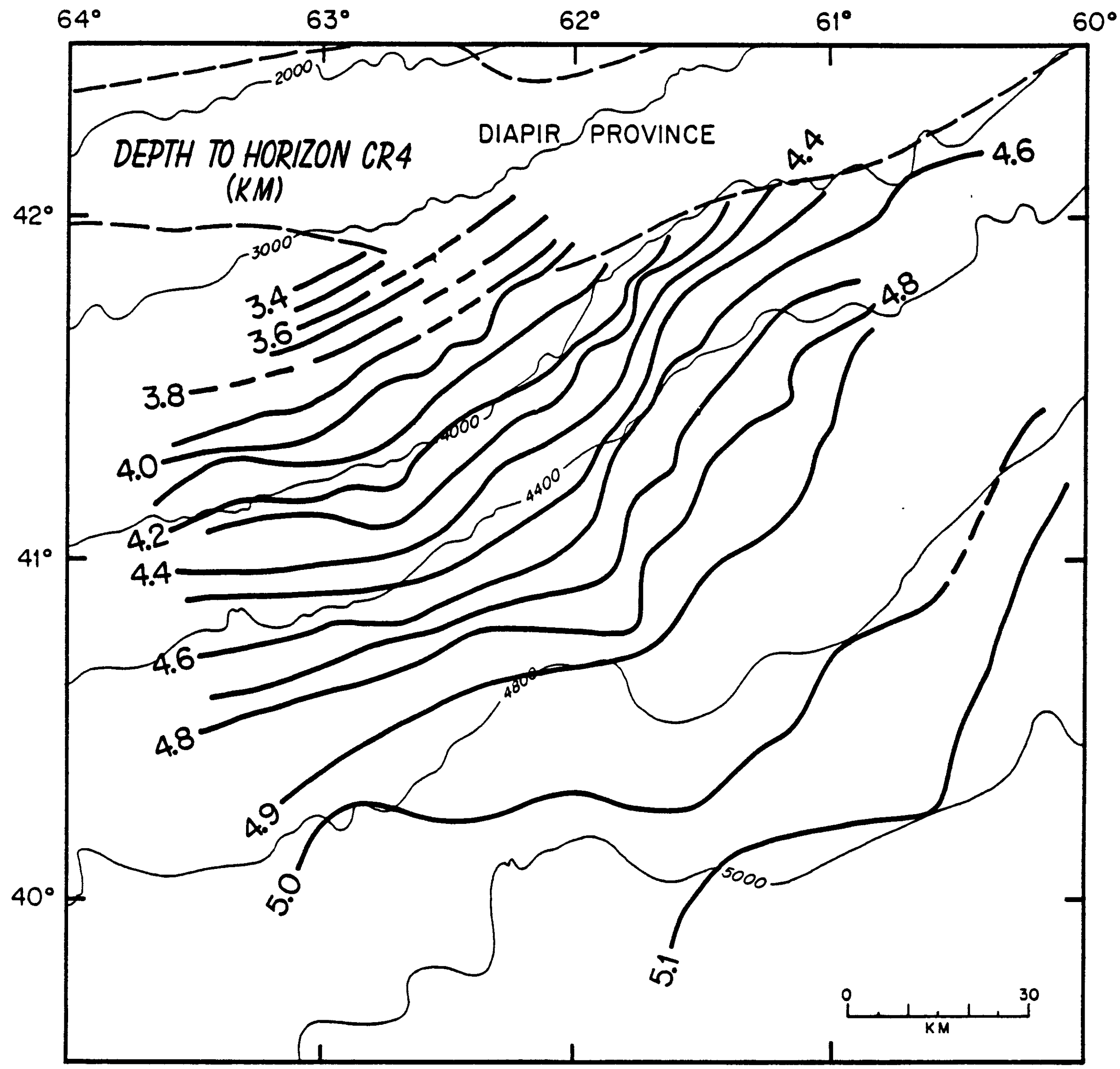

64°
63°
62°
61°
60°
42°
41°
40°
DEPTH TO HORIZON CR4
(KM)
DIAPIR PROVINCE
2000
3000
4000
4400
4800
5000
3.4
3.6
3.8
4.0
4.2
4.4
4.4
4.6
4.6
4.8
4.8
4.9
5.0
5.1
0
30
KM

Figure 12d

the continental rise (Oligocene). Elsewhere in the North Atlantic, abyssal current erosion (Horizon A^u, Oligocene) preceeds major canyon cutting (middle Oligocene; Miller et al., 1985; Mountain and Tucholke, in press).

Early Miocene

Seismic unconformities of Miocene age have not been noted on the shelf (eg. King et al., 1974). Between 61°-64°W, foreset beds prograded to the paleo-shelf edge (Fig. 8). At the Oneida O-25 well seismic profiles and lithostratigraphy suggest that a facies change associated with progradation produced coarsening upwards of sediment grain size, a trend noted for the Scotian shelf in general by Given (1977) and Hardy (1975). Lower Miocene rocks are thin or missing on the paleoslope; current erosion (Horizon CS2, see below) probably removed this section.

Seaward in the Schubenacadie basin, lower(?) Miocene mudstones were deposited above the Oligocene hiatus (Fig. 7). Seismic reflectors are continuous and diverge slightly seaward. Divergence of reflectors seaward may be due to ponding of turbidites on the upper rise landward of raised seafloor uplifted by early salt movement (see below). It is unlikely that this sediment wedge is the updip remnent of a sediment drift deposited by abyssal currents because major drifts in the North Atlantic (McCave and Tucholke, in press; Mountain and Tucholke, in press) do not typically extend landward to the base of slope.

Erosion interrupted mudstone deposition in the Miocene forming Horizon CS2, a prominent angular seismic unconformity (Figs. 7, 8). In the Schubenacadie basin, Horizon CS2 is flat (Fig. 10b) and truncates Eocene-

Miocene beds that dip 1-2° seaward. Erosion, however, did not cut deep enough to truncate Horizon CS1 (Figs. 8, 10d). On the paleoslope, Horizon CS2 rises landward; reflectors above Horizon CS2 also increase in dip but still lap out landward on Horizon CS2 at 1-2°. Lapout onto Horizon CS2 continues landward to the paleoshelf edge where the horizon can be traced as a reflector with indistinct truncations. Horizon CS2 ties to lower(?) Miocene siltstone and sandstone (Esperanto bed) at the Oneida O-25 well and to Miocene mudstone at the Schubenacadie H-100 well (Fig. 7). An estimate of 75 m of section removed at the Schubenacadie H-100 well can be obtained by tracing the deepest dipping reflector truncated by Horizon CS2 seaward to the first diapir. A decrease in this estimate due to the slight seaward divergence of beds is probably offset by an unknown increase due to total removal of beds.

Horizon CS2 is unusual in that it incises stratigraphically deeper in a landward direction for about 40 km and forms a flat surface "perched" on the upper paleorise (Fig. 10b). Tectonic origins were considered but rejected for the following reasons: (1) Horizon CS2 is unlikely to be a growth fault because of its large extent (minimum 40-50 km by 100-125 km), the sharp angular contacts, and the consistently low angle of both the dip and stratal contact (2-5°) when traced landward onto the shelf. (2) Rotation due to salt withdrawal in a seaward direction is unlikely because such a process should have flattened strata beneath as well as above Horizon CS2. (3) Gravity-slide failure seems unlikely because of its large areal extent and broad depth range (from shelf to upper rise).

At least two explanations are possible for deeper stratigraphic exposure near the paleoslope/rise boundary. Currents might have eroded deeper at the change in slope because changes in hydrodynamics or seafloor gradient at that paleodepth intensified flow. The significant difference in dip between Horizon CS2 and the upper rise equivalent of Horizon A^u may indicate a change in abyssal hydrodynamics or water masses between the Oligocene and Miocene. The erosion episode forming Horizon CS2 may correlate with abyssal current erosion off the eastern United States in the late middle Miocene that formed Horizon Merlin (Fig. 4; Mountain and Tucholke, in press).

Alternately, depth of erosion was uniform across the uppermost rise, and deeper stratigraphic exposure updip was due to thinner beds closer to the slope. In this case, the difference in dip between Horizons CS2 and A^u indicates that the seafloor gradient of the upper rise was leveled in the early Miocene by thickening deposition basinward. To produce the nearly horizontal seafloor indicated by Horizon CS2 (Fig. 10b), sediments may have ponded landward of linear seafloor ridges or pillows raised by early stages of halokinesis (eg. Trusheim, 1960; Seni and Jackson, 1983), although flattening due to sediment drift development cannot be ruled out.

Seaward of the diapir province, the age of the oldest sediments above Horizon A^u is uncertain. An early Miocene age is assumed because the oldest sediments recovered above Horizon A^u on the lower United States continental rise are of that age (Tucholke and Mountain, 1979). In general, incoherent reflectors occur immediately above Horizon A^u landward of the 4700 m isobath (~6.3 s reflection time), and continuous, closely-spaced reflectors occur seaward (compare Figs. 13a and 13b). This change in seismic character may be due to a lateral change in depositional process or

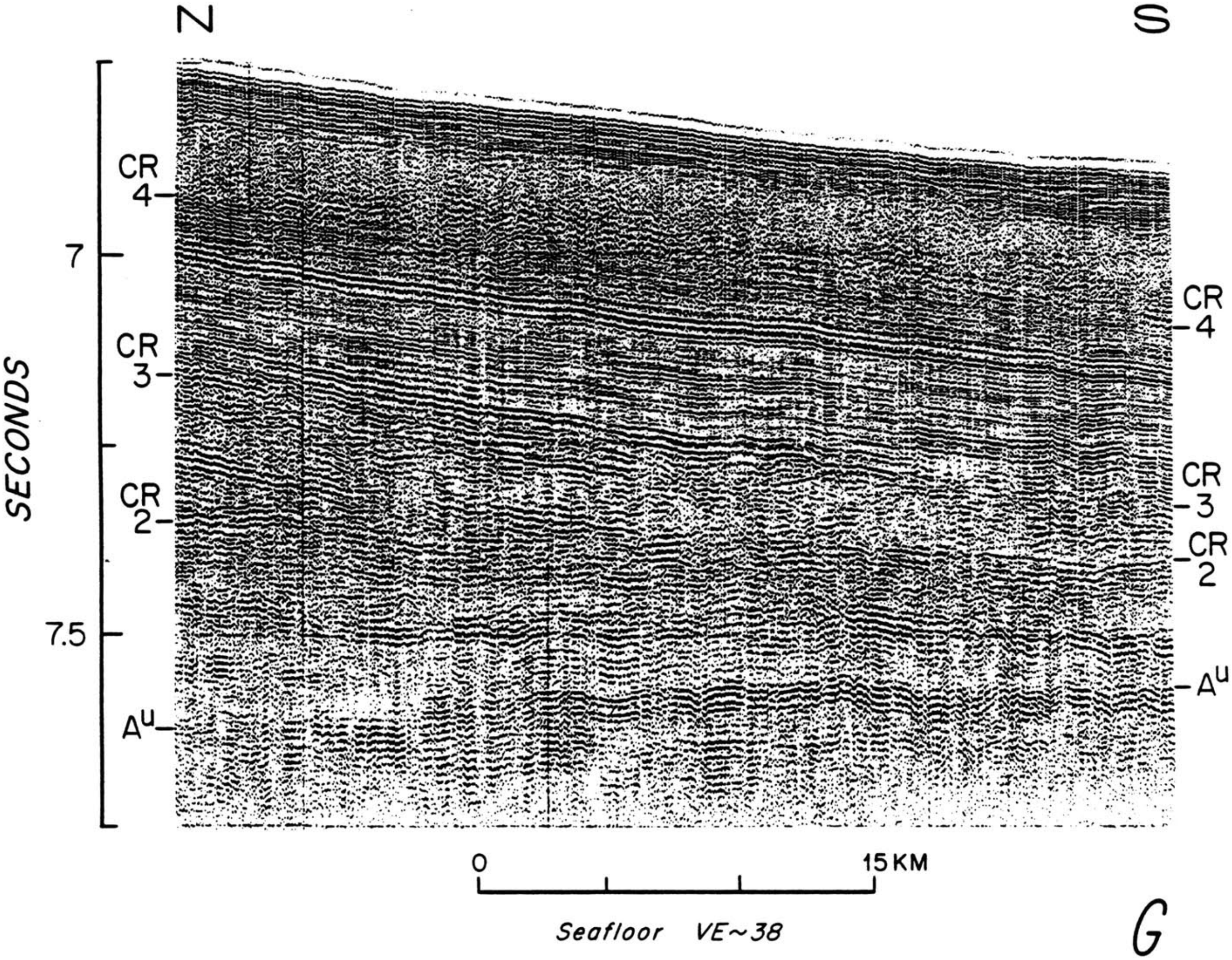


Figure 13 Single-channel, water-gun reflection profiles (80-170 Hz) illustrating stratigraphy on the lower rise seaward of diapir province. Locations in Fig. 5. Continental rise (CR) horizons (1-4) discussed in text are shown. (a) RC2408 line 25, Profile G. (b) RC2408 line 25, Profile H. "HR" indicates hyperbolic reflections correlated with Horizon X. (c) RC2408 line 31, Profile I. Note sediment waves between Horizons CR3 and CR4 and northward change in seismic character of interval between Horizons CR1 and CR2.

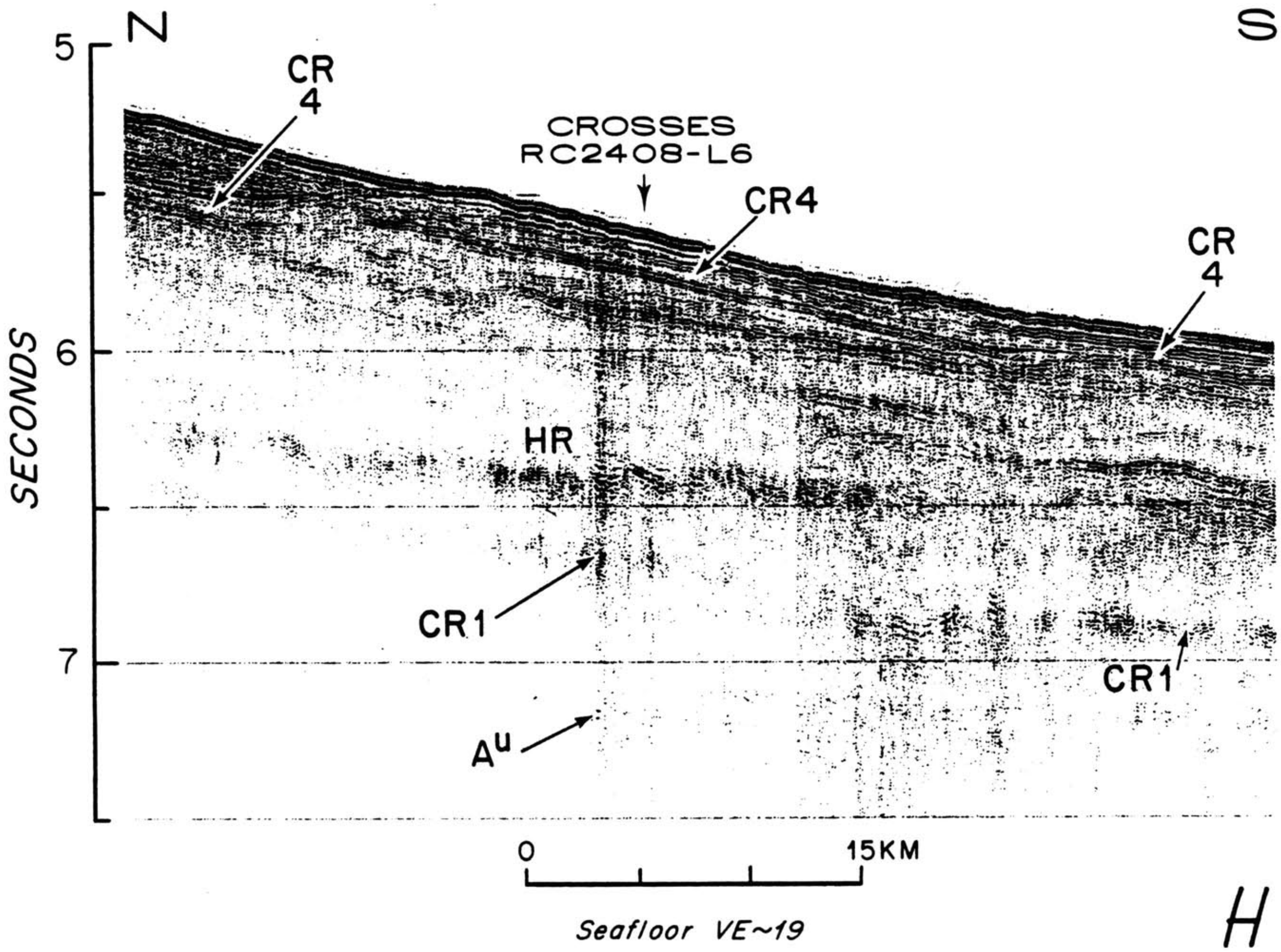

N
S
5
6
7
SECONDS
CR 4
CROSSES RC2408-L6
CR4
CR 4
HR
CR1
CR1
A^u
0
15KM
Seafloor VE~19
H

Figure 13b

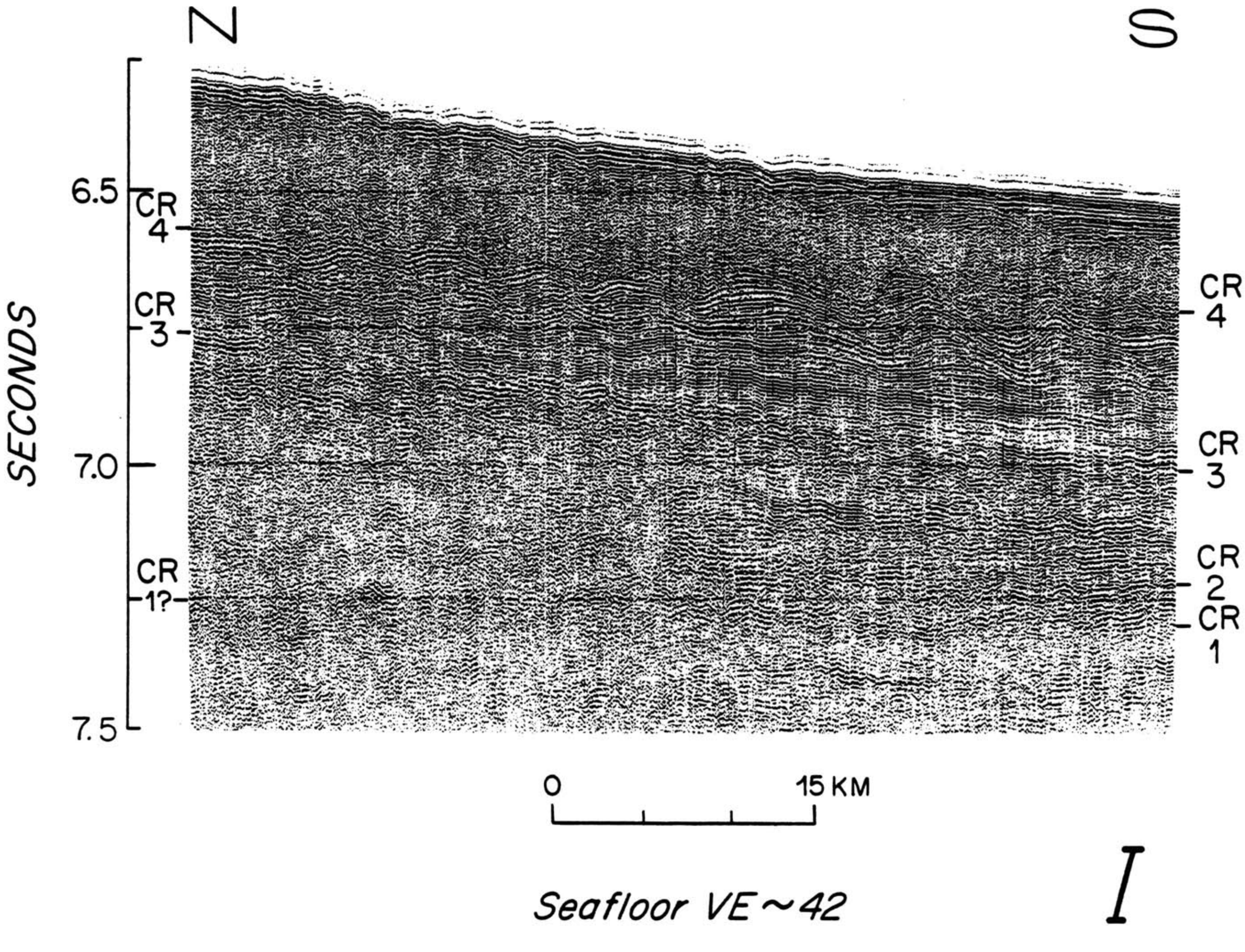

N
S
SECONDS
6.5
7.0
7.5
CR
4
CR
3
CR
1?
CR
4
CR
3
CR
2
CR
1
0
15 KM
Seafloor VE ~ 42
I

Figure 13c

to a seaward increase in signal strength caused by thinner overlying sediments. The interval between Horizons A^u and CR1 (lower Miocene) thins seaward by a factor of 3-4, and isopachs bulge seaward in a tongue that is 80 km across by 150 km long with a relief of about 0.2 km (Fig. 14b). A fan interpretation for this interval on the lower rise is consistent with turbidity current ponding on the upper rise suggested above.

Horizon CR1 is a coherent reflector along GSI 108 (Fig. 9) and in single-channel seismic records from the lowermost rise, but it is indistinct in single-channel records close to the diapirs (Fig. 13b). Downlapping reflectors, a coincident velocity discontinuity in 3 of 5 sonobuoys, and a change in velocity gradient (Table 2) suggest that the reflector is an erosional unconformity. Horizon CR1 mimics Horizon A^u topography, dipping seaward from the diapirs at a somewhat steeper slope and northeastward along the lowermost rise (Fig. 12). Similarity of form suggests erosion by a current system similar to that which cut Horizon A^u. However, smaller changes in the velocity-depth profile (Table 2) and a lower amplitude reflection (Fig. 9) at Horizon CR 1 compared to Horizon A^u suggest a current of shorter duration or slower velocity.

Lower rise Horizon CR1 may correlate with slope Horizon CS2. Both horizons are widespread, planar features interpreted as surfaces eroded by abyssal currents and both occur just above Horizon A^u. Alternately, Horizon CS2 may correlate with Horizon X (see below), but differences in seismic character suggest that Horizon X is a different event.

In summary, during the lower Miocene the shelf sediment wedge reached the shelf edge in the lower Miocene. On the upper rise, turbidity currents

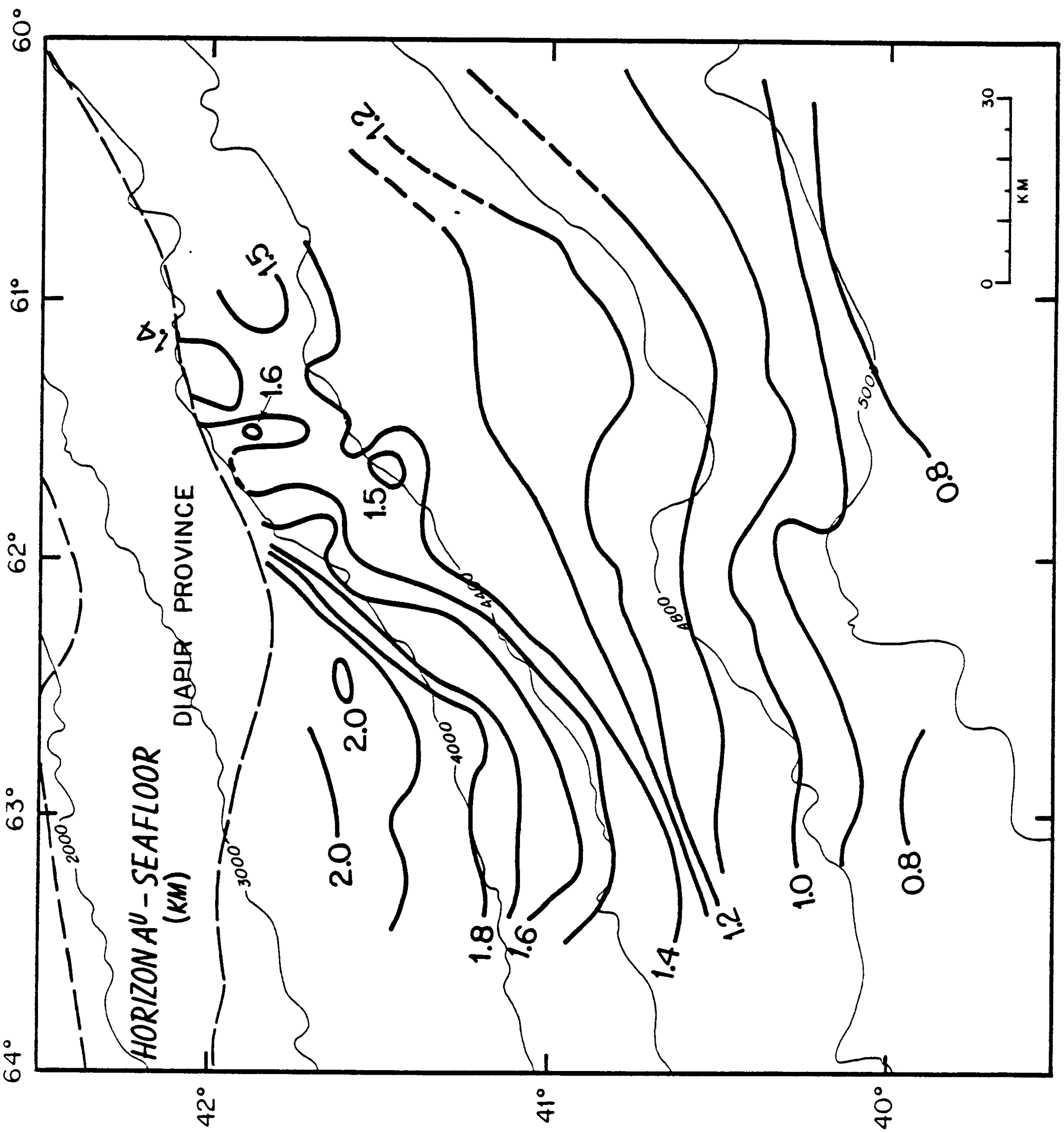


Figure 14 Isopachs (km) for intervals between horizons on the lower continental rise. Data coverage is approximately that shown in Fig. 12a. (a) Interval between Horizon A^u and seafloor. (b) Interval between Horizons A^u and CR1. (c) Interval between Horizon CR1 and the next younger major unconformity: Horizon CR3 south of the ~4600-4700 m isobaths and Horizon CR4 to the north (note extent of Horizon CR3 in Fig. 12c). (d) Interval between Horizons CR3 and CR4. (e) Interval between Horizon CR4 and seafloor.

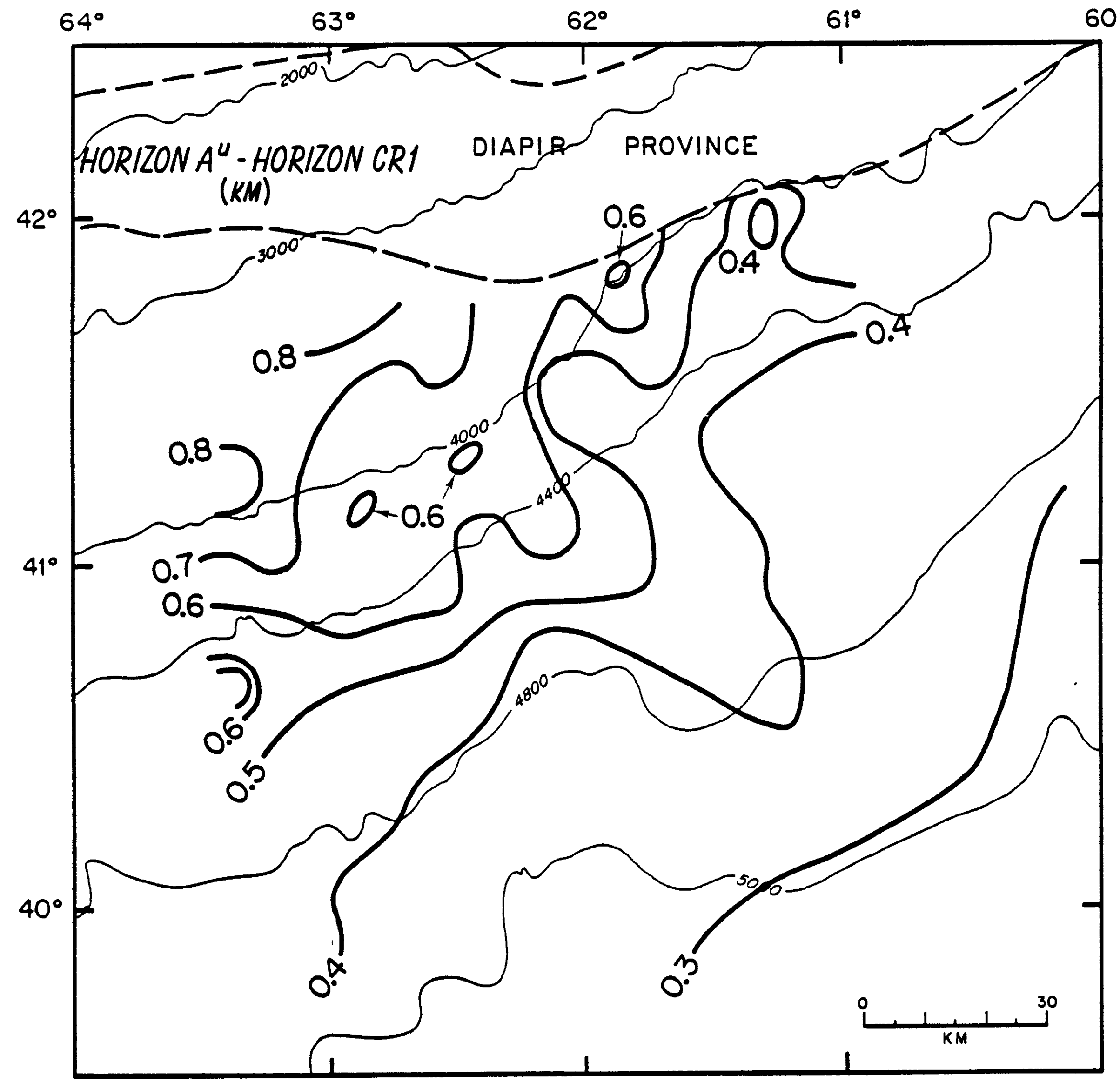

64°
63°
62°
61°
60°
42°
41°
40°
HORIZON A^u - HORIZON CR1
(KM)
DIAPIR PROVINCE
2000
3000
4000
4400
4800
5000
0.8
0.8
0.7
0.6
0.6
0.6
0.6
0.5
0.4
0.4
0.4
0.3
0
30
KM

Figure 14b

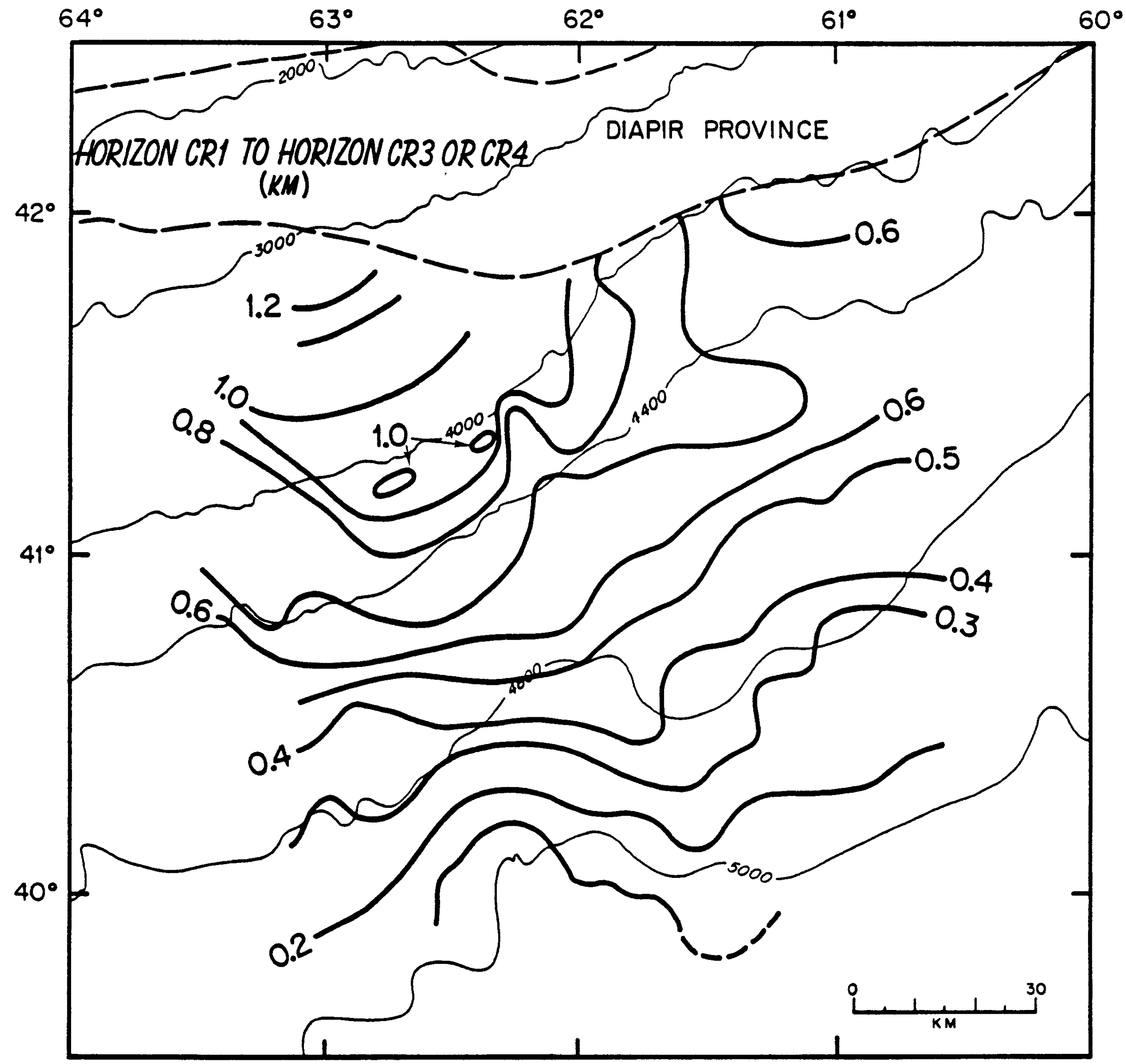

64°
63°
62°
61°
60°
42°
41°
40°
2000
3000
4000
5000
1400
DIAPIR PROVINCE
HORIZON CR1 TO HORIZON CR3 OR CR4
(KM)
1.2
1.0
0.8
1.0
0.6
0.6
0.5
0.4
0.3
0.6
0.4
0.2
0
30
KM

Figure 14c

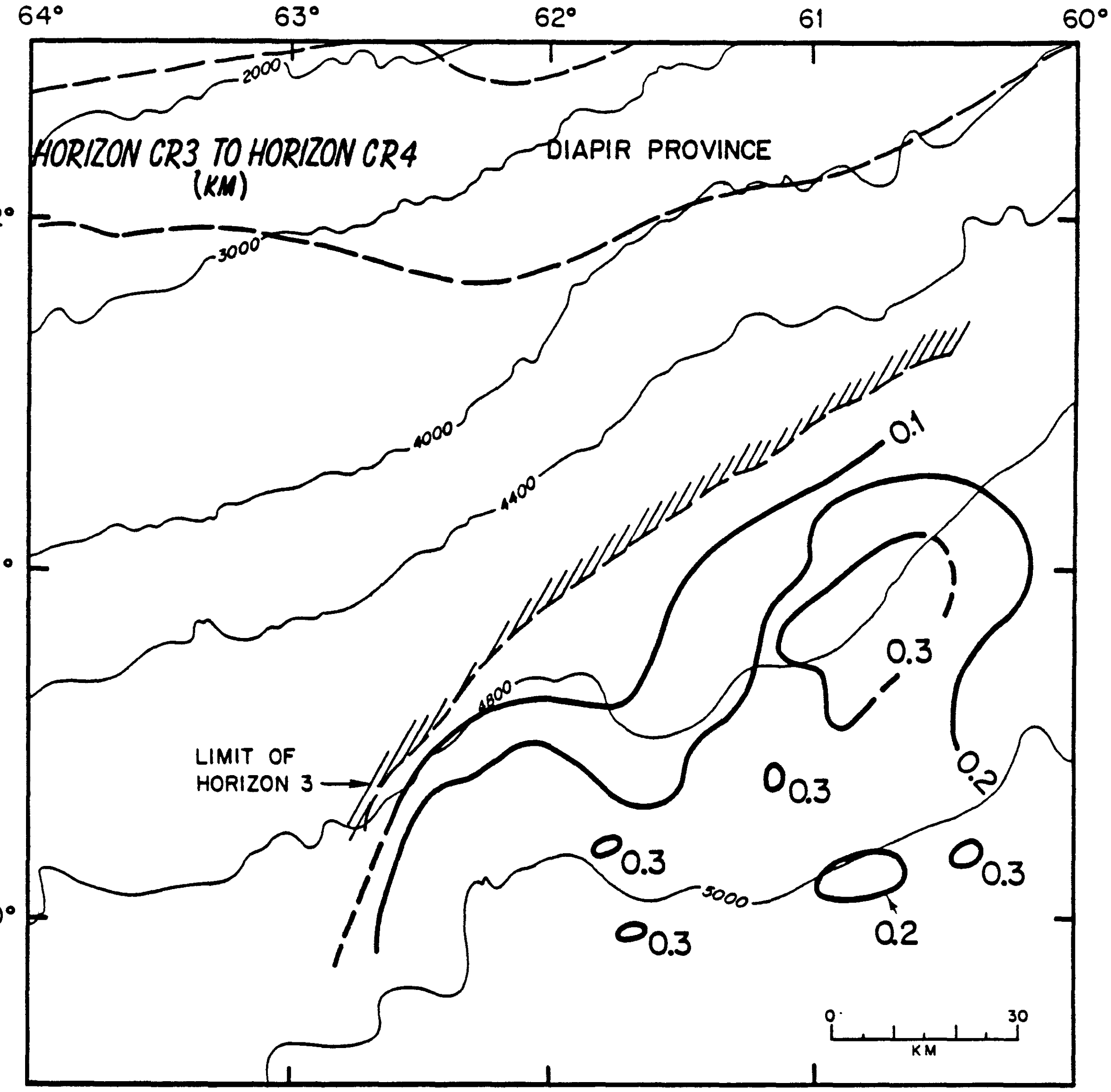

64°
63°
62°
61
60°
42°
41°
40°
HORIZON CR3 TO HORIZON CR4
(KM)
DIAPIR PROVINCE
2000
3000
4000
4400
4800
5000
0.1
0.2
0.3
0.3
0.3
0.3
0.3
0.2
0.3
0.2
LIMIT OF
HORIZON 3
0
30
KM

Figure 14d

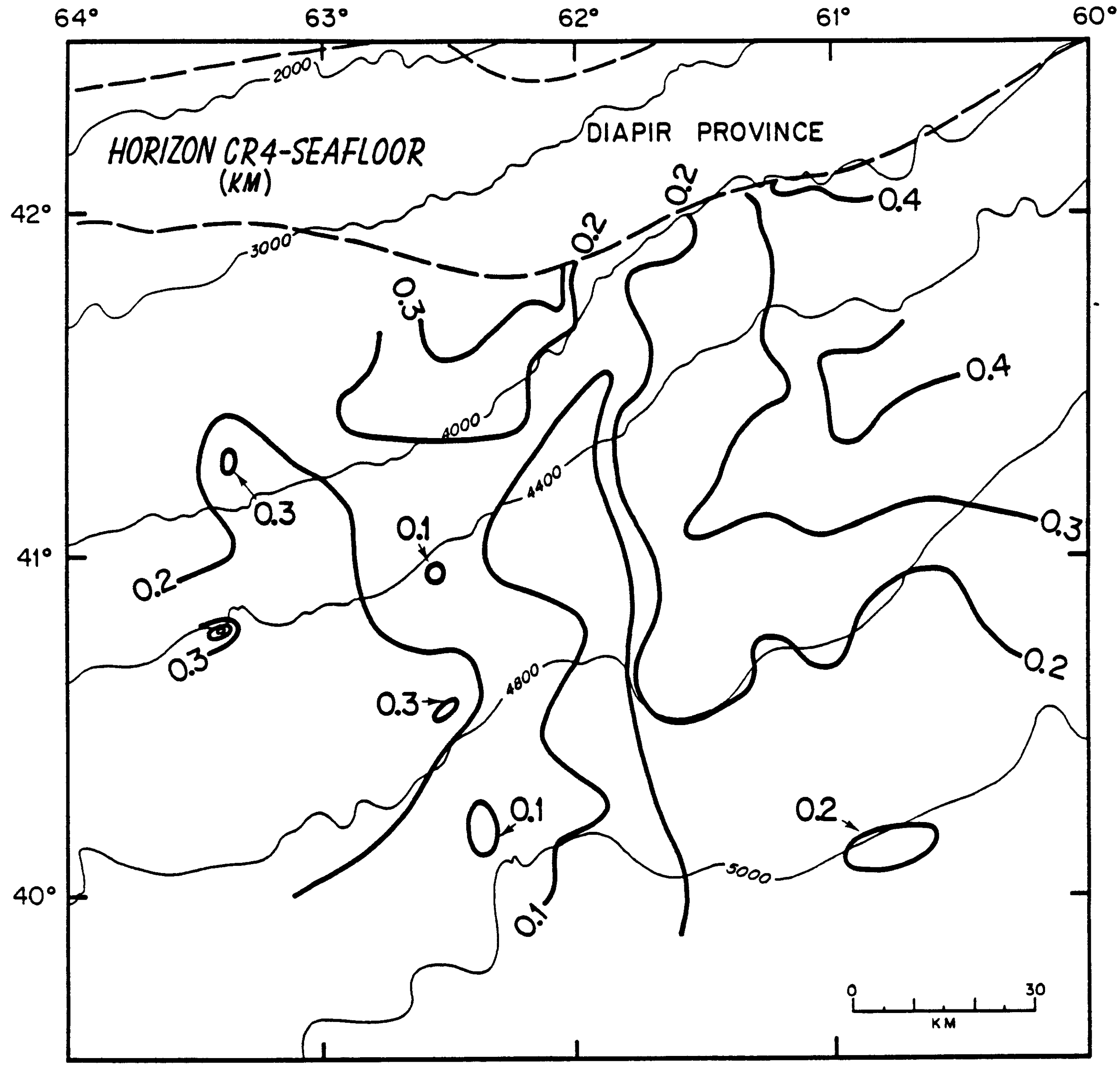

64°
63°
62°
61°
60°
42°
41°
40°
HORIZON CR4-SEAFLOOR
(KM)
DIAPIR PROVINCE
2000
3000
4000
4400
4800
5000
0.1
0.2
0.3
0.4
0
30
KM

Figure 14e

deposited fine-grained sediments landward of seafloor ridges or pillows uplifted by early stages of halokinesis. Seafloor uplift probably also shed mass movements and turbidity currents onto the lower rise. Later, abyssal currents eroded the entire rise, smoothed relief on the continental slope, and sharpened the change in gradient at the base of slope.

Middle-late Miocene

As the shelf sediment wedge prograded during the Tertiary, sediment accumulated in the heads of shelf edge canyons filling them progressively in a seaward direction. Mohican I-100 well, drilled in the axis of Mohican Canyon near the paleo-shelf edge, recovered 1410 m of Pliocene-Pleistocene and 80 m of middle-to-upper Miocene shale with sandstone interbeds lying uncomformably upon Cenomanian shale (Poag, 1982). This sequence indicates that filling began here in the middle Miocene.

After erosion of Horizon CS2, the locus of deposition on the upper rise shifted landward. Seismic reflectors immediately overlying Horizon CS2 are flat, continuous, and parallel. These reflectors diverge landward and lap onto Horizon CS2. With time, they progressively reduced the break in gradient at the base of the slope (Fig. 8). This section probably was deposited from turbidity currents derived from numerous sources along the shelf-edge. Remnent shelf edge canyons cut in the Paleocene and Eocene-Oligocene may have funneled sediment seaward, however, continuation of these structures onto the paleoslope could not be resolved with the seismic data available. Seaward, reflectors above Horizon CS2 are truncated by later erosion but spatial relationships suggest that they extended no farther basinward than the diapir province. In the diapir province, the seafloor gradient probably increased, and as a result turbidity currents bypassed this portion of the rise.

The landward shift in depocenter indicated by the reversal of direction of reflector divergence across Horizon CS2 was probably due to changes in seafloor gradient rather than transport mechanism (primarily turbidity currents, see above) or sediment grain size (mudstone; Fig. 7) because the latter are uniform across Horizon CS2. Deeper erosion on the upper rise rather than on the slope may have increased the change of gradient at the boundary between continental slope and rise, and, thus, caused turbidity currents to deposit their load at the base of slope.

Just seaward of the diapirs, hyperbolic reflectors form a horizontal, fuzzy reflector in both single and multi-channel profiles immediately above Horizon CR1 (Figs. 9, 13b; also Emery et al., 1970, their Fig. 17). Based on similar seismic character and stratigraphic position this horizon may correlate with Horizon X, a time transgressive Miocene reflector identified off the United States east coast (Fig. 4; Markl and Bryan, 1983; Mountain and Tucholke, in press).

Horizon CR2 is an erosional seismic unconformity and a relatively high-amplitude, continuous reflection with an occasional wavy form (amplitudes of 80-100 m and apparent wavelengths of 1-2 km). Horizon CR2 can be traced only on the lowermost rise (Fig. 13). Landward of the 4800m isobath (~6.4 s reflection time), Horizon CR2 increases dip and loses amplitude and continuity until it can no longer be traced within a layer of incoherent reflections (Fig. 13c). Horizon CR2 may correlate with Horizon X observed higher on the lower rise (Fig. 4).

About half of the sediment wedge above Horizon A^u is formed by the layer between Horizon CR1 (lower Miocene?) and Horizons CR3 (Pliocene?) or CR4 (top of Pliocene?) (compare Figs. 14a and 14c). This interval was probably formed by deposition from sediment mass movements and turbidity currents originating primarily from uplift of the seafloor in the diapir province. The relatively low seismic velocity gradient for the interval (Table 2) suggests rapid deposition. On the lowermost rise, this interval fills a low in Horizon CR1 (Fig. 12b). Isopachs for the interval (Fig. 14c) and structural contours on Horizon CR3 (Fig. 12c) bulge southeastward suggesting a fan. Deposition by abyssal currents is unlikely because deposits from currents (eg. Markl and Bryan, 1983; Mountain and Tucholke, in press) are typically much less chaotic and do not have the fan-shaped form of this layer. Coherence of reflectors within the intervel increases seaward of the 4700m isobath (~6.3 s reflection time; Fig. 13c) possibly marking a change from mass movement to turbidite deposition.

Within the chaotic interval, "fuzzy" reflectors of short lateral extent along strike (15-35 km) occur having flat bottoms and narrow peaks 5-8 km wide (Fig. 15). These structures are interpreted as abandoned distributory channels subsequently covered by fine-grained hemipelagic sediments and, later, by more channel deposits (compare with Droz and Bellaiche, 1985, their Figure 10b; Graham and Bachman, 1983, their Fig. 10b; Damuth et al., 1983, their figs. 2 and 3). Most of these reflectors cannot be traced confidently between strikelines set ~35 km apart.

The difference in seismic character between Miocene sequences on either

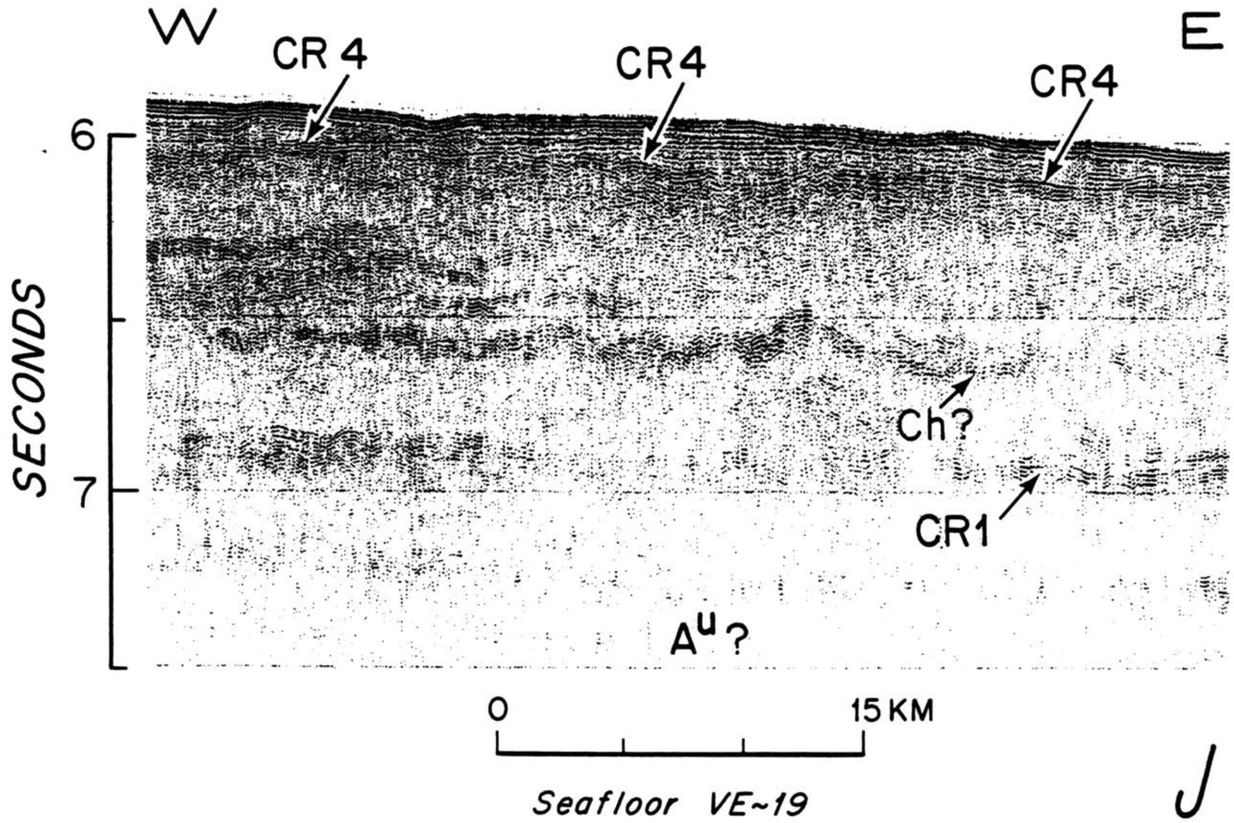


Figure 15 Single-channel water-gun profile (80-170 Hz) seaward of diapir province. Location in Fig. 5 (Profile J). Seismic horizons discussed in text are indicated. "Fuzzy" reflector at ~6.6 s is interpreted as a buried leveed channel (see text).

side of the diapir province indicates different sedimentary environments. Landward of the diapirs, turbidity currents deposited sediment primarily at the base of slope, thus, building an upper rise sediment wedge that lapped onto and eventually buried the continental slope. Seaward, instability due to laterally and temporally variable uplift in the diapir province caused pulses of mass movements and turbidity currents. Brief episodic erosion by abyssal currents occurred during fan construction to form hyperbolic reflections of Horizon X (periodic bedforms?) and wavy reflections on Horizon CR2.

Pliocene

Pliocene sediments are poorly sampled by boreholes on the Scotian shelf (Hardy, 1975). Seismic data between 61°-64°W suggest that seaward progradation of the shelf clinoform sequence continued through the Pliocene. Thin topset beds, probably muddy sandstones deposited in shallow water, are found near the Oneida O-25 well but were not sampled during drilling. Seaward of the paleo-shelf, fine-grained turbidites and hemipelagic sediments mudstones at the Schubenacadie H-100 well accumulated on the paleoslope and uppermost rise (Fig. 7). On the lower rise, deposition from mass movements and turbidity currents continued into the Pliocene.

The first change in Pliocene accumulation patterns occurred on the lowermost rise. Horizon CR3 marks a prominent change in seismic character. Reflectors below are typically chaotic while reflectors above are thin, closely spaced, seaward dipping, and are continuous over relatively large distances (Fig. 13a and c). Underlying reflectors are occasionally truncated, and overlying reflectors terminate on Horizon CR3 by both onlap and downlap. Horizon CR3 dips gently seaward with a southerly bulge of about 50 km width and 200-300 m height (Fig. 12c).

There is no age control for Horizon CR3. Because seismic evidence suggests that the horizon and the overlying interval were formed by abyssal current deposition (see below), the horizon may correlate with an increased influx of sediment to the continental rise upstream (northeast) caused by a shift of shelf and slope sediment depocenter eastward to the Laurentian channel/fan system in the Pliocene-Pleistocene (Jansa and Wade, 1975) Thus, Horizon CR3 and the overlying current deposits are tentatively assigned an age of Pliocene (Fig. 4).

Numerous internal unconformities and periodic sediment waves occur between Horizons CR3 and CR4 (Fig. 13c). Wave crests migrate updip. These waves have typical amplitudes of 30 m (maximum about 55 m) and apparent wavelengths of 6 km. Shortest observed wavelengths were on a north-south line, but the data are insufficient to reliably measure wave orientation. The thickness of the layer between Horizons CR3 and CR4 (Fig. 14d) increases northeastward and southwestward of the bulge in the Horizon CR3 surface (Fig. 12c). Landward, CR3 merges with CR4, and the interval between them pinches out on incoherent reflections of the interval below (Fig. 16).

Abyssal currents probably deposited the sediments between Horizons CR3 and CR4. Thin, continuous reflectors and migrating sediment waves are common characteristics of current deposits (Markl and Bryan, 1983; Tucholke and Laine, 1983). Thin reflectors and waves are also known from turbidite

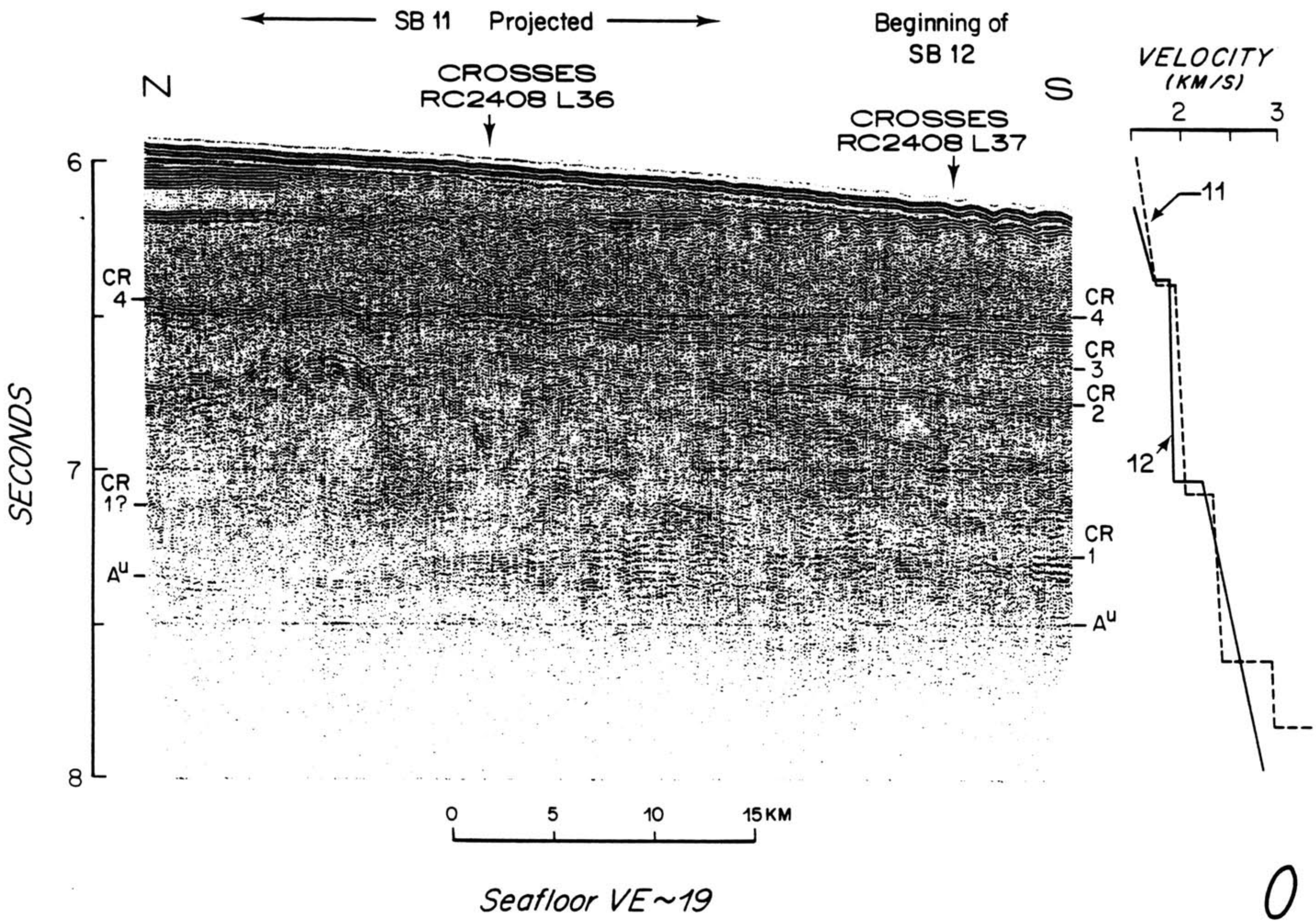


Figure 16 Reflection profile on lower rise illustrating seismic horizons and upper portion of seismic velocity functions from nearby sonobuoys (11 and 12). Location in Fig. 5 (Profile O). Single-channel profile shot with water-gun and filtered at 80-170 Hz. Sonobuoys were shot on crossing lines where indicated. Velocity functions plotted vs. travel time below seafloor where reflection profile crossed sonobuoy shooting line. Flat, incoherent units above Horizon CR4 are interpreted as olistostromes.

levee deposits (eg. Normark et al., 1980), but such an origin for the layer between Horizons CR3 and CR4 is unlikely because reflectors consistently lap onto the underlying interval in an updip direction. The bulge of chaotic reflectors that separates the layer into two ponds could not have been a channel source because sediments eventually overtop the bulge (Fig. 17).

The next change in Pliocene accumulation pattern occurred on the upper rise. Erosion of the seaward edge of the Miocene-Pliocene depocenter in the Schubenacadie basin began during the late Pliocene forming Horizon CS3 (Fig. 8). Horizon CS3 dips seaward at 1-2°, roughly parallel with the present seafloor, and merges with Horizon CS2 near the landward edge of the diapir province. Landward, Horizon CS3 is conformable with deeper beds and merges with shelf clinoform beds beneath the present shelf edge. Structural contours on Horizon CS3 do not follow the trend of deeper horizons, and the position of the continental slope at Horizon CS3 is not clearly recognizable (Fig. 10c). In strike profiles, irregular dips of Horizon CS3 (Fig. 18b) differ from the smooth surface of Horizons CS2 or A^u suggesting that abyssal currents probably were not the cause. Cut-and-fill structures are also absent in strikelines indicating that canyon cutting is also an unlikely explanation. Sharp changes in stratigraphic level of Horizon CS3 on scales of 10-20 km along strike are consistent with scales of Quaternary mass sediment failures along bedding planes (eg. Embley, 1980). Erosion of Horizon CS3 began on the middle rise and progressed landward. Failure probably began with slumping seaward off local highs above salt ridges or diapirs leaving an unstable scarp. Subsequent failure moved the local scarp and the locus of erosion upslope decreasing the sharp change in gradient

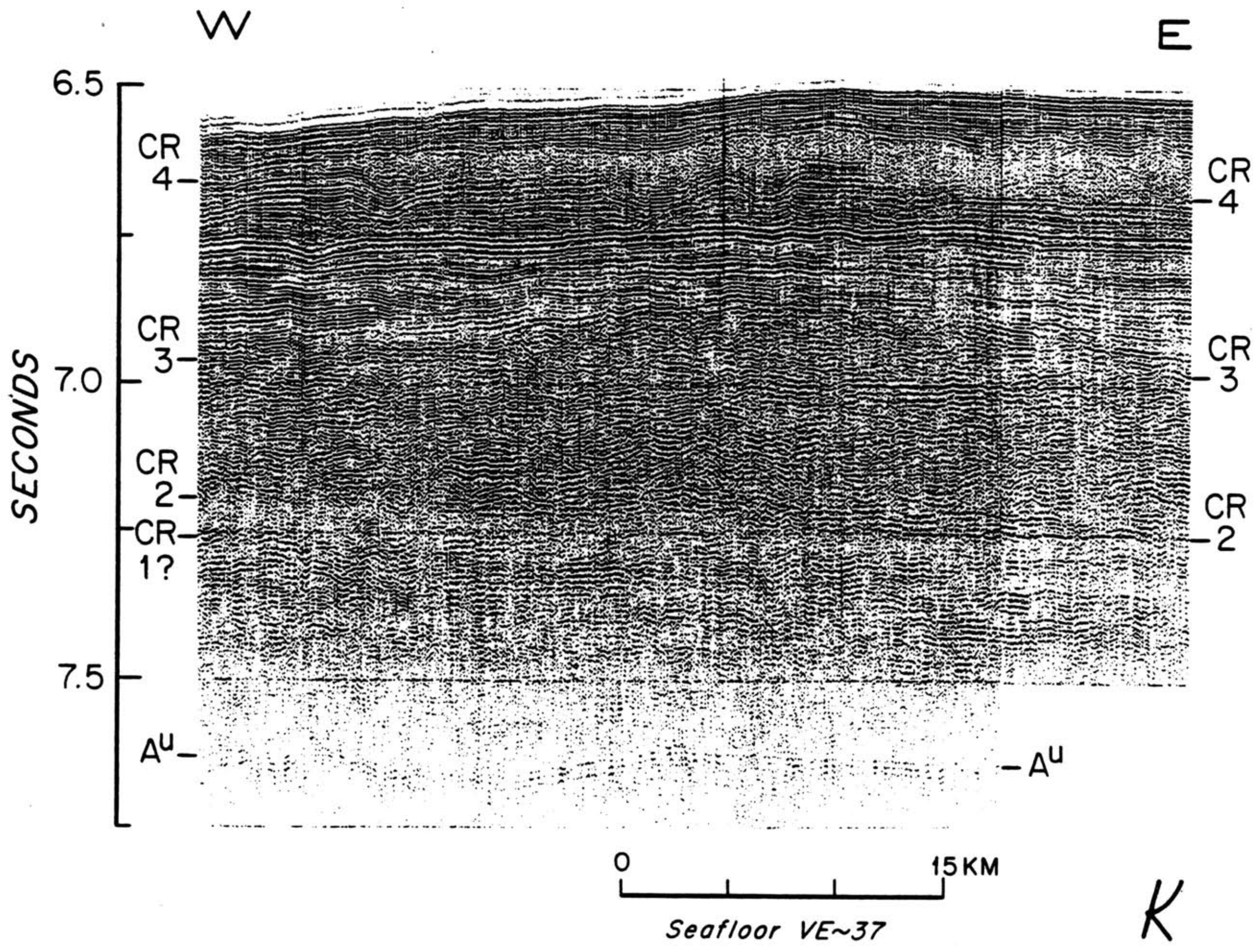


Figure 17 Single-channel, water-gun reflection profile (70-170 Hz) illustrating seismic horizons on lower rise. Location in Fig. 5 (Profile K). Note onlap of reflectors between Horizons CR3 and CR4 onto broad mounded region of incoherent reflectors below Horizon CR3. Horizon CR2 deepens west to east and truncates Horizon CR1 (see also Fig. 11b). On left side of profile, incoherent interval above Horizon CR4 pinches out westward.

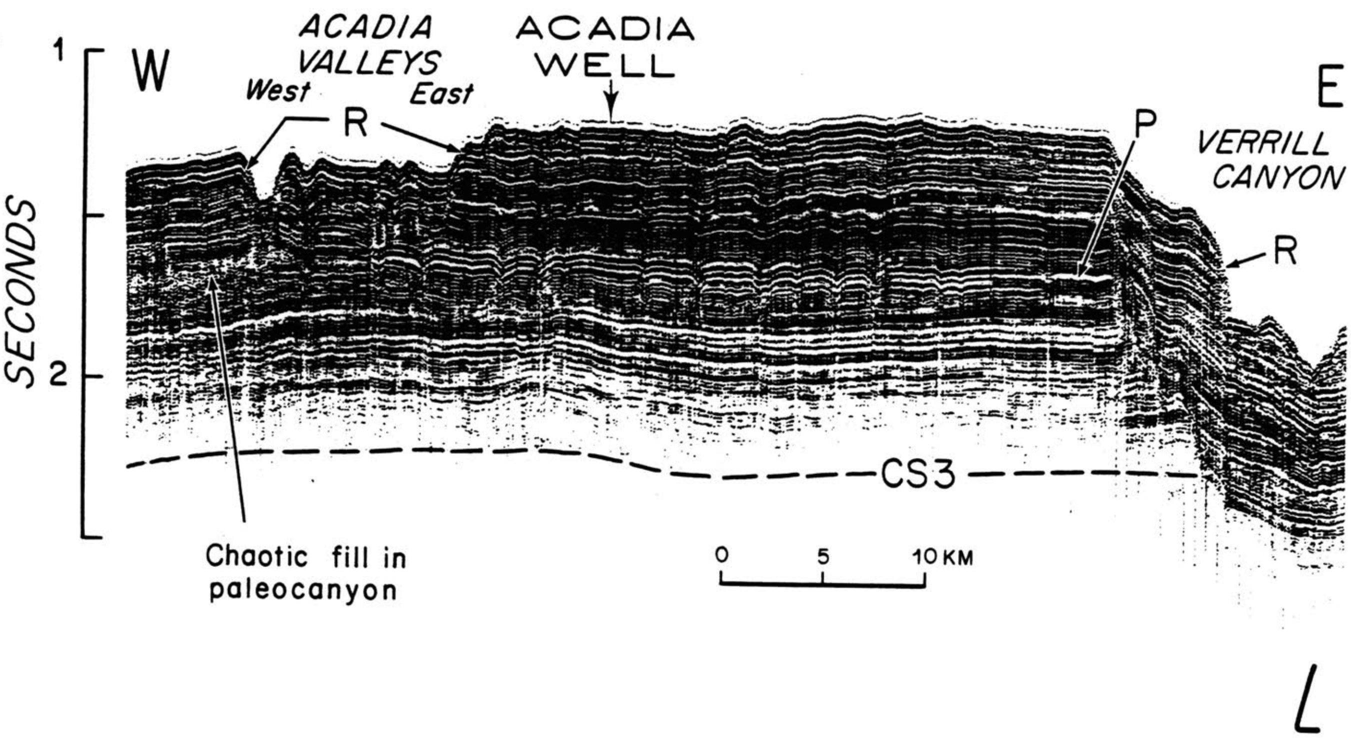


Figure 18 Single-channel water-gun profiles (80-170 Hz) showing unconformities cut in Pleistocene. Locations in Fig. 5. Vertical exaggeration of seafloor about 16. (a) Profile L across slope region studied in more detail by Piper et al. (in press). Note draped seismic character and reflectors (R) cropping out on valley walls. P indicates unconformity dated to Pliocene-Pleistocene boundary inferred to be time of cutting of Verrill Canyon (see text and Fig. 19). Position of Horizon CS3 based on crossing dip lines. (b) Profile M across partially filled Pleistocene canyons on upper rise. (See also Hill, 1983, 1984). Solid line above Horizon CS3 indicates approximate walls of buried canyon. Canyons lose relief landward and cannot be traced confidently to either Mohican or smaller canyons.

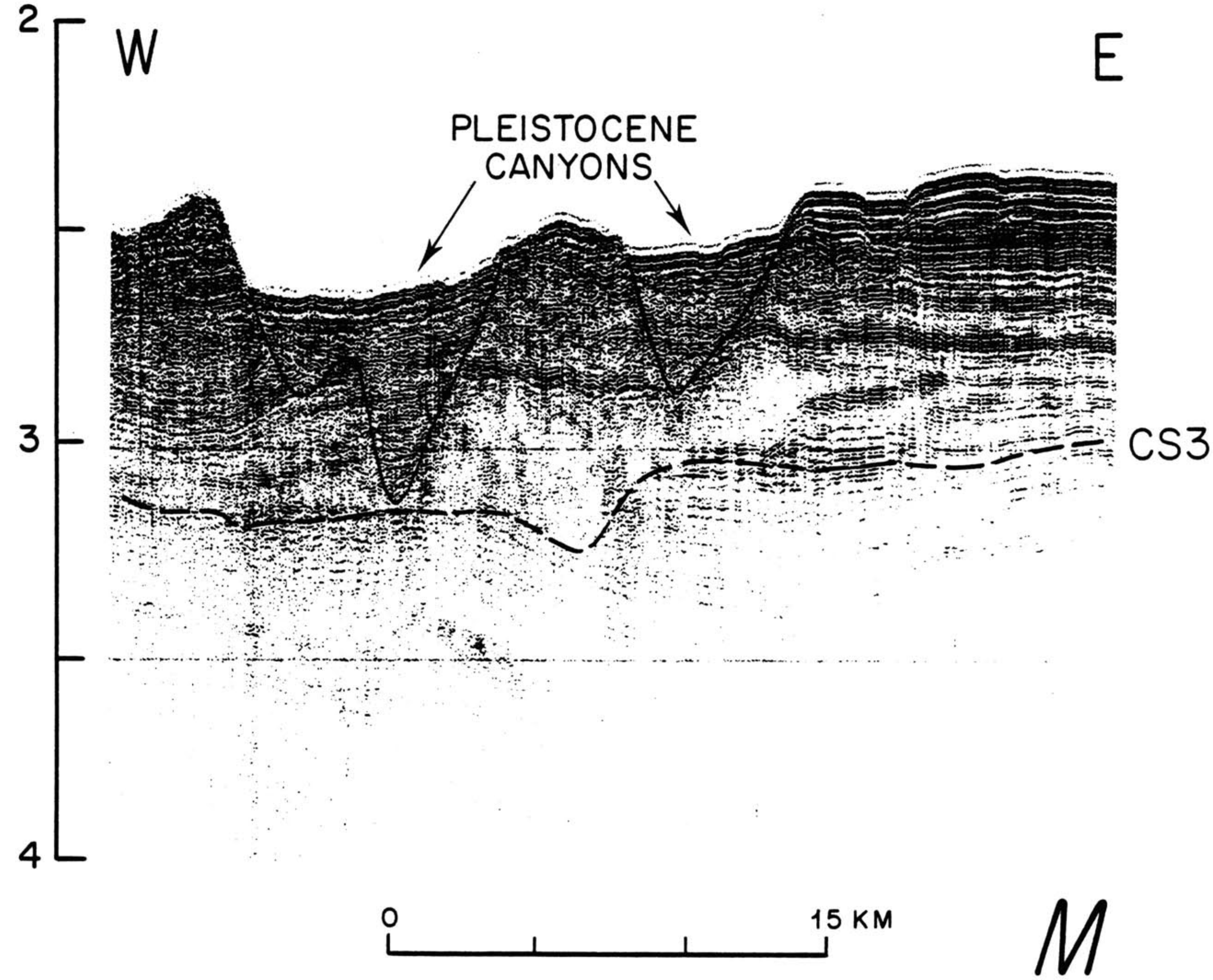


Figure 18b

caused by uplift in the diapir province. Cashman and Popenoe (1985) describe a similar erosional pattern associated with diapir movement off North Carolina.

Seaward of the diapir province, Horizon CR4 marks a prominent upward change in seismic character from continuous reflectors to seismically incoherent layers interpreted below as olistostromes (Figs. 13a and c, 16, 17). Horizon CR4 is traceable on the lower rise as an irregular unconformity with relief of 50-100 m over less than 10 km and, closer to the diapirs, as a distinct, smooth reflector. The surface dips gently southeastward and flattens seaward of the 4600 m isobath (~6.1 s reflection time; Fig. 12d). Rise Horizon CR4 is tentatively correlated with slope Horizon CS3 (Fig.4) based on travel time below seafloor, similar seismic character, ties (however uncertain) through the diapir province, and similar sedimentary processes (see below). Horizon CR4 correlates along GSI 108 (Swift et al., in press) with Horizon L, the base of the latest sequence in the Laurentian Fan, tentatively dated to the Pliocene-Pleistocene boundary (Uchupi and Austin, 1979; Normark et al., 1983; Piper et al., 1984).

In summary, during the Pliocene progradation of the shelf clinoform sequence onto the slope began to move the shelf edge seaward from its position established in the Late Cretaceous. Onlap by upper continental rise turbidites buried the Paleogene paleoslope and formed the smooth physiographic transition from shelf to rise that is apparent in the present seafloor west of 61°30'W. On the lowermost rise, sediments began to accumulate in migrating bedforms from abyssal currents. Near the end of the Pliocene, a tectonic pulse in the diapir province caused a rapid increase in seafloor instability and erosion by mass movements. Erosion, affecting sediments as old as lower(?) Miocene, migrated updip onto the upper rise. Eroded sediments were deposited on the lower rise as olistostromes.

Pleistocene

Upper Pliocene-lower Pleistocene deposits are sparse in Nova Scotia and on the inner Scotian shelf (King, 1970; King and MacLean, 1976; Grant, 1977; King and Fader, undated). On the outer shelf off western Nova Scotia, Quaternary sediments reach thicknesses of 200 m (King et al., 1974), and most were deposited in the last 150,000 yrs (King and Fader, undated). Although continental drainage patterns probably diverted seaward transport of terrigenous sediment through Laurentian and Northeast channels, Pliocene-Pleistocene deposits up to 1 km thick seaward of the shelf edge (see below) indicate significant sediment transport directly across the Scotian shelf.

In the Late Tertiary-early Pleistocene, fluvial and glacial erosion cut into the middle Scotian shelf following the surface subaerially eroded during the late Eocene (Fig. 4; King et al., 1974). Isolated cuts were breached through outer shelf banks. The morphology of this surface suggests that fluvial sediment could not reach the shelf edge during Pleistocene low sea level stands except through the widely spaced breaks in the outer shelf. West of ~61°W, erosion of the tops of outer banks may have supplied sediment to the continental slope, while to the east drill data from Sable Island Bank (Boyd, 1985) indicate that the outer shelf was a depocenter.

During the Pleistocene, sediment probably reached the shelf edge via grounded ice sheets or thin, floating ice shelves derived from local ice domes similar to that proposed for the Wisconsin by King and Fader (undated). Cross-shelf transport by glaciers during the Wisconsin has been inferred from distribution of gravel on the outer shelf (Stanley and Cok, 1968; Emery and Uchupi, 1972; Schlee, 1973), from extensive seismic, core, and dredge surveys of Wisconsin deposits (King, 1970; King and Fader, undated), and from core and 3.5 kHz stratigraphy on the continental slope and rise (Piper, 1975; Stow, 1977; Swift, 1985). Thick Pleistocene deposits along the shelf edge (Fig. 10f) and on the flanks of slope canyons to the east probably represent hemipelagic accumulation of glacial deposits reworked during rising and falling sea level.

Shelf progradation during the Pleistocene modified Mohican Canyon. Pleistocene sediments filled the head of Mohican Canyon at the Mohican I-100 well (Poag, 1982) while rapid deposition on flanking slopes extended the canyon physiography seaward. Between 61°20' and 62°20'W, the interval between Horizon CS3 and the seafloor thickens basinward from less than 0.50 km to more than 1 km at the shelf edge (Fig. 10f). The present form of Mohican Canyon, thus, had a composite origin (Shepard, 1952; Rona, 1970).

Seismic character, isopachs, and absence of major canyons suggest that hemipelagic sedimentation and low density turbidity currents (sensu Rupke, 1978) probably deposited the Horizon CS3-seafloor interval between 61°20-62°30'W. Reflectors are generally parallel, continuous, and drape over topography (Fig. 18a). Although the depocenter thins farther seaward by surface and internal pinchout (Fig. 10f), reflectors extend seaward more than 50 km from the paleo-shelf edge across erosional features on Horizon

CS3 in the Schubenacadie basin and well into the diapir province. High resolution studies of the upper portion indicate that thin slide intervals also are present (Piper et al., in press). West of 62°20'W, well-stratified intervals dip irregularly or occur in isolated blocks, and chaotic reflections fill deep V-shaped canyons (Fig. 18b; Hill, 1983, 1984). The stratigraphic position of the unconformity(s) cutting these canyons is uncertain.

At least two episodes of slope canyon cutting occurred after Horizon CS3. Erosion during the first episode formed small V-shaped canyons and cut-and-fill structures within the drape interval (Horizon P in Fig. 18a). This unconformity can be traced to the edge of Verrill Canyon where it dips sharply eastward cutting Tertiary intervals; subsequent intervals drape over irregularities formed at the time of unconformity (Fig. 19). West of 61°20'W, hemipelagic drape (~500 m thick) subsequently filled and covered these features. An age of 1.6-1.7 Ma (Pliocene-Pleistocene boundary) is estimated for this unconformity using an accumulation rate of 300 m/my consistent with Wisconsin accumulation rates (Stanley et al., 1972; Piper, 1975; Stow, 1979; Hill, 1983, 1984) and the thickness of Pliocene-Pleistocene fill in Mohican Canyon. This event may correlate with shelfwide erosion in the late Tertiary-early Pleistocene (King et al., 1974).

During the second episode, erosion truncated reflectors in the drape interval estimated to be upper Pleistocene on the basis of accumulation rates (Horizon R in Fig. 18a). Small V-shaped canyons were eroded in the drape interval and deeper canyons were re-excavated. The erosion predates a drape interval less than 90,000 yr old that can be mapped in 3.5 kHz echograms (Swift, 1985).

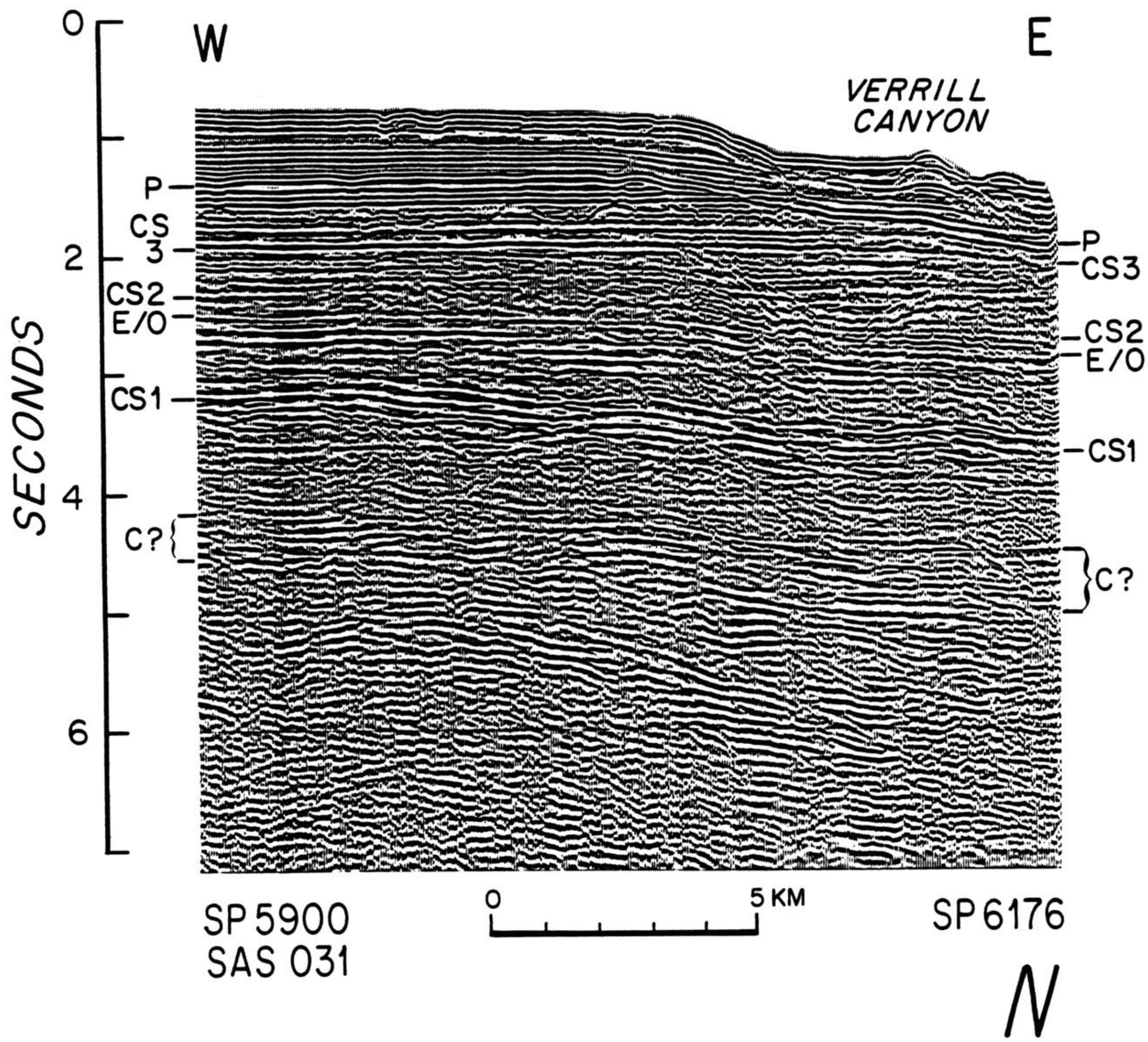


Figure 19 Multi-channel profile across west wall of Verrill Canyon. Location in Fig. 5 (Profile N). P indicates dipping unconformity (Pliocene-Pleistocene boundary) interpreted to be trace of the major canyon erosion episode (see also Fig. 18a). Note gentle dip caused by oblique crossing of Horizon CS1 and reflectors traced to base of Neocomian carbonate escarpment (C). Vertical exaggeration of seafloor about 3.

The morphology of Pleistocene unconformities suggests that erosion was caused by turbidity currents provoked by high sediment flux delivered to the shelf edge by continental ice sheets or floating ice shelves. Lack of high-relief slope canyons west of 61°W suggests that the western Scotian shelf edge was fed by a smaller ice source or was more distant from the ice edge.

Seaward of the diapir province, Pleistocene thickness variations (Horizon CR4-seafloor interval) trend north-south, unlike deeper layers (Fig. 14e). Three deposits with thicknesses of 300-400 m are separated by deposits of only 50-150 m thickness. This interval is formed by one or more homogeneous seismic units separated by thin, relatively high amplitude, continuous reflectors (Fig. 16). Thicker deposits have more homogeneous layers. Each unit flattens seaward. The layers in the eastern deposit appear to continue towards the east, but terminate abruptly to the west and southwest (Fig. 16, 17). Although the units are similar in seismic character, abrupt edges, irregular tops and bulging contours to Quaternary slide deposits originating by mass sediment failure on continental slopes (eg. Stuart and Caughey, 1977; Embley, 1980; Damuth and Embley, 1981), they are generally thicker and broader. These deposits are analogous in size to olistostrome deposits in the eastern Atlantic that were activated by Alpine tectonics in northern Morocco and southern Spain (Watkins and Hoppe, 1979; Emery and Uchupi, 1984). Off Nova Scotia, the slide units probably originated as a result of rapid uplift in the diapir province.

Other explanations for the homogeneous units cannot explain all their features. Channel deposits often return poorly coherent reflections, but migrating channels are unlikely to have formed deposits as broad as the homogeneous units. Current deposits may lack coherent reflectors, especially in single-channel profiles (eg. Ewing and Hollister, 1972), but abyssal currents could not form abrupt edges trending down slope. Erosion of the western edge by turbidity currents is possible, but it seems unlikely that turbidity currents traversing the rise in such a narrow zone could also be so persistently erosive while depositional elsewhere.

Continuous, thin reflectors draping underlying structures form the bulk of the northern deposit as well as the Horizon CR4-seafloor layer between principal deposits (Figs. 13b, 15). Reflectors within the northern deposit continue landward into the diapir province where they are deformed by faulting and folding, but to a lesser degree than deeper horizons (Fig. 3). The continuous, draped reflector character of the northern deposits correlates across the diapir province to the Pleistocene section on the upper rise and slope. Low-density turbidity currents spilling over banks of shallow channels and hemipelagic processes probably deposited these sediments.

The western deposit is topped by a well-developed, low-amplitude leveed channel (Shor and Lonsdale, 1981; Shor et al., 1984). This surface morphology is probably a late Pleistocene overprinting by turbidity currents (Swift, 1985). Reworking by late Pleistocene abyssal currents probably formed "contourite-fan" features observed in 3.5 kHz echograms from the eastern deposit (Swift, 1985).

In summary, fluvial and glacial processes eroded the present shelf morphology in the early Pleistocene. Later, this morphology controlled the movement of grounded ice sheets and floating ice shelves as they crossed the shelf. Deep canyons were eroded east of 61°W where glacial sediment was input directly across the shelf edge. Elsewhere, outer banks prevented most glacial sediment from directly reaching the slope, but submarine reworking of these banks formed a thick drape interval on the slope and upper rise. In the diapir province, seafloor uplift continued episodically shedding olistostromes seaward onto the lower rise. Turbidity currents originating in slope canyons built leveed channels on the lower rise. Abyssal currents reworked but did not significantly redistribute mud along the lower rise.

CONCLUSIONS

Whereas off eastern Nova Scotia the Late Jurassic-Neocomian carbonate platform was buried by the Early Cretaceous Sable Island delta, on the LaHave platform the seaward edge of the carbonate platform remained exposed through the Late Cretaceous. Between Berriasian and Maestrichtian the shelf edge retreated 10-20 km landward of the platform edge. This retreat resulted from a slow rate of sediment supply. Concurrent with this retreat, terrigenous mudstones, probably transported by turbidity currents through gaps in the carbonate platform, lapped onto the seaward side of the carbonate platform.

After deposition of the Wyandot chalk in late Campanian-early Maestrichtian, a sediment wedge intermittantly prograded seaward across the outer shelf reaching the shelf edge in the Miocene-Pliocene. Headward erosion of Mohican Canyon, a paleo-shelf edge canyon with up to 1.5 km relief, reached a maximum in the late Maestrichtian-early Paleocene.

Although seaward filling of Mohican Canyon began immediately, the slopes flanking the mouth of the canyon also built up preserving a vestigal canyon form to the present.

On the upper rise, chalks interbedded with mudstones were deposited during the Paleocene-Eocene in fans 5-10 km in width and up to 0.5 km in thickness. Slope canyons linking the fans to the shelf edge are not present; their relief was probably smoothed by current erosion in the Miocene.

In the late Eocene, subaerial erosion cut fluvial morphology on the shelf. Somewhat later in the Oligocene, abyssal currents flowing along bathymetric contours eroded and smoothed the continental rise between ~2000 and 5100 m present water depth forming an unconformity which can be correlated to Horizon A^u southwest of the New England Seamounts. The relationship to shelf erosion in the late Eocene-Oligocene is unclear.

Seismic facies above Horizon A^u differ on either side of the diapir province (~2200-3800 m present water depth) suggesting that movement of salt (and shale?) began as late as Miocene. Landward of the diapirs, thinly bedded, continuous mudstones were deposited in the Miocene by turbidity currents from low relief channels. Seaward, well-defined channel facies in seismic profiles have relief of 100-300 m and widths of 5-10 km.

During the Miocene, abyssal currents again eroded the rise on either side of the diapirs. Seaward of the diapirs, the erosion surface mimicked Horizon A^u, but probably did not eroded as thick a stratigraphic section. Landward of the diapirs, erosion incised the base of slope-upper rise. Afterwards, upper rise turbidites lapped onto the current-cut unconfomity and smoothed the slope-rise transition by the Pliocene.

Seaward of the diapirs, Neogene and Quaternary sediments were deposited principally by down slope processes. Although evidence of current erosion occurs throughout the section, only a thin (<300 m) Pliocene(?) interval on the lowermost rise can be unambiguously ascribed to current deposition.

In the Pleistocene, up to 1 km of hemipelagic or unchanneled turbidity current deposits derived from reworked glacial shelf deposits accumulated on the continental slope and upper rise. Accumulation was interrupted at least twice in the earliest Pleistocene and in the late Pleistocene by erosion of small, V-shaped canyons when glacial ice advanced across the outer shelf banks and delivered sediment directly to the continental slope. Diapirism accelerated in the Pleistocene and shed olistostromes up to 300 m thick seaward onto the lower rise. Abyssal currents and channel-forming turbidity currents reworked deposits on the lower rise during the latest Pleistocene.

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TABLE 1. Interval velocity and depth vs. travel time functions

Oneida O-25

$V = 1.65^{*} + 0.90\ T$ (seafloor to top of Logan Canyon)

$D = 0.83\ T + 0.11\ T^2$

Schubenacadie H-100

$V = 1.66^{*} + 0.29\ T$ (seafloor to top of Paleocene)

$D = 0.83\ T + 0.073\ T^2$

General function for lower rise from sonobuoys (Table 2; Swift, 1986)

$V = 1.55^{*} + 0.52\ T$ (seafloor to horizon A^u)

$D = 0.78\ T + 0.13\ T^2$

V = interval velocity (km/s)

T = two-way travel time (s) below seafloor

D = depth below seafloor (km)

* assumed surface sediment velocity

TABLE 2. Seismic velocity models for RC2408 sonobuoys derived by ray tracing fits to travel times.

			SONOBUOYS														
			9			11			12			15			16		
AYER		REFLECTOR	DEPTH km	VELOCITY km/s	GRAD s^{-1}	DEPTH km	VELOCITY km/s	GRAD s^{-1}	DEPTH km	VELOCITY km/s	GRAD s^{-1}	DEPTH km	VELOCITY km/s	GRAD s^{-1}	DEPTH km	VELOCITY km/s	GRAD s^{-1}
1	Top		0.00	1.50		0.00	1.51		0.00	1.50		0.00	1.51		0.00	1.52	
	Bottom	Seafloor	3.94	1.50		4.43	1.51		4.48	1.50		4.57	1.51		4.64	1.52	
2	Top		3.94	(1.55)	0.7	4.43	(1.55)	0.6	4.48	(1.55)	0.9	4.57	(1.65)	0.4	4.64	(1.65)	0.4
	Bottom		4.19	(1.73)		4.79	(1.75)		4.69	(1.73)		5.06	(1.85)		5.08	(1.85)	
3	Top		4.19	1.74	0.1	4.79	1.94	0.1	4.69	1.90	0.08	5.06	1.92	0.1			
	Bottom	Horizon 1	4.81	1.82		5.51	2.04		5.34	1.95		5.48	1.97				
4	Top		4.81	2.13	0.2	5.51	2.33	0.2				5.48	2.32	0.1			
	Bottom	A^U	5.42	2.25		6.19	2.45					5.98	2.39				
5	Top		5.42	2.90	0.3	6.19	2.97	0.06				5.98	2.90	0.3	5.08	2.00	0.2
	Bottom	Beta?	6.33	3.18		6.51	2.99					6.89	3.15		6.48	2.32	
6	Top		6.33	3.42	0.4	6.51	3.38	0.3	5.34	2.25	0.5				6.48	3.15	0.2
	Bottom	J1	7.36	3.80		7.59	3.70		7.14	3.10					7.12	3.28	
7	Top		7.36	4.10	0.3				7.14	3.80	0.4	6.89	3.73	0.4	7.12	4.05	0.2
	Bottom		7.88	4.24					7.86	4.10		8.48	4.35		7.95	4.24	
8	Top		7.88	4.30	0.6	7.59	4.01	0.4	7.86	4.12	0.4						
	Bottom	Basement	9.76	5.51		9.30	4.62		9.57	4.80							
9	Top		9.76	5.60	0.5	9.30	4.65	0.5	9.57	5.0	0.6				7.95	4.27	0.6
	Bottom		12.08	6.81		12.01	6.0		10.77	5.7					11.90	6.80	

( ) = assumed velocity

APPENDIX

SEISMIC VELOCITIES

INTRODUCTION

Seismic velocities are used in this thesis to calculate depth from travel time, to infer sediment composition and history, and to correlate seismic profiles with well logs. The purpose of this appendix is to review published seismic refraction results, to describe results of new velocity measurements, and to describe the analytical techniques used.

Data for new velocity measurements are from (1) seven of 34 sonobuoys shot during 1983 on a R/V ROBERT CONRAD (RC2408) survey off western Nova Scotia (Fig. 1, Table 1), (2) sonic logs of 5 wildcat wells (Fig. 1, Table 2), and (3) stacking velocities computed from industry multi-channel reflection data (Fig. 1, Table 3).

PREVIOUS MEASUREMENTS

Houtz (1984a,b) compiled seismic refraction results off Nova Scotia (Fig. 1). From this review, the only previously published data in this study area are reversed two-ship measurements shot with explosives along bathymetric dip lines in the 1950's by Lamont Geological Observatory (Officer and Ewing, 1954) and sonobuoys shot in 1968 with a sparker-40 cu in air gun array (Emery et al., 1970). Farther east, refraction data were obtained along bathymetric strike lines using a 1000 cu in air gun and explosive charges by reversed shooting to tape-recording sonobuoys (Keen et al., 1975; Jackson et al., 1975), and by unreversed shooting to an ocean bottom seismometer/hydrophone (Keen and Cordsen, 1981). In a survey conducted to the southeast, Jackson et al. (1975) report results using a 1000 cu in air gun shot to expendable sonobuoys. Houtz (1980) reduced

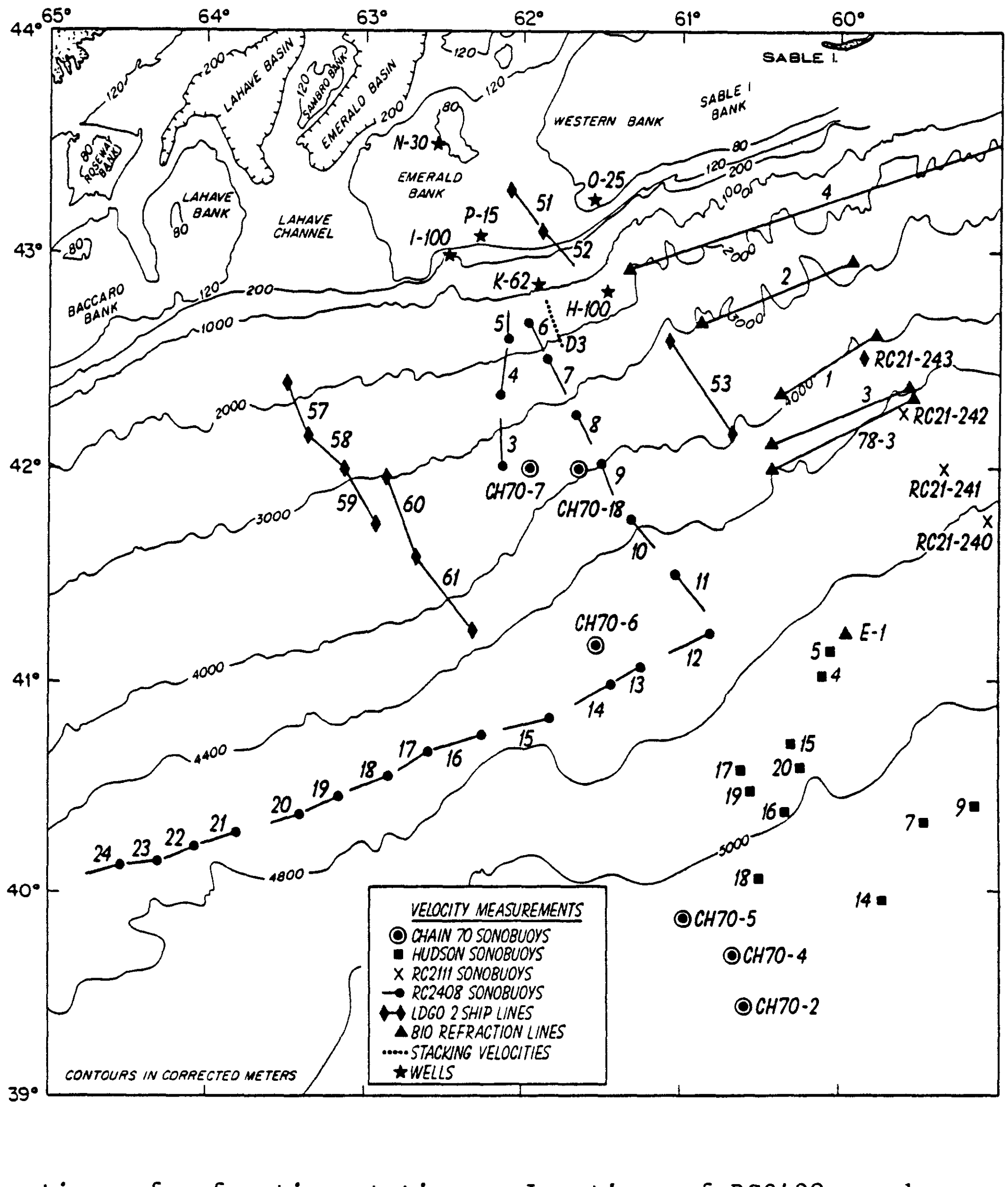


Figure 1 Locations of refraction stations. Locations of RC2408 sonobuoys (Table 1) and wells (Table 2) from this chapter. All others from Houtz (1984a).

TABLE 1. Sonobuoys deployed off western Nova Scotia on RC2408.

No.	Date	Time (GMT)	Deployment Location	Type	Hydrophone Depth (ft)	Water start (m)	Depth end (m)	Heading made-good	Range-made-good (km)	Surface velocity (m/s)	Sounding velocity (m/s)	No. air guns
3	6 July	1555-1804	42°0.8'N 62°8.4'W	57A	60?	3405	2870	58°	25			3
4	6 July	1909-2129	42°21.4'N 62°9.0'W	57A	60	2451	1702	6°	27			4
5	6 July	2135-2238	42°36.5'N 62°6.2'W	41B	60	1667	1285	0°	11			4
6	7 July	0317-0531	42°41.3'N 61°58.1'W	41B	60	1473	2312	153°	21	1519	1482	4/3
7	7 July	0538-0743	42°30.8'N 61°51.3'W	57A	60	2328	2898	153°	20	1516	1482	3
8	7 July	0853-1037	42°15.2'N 61°40.5'W	57A	60	3270	3806	151°	17			3
9	7 July	1149-1415	42°2.0'N 61°30.4'W	57A	60	3981	4190	156°	18	1516	1498	3
10	7 July	1538-1711	41°45.6'N 61°19.4'W	57A	60	4344	4442	138°	18			3
11	7 July	1854-2055	41°30.0'N 61°1.8'W	57A	60	4497	4596	138°	24	1527	1506	3
12	7 July	2248-0100	41°13.6'N 61°49.4'W	57A	60	4641	4631	244°	22	1527	1504	3/4
13	8 July	0238-0352	41°4.1'N 61°15.9'W	41A	300	4665	4661	242°	15			4
14	8 July	0408-0551	40°59.1'N 61°26.9'W	57A	60	4669	4670	239°	21			4
15	8 July	0711-0908	40°49.7'N 61°50.3'W	41B	60	4747	4746	256°	23	1533	1511	4
16	8 July	1017-1226	40°45.0'N 62°15.9'W	57A	60	4738	4642	253°	25	1534	1510	4

TABLE 1. (continued). Sonobuoys deployed off western Nova Scotia on RC2408.

No.	Date	Time (GMT)	Deployment Location	Type	Hydrophone Depth (ft)	Water start (m)	Depth end (m)	Heading made-good	Range-made-good (km)	Surface velocity (m/s)	Sounding velocity (m/s)	No. air guns
17	8 July	1251- 1429	40°40.2'N 62°35.5'W	41A	300	4626	4626	236°	19			4
18	8 July	1503- 1641	40°33.1'N 62°51.8'W	41A	60	4629	4632	247°	19			4
19	8 July	1724- 1917	40°27.3'N 63°9.7'W	41B	60	4632	4660	245°	21			4
20	8 July	1924- 2059	40°22.3'N 63°24.8'W	41B	60	4659	4645	252°	17			4
21	8 July	2301- 0051	40°16.7'N 63°48.8'W	41A	60	4668	4717	253°	21			4
22	9 July	0052- 0239	40°13.3'N 64°3.2'W	41A	60	4716	4725	246°	20			4
23	9 July	0251- 0430	40°8.6'N 64°18.5'W	41A	300	4723	4702	263°	19			4
24	9 July	0439-	40°7.3'N	41A	60	4703	4575	254°	18			4

TABLE 2. Exploratory wells on Nova Scotian outer shelf and slope used for stratigraphic correlation.

Well	Location	Water Depth (m)	TD (m) below sealevel	Oldest rock recovered	Sonic log digitizing interval (ms)	Date Completed
Schubenacadie H-100	42°49'28.43"N 61°28'42.81"W	1476	4174	Albian, Dawson Canyon Fm, shale	1	2/11/1983
Oneida O-25	43°14'57.49"N 61°33'36.38"W	82	4085	Callovian, Mohican Fm, brown-brown sandstone and shale	5	2/1/1970
Acadia K-62	42°51'42.54"N 61°55'00.21"W	866	5275	L. Jurassic, Iroquois Fm, oolitic limestone	5	8/2/1978
Moheida P-15	43°4'56.32"N 62°16'44.32"W	112	4269	U. Triassic, Eurydice Fm. red-brown shale and siltstone	5	2/13/1977
Mohican I-100	42°59'39.04"N 62°28'51.32"W	153	4365	U. Hettangian-L. Sinemurian, Argo Fm, salt	5	3/6/1972

TABLE 3. Shotpoints along multichannel dipline 3 at which interval velocities were calculated.

Shotpoint No.	Location	Water Depth (km)
SAS 044 line 212		
4307	42°46.96'N 61°52.40'W	1.29
4373	42°44.91'N 61°51.43'W	1.41
SAW 054 line 212		
4370	42°45.07'N 61°51.58'W	1.43
4403	42°44.02'N 61°51.12'W	1.49
4502	42°40.92'N 61°49.62'W	1.74
4536	42°39.87'N 61°49.12'W	1.79
4700	42°34.83'N 61°46.75'W	2.19
4790	42°31.98'N 61°45.38'W	2.42

seismic velocity measurements in sediments along the continental slope and rise of Nova Scotia and Newfoundland to a single linear regression equation: V(km/s) = 1.62 + 1.19 T, where T = one-way travel time (s).

SONOBUOY ANALYSIS

Eighteen sonobuoys were deployed on RC2408 during survey operations within the study area (Fig. 1). The digitally recorded signal from seven of these buoys was later replayed and displayed on a travel time vs. range plot. Arrival times of coherent wide-angle reflections and refractions were marked and digitized. Layer thicknesses were calculated using the slope-intercept method. Velocity-depth functions were also obtained by modeling travel time arrivals using a forward modeling (ray tracing) method. Ray tracing was used because layer thicknesses, velocity offsets, and velocity gradients could be well-constrained by modeling both reflected and refracted arrivals. Inversion techniques currently available for either reflected or refracted arrivals do not utilize the information around the critical point. It would have been prohibitively expensive to produce models with the same detail obtained with ray tracing by iterative comparison of amplitudes in recorded sections with those in synthetic sections. In addition, the effectiveness of this method would have been poor given the low signal-to-noise ratio in the recorded sections.

Data

Expendable sonobuoys (military SS types 41A, 41B, and 57A) were used with the hydrophone usually set at 60' (Table 1). Three or four 466 cu in air guns were fired at 1800 psi every 20 s (about 60 m). Both fixed gain (about 47 db) and time-varying gain were applied to the sonobuoy signal. The signal was band-pass filtered at 15-200 Hz and recorded on a 19" dry paper recorder at 5 s sweep and 3 "/hr. 12-12.8 s of signal were also low-pass filtered at 300 Hz, digitized at 833 Hz (1.2 msec), and recorded on tape.

Buoy locations were determined by LORAN C corrected by average offset to simultaneous satellite fixes. Range to sonobuoy was determined by direct wave calculation for sonobuoys 6 and 7. For the other buoys, range was determined for the first 10 min. of operation from the direct wave travel time and for the remainder by the seafloor reflection travel time. LORAN C ranges do not account for significant drift away from the direction of ship travel by buoys 6-11. Surface water velocities (Table 1) were determined for each buoy from Mathew's Table 3 correcting for temperature but not pressure. Average surface water temperature was determined by either expendable bathythermograph (XBT) or by hourly engine room intake temperatures corrected by XBT measurements. Sounding velocities were calculated by dividing the Mathew's Table corrected echo-sounding depth by the echo-sounding one-way travel time. Seafloor reflection travel time was digitized from shipboard records every 2 minutes (6 shots) to a resolution of ± 5 ms. The range calculation took into account linear dips of the seafloor where appropriate but ignored seafloor topography, which was generally less than 100 m.

Travel time of the digital sonobuoy signal was reduced using 2.2 km/s. The signal was bandpass filtered (2/8-50/60 Hz) and plotted as true amplitude displayed in filled wiggles at 2"/s and 1.25"/km.

The record section was interpreted by marking wide-angle reflection (both pre- and post-critical) and refracted arrivals. These picks were then digitized on a standard digitizing table to an accuracy of $\pm$ 2 milliseconds travel time and $\pm$ 5 meters in range. In general, few near vertical incidence arrivals could be picked deeper than 2.5 s and few arrivals could be traced beyond 20 km.

Picks of stratigraphically significant reflectors in vertical incidence records were used as a guide in marking wide-angle reflections, but, generally, coincidence between such picks and coherent reflections to better than 0.15 s was uncommon. White (1979) suggests three possible sources of error: (1) decreasing reflectivity of interfaces at lower angles of incidence, (2) ship noise at very short ranges, and (3) the difference between noise and frequency characteristics of sonobuoys and vertical profiling systems. In addition, low signal-to-noise ratio in these sonobuoy data is a source of error. Also, bubble pulse arrivals with magnitudes as great as the first pulse could clearly be identified for the seafloor return and some refracted arrivals. Coherent reflectors arriving at times when vertical incidence records are incoherent suggests that internal multiples are an additional source of noise. The reflection off many of the late Cretaceous to Cenozoic stratigraphic markers used in this study are often lost in multiples or bubble pulse trains off shallower reflections. Reflection amplitude for these markers is probably low because the sequences studied contain few significant lithologic changes except in the Mesozoic section. Unless an unconformity was produced by deep erosion (eg. Horizon A^u), a significant reflection is unlikely and would be difficult to trace. The apparent signal-to-noise ratio was lower in deeper water, perhaps as a result of coarser sediments and rougher microtopography.

Slope-intercept calculation

Velocity-depth functions were calculated using the slope-intercept method for flat, homogeneous, uniform layers (Dobrin, 1976, p. 296-298). Velocity was assumued to increase monotonically downward across discrete interfaces, and all coherent phases which broke through the seafloor arrival were assumed to be arrivals of head waves. The travel time picks were fitted with a straight line by linear regression. The layer thicknesses were then iteratively calculated downward from the apparent velocities and intercepts using the sounding velocity for the first layer velocity. A second calculation was done assuming the thickness and velocity of the water and uppermost sediment layer was known. The results (Table 4) were used as a guide in finding a forward model solution.

Forward modeling

An interactive computer program* was used to determine a velocity-depth model that best matched travel time picks of reflected and refracted arrivals. For a given model, the program uses ray tracing to calculate travel times as a function of range and overlays modeled and picked arrivals on a video screen. The old model can be modified to correct discrepancies, and the program run again until a satisfactory fit is achieved. Shallow layers are modified first, and deeper layers are added incrementally. As a

* L. Gove and G.M. Purdy (WHOI) wrote TTIME.

TABLE 4. Layer thicknesses from slope-intercept calculation.

Sonobuoy	Refraction Layer	Apparent Velocity (km/s)	Intercept (s)	Thickness Model 1 (km)	Thickness Model 2 (km)
6	1	(1.50)		1.50	(1.47)
	2	1.74	1.01	0.21	0.27
	3	1.82	1.20	0.52	0.50
	4	2.21	1.94	0.59	0.59
	5	2.42	2.33	1.17	1.17
	6	2.85	3.18	0.67	0.67
	7	3.64	3.97		
7	1	(1.50)		2.59	(2.29)
	2	1.73	1.72	0.52	1.15
	3	1.82	2.14	0.09	-0.14
	4	2.05	2.72	-0.78	-0.80
	5	2.14	2.65	1.44	1.41
	6	2.45	3.46	1.82	1.80
	7	5.00	5.77		
9	1	(1.50)		4.21	(3.98)
	2	1.82	3.18	0.94	1.30
	3	2.30	4.89	0.42	0.34
	4	3.15	6.03	1.00	0.99
	5	3.83	6.73	0.68	0.68
	6	4.22	7.06	1.68	1.67
	7	5.51	8.00	1.90	1.76
	8	6.83	8.68		

TABLE 4. Layer thicknesses from slope-intercept calculations (continued).

Sonobuoy	Refraction Layer	Apparent Velocity (km/s)	Intercept (s)	Thickness Model 1 (km)	Model 2 (km)	Model 3 (km)	Model 4 (km)
11	1	(1.51)		(4.45)	(4.44)	4.82	
	2	(1.85)		(0.26)			
	3	2.02	4.24	1.16	0.80	0.78	
	4	2.44	5.45	0.26	0.39	0.41	
	5	2.96	6.25	0.91	0.96	0.94	
	6	3.74	7.14	0.43	0.45	0.43	
	7	4.22	7.48	1.50	1.51	1.52	
	8	4.67	7.97	2.16	2.16	2.15	
	9	6.34	9.16				
12	1	(1.50)		(4.62)	(4.62)	(4.75)	4.75
	2	(1.65)		(0.25)	(0.25)		
	3	1.95	4.05	1.22	1.22	1.34	1.34
	4	3.03	6.56	3.80	0.93	0.92	3.80
	5	3.12	7.23	-3.08			-3.08
	6	4.08	7.51	1.87	1.76	1.76	1.87
	7	5.06	8.32	1.14	1.12	1.12	1.141
	8	5.63	8.70	2.21	2.19	2.18	2.21
	9	8.31	9.81				
15	1	(1.51)		5.03	(4.57)		
	2	(1.65)			(0.21)		
	3	2.02	4.42	0.78	1.33		
	4	2.40	5.59	0.16	-0.02		
	5	3.10	6.48	1.09	1.05		
	6	4.20	7.48	1.00	1.00		
	7	4.37	7.67	1.97	1.96		
	8	6.51	8.99				
16	1	(1.50)		5.32	(4.68)		
	2	(1.65)			(0.21)		
	3	2.32	5.41	1.05	1.94		
	4	3.31	6.97	0.62	0.44		
	5	4.24	7.63	3.04	2.96		
	6	7.06	9.27				

check, a plot of the travel times for the complete model solution can be overlain on the record section.

Travel time calculations

The program assumes flat, laterally uniform layers. The velocity within a layer (i) can be homogeneous ($V=V_i$, where V_i= velocity at top of layer) or a linear gradient ($V= V_i + a_i Z$, where a= velocity gradient, Z= depth within the layer). Only compressional wave travel times are calculated. Relative amplitudes and wave forms are not calculated.

The travel time curves for wide angle reflection (both pre- and post-critical) and diving waves are calculated incrementally downward from the surface layer using parametric equations derived by Slotnick (1936). Arrival times of head waves refracted within layers of zero gradient are not calculated. (Model studies suggest head wave amplitudes are much lower than those of diving waves for crustal velocities (Kennett, 1977)). Thus, the program implicitly assumes that waves incident on an interface at angles greater than the critical angle ($i_c= \sin^{-1} (V_i+a_i h_i)/(V_{i+1})$, where h=layer thickness) are totally reflected.

Usage

Additional assumptions were made in this study. In general, attenuation effects were ignored, ie., a decrease in amplitude of a phase arriving before the seafloor arrival was modeled as an outward pointing cusp between a post-critical reflected arrival and diving waves turning in the overlying layer. No decreases in velocity across interfaces were allowed. With one exception (see below) all layers that were included in the model had arrivals which broke out from the seafloor arrival.

A standard modeling procedure was followed. The seafloor was fitted using a constant velocity layer. The uppermost layer was assumed to be a velocity gradient having a thickness of 0.25-0.5 km and an upper velocity of 1.55-1.65 km/s. Layers based on pre-critical wide-angle reflections alone were rarely considered because experience showed that velocities required to fit most such picks were anomalously low suggesting that they were multiple or bubble pulse arrivals). The thicknesses of layers were constrained by the travel time to post-critical reflections and by the range to which reflections could be picked. Velocity gradients were used in almost all layers because almost all reflections terminated within buoy range. The velocity at the bottom of a layer was adjusted to best fit the pre- and post-cross over picks of the wide-angle reflection. The velocity at the top of the layer was choosen to produce a gradient which (with the thickness) would terminate the reflection at the range indicated by the picks and would also satisfy constraints on the position of the critical point.

Dipping layers

Two sonobuoys analyzed on the upper continental rise landward of the salt diapir province (6 and 7) were shot while traveling down bathymetric dip. The maximum seafloor dip (6) is about 2°; vertical incidence records indicate similar dips on near surface and deeply buried horizons. Such dips should give low apparent velocities (Ewing, 1963).

Two calculations were done in order to assess whether the effect of such dip was significant. First, travel times to a flat seafloor and a seafloor dipping 2° were calculated using single layer formulas in Dobrin (1976) (Fig. 2, Table 5). The greatest change in travel time differences occurred between 0 and 4 km. Beyond 4 km the differences were almost constant, averaging 67 ms. These travel times were fitted using the ray tracing program. The flat seafloor could be perfectly fit with the model parameters. The dipping seafloor travel times could be less well fit using parameters slightly different from the model: V=1.49 km/s and h=2.03 s. The differences of about 0.6% in velocity and 1.5% in thickness are insignificant. Since, however, modeling experience shows that minor errors, whether random errors or systematic errors due to violation of model assumptions, in fit to travel times shallow layers can cause major discrepencies in fitting deeper layers, an additional calculation was done.

A specific 3 layer model (Fig. 3) similar to the inferred sedimentary structure for buoy 6 was used. No attempt was made to test in a more general fashion the dip limits of the travel time modeling program. Travel times for reflections from the top of layer 4 for the flat bed model were calculated using equations in Slotnick (1959, p. 195; Table 6). Presuming that the shallow layers were known, ray tracing of layer 4 reflection times gave back the model parameters (Table 6). Travel times for pre-critical reflections from the dipping model were calculated using ray tracing equations in Sattlegger (1966; Table 6). Travel times were calculated only to a range of 3.7 km, since the modeling of seafloor travel times indicated that changes in error due to dip were insignificant at greater ranges. When the velocities of overlying layers are reduced by ~1% below model

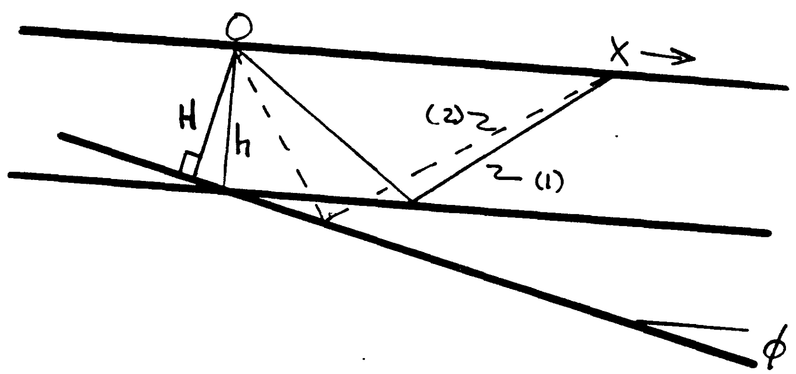


Figure 2 Ray geometry for travel time through a single bed. Calculations done using equations from Dobrin (1976): (1) flat bed, p. 203; (2) dipping bed, p. 207.

TABLE 5. Travel time (s) vs. range (km) for reflections off (1) flat and (2) dipping surfaces.

	RANGE (km)									
	0	1	2	4	6	8	10	12	14	18
(1) Flat	2.001	2.109	2.405	3.334	4.473	5.696	6.960	8.246	9.545	12.166
(2) Dipping 2^{o}	2.000	2.130	2.442	3.389	4.534	5.761	7.027	8.314	9.613	12.234
Difference (km)	0.001	0.021	0.037	0.055	0.061	0.065	0.067	0.068	0.068	0.068

Parameters

V = 1.5 km/s
H = 1.5 km
h = H/cos { = 1.501 km
{ = 2^{o}

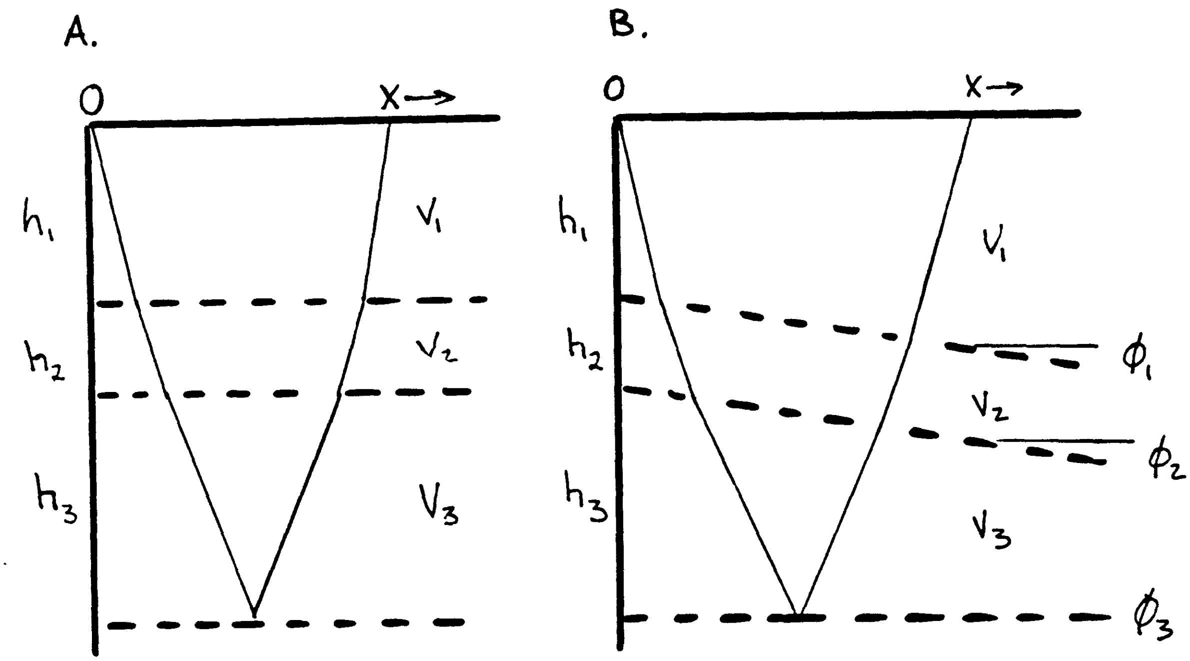


Figure 3 Ray geometry for travel time calculations through multiple beds. Calculations done using equations from: (a) flat beds, Slotnick (1959), p. 195; (b) dipping beds, Sattlegger (1966). Parameter values in Table 6.

TABLE 6. Travel times calculated for multiple bed model.

Departure angle (degrees)	Ray parameter, p	Distance, x (km)	Travel time, t (s)
Flat beds			
5°	0.0581	0.624	3.483
10°	0.116	1.266	3.538
20°	0.228	2.702	3.788
30°	0.333	4.639	4.341
40°	0.428	9.384	6.208
Dipping beds			
5°		0.707	3.594
8°		1.075	3.529
15°		2.041	3.668
20°		2.802	3.833
25°		3.687	4.076

Layer	Thickness, h (km)	Velocity, V (km/s)	Dip of base of layer, θ (degrees)
Flat beds			
1	1.5	1.5	0°
2	0.5	1.8	0°
3	1.0	2.2	0°
Dipping beds			
1	1.5	1.5	2°
2	0.5	1.8	2°
3	1.0	2.2	0°

velocities, ray tracing of layer 4 gives a velocity of 2.1 km/s which is 5% low (Table 6). The effect of dip can be expected to increase for deeper layers. Error of this magnitude does not significantly affect accuracy of the geological maps made with these velocities. Systematic errors for sonobuoys 7-16 in deeper water, where dip on the seafloor and subbottom layers is less than 1°, are probably much less than 5%.

Results

Velocity-depth results for the 7 sonobuoys analyzed (Table 7 and Fig. 4) are discussed in Chapter 4.

WELL LOG ANALYSIS

Analog copies of sonic logs for five wells (Table 2) were digitized and filed on computer as velocity-travel time and velocity-depth tables for convenient plotting. Sonic logs were made using long-spaced (2.4-3.0 m), borehole compensated tools operating at 25 kHz. Each sonic log includes a plot of inverse velocity (usually microseconds/ft) and tic marks indicating the integrated travel time of the sonic signal (milliseconds). Sonic velocity was calculated by dividing the depth differences between tics by the time between tics. Sonic velocity was interpolated across sections with anomalous velocity (eg. large hole diameters) and across breaks in the well log at casing terminations.

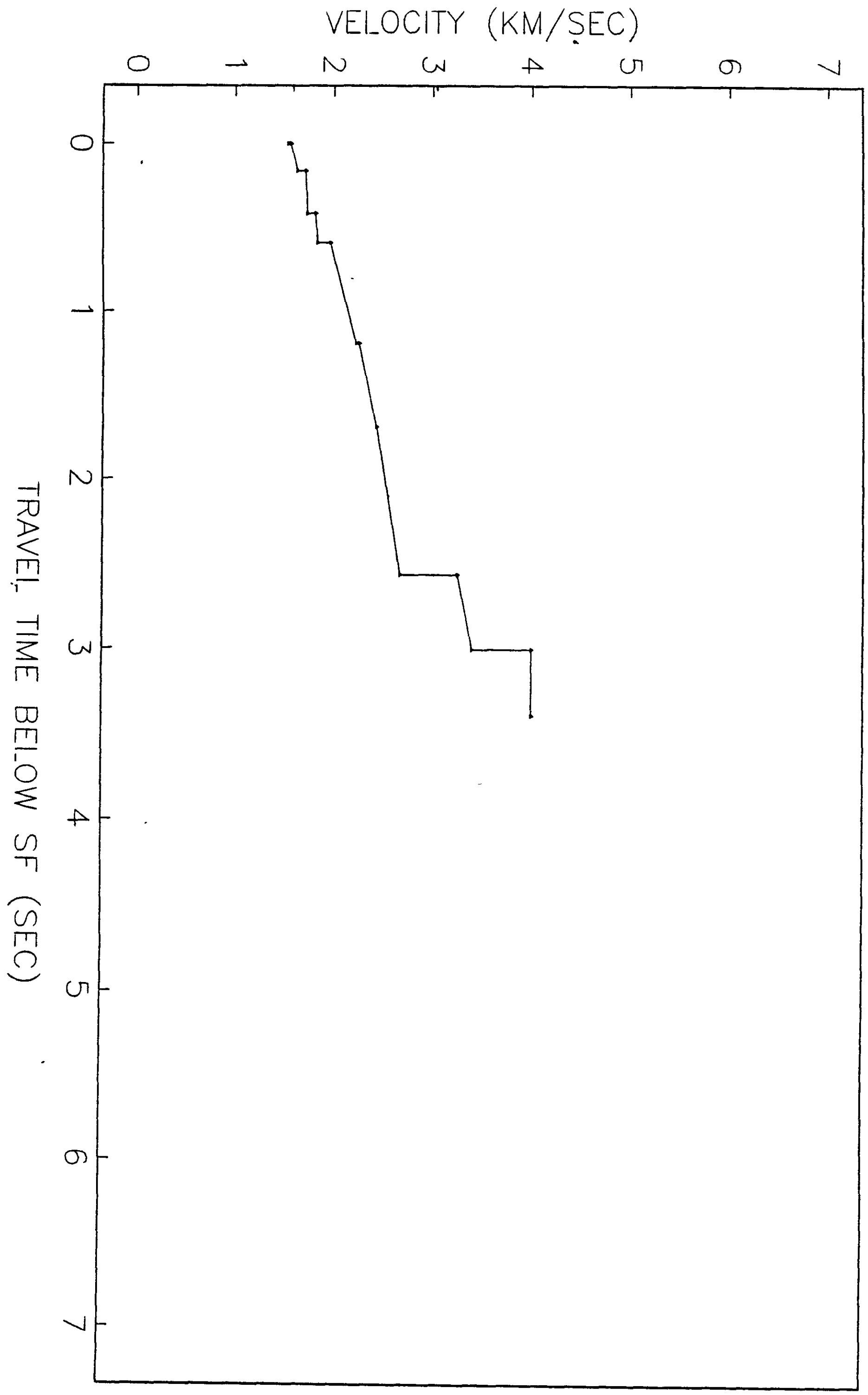
MULTICHANNEL VELOCITIES

Analog seismic profiles, shot and processed by SHELL in the early

TABLE 7. Travel time-depth functions and velocity gradients.

Source	Velocity vs. Travel Time Function	Depth vs. Travel Time Function	Gradient (s^1)	Depth Where Applicable: Travel time (s) below sealevel	Depth (km) below sealevel	Stratigraphic base level
Shelf						
Oneida O-25	$V=1.66+0.905T$	$D=0.83T+0.226T^2$	0.85	1.8	1.9	Base of Dawson Canyon Fm
Moheida P-15	$V=1.66+0.985T$	$D=0.83T+0.246T^2$	0.92	1.5	1.7	Base of Dawson Canyon Fm
Mohican I-100	$V=1.66+0.567T$	$D=0.83T+0.142T^2$	0.54	1.5	1.7	Base of Banquereau Fm
Acadia K-62	$V=1.67+1.151T$	$D=0.83T+0.288T^2$	0.95	1.9	2.6	Base of Banquereau Fm
Slope and upper rise						
Schubenacadie H-100	$V=1.66+0.293T$	$D=0.83T+0.0732T^2$	0.28	2.6		
D3 stacking velocities (ave)	$V=1.53+0.618T$	$D=0.77T+0.154T^2$	0.59	3.0		
RC2408 SB 6	$V=1.58+0.489T$	$D=0.79T+0.122T^2$	0.47	2.6		
RC2408 SB 7	$V=1.41+0.655T$	$D=0.71T+0.164T^2$	0.63	2.0		
Lower rise						
RC2408 SB 9	$V=1.55+0.468T$	$D=0.78T+0.117T^2$	0.47	1.75		A^U
RC2408 SB 11	$V=1.55+0.56T$	$D=0.78T+0.139T^2$	0.54	2.0		A^U
RC2408 SB 12	$V=1.55+0.687T$	$D=0.78T+0.172T^2$	0.58	1.5		A^U
RC2408 SB 15	$V=1.58+0.572T$	$D=0.79T+0.143T^2$	0.58	1.45		A^U
RC2408 SB 16	$V=1.65+0.431T$	$D=0.82T+0.108T^2$	0.46	1.37		A^U

V= velocity (km/s)
D= depth below seafloor (km)
T= two-way travel time below seafloor (s)



<u>Figure 4</u> Velocity vs. depth profiles determined on RC2408 sonobuoys. (a) sonobuoy 6, (b) sonobuoy 7, (c) sonobuoys 9, 11, 12, 15, 16.

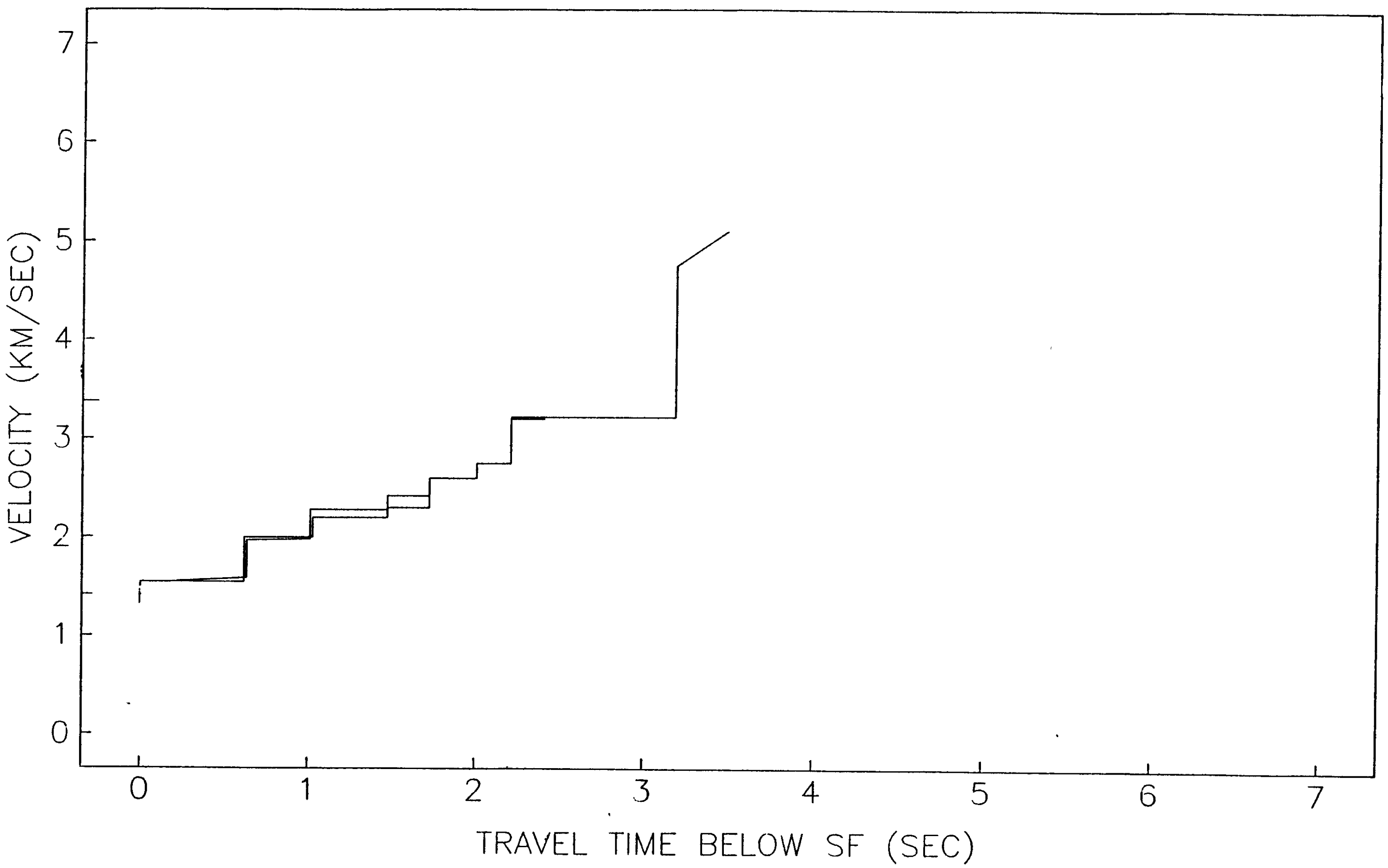


Figure 4b

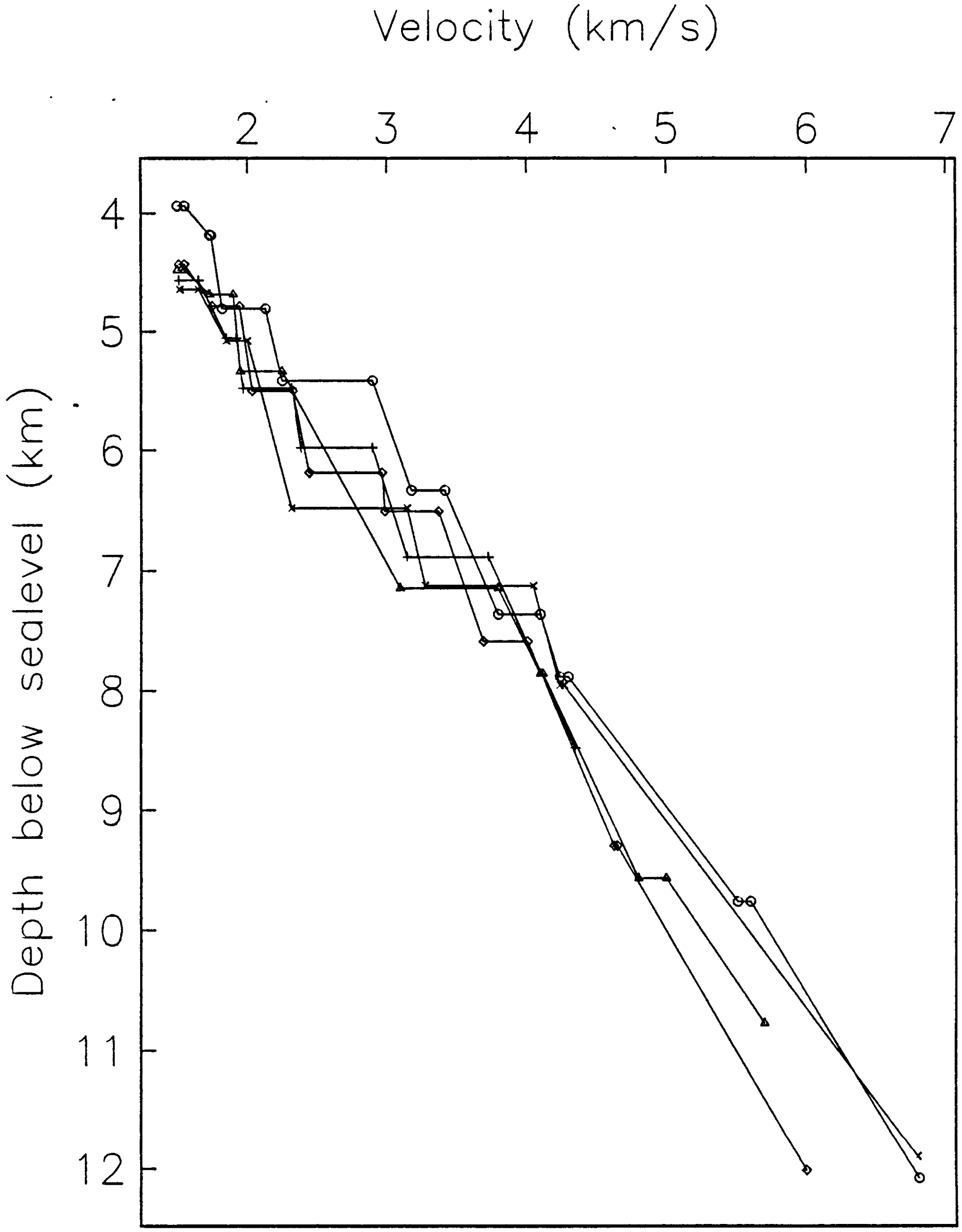
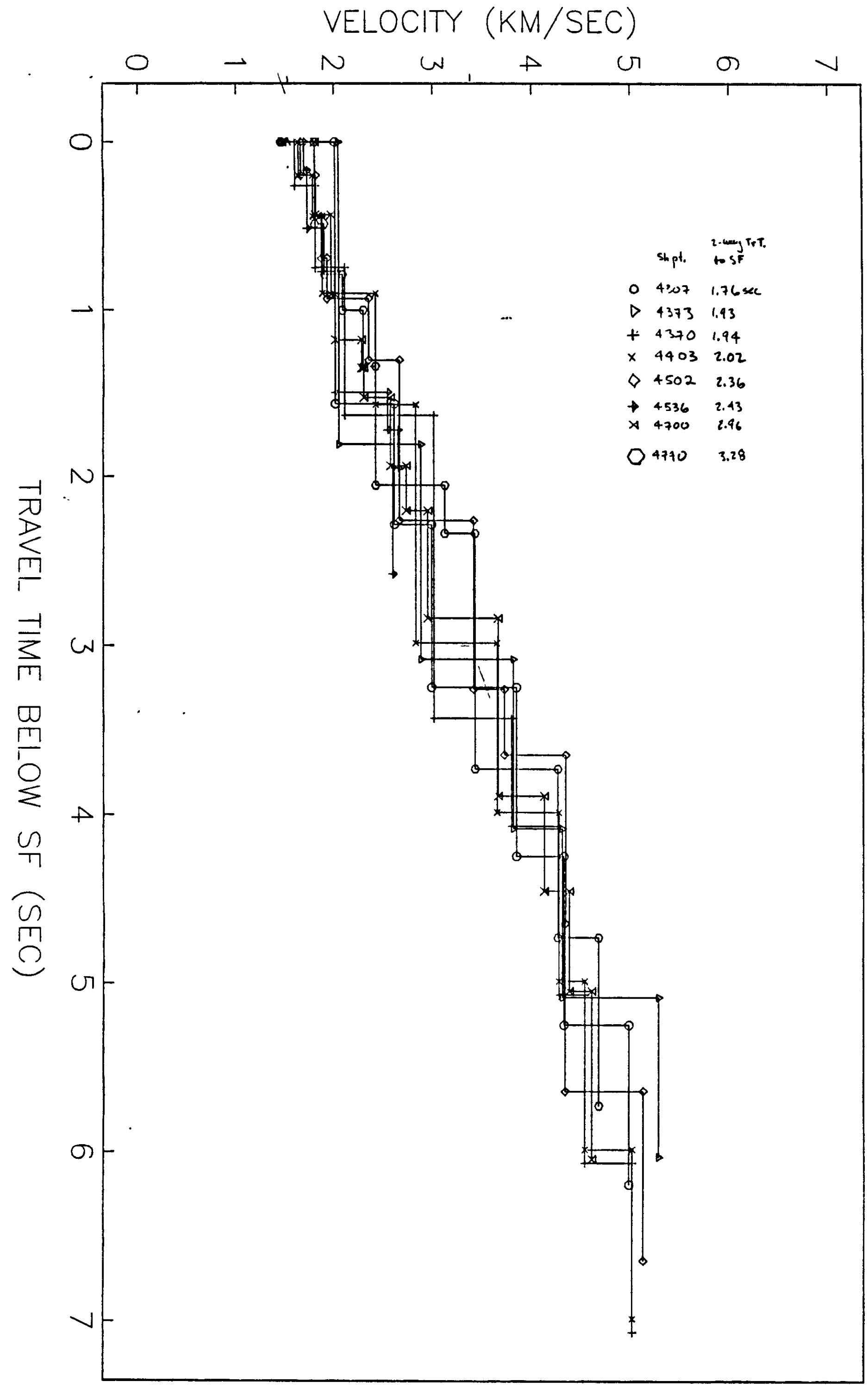


Figure 4c



<u>Figure 5</u> "Interval" velocity vs. depth profiles along line D3; location in Fig. 1. "Interval" velocities calculated using Dix's formula from stacking velocities.

1970's, were purchased from Canadian Oil and Gas Lands Administration (COGLA) for stratigraphic studies. Tables of stacking velocities at spaced shot points were printed on the leaders to some profiles.

Stacking velocities can be reasonably assumed to be root mean square velocities when the range over which the velocity scan was made (nominally the gun-streamer separation) is small relative to the depth of the common mid-point. Interval velocities can than be calculated using the Dix formula (Dobrin, 1976). In order to assess relationships between velocity-depth functions and stratigraphic horizons on the upper contintental rise, interval velocity calculations were made on a seismic dipline shot with a 180 m streamer-gun separation and a 2150 m streamer in water depths of 1.43 km to 2.42 km (Fig. 5).

TRAVEL TIME-DEPTH FUNCTIONS

One of the principal reasons for these velocity analyses is to provide a means for calculating depth to horizons and layer thicknesses within the Cenozoic section. In order to do such calculations, continuous linear functions of velocity vs. travel time below seafloor were derived and integrated to give an expression for depth as a function of travel time. A layer approach was not choosen because the seismic horizons that define such layers are not present everywhere and because the velocity analyses indicated that velocity on the slope and upper rise varied as a function of depth of burial rather than seismic stratigraphic layer. Linear functions were choosen because velocity vs. travel time plots over the depth of geological interest looked linear. Quadratic functions have been used by others (eg. Houtz et al., 1968; Hamilton et al., 1977; Bachman et al., 1983). However, plots of residuals from linear fits vs. travel time, in general, show random distributions about zero suggesting that a linear fit is a reasonable assumption. Use of linear functions introduces systematic errors in depth estimation near the seafloor and across velocity discontinuities. Although seafloor velocities estimated by the linear functions are probably too high by about 5-15%, velocity gradients near the seafloor are high (~1.3 s^{-1}, Hamilton et al., 1977) and the near surface layer, for which data is lacking, is thin (about 0.25 s), so depth estimates are low by an amount (< ~0.03 km) small compared to both error produced by variability in velocity between sonobuoy stations and to the interval used to contour the results (0.25-0.5 km on shelf to upper rise and 0.1 km on lower rise).

Methodology

From plots of velocity vs travel time, a linear function of the form:

$$V=a+bT, \quad \text{where } V(\text{km/s})= \text{"instantaneous" velocity at a depth } T$$
$$T(\text{s})= \text{two-way travel time below seafloor}$$

was estimated by eye. The thickness, h(km), of a thin layer <u>i</u> is given by:

$$h_i=v_i t_i, \quad \text{where } v_i= \text{"interval" velocity for layer } \underline{i}$$
$$t_i= \text{one-way travel time through layer } \underline{i}$$

The depth to a specific horizon n with two-way travel time(s) T_n is the sum of the layers from the surface to n:

$$D_n= \sum_{i=0}^{n} h_i = \sum_{i=0}^{n} v_i t_i, \quad \text{where } D_n= \text{depth(km) to horizon } \underline{n}.$$

To change to two-way travel time, we substitutue variables:

$t'_i = 2t_i$ where t'_i = two-way travel time(s) through layer i.

Thus, $$D_n = \sum_{i=0}^{n} (v_i t'_i)/2.$$

In the limit as $t'_i \rightarrow 0$,

$$D_n = \int_0^{T_n} V/2 \, dT.$$

Substituting, $$D_n = \int_0^{T_n} (a/2 + bT/2) \, dT$$

$$= (a/2)T_n + (b/4)T_n^2.$$

Note that both D_n and T_n are depths below seafloor. The functions derived are given in Table 7 and in Chapter 4.

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