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3. FRONTAL ACCRETION AND PIGGYBACK BASIN DEVELOPMENT AT THE SOUTHERN EDGE OF THE BARBADOS RIDGE ACCRETIONARY COMPLEX[1]

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ABSTRACT

About 4000 m of sediments are accreted at the front of the Barbados Ridge between 10° and 11°S. As a result, the structures are wide enough to be clearly depicted by seismic reflection and Seabeam data in contrast to the Leg 110 transect to the north, where stuctures are poorly resolved by geophysics. This is therefore a convenient area for studying the tectonic processes such as the thrusting and folding prevailing at the toe of an accretionary complex. Piggyback basins have developed as a result: (1) of a high rate of sedimentation from the Guyana margin through a network of submarine canyons, and (2) of the relatively high angle of thrust faulting at the deformation front as deduced from section balancing modeling. Most of the differences in structural style along the deformation front of the Barbados Ridge from north to south can be related to variations in the distribution, facies, and thickness of sediments on the Atlantic oceanic crust.

INTRODUCTION

South and east of Barbados Island, the Lesser Antilles active margin is characterized by the great development of the Barbados Accretionary Complex. The prism here is about 300 km wide, compared to only 150 km at the latitude of ODP leg 110 (Valery et al.,1985; Brown and Westbrook, 1987; Fig. 1). This is the result of large amounts of clastic sediment coming from the Orinoco river in the Neogene (Leonard, 1983) and maybe even earlier from the northern South America margins. The sedimentary cover of the Atlantic abyssal plain is also older to the south as the oceanic crust is assumed to be of late Jurassic or Early Cretaceous age in accordance with the pattern of magnetic anomalies (Speed, Westbrook et al., 1984; Westbrook et al., 1984) and the structure of the Guyana margin farther south (Sancho, 1985; Gouyet, 1988). Conversely, the oldest sediments on the oceanic crust to the north have been dated as Senonian age only at Site 593 of DSDP Leg 78A (Moore, Biju-Duval et al., 1982).

Because large volumes of sediments are involved in the frontal accretion of the Barbados Ridge, the resulting structures are broad enough to be clearly imaged by seismic profiling and Seabeam seafloor mapping (Biju-Duval et al., 1982; Mascle et al., 1986). It is thus possible to ascertain the geometry of fault planes better at the deformation front and to estimate the actual amount of displacement along each individual thrust. The presence of syntectonic sedimentation in some piggyback basins should, furthermore, give some insight on the actual rate and kinematics of thrusting (i.e., continuous vs. discontinuous). These, however, will be only rough estimations because the age of seismic sequences is poorly known as inferred from age correlations with the Trinidad and Venezuela offshore areas along regional seismic profiles (Leonard, 1983; Cedraro, 1987).

The present paper gives some results of a multichannel high-resolution seismic survey shot in 1982 by the Institut Français du Pétrole (IFP) and Société Nationale Elf Aquitaine over an area previously covered by a Seabeam survey (Valery et al, 1985). The tectonic and sedimentary processes that we will describe and discuss in the following chapters can be of use for understanding similar processes in areas where the geophysical data are more difficult to interpret, such as the Leg 110 area. We must keep in mind, however, that the particular setting of the southern Barbados Ridge, close to the southern Caribbean transform margin, may lead to the development of additional structures such as lateral transcurrent faults.

FRONTAL ACCRETION

Frontal thrusts and folds are spectacularly depicted on both Seabeam and seismic profiles. To the north of the area studied (Figs. 2 and 3), the frontal thrust produces a scarp rising about 100 m up from the seafloor. The fault plane deepens to the west to an inferred décollement (not seen on the profiles) with a dip of about 20° (Fig. 4). This simple arrangement contrasts with the more complex structures seen on the second thrust, where the shear zone is not restricted to a single fault plane but involves a broad zone about 2 km wide at the surface where the structures seem to be an order of magnitude smaller than the major ones (0.5 km vs. 5 km). A similar superficial diverticulation of thrust planes has already been noted farther north (Biju-Duval et al., 1982) at the latitude of Barbados Island and has been proposed to explain some of the thrust faults of the Leg 110 transect (Moore, Mascle, et al., in press). These structures can be interpreted in two different ways. They could be minor thrusts having limited lateral extent and branching off the major thrust. This could explain the rapid disappearance of their surface expression south of line CRV 110 (Fig. 3). Alternatively, the fault pattern observed on the section could represent the southward coalescence of several thrust faults above a transverse ramp. This hypothesis would explain the rapid increase in the wavelength of the folds (which reflect a deepening of the décollement) from the north to the south of the canyon intersecting the deformation front at 10° 30′ N. Farther south, another canyon crosses the ridge at the exact place where the deformation front is offset by nearly 2.5 km, thus suggesting that it is superimposed on a dextral transverse fault.

The cumulative displacement on the two frontal thrusts of line CRV 110 is about 2 km, i.e., 10% of the total length of the two accreted packets. This line, however, is not perpendicular to the trend of structures, and the displacement along the fault plane is probably overestimated. Taking into account this obliquity, the total displacement would be only about 1.5 km, i.e., roughly similar to the one deduced from balanced sections on the two major frontal thrusts of the Leg 110 transect (Brown, Mascle, and Berhmann, this volume). The vertical elevation on

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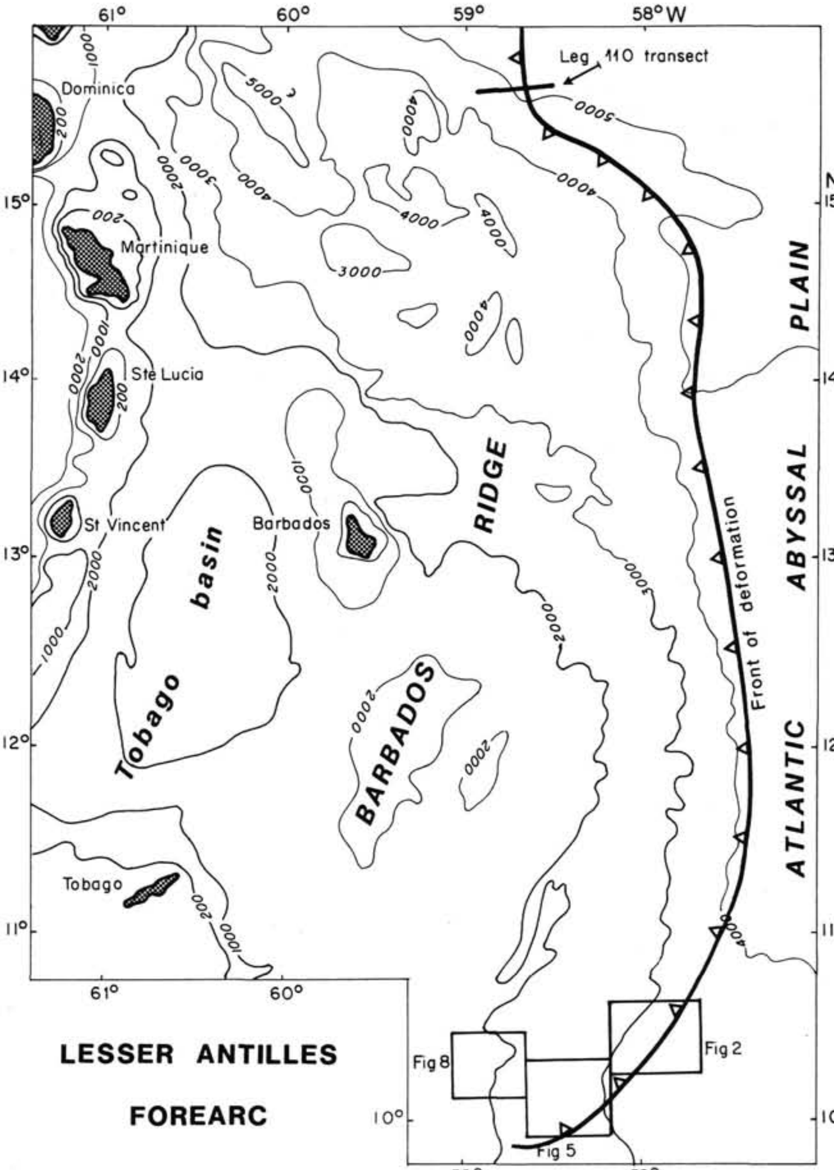


Figure 1. Map of the eastern Caribbean showing the southern edge of the Barbados Ridge and the Leg 110 transect. Contour interval, 1000 m.

top of the second ramp anticline is approximately 700 m with respect to the nearby abyssal plain. This is much more than the elevation of about 100 m we observed on top of the second anticline along the Leg 110 transect. This difference can be explained as resulting predominantly from the greater dip of thrusts to the south than to the north (20°W vs. 10°W).

The southern part of the area under study (Figs. 5 and 6) is located very close to the southern end of the Barbados Ridge. The frontal thrust is at the origin of a prominent scarp rising about 200 m from the seafloor. West of the ramp anticline and as depicted by the diverging reflectors above horizon A (Fig. 6), syntectonic clastic sedimentation has almost completely infilled the synclinal trough. Part of this infilling probably resulted from redeposition of eroded material from the summit of the next anticline to the west. A maximum thickness of 650–700 m is encountered at the foot of the second thrust. We do not know with precision the age of these sediments but they very probably are only of Pleistocene age. They represent the time needed by the anticline to grow progressively or, in other words, the duration of slip along the frontal thrust for a total displacement of about 400 m. Assuming an age of 0.5 or 1 m.y. for the piggyback basin, the rate of slip would be 0.08 or 0.04 cm/yr, respectively, on this single fault plane.

The horizontal shortening over the first 55 km from the deformation front is about 6 km, i.e., 10% of the initial length of the accreted packets. This is much smaller than the 25–90% shortening deduced from section balancing along the Leg 110 transect for the last four accreted packets (Brown, Mascle, and Behrmann, this volume). If we assume that subduction rates are similar north and south of the Lesser Antilles Margin, this would imply that, at the southern extremity of the Barbados Ridge, part of the relative motion between the two converging Atlantic and Caribbean plates is already taking place along transform faults as already suggested by local offsets observed in the deformation front. Additionally, homogeneous shortening related to sediment compaction and water expulsion could

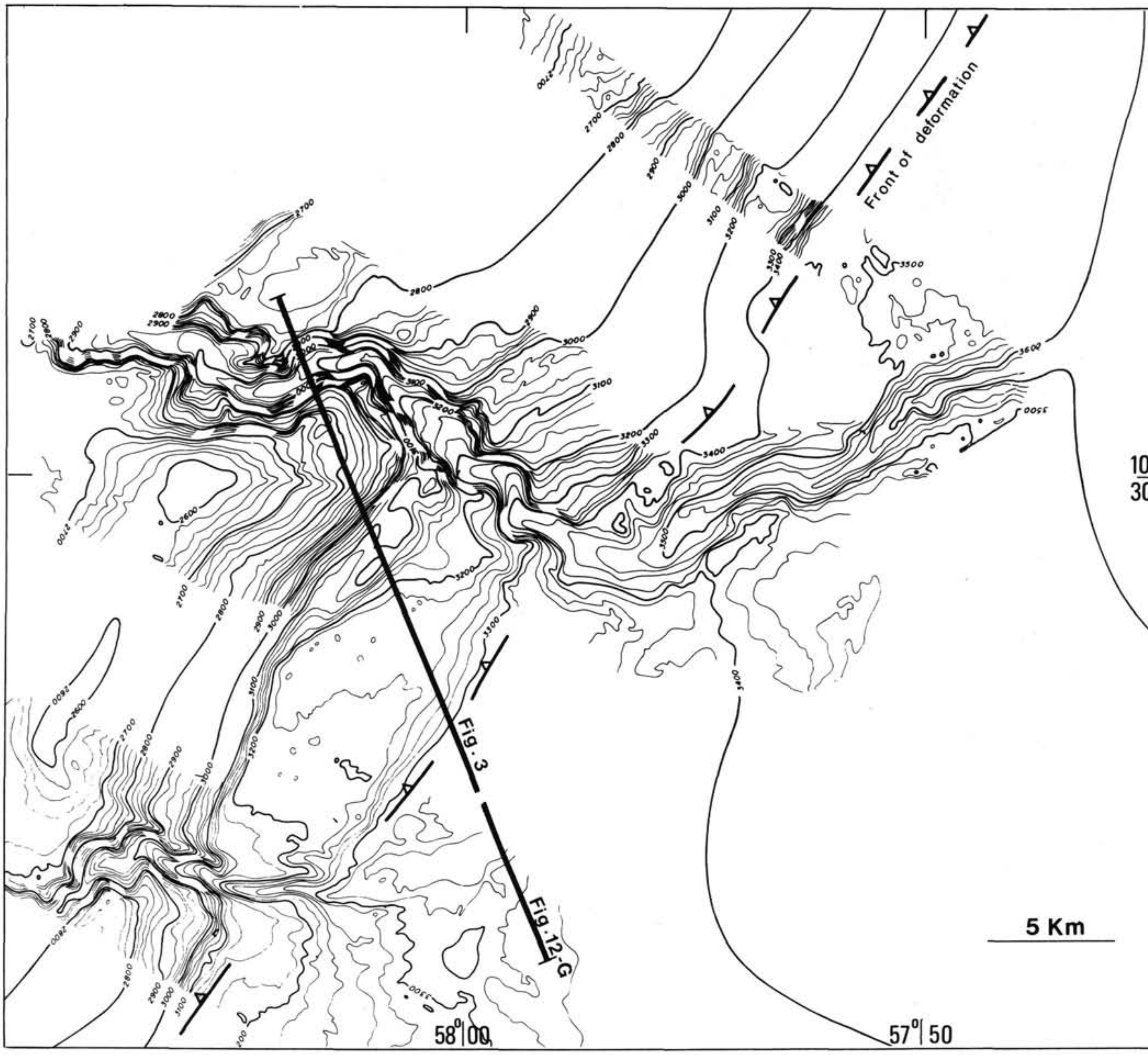


Figure 2. Seabeam bathymetric map. Depths are in meters. Location shown in Figure 1.

be more developed there than on the northern transect due to the presence of permeable clastic sequences in the Pliocene and Pleistocene section (distal edge of the Orinico deepsea fan), which are believed to make up most (if not all) of the accretionary complex.

The deformation front, as has been defined above, is the first continuous tectonic feature seen on the Seabeam map and on the seismic profiles. A new thrust is developing, however, 8 km to the east (seaward) of this deformation front with only limited expression on the seafloor morphology (Figs. 4 and 7). This can be compared on the Leg 110 transect with a similar thrust emerging on the seafloor 1.5 km east of the deformation front and branching off from the main décollement just west of Site 676. These two examples provide two possible explanations for the mechanisms of seaward propagation of the deformation front. The frontal thrusts could appear either as several discrete faults having limited lateral extent that will later join together to make a single front; or, alternatively, one single thrust will locally develop and will progressively extend laterally as the horizontal displacement along the initial fault increases. In both cases the thrust faults will propagate seaward and upward from the stratigraphic layer corresponding to the décollement where evidence of horizontal shearing has been found at Sites 672 and 676.

STRUCTURES AT THE BACK OF THE DEFORMATION FRONT

Piggyback basins are also present a few tens of kilometers west of the deformation front where active thrusting is still occurring. In such an area (Figs. 8 and 9B), the syntectonic nature of sedimentation can be deduced from both the external and internal configuration of seismic sequences (Chennouf, 1987). Continuous uplifting of the anticline to the east has induced the progressive tilting of strata which are characterized by westward diverging and eastward onlapping reflectors. The chaotic configuration at the eastern edge of some sequences is interpreted

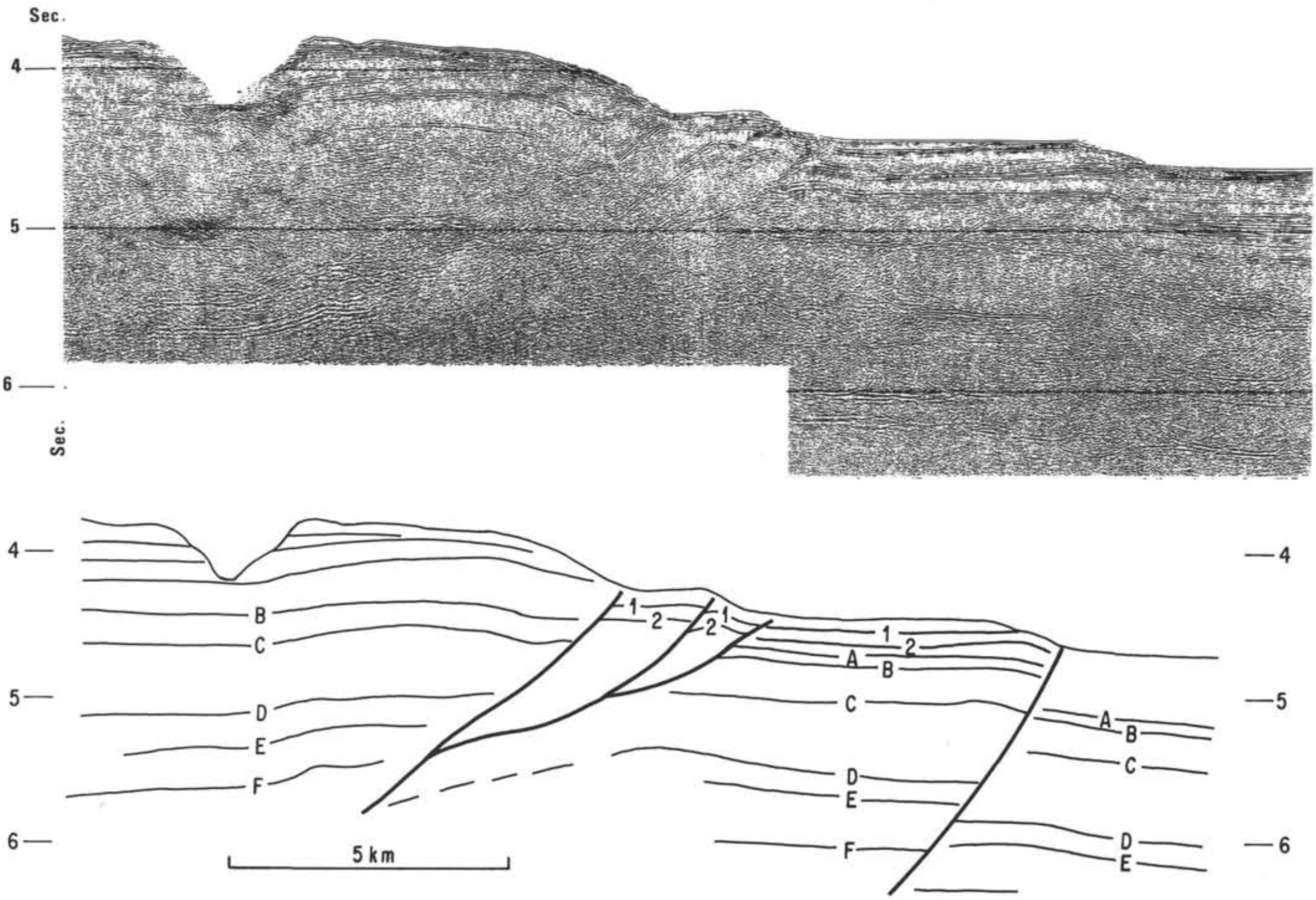


Figure 3. High-resolution multichannel seismic line CRV 110 and corresponding line drawing. Location shown in Figures 2 and 4. Seismic reflectors that have tentatively been correlated on each side of faults are marked 1–2 and A–F.

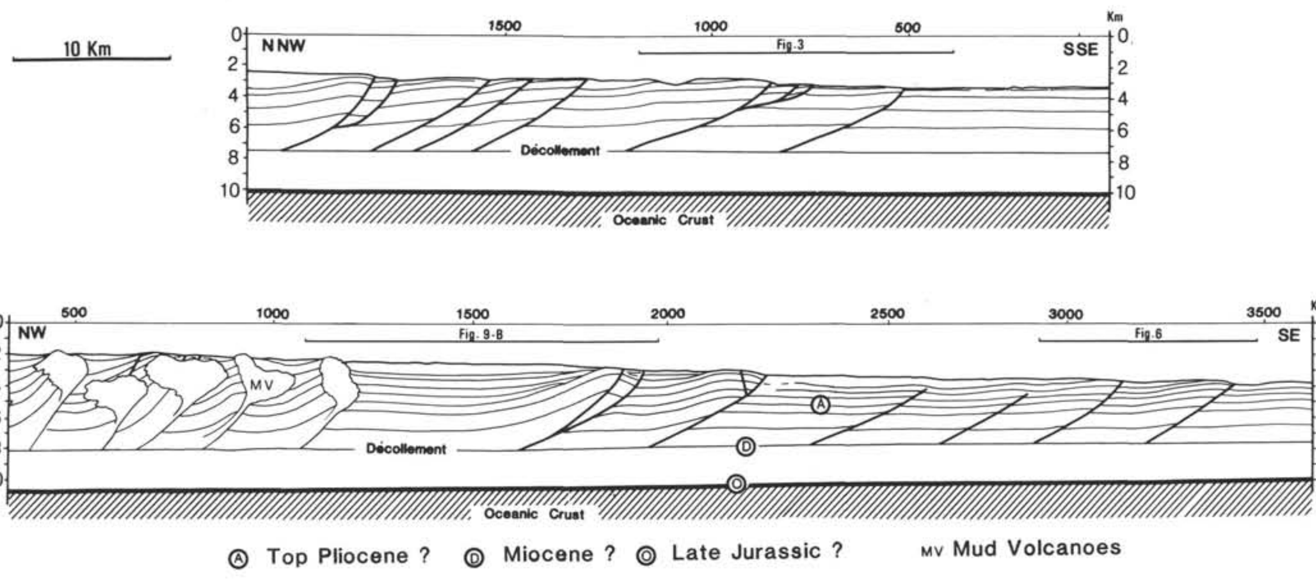


Figure 4. Depth sections of lines CRV 110 and CRV 105. Location shown in Figure 13.

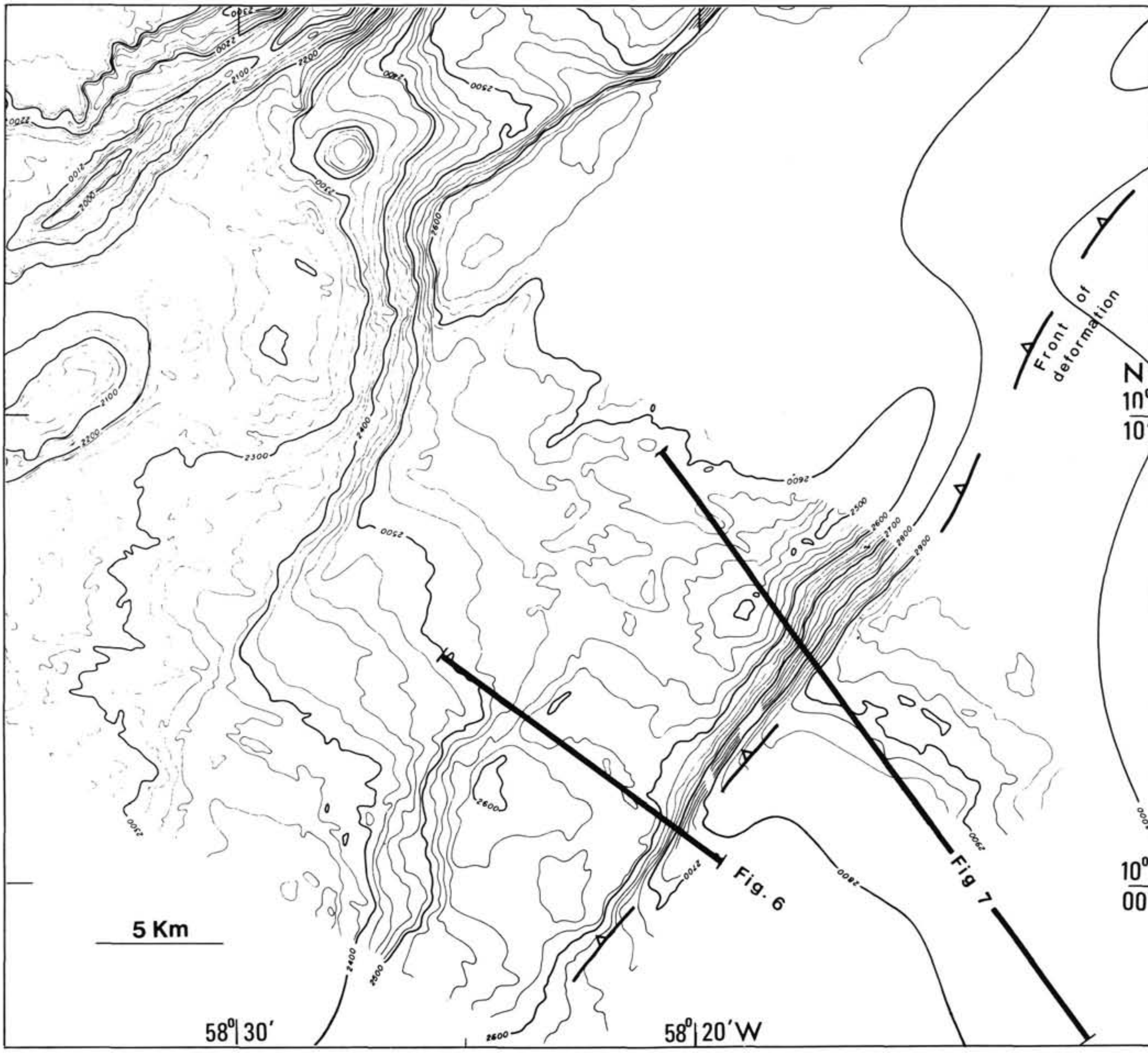


Figure 5. Seabeam bathymetric map. Depths are in meters. Location shown in Figure 1.

as being related to slumping of unconsolidated material on the western flank of the rising anticline. This 2.5- to 3-km thick sedimentary infilling was thus deposited during the lapse of time needed for the anticline to grow above a thrust fault where the total displacement can be estimated to be about 2–2.5 km. This is much more than the displacement of 400 m observed on the frontal thrust and confirms that present-day horizontal shortening is not restricted to the deformation front but extends far to the back (as predicted by numerical modeling; Ngokwey, 1984). In the present case, and if the syntectonic infilling of the basin is only of Pleistocene age, i.e., 2 m.y. old, the rate of displacement along the thrust would be about 0.1 cm/yr.

To the west (Figs. 8 and 9B), the culmination of a mud volcano is found 200 m above the basin. The morphology of the seafloor suggests that this volcano is located on the top of the anticline on the western boundary of the basin. These mud volcanoes become more numerous to the west, and many of them are associated with anticlinal structures (Figs. 4 and 8). We interpret them as originating from the accreted sequences (probably from Miocene deep marine clays if we refer to mud volcanoes known in Trinidad) in response to the continuous loading by the piggyback basins. Whereas this loading of Miocene clays is a necessary condition for the formation of an overpressured zone and mud volcanoes, this is probably not a sufficent one as the same clays have been buried to a similar (or even greater) depth below the Orinoco delta on the Guyana passive margin, where no mud volcanoes have so far been encountered (Leonard, 1983). Such mud volcanoes reappear onshore farther north in the southern Trinidad and eastern Venezuela thrust belt (Higgins and Saunders, 1979; Rossi et al., 1987). They thus appear to be restricted to areas under compressional stresses. It is proposed that this tectonic regime radically interrupts the hydraulic continuity in the Miocene clays, thus allowing high pressures to develop until fractures or faults may allow water and mud to escape to the surface. The thrust planes could be preferential pathways for such migration of water and mud. As a matter of

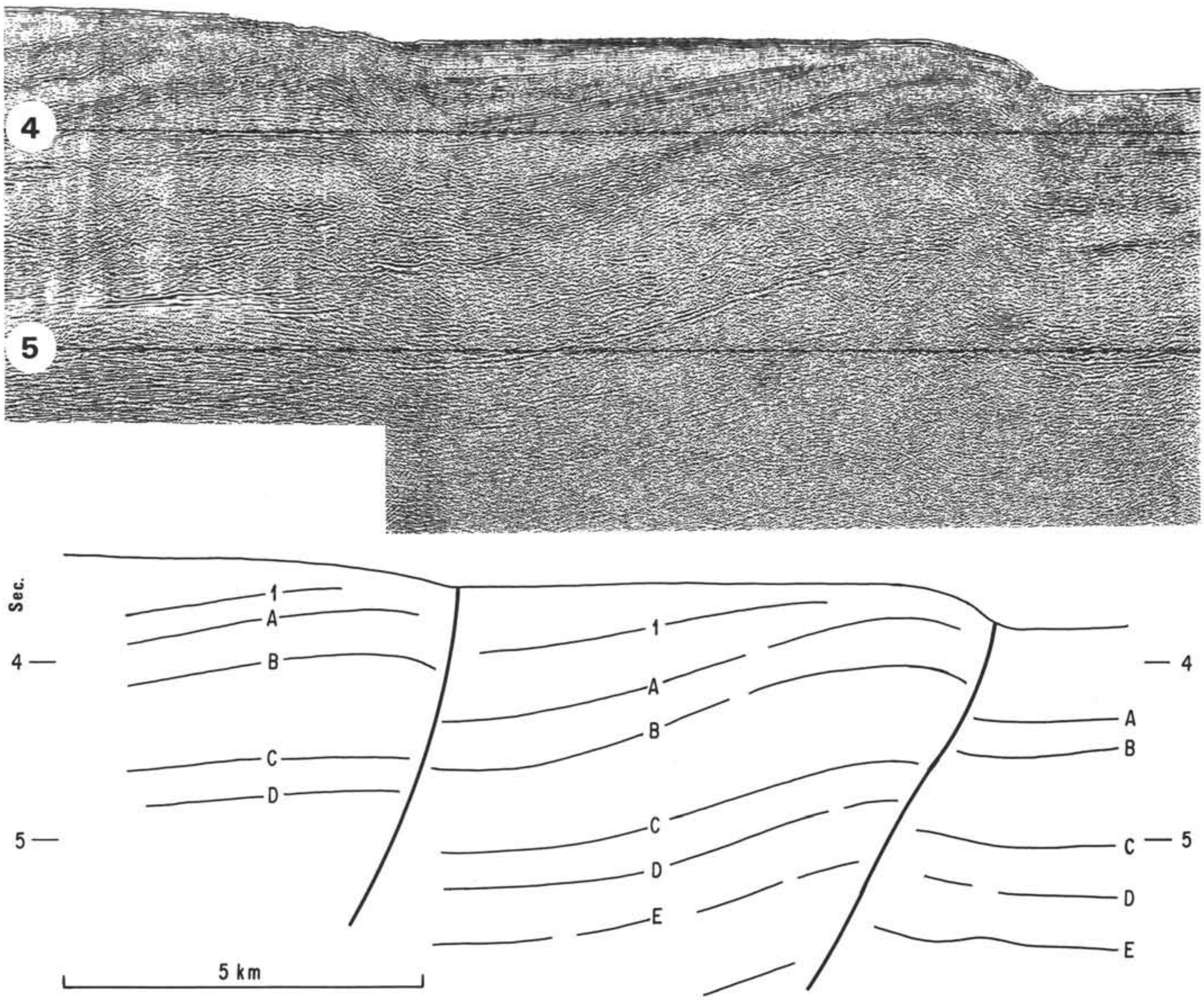


Figure 6. High-resolution multichannel seismic line CRV 105 and corresponding line drawing. Location shown in Figure 4. Seismic reflectors which have been tentatively correlated on each side of faults are marked 1 and A–E.

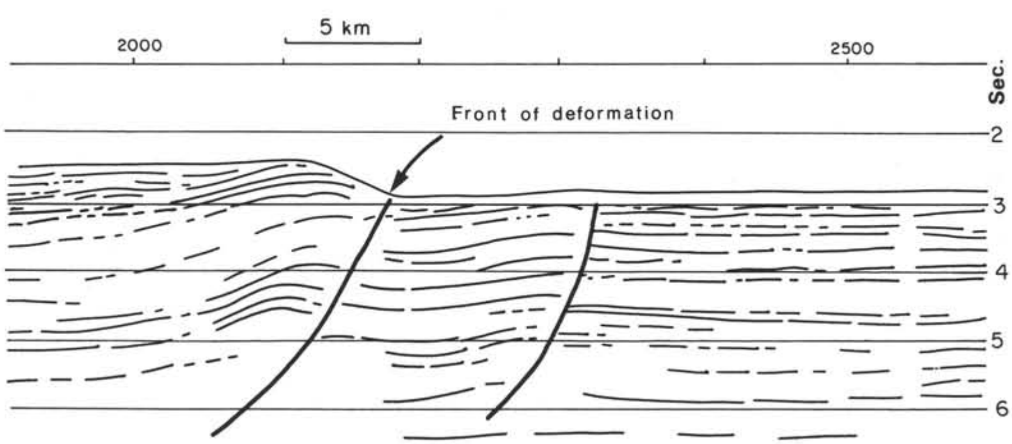


Figure 7. Line drawing of seismic profile showing the initiation of a thrust fault seaward of the deformation front. Location shown in Figure 5.

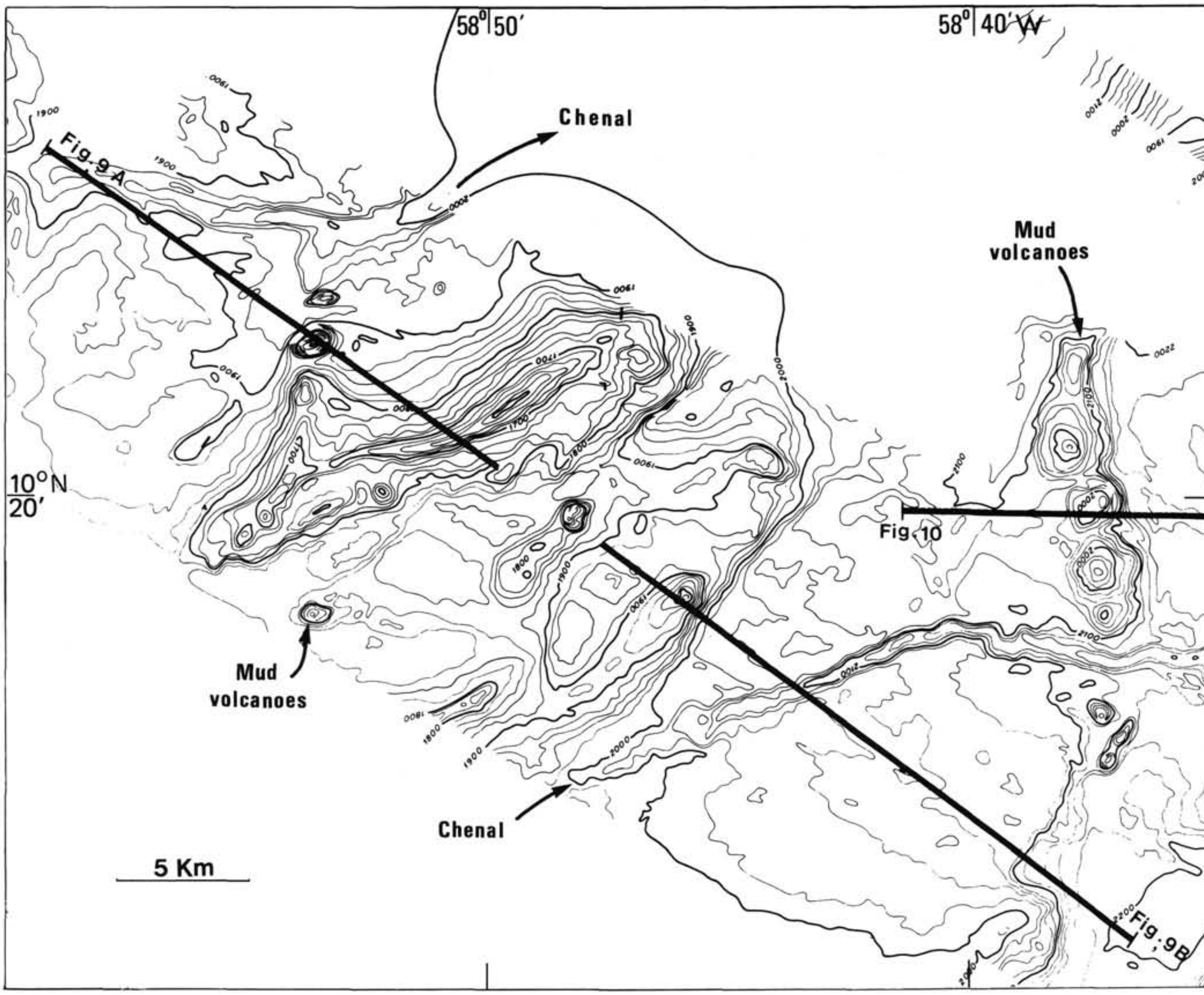


Figure 8. Seabeam bathymetric map. Depths are in meters. Location shown in Figure 1.

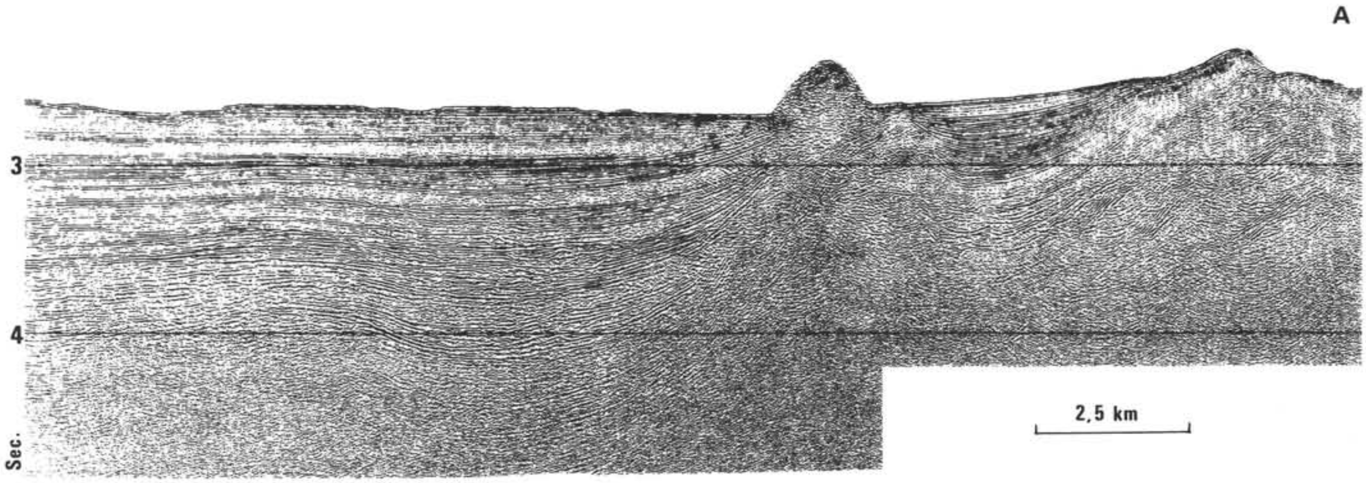
fact, the present activity of these faults as fluid conduits is confirmed by the presence of living benthic communities observed 5 km west of the seismic CRV 105 (Faugères et al., 1987; Faugères et al., this volume).

Some of the mud volcanoes also appear as isolated seamounts in sedimentary basins (Figs. 9A and 10) with local alignment as depicted with a north–south trend on the eastern and western sides of Figure 8. Such alignments strongly suggest fault control even if such faults have not been recognized on the seismic profile (because of the lack of resolution below the mud volcano). The difference in water depth on each side of the eastern alignement (Fig. 10) is not related to any vertical displacement along the inferred fault, as corresponding deep reflectors are at similar depths. The reason is rather the difference in thickness of the transparent seismic sequence above, which will be related to: (1) a partial dam effect from the seamounts alignment because clastic inputs come mainly from west to east (see next section), and/or (2) larger mud flows to the west from the mud volcanoes as suggested by a few westward-dipping reflectors.

SEDIMENTARY PROCESSES

The most prominent sedimentary process now occurring is the transit of clastic materials from the Guyana Margin to the Atlantic Abyssal Plain, mainly through three canyons intersecting the structures of the southern Barbados Ridge. The course of these canyons was followed from the shelf break down to the abyssal plain using seismic data, Seabeam, and Gloria seafloor mapping (Fig. 11). Most of the time they are parallel to the direction of the average maximum slope, even when they abruptly change their direction from WSW-ENE to WNW-ESE. At about 10° 30′, this area could represent the place where dextral transcurrent faulting is taking place along E-W trending faults. We have already noted that it is an area where frontal structures are becoming wider and less continuous to the south. This also follows the trend of the El Pilar fault system onshore, although we do not believe that this continental feature extends as such through the accretionary wedge.

Locally the course of the canyons is strongly influenced by structures having prominent relief on the seafloor, such as mud diapirs (Fig. 8) and folds (Fig. 2). The canyon segments intersecting the anticlines have the highest slope on their flanks (depth up to 400 m with slopes of about 10° locally). Such erosion was probably made possible because of the relatively slow and progessive uplift of structures. The two southern canyons merge in the abyssal plain where they cross the deformation front. The resulting channel then runs northward in front of the



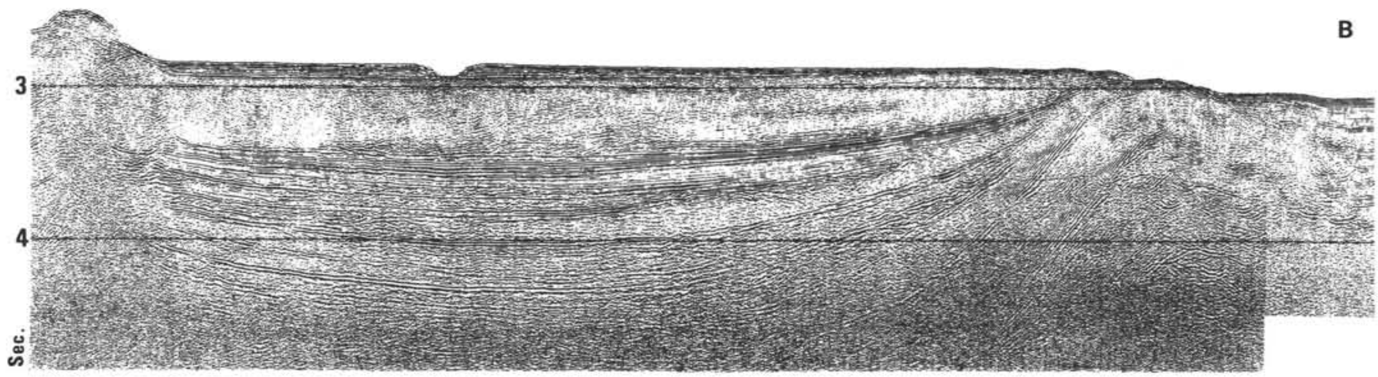


Figure 9. High-resolution and multichannel line CRV 10. Location shown in Figure 8. These two sections show two large piggyback basins developing to the west of growing anticlines. The mud volcano to the west of the section B is located on top of the next anticlinal structure when the volcano on section A appears as an isolated feature isolated in the basin (see Fig. 8). Present-day channels can be either erosive and widespread over a large area as on section A, or restricted to a single distributory canyon with lateral levees as on section B.

accretionary complex up to at least 13°N (Brown and Westbrook, 1987). There are currently no data that are detailed enough to estimate how much clastic material is being carried through this drainage system from the shelf break to the abyssal plain. The morphology of the channels with lateral levees and related lenticular seismic sequences suggests that some of the material is deposited here and there on the continental slope and in the piggyback basins (Fig. 12). Conversely, erosion in the V-shaped segments crossing the anticlines should result in new material being transported and distributed in the abyssal plain.

CONCLUSIONS AND DISCUSSION

The overall structure of the southern edge of the Barbados Ridge is shown in Figure 13. A major difference with the area located farther north (Biju-Duval et al., 1982, Brown and Westbrook, 1987) is the small longitudinal extension of anticlinal structures and the curvature of the fold axes. Typically, the anticlinal axis at the deformation front can be followed in a straight direction over several dozen km at the latitude of Barbados Island. Conversely, to the south most of these fold axes are between 10 and 20 km long. This structural pattern is interpreted as resulting from the presence of E–W dextral transcurrent faulting, which develops as the southern Caribbean Plate boundary is approached. As already proposed (Mascle et al., 1979), the structural arrangement observed is compatible with a local shortening direction of N 040°W, in agreement with the vector of plate convergence proposed by Perez and Aggarval (1981), and Aggarval (1983) in this area.

In spite of this local geological setting, it is of some interest to compare the frontal structures of the Barbados Ridge from south to north, to try to explain the variability in the tectonic style of initial accretion. Figure 14 is an attempt to make such a comparison using automatically balanced sections. Details of the processing can be found in Endignoux and Mugnier (in press). Starting from the undeformed and isopach strata, giving the location and dip of thrust faults and the amount of displacement along each of them as deduced from seismic data, the program successively reconstructs thrusting and folding from the inner zone to the outer zone (Fig. 15). Such a process will depict the first-order structures alone, as observed for the CRV 128 section (Leg 110 transect), but this overall structure fits reasonably well with the hand-drawn section where secondary structures have also been taken into account (Moore, Mascle, Taylor et al., 1988), and a rough estimate of shortening and deepening of the décollement is immediately given. It is clear, however, that the kink method used for section balancing is not relevant to deformation of relatively uncompacted sediments. As a consequence the geometry of ramp anticlines with angular points is incorrect and has to be smoothed out.

This technique has been applied for lines CRV 105 and 108A. The thicknesses of accreted sequences are similar and much thicker than to the north, but only on line CRV 105 have piggy-

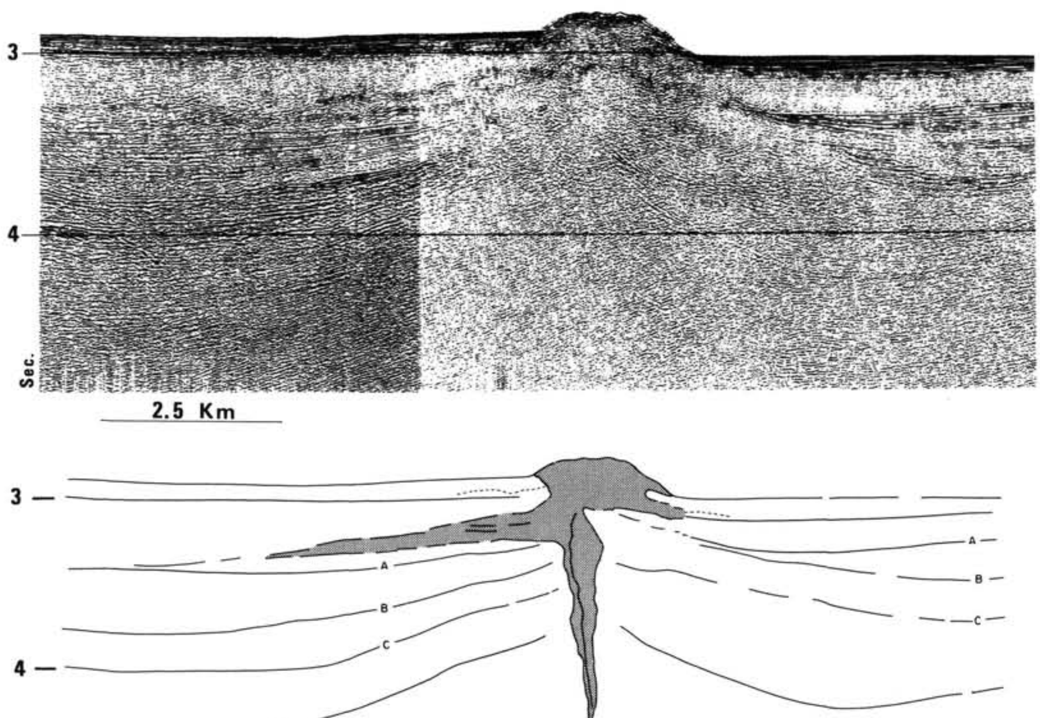


Figure 10. High-resolution and multichannel seismic line CRV 107 and corresponding line drawing. A, B, and C are seismic reflectors that have been tentatively correlated on each side of the mud volcano. Location shown in Figure 8.

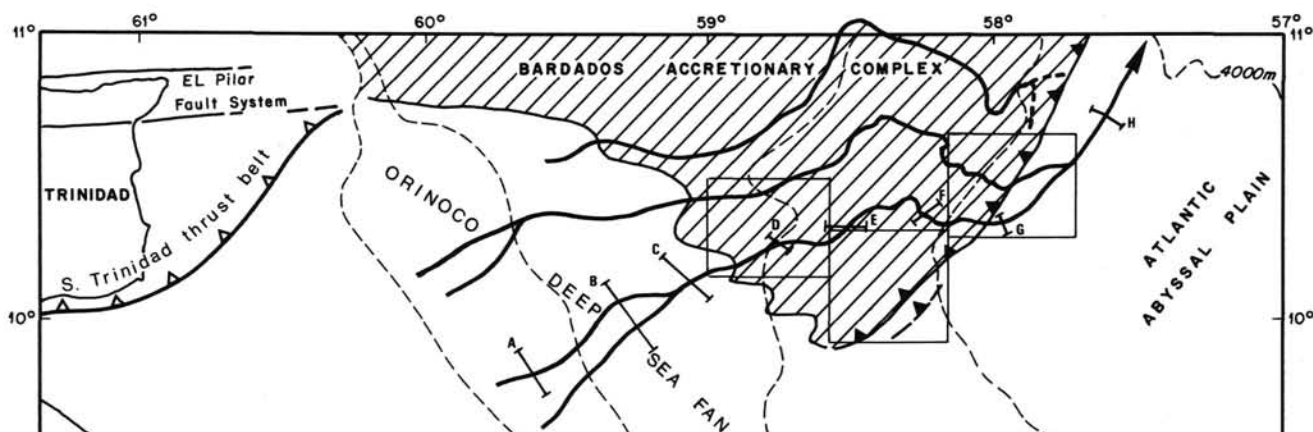


Figure 11. Map showing the course of the main distributory channels and canyons of the Orinoco deep-sea fan from the shelf break to the Atlantic Abyssal Plain.

back basins developed. This could be due in part to the close vicinity of clastic influxes from the Guyana margin as described previously, but there is obviously no place for such a basin to develop along line 108 A. Conversely, troughs a few hundred meters deep appear on line CRV 105 at the back of anticlines where piggyback basins are actually found. Such a result has been obtained using steep dips (45°) for the thrust faults. This allows a sufficient distance to be maintained between successive ramp anticlines, and consequently the synclinal trough can be preserved. Conversely, the smaller dip of frontal thrusts (20°) on line 108A leads to the close juxtaposition of ramp anticlines and prevents the development of any significant basin (Fig. 15). The low angle of thrusting along the Leg 110 transect (10°–20°) leads to still more stacking of the accreted packets.

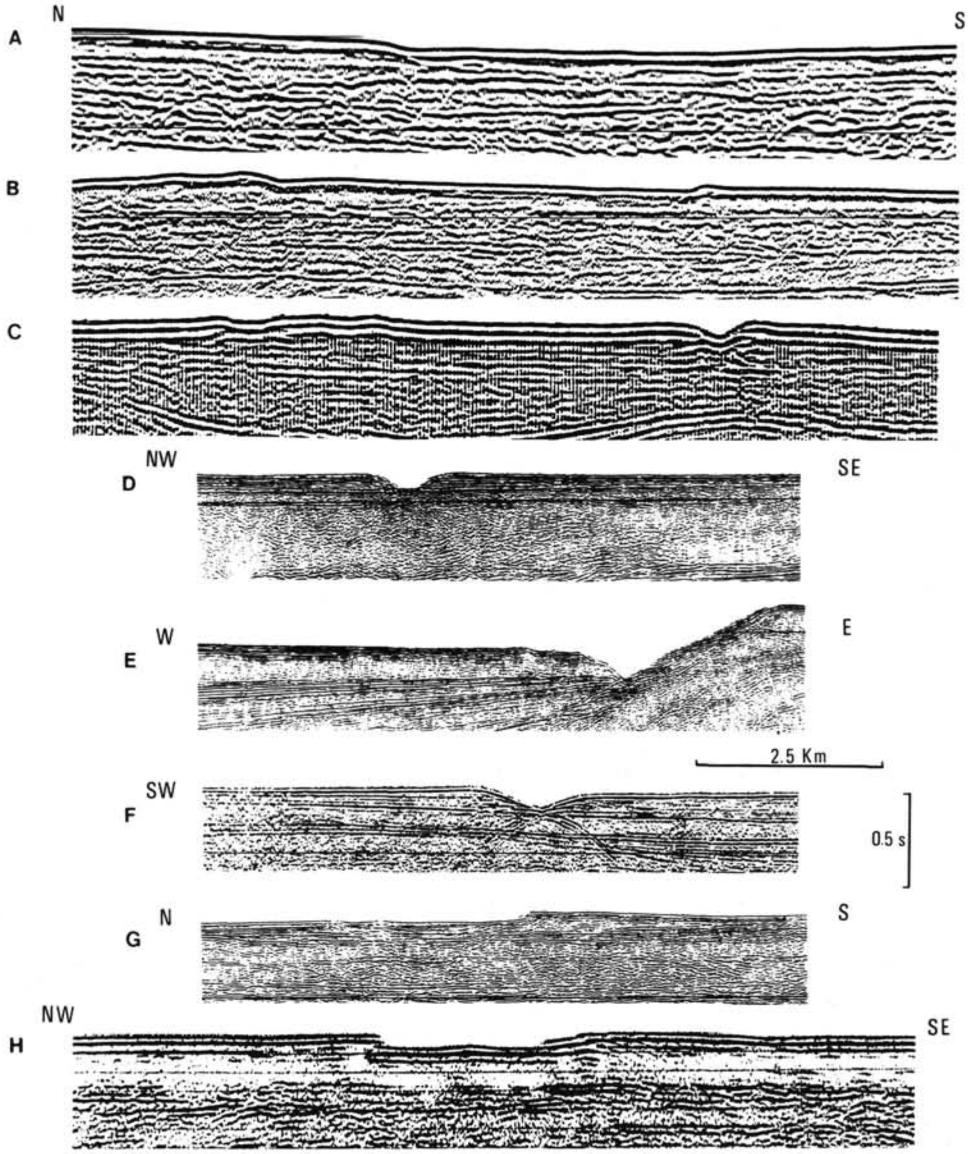


Figure 12. Seismic sections showing the seafloor expression of one of the distributory channels and some of the seismic sequences related to recent erosional and depositional processes. From section A to section H the average water depth is: 600 m, 1300 m, 1600 m, 2000 m, 2900 m, 2800 m, 3150 m, and 3200 m. Sections D to G are high-resolution seismic profiles, section A, B, C, and H are conventional lines. Location shown in Figure 11.

These three examples emphasize the variety of structures that have developed in front of the accretionary complex. As depicted in Figure 16, the length of the initially accreted packets is dependent on the depth of the décollement, even if, for the moment, data are too scarce to propose a definitive correlation law. We can also assume a similar behavior for the dips of frontal thrusts which get steeper with increasing sediment thicknesses. It has already been noted that the present location of the décollement on the Leg 110 transect is due to the presence in the early Miocene interval of radiolarian-rich mudstones with relatively high porosity compared to overlying and underlying mudstones. Therefore the tectonic style of frontal thrusts and folds appear to be strongly dependent on the distribution, facies, and thickness of sediments on the oceanic crust.

The northern and southern sections actually represent two end-members of such a "foreland" environment. To the north, a "thin skin" accretion has developed at the expense of thin and predominantly pelagic series with little permeability. Conversely, to the south the sedimentary cover of the oceanic crust is at least 6000 m thick and includes, in its Pliocene and Quaternary section, the eastern edge of the Orinoco deep-sea fan where clastic facies can be expected. The deformation front is formed of 7-km-wide ramp anticlines with inserted piggyback basins. Widespread mud volcanoes become prominent features a few dozen km to the west as the result of the progressive loading of the Miocene sequences. These series are probably of deep marine and pelagic origin with low permeability if we refer to rocks of similar age outcropping or having been drilled on Trinidad.

Legs 78A and 110 provided much data about tectonic and hydrogeological processes occurring along the northern transect. More data, including one or several deep holes and *in-situ* measurements of physical properties and temperatures, will be necessary for a better understanding and quantification of all of these processes. We also suggest that a few holes be drilled to the south to document the kinematics of thrusting and the syntectonic infilling of piggyback basins. A second major objective would be the compaction and water migration processes in an environment dominated by horizontal stresses without any intervening large vertical stacking, as is present along the northern transect.

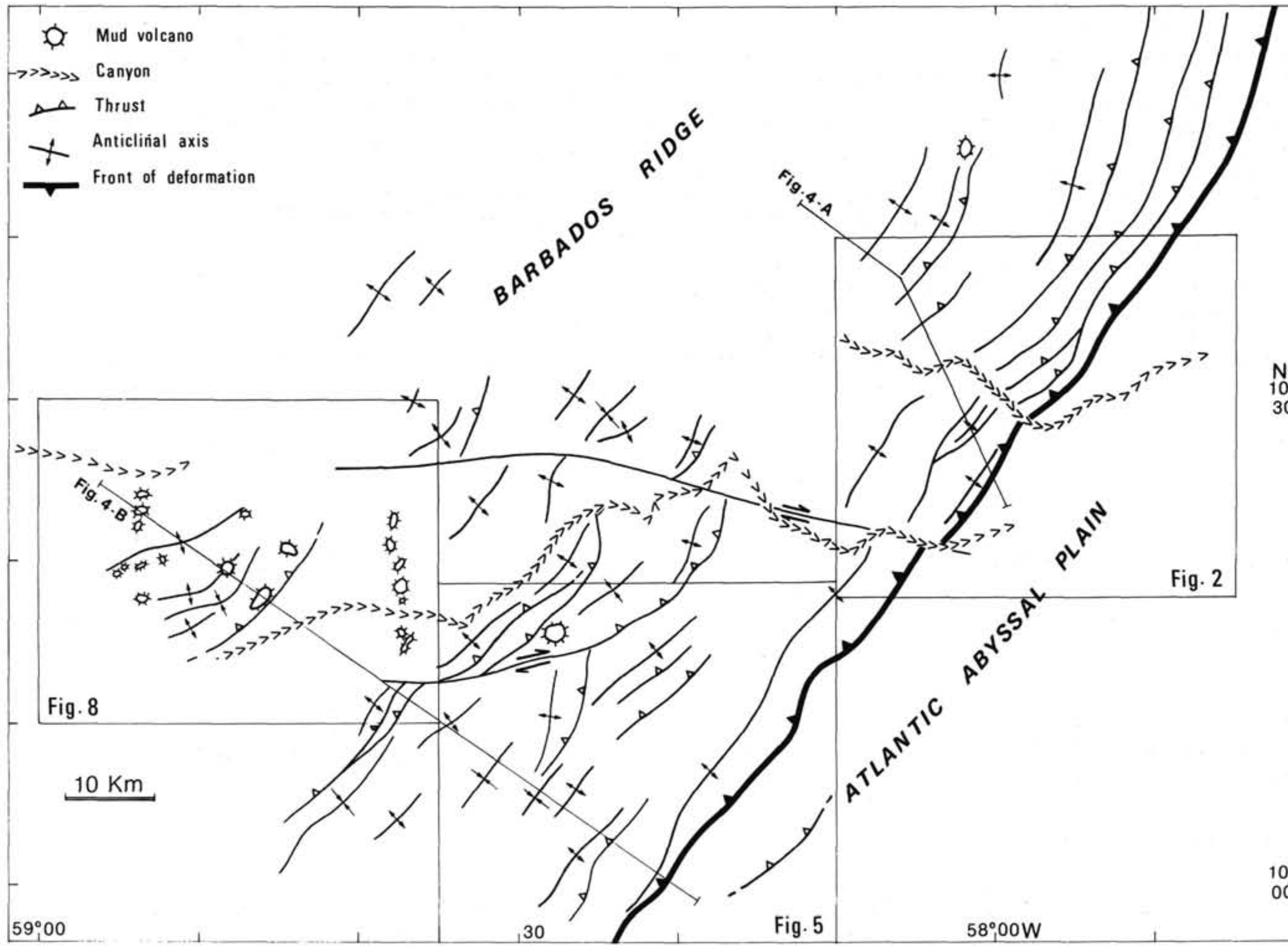


Figure 13. Structural map of the southern edge of the Barbados Ridge.

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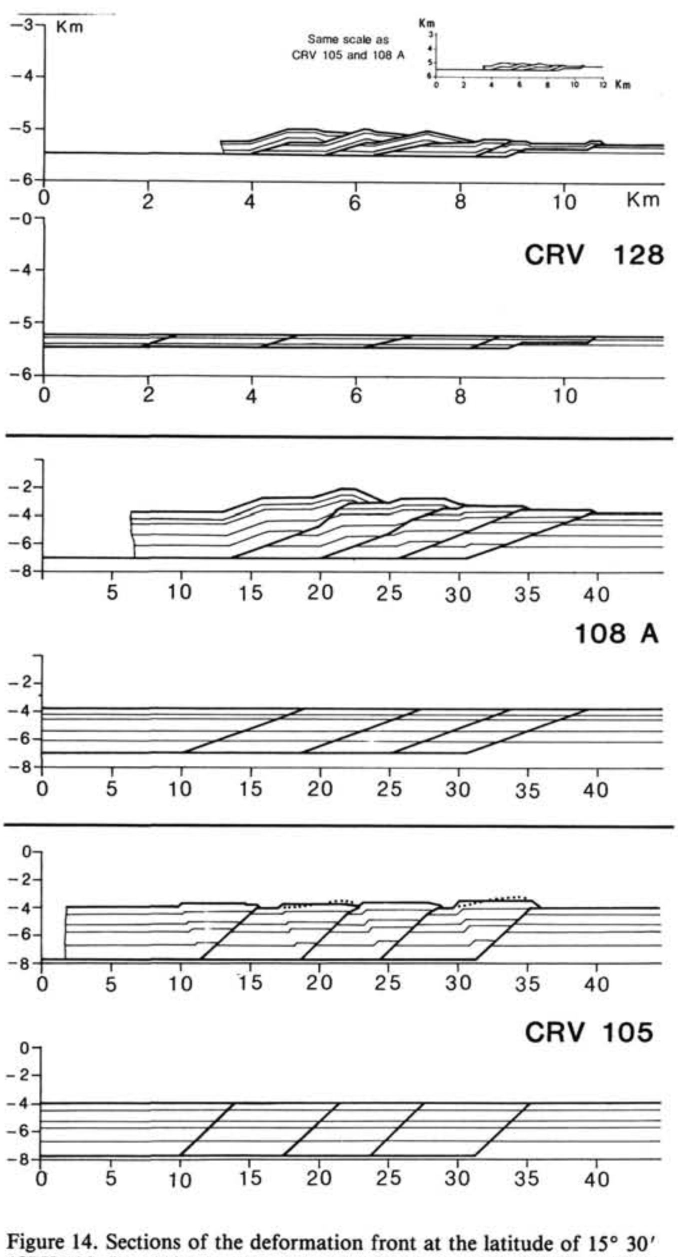

Figure 14. Sections of the deformation front at the latitude of 15° 30′ (CRV 128, Leg 110 transect), 10° 50′ (108A), and 10° 00′ (CRV 105). On line CRV 105 the dotted line on the seafloor represents a smoothed profile because the kink method used for the computation of folding is not suitable for relatively uncompacted sediments.

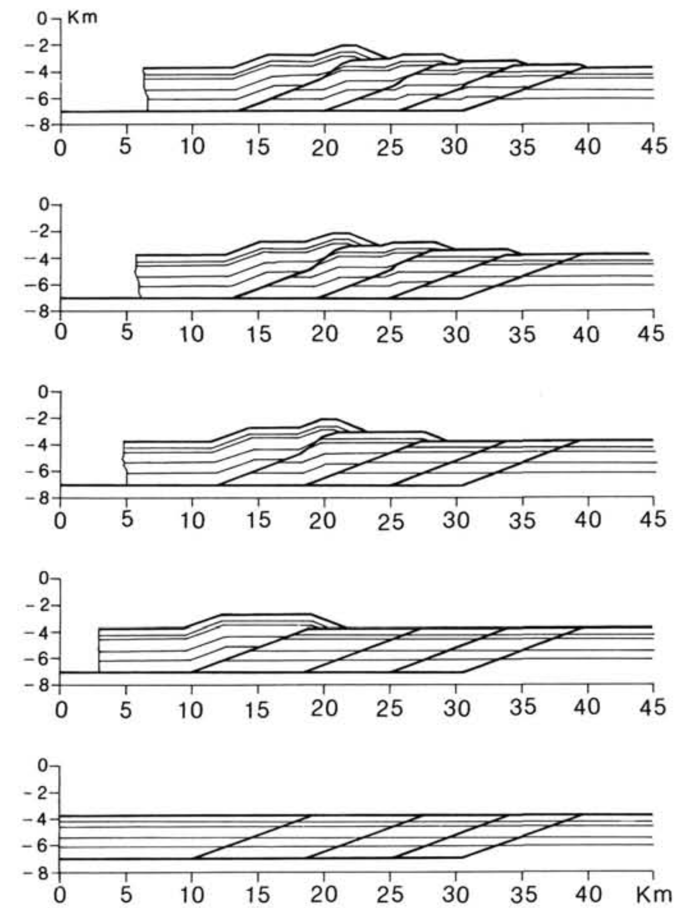

Figure 15. Forward modeling of frontal accretion on line 108A.

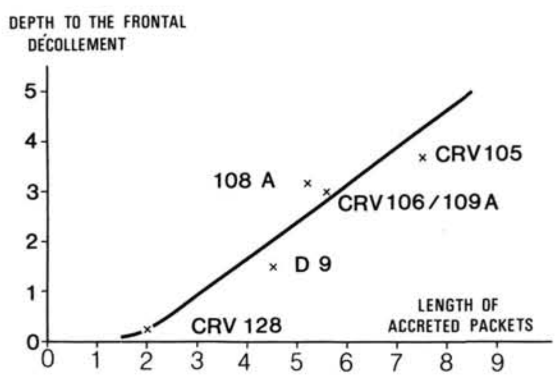

Figure 16. Relationship between the length of accreted packets and the depth of the décollement at the deformation front.