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# *Lateral Changes of Frontal Accretion and Mud Volcanism Processes in the Barbados Accretionary Prism and Some Implications*

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## ABSTRACT

This paper focuses on tectonic, sedimentary, and hydrogeologic processes presently occurring at the eastern leading edge of the Barbados accretionary prism. Thrusts and folds develop over a décollement hosted at the deformation front in weak and probably undercompacted sedimentary layers of Miocene age. The volume of the anticlinal potential traps is changing from north to south and is related to the thickness of accreted sediments, i.e., of the Neogene-Quaternary section. This thickness is increasing drastically from north to south as it approaches the South American continent and the Orinoco deep-sea fan. The middle and distal parts of this fan are now being incorporated into the prism, supplying the piggyback basins developing between the growing folds and the immediate foreland in front of the prism with deep-water clastic sediments. These potential reservoirs in the Neogene-Quaternary section are possible targets for future exploration, subject to their geophysical identification, in such areas where the seismic resolution may be altered seriously by the complexity of structures. The distribution of potential source rocks is poorly known, but marine middle–Late Cretaceous source rocks are widespread on the northern South American continent, and also may have been deposited in the near-deep marine environment. Efficient paths for the migration of hydrocarbons originating and expelled from such source rocks are provided by deep-water clastic layers, active faults, and widespread mud volcanoes. The success of efficient entrapment of hydrocarbons could, in some cases, be hampered seriously by the efficiency of these vertical conduits connecting deep stratigraphic layers and the sea bottom.

## INTRODUCTION

The eastern boundary of the Caribbean Plate corresponds to the Lesser Antilles active margin, which includes at its leading edge the Barbados accretionary complex, an archetype of a mature accretionary prism (Figure 1). This huge tectonic prism is located in water depth generally in excess of 1000 m. This is still a frontier area in which petroleum exploration has started recently and is likely to develop in the near future. Indeed, the growing interest in exploring continental margins mainly has focused during recent times on nonvolcanic passive margins, especially where already productive areas are present upslope with well-known petroleum systems assumed to extend from the shelf to deeper water. On the other hand, active margins have received little attention because most of them are not associated directly with productive areas upslope. Petroleum systems are generally unknown or, in better cases, only inferred from surrounding basins. Large uncertainties remain, therefore, concerning the distribution of source rocks and reservoirs. Furthermore, such margins are located near active volcanic arcs or cordilleras; i.e., near areas with a supply of expected low-quality clastics (from a petroleum reservoir point of view). In addition, seismic imagery is a problem because severe tectonic deformations in the accretionary wedges result in very complex geometries of basins and potential traps and consequent poor seismic resolution.

The forearc of such active margins is characterized by high shortening rates, and by frequent development of tectonic overpressure associated with shale diapirism and mud volcanism. These margins, however, can contain enormous volumes of sediments when they are located in the vicinity of continents. Indirect evidence for hydrocarbons often exists (organic-rich sediments cored or dredged on top of mud volcanoes, gas hydrates). Some of them are close to producing areas, such as, for instance, Trinidad and Eastern Venezuela to the south of the Lesser Antilles active margin (Rodriguez, 1988; Talukdar et al., 1988; Heppard et al., 1998). Oil also has been produced for several decades in onshore Barbados Island, the emerged summit of the Barbados accretionary prism (Payne et al., 1988; Speed et al., 1991; Babaie et al., 1992).

Many academic studies have been devoted to this accretionary wedge. See Biju-Duval et al. (1982), Speed et al. (1984), and Brown and Westbrook (1987) for a comprehensive review. Many works have focused on the tectonic front to the north where four DSDP/ODP legs (78A, 110, 156, and 178A) have been drilled. Active mud volcanism processes have been studied in detail in the abyssal plain east of the prism (Henry et al., 1990). Piggyback basin development and interaction between tectonic deformation, sea-floor topography, and deep-marine clastic sedimentation related to the Orinoco deep-sea fan have been investigated at the southern edge (Mascle et al., 1990; Faugères et al., 1993; Huyghe et al., 1999). In this paper, we provide an overview of the recent to present deformation processes at the leading edge of the Barbados Ridge, together with resulting implications for fluid dynamics and potential hydrocarbon entrapments.

## REGIONAL SETTING OF THE LESSER ANTILLES ACTIVE MARGIN

An oceanic seaway was initiated as recently as the Late Jurassic between the North and South American continents as a result of the southwest propagation of the Central Atlantic Rift and Ocean. This seaway subsequently was subducted below the Caribbean Plate, an oceanic domain supporting volcanic plateaus of Late Jurassic–Early Cretaceous age, probably originating in the South Pacific area. The Caribbean Plate has migrated successively northwestward, then westward, since the Late Cretaceous (Dercourt et al., 1993; Mauffret and Leroy, 1997; Pindell et al., 1998; Mann, 1999). The convergence rate and direction of the Caribbean and Atlantic Plates for the last 10 m.y. are not accurately known, but are believed

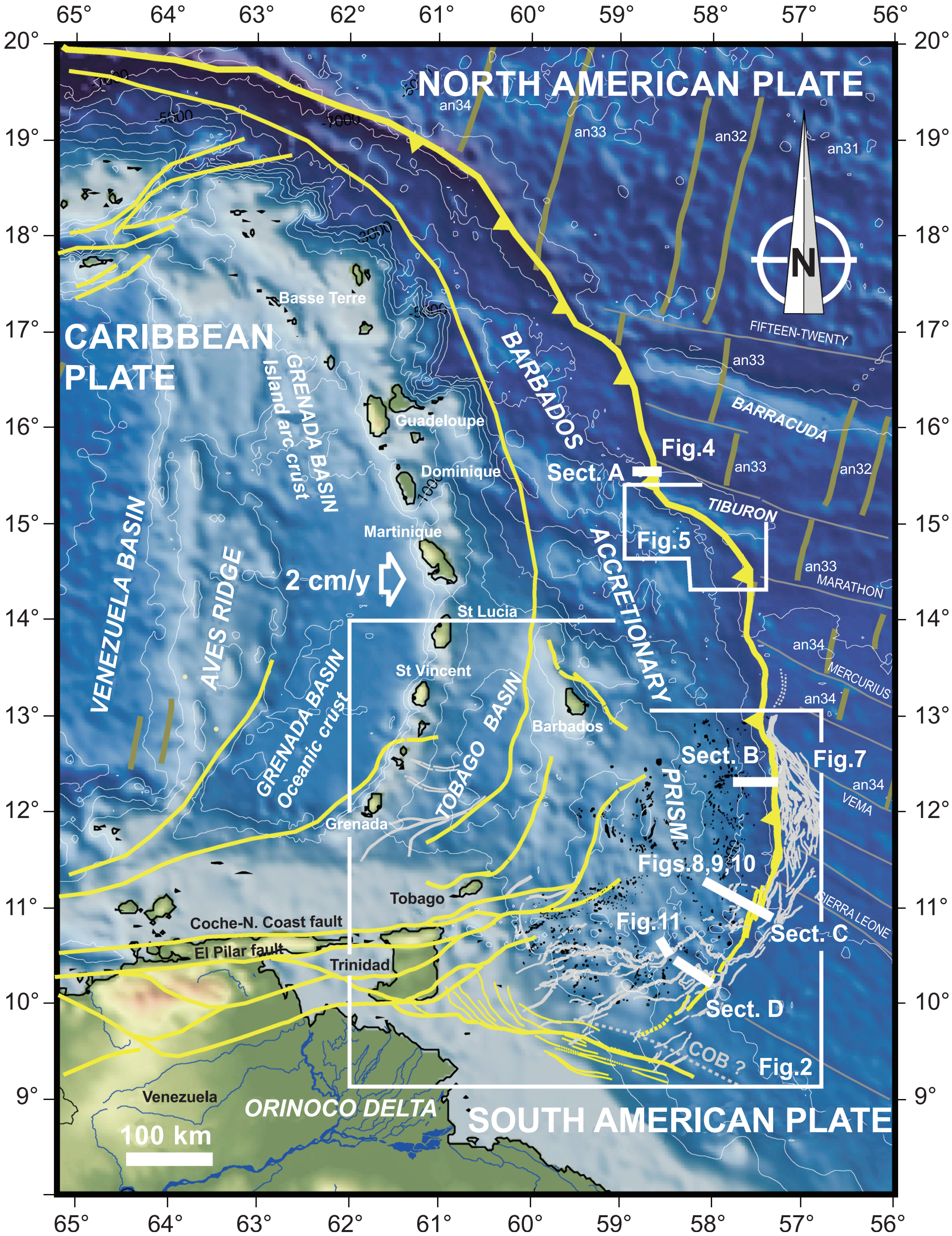



**Figure 1.** Simplified sketch map of the Barbados prism and location of figures. an = magnetic anomalies; COB = continent-ocean boundary.

to approach 2 cm/yr in an east-west direction (Pindell et al., 1998; Dixon et al., 1998; Weber et al., 2001). Although still poorly constrained, a slight convergence (2 mm/yr) is expected generally between the South and North American Plates near Barbados Ridge, between 18° and 150°N, probably reactivating east-west-trending oceanic transform faults (Summer and Westbrook, 2001).

Several active margins and island arcs have developed at the leading eastern edge of the Caribbean Plate since the Cretaceous. The Cretaceous volcanic arc is believed to be present in the Aves Ridge, a remnant arc now being separated from the present arc by the Grenada back-arc basin of inferred early Tertiary age (Bouysse, 1988; Bird et al., 1999). Two overimposed Tertiary volcanic arcs (late Paleocene to early Miocene and middle Miocene to Holocene) can be differentiated from field evidence (Bouysse et al., 1990; Bouysse and Mascle, 1994). Both are well exposed in the Lesser Antilles islands. The Lesser Antilles margin is seismically active, and the spatial distribution of epicenters shows that the subducting slab is deepening from east to west under the active volcanic arc to depths of more than 200 km. To the south, the slab is being subducted below the northern part of the island of Trinidad and the Paria Peninsula of Venezuela (Russo et al., 1993).

In front of the volcanic arc, a large part of the sedimentary cover of the subducting Late Jurassic (?) to Late Cretaceous Atlantic oceanic crust has been scraped off to form the Barbados Ridge. Because of the immediate vicinity of the South American continent, large influxes of siliciclastic sediments have been deposited in the oceanic foreland, resulting in development of a very broad and thick accretionary prism to the south, but narrowing to the north (Speed et al., 1984). Two main stages of voluminous sedimentation and subsequent accretion have been recognized so far, one in the Eocene and a second in the Miocene to Holocene. Eocene accreted turbidites are well exposed on the island of Barbados, and sandy turbidites have been cored farther north at site ODP leg 110. More recent influxes of sediments originate from the Orinoco River and, to a lesser extent, from the Amazon Delta. The Orinoco deep-sea fan now is being incorporated into the southernmost part of the Barbados Ridge.

## TECTONIC PROVINCES OF THE BARBADOS ACCRETIONARY PRISM

East of the prism front, the Atlantic abyssal plain is segmented by multiple west-northwest–east-southeast-trending fracture zones and by prominent ridges (particularly Tiburon and Barracuda) in the area of the diffuse plate boundary between North and South America. These ridges have been interpreted to be the result of transpressive movements during the Tertiary (Mueller and Smith, 1993). South of Tiburon, the Atlantic Plain is covered in the vicinity of the prism by a braided turbiditic system (channel-levees) of the Orinoco deep-sea fan (Belderson et al., 1984 ; Faugères et al., 1993; Figure 1).

In its northern part, the accretion prism is composed of a single morphological ridge, whereas south of the island of Barbados, the prism comprises two different ridges, designated generally as the Barbados Ridge (the eastern ridge), and the Barbados Crest (the western ridge). These two ridges are separated by the Barbados Basin, which corresponds to an isolated depression on the top of the accretionary complex (Biju-Duval et al., 1982; Valery et al., 1985).

In the southern part of the Barbados prism, the continuous terrigenous flux, sourced mainly from the south by the Orinoco River at least since the Eocene, induces lateral changes in the syntectonic sedimentation above the accretionary prism and in the sedimentation of the Orinoco deep-sea fan, both of which contribute to the growth of the prism (by piggyback basin deposition and by frontal accretion, respectively). This influences the evolution of the superficial morphology, the geometry of the structures, and the fluid dynamics within the prism. Furthermore, the southern part of the Barbados prism that is close to clastic sources shows evidence of interactions between tectonic and sedimentation processes and expulsion of fluids associated with mud diapirism.

In the southern part of the prism, we can distinguish the following tectonic provinces (Biju-Duval et al., 1982; Brown and Westbrook, 1987): (1) the front of the prism, which is characterized by an imbricate thrust system partly covered by piggyback syntectonic turbiditic basins with fan geometry and partly eroded by a system of canyons in the southernmost part of the Barbados Ridge; (2) a large province of active mud volcanoes and shale domes (associated with seeps of methane-rich waters; Jollivet et al., 1990) with sedimentary sequences having very complex geometries. This belt of mud volcanism extends toward the south onshore in Trinidad and Venezuela; and (3) the Barbados basin, which appears as a closed basin in the tectonic prism and is characterized also by mud volcanism. The southern part of the prism (south of Barbados island) is characterized both by synsedimentary extensional structures (active normal faults; Deville, 2000) and by coeval backthrusting at the western edge of the prism bounding the Tobago Basin. Below the superficial sequences of the Barbados Crest that record extensional tectonics is a sedimentary basement that is strongly deformed by thrust tectonics and where fossil mud volcanism is known (Joe's River Formation; Senn, 1940).

## FRONTAL ACCRETION PROCESSES

### *Sole Thrust*

In the northern area of the accretionary prism, legs DSDP/ODP 78a, 110, 156, and 171A have demonstrated that the décollement at the toe of the prism is hosted in the early Miocene hemipelagic radiolarian claystones. A proto-décollement 4 km east of the deformation front is hosted in the same lithostratigraphic interval. Horizontal shearing and high preserved porosity have been encountered in this interval (60–80% at a depth of 200 m, i.e., with lower strength than the surrounding rocks), as well as geochemical anomalies in pore fluids (high methane content). These anomalies are similar to those found in the décollement below the accretionary wedge (Gieskes et al., 1990). These data strongly suggest that the décollement is propagating far ahead of the deformation front, which, in turn, would imply that the Neogene sediments above already are undergoing slight horizontal shortening (porosity loss, microfaults, and folds below seismic resolution). Similar results from other active margins (e.g., Nankaï Trench; Moore et al., 1991) have led to definition of a proto-accretionary zone in front of the main accretionary wedge.

The décollement is defined at the base of the accretionary complex as a 20–40-m-thick interval of scaly fabric. The related bulk density is low in the foreland and also is low in some places below the accreted complex, which suggests that sediments are overpressured at these locations (Moore et al., 1998; Moore, Klaus et al., 1998). Mineral veins (zeolite, rhodocrosite) and geochemical pore-fluid anomalies (methane, chloride) argue for enhanced fluid flows in this décollement (Moore et al., 1988). Reflection profiles revealed spatial variations in reflection polarity of the décollement (Shipley et al., 1994), for which the negative polarity has been interpreted as planar dilatancy related to elevated fluid pressure along the décollement (Bangs et al., 1996; Moore et al., 1997), and/or abnormally high-porosity matrix (under compacted layers) (Moore et al., 1998; Moore, Klaus et al., 1998; Bangs et al., 1999). A significant increase in porosity below the décollement at the edge of the Nankaï Trough, Japan (ODP Leg 190), has been interpreted similarly as resulting from undercompaction of the rapidly buried underthrust section (plus enhanced tectonic compaction caused by subhorizontal stress).

Sediments of Neogene age above the décollement are accreted in sequence and form narrow (1.0- to 1.5-km-wide) and thin (a few hundred meters) ramp anticlines. Late Cretaceous to Oligocene sediments below the décollement are underthrust initially without evidence of deformation over distances of several tens of kilometers for the youngest, and possibly up to the crustal plate boundary for the oldest, as suggested by regional deep seismic profiles (Westbrook et al., 1988). ODP leg 110 actually has demonstrated that the décollement is deepening rapidly to the west in older strata. Severely deformed Oligocene and Eocene rocks, quite similar to those found in the foreland, have been discovered far above the décollement at site 674, located only 18 km from the eastern edge of the prism. Deformed Eocene sandy turbidites also are well exposed on the island of Barbados. The deepening of the décollement also is well evidenced on the seismic data (Biju-Duval et al., 1982).

From a more general point of view, according to our interpretation of available seismic data, the Miocene horizon corresponds to the main décollement level in the entire prism toe, except locally where upper secondary décollement levels occur, notably in the Pliocene. Indeed, for instance, on the site of the DSDP/ODP wells, a complication of the frontal thrust can be seen with the development of a flat in the Pliocene (Brown et al., 1990; Figure 5). Also, in several places in the Barbados prism toe, the thrust faults tend to bifurcate in several branches toward the surface, and microprisms develop in the Pliocene-Quaternary formations. Nice examples can be seen in the area of 12° N (Biju-Duval et al., 1982). In the southern part of the prism, the sole thrust develops at much greater depths than in the DSDP/ODP sites and is not visible on the available scientific seismic data.

Moreover, only Tertiary formations are involved in the front of the Barbados prism. In the northern part of the prism, even in its inner parts, there is no known evidence or implication of Cretaceous formations in the prism. In the inner parts of the prism on Barbados island, the older formations that are outcropping are of Tertiary age (Eocene turbidites of the Scotland group). The Cretaceous fossils described on Barbados are reworked in the Eocene turbidites (Senn, 1940). Even in the fossil mud intrusions (Joe's River Formation), the older exotic blocks are of Paleocene age (Biju-Duval et al., 1985). However, toward the south, the Cretaceous clearly is involved in the tectonic wedge at the vicinity of the island of Trinidad.

The topography of the prism front (critical taper of the tectonic wedge) exhibits changes that could be related to mechanical behavior of the décollement. Mean slope values of 2–3° are observed in the frontal folds area, which correspond to a low friction condition for the basal décollement, as can be expected for undercompacted clay. Farther west (in the mud volcano and shale diapir area), the mean slope decreases again and tends to be lower than 1° (Figure 4). This suggests a relatively rapid change of properties of the sole in this area, with extremely low friction conditions. In the southernmost part of the prism, the topography is influenced largely by sedimentary input from the Orinoco River, which covers the tectonic prism (Figures 1, 2, and 3).

### *Structure of the Frontal Thrust System*

The structure of the prism front immediately north of 15°N is well constrained by wells from the four DSDP/ODP legs (78A, 110, 156, and 178A). This is an area of low clastic sediment supply and relatively thin (less than 1000 m) sedimentary cover on the incoming Atlantic crust (Moore et al., 1997). In this northern part of the prism, a stack of imbricate thrusts characterizes the frontal zone. The available balanced structural sections have been published after the results of leg 110 (see, notably, Brown et al., 1990). An updated section is proposed in Figure 5, which takes into account results of the wells of the ODP legs 156 and 178A. More thrust faults are proposed, but with fewer throws along each fault. A total shortening of 4 km is deduced from the restored, initially long section. These deformations are of Quaternary age (Pleistocene sediments are found below late Miocene sediments along the upper ramp, cored at site 671). Lack of seismic resolution and scarcity of wells prevent any detailed structural analysis farther west in the area of the DSDP/ODP wells (at 15°30′N).

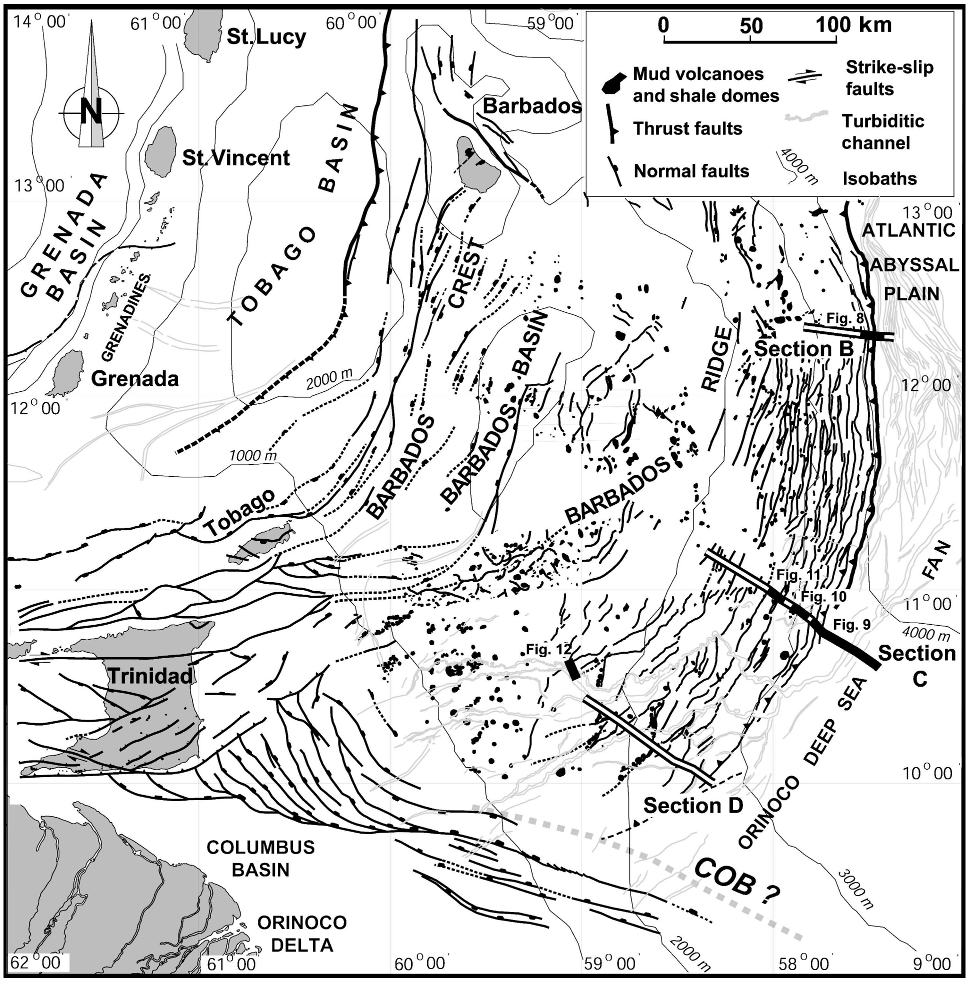



**Figure 2.** Structural sketch map of the southern area of the Barbados accretionary prism. COB = continent-ocean boundary.

The deformation front between 11° and 15°N is well organized and well imaged by seismic and multibeam bathymetric data (Biju-Duval et al., 1982). Ramp anticlines are about 5-km wide, and their initial longitudinal length is about 25–30 km. The stratigraphic control in this area is poor, because the only way to date seismic reflectors is by long-distance ties from wells in the vicinity of Trinidad. The proposed ages thus have to be considered as rough estimates only. For instance, the Pliocene and Pleistocene interval could be much more developed and, conversely, the Miocene section could be thinner. These ages will be used, however, to estimate a lower limit for the rate of deformation. High-resolution seismic profiles provide evidence for the sequential propagation of ramp anticlines. A well-defined transparent seismic unit of Pliocene (or younger ?) age is progressively buried below younger deposits from west to east; i.e., from the inner part of the deformation front to the foreland. Keep in mind that most of the present sedimentation is related to channelized turbiditic currents on the deepest basin floor (Faugères et al., 1993); this implies that the ramp anticlines rise successively above the level of the

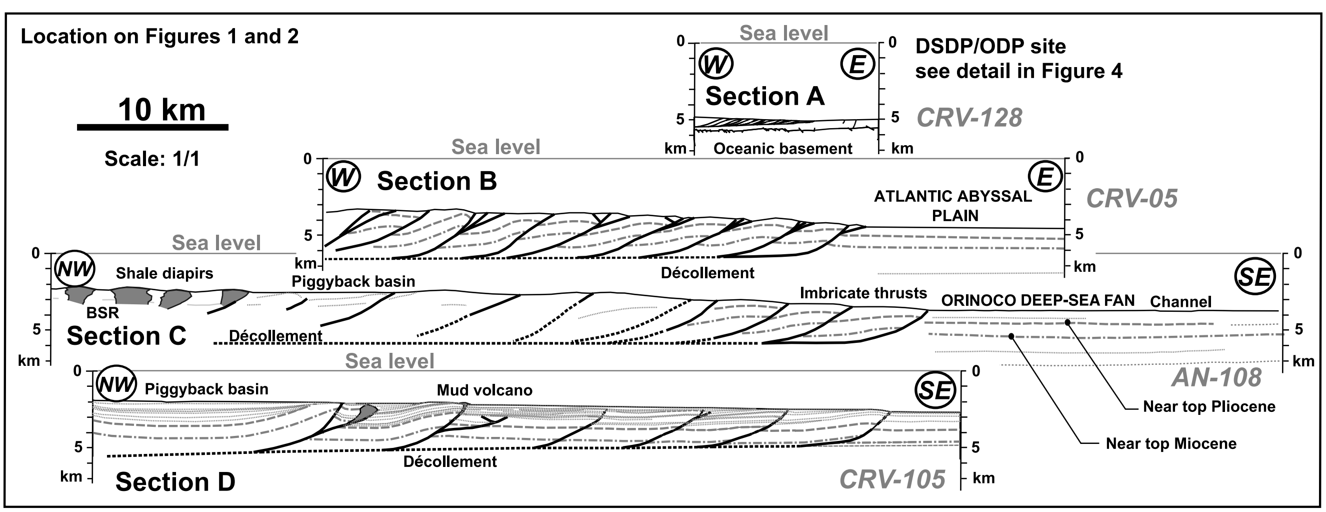



**Figure 3.** Depth sections interpreted from several seismic profiles in the front of the Barbados prism.

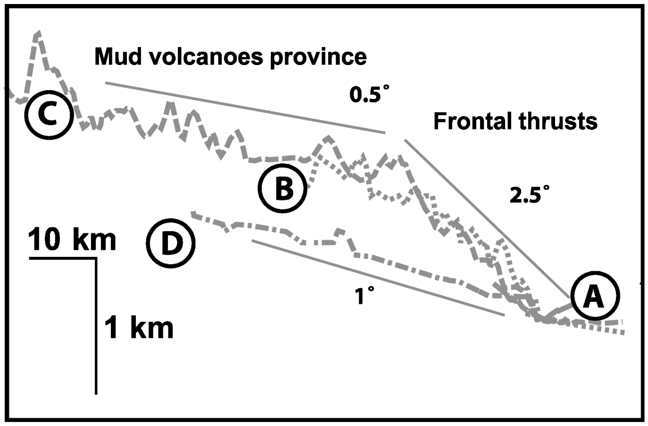



**Figure 4.** Topographic profiles along geological sections shown in Figure 3, locations of which are shown in Figures 1 and 2.

foreland sea bed from west to east. This does not imply, however, that propagation of a new ramp will cause the inactivity of the previous one. The progressive uplift of successive ramp anticlines from east to west and related observed increase of displacement along these ramps testify to a simultaneous, wide group of folds forming the deformation front and slope, which allows the morphological (taper) profile of this edge of the accretionary prism to be maintained (for as long as boundary conditions, such as basal friction, remain the same).

The deformation front around the latitude of 11°N shows tectonic features significantly more complex than those to the north (Figure 2). Four more external ramp anticlines developed above ramps originating from a 2500-m-deep (below seafloor) décollement. These anticlines are 4.5-km wide and lead to a 750-m elevation of the seafloor with respect to the foreland. A wider anticline develops immediately at the back, with a single elevation of 450 m. This geometry should reflect a different deformation process. The development of this anticline is believed to have occurred over a deeper décollement level, with a subsequent passive transportation over the shallower décollement. Other interesting features are the occurrence of mud volcanoes, bottom-simulating reflectors (BSR), and thin piggyback basins along this transect. Present sedimentation observed in piggyback basins requires that basins (topographic lows) are being formed at the back of rising ramp anticlines, but also that clastic influxes can reach these basins (otherwise pelagic sedimentation will uniformly drape over the sea-floor morphology). This is indeed the case, as the accretionary wedge is developing south of 11°N at the expense of the Orinoco deep-sea fan (Figure 1).

From 10°30′ southward, there is also a distinctive change in the geometry of ramp anticlines. Instead of the well-organized trends of linear anticlines described previously, a more complex pattern of sigmoidal folds, submarine canyons, and mud volcanoes is depicted on the multibeam bathymetric maps (Valery et al., 1985; Mascle et al., 1990). A representative section across the frontal ramp anticlines is given in Figure 2. The total amount of shortening approaches 8 km. The rate of shortening could be proposed if the age of the syntectonic infilling of the thick piggyback basins were known, which is not the case, unfortunately. Long-distance correlations from the Venezuelan shelf (Di Croce, 1995) are not very accurate but can be used to propose a Pleistocene or late Pliocene–Pleistocene age for the thickest basins. Accordingly, the rate of shortening of this 55-km-long section would be 2.5 to 5 mm/yr.

In the southern area of the Barbados prism, where thrust tectonics interact with intense turbiditic sedimentation, faults and folds develop according to the following sequence: (1) creation of a thrust fault that affects the entire stratigraphic series from the décollement level (so this obviously is not a mechanism of fault-propagation folding); and (2) progressive development of an anticline in the hanging wall above the thrust fault, whereas the thrust has a moderate throw (not enough to consider the ramp anticline as generated by fault-bend folding).

### *Transfer Zones*

The segmentation of the Atlantic Plain by west-northwest–east-southeast fracture zones has generated step geometry in the Barbados prism foreland that has drastically controlled the thickness of the accreted sediments in the accretionary complex. This has influenced the development of very spectacular transfer zones in the prism in the western continuation of the Barracuda and Tiburon rises and in the Marathon and Mercurius transform faults, particularly since examples of lateral ramps are present in the west of the Tiburon rise (Figure 6). The rise has acted as a barrier to turbiditic sediment supply issued from the South American shield (a thick sedimentary pile to the south and a thin cover to the north). Consequently, a very significant jump is observed in the location of the outer front of the Barbados (Mascle and Moore, 1990; Ladd et al., 1990). In order to understand how this foreland geometry has

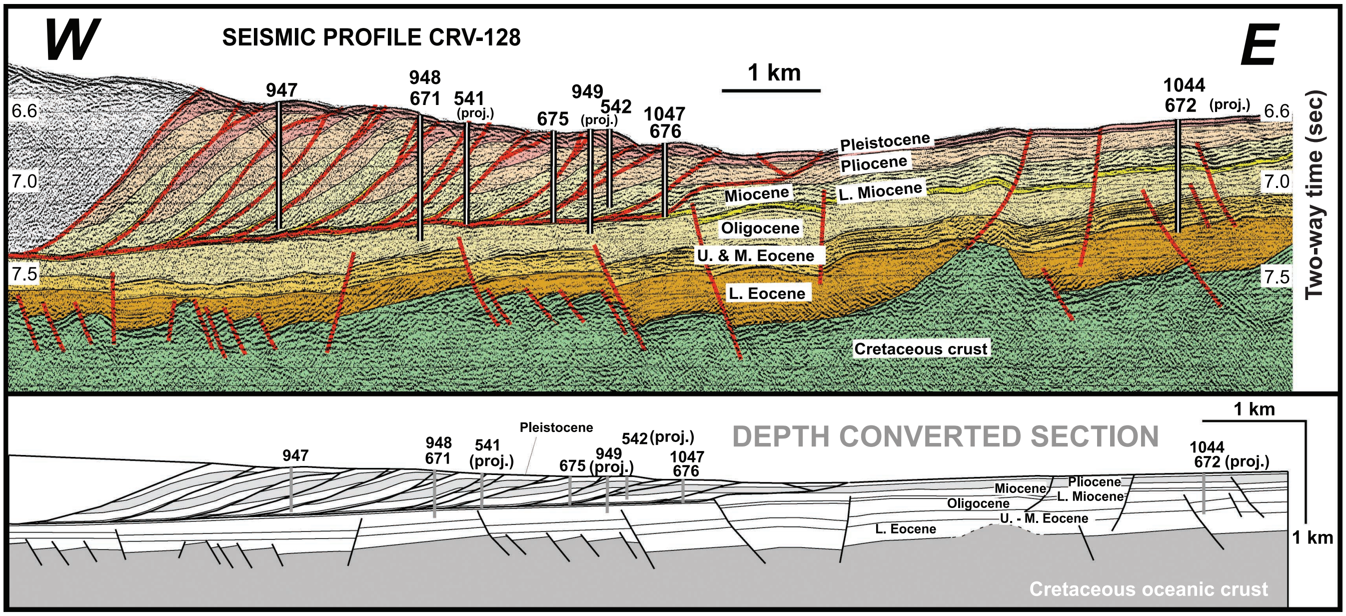



**Figure 5.** Updated seismic interpretation of profile CRV-128 in the area of the ODP legs (at 15°N) and depth-converted balanced cross section.

affected the propagation of thrusts, sand-box experiments have been conducted with an equal amount of shortening applied to a box with contrasting sand thicknesses. The results, shown in Figure 7, support the direct link between sedimentary thickness, spacing between ramps, and quicker forward propagation of the deformation front. These results also suggest that the regularly spaced elevations of the sea bed observed on the seismic section across the transform zone should be related to the emergence of lateral ramps (Figure 5).

The southern edge of the Barbados prism corresponds to en-echelon terminations of folds complicated by the normal growth fault system of the Columbus Basin (Figure 1). The tectonic front then is transferred to southern Trinidad where marine Late Cretaceous to Pliocene sediments are severely deformed in a south-verging fold and thrust belt facing the South American shield. But how the deformed inner Barbados accretionary wedge passes laterally to the south into the continental rise west of Trinidad still is poorly understood and a matter of debate. This is obviously not achieved across a single dextral tear fault. The available multibeam imagery (Valery et al., 1985) shows an east-west-trending area about 30-km wide and slightly more than 100-km long, with a complex pattern of highs and deeps with numerous mud volcanoes. The rugged morphology of the sea floor and the widespread occurrence of mud volcanoes and active fluid vents have been observed from deep diving (Vincent and Faugères, 1993). Pindell and Kennan (2001) propose that the locus of primary plate-boundary displacement jumped from the Cuche–North Coast fault (Figure 1) to the Southern and Central Ranges of Trinidad at about 5 Ma, such that we should not expect much deformation north of Trinidad since that time.

## SYNTECTONIC SEDIMENTATION AND TIMING OF THE DEFORMATION

Formed and carried syntectonic basins above the tectonic wedge in the southern part of the accretionary prism have been described in a few articles (Chennouf, 1987; Mascle et al., 1990; Huyghe et al., 1999). The sedimentary infilling of these basins can be a result of either (1) external sources other than the prism, such as terrigenous flux by gravity deposition (e.g., turbidites, grain-flows, debris-flows) and hemipelagic sedimentation, or (2) recycled sediments from within the prism by mud intrusions or superficial mud effusions around mud volcanoes and gravity mass flows from the surrounding slopes.

The age of these sediments is very poorly constrained because of lack of any drilling. Seismic stratigraphy analyses by correlation with well data from the Orinoco platform (Columbus Basin) provides indications about the age of the syntectonic sequences and suggests that most of the syntectonic sediments are of Quaternary age (Di Croce, 1995; Di

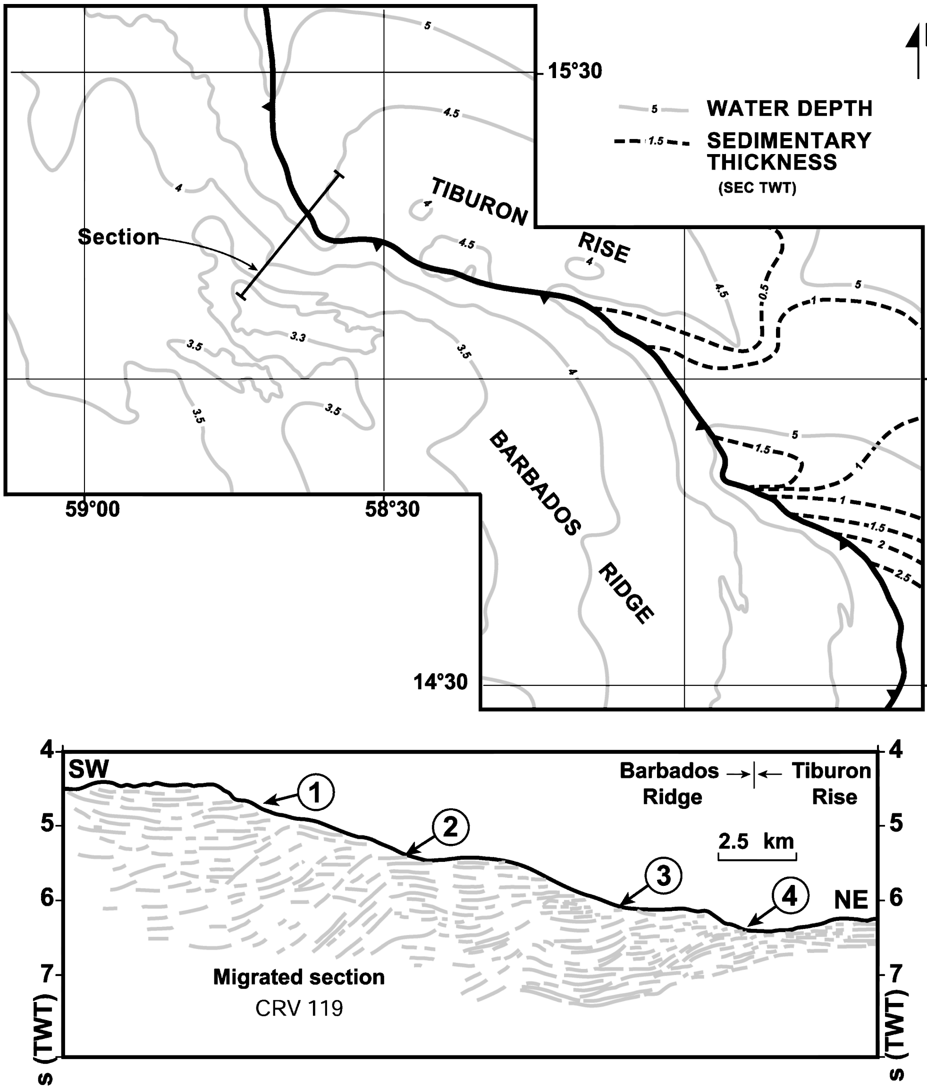



**Figure 6.** Bathymetric map of the South Tiburon Ridge Tranverse Zone (top), and line drawing of a seismic section across the inferred lateral ramps (arrows) of the Barbados Ridge front at 15°15′ N.

Croce et al., 1999). Consequently, rates of sedimentation also are poorly constrained. Sedimentation is coeval with erosion processes and sedimentation bypass, as evidenced by the canyon network intersecting the prism (Valery et al., 1985; Mascle et al., 1990) and by the deep-sea fan sedimentation in the Atlantic abyssal plain at the toe of the prism (Belderson et al., 1984; Faugères et al., 1993) (Figure 9).

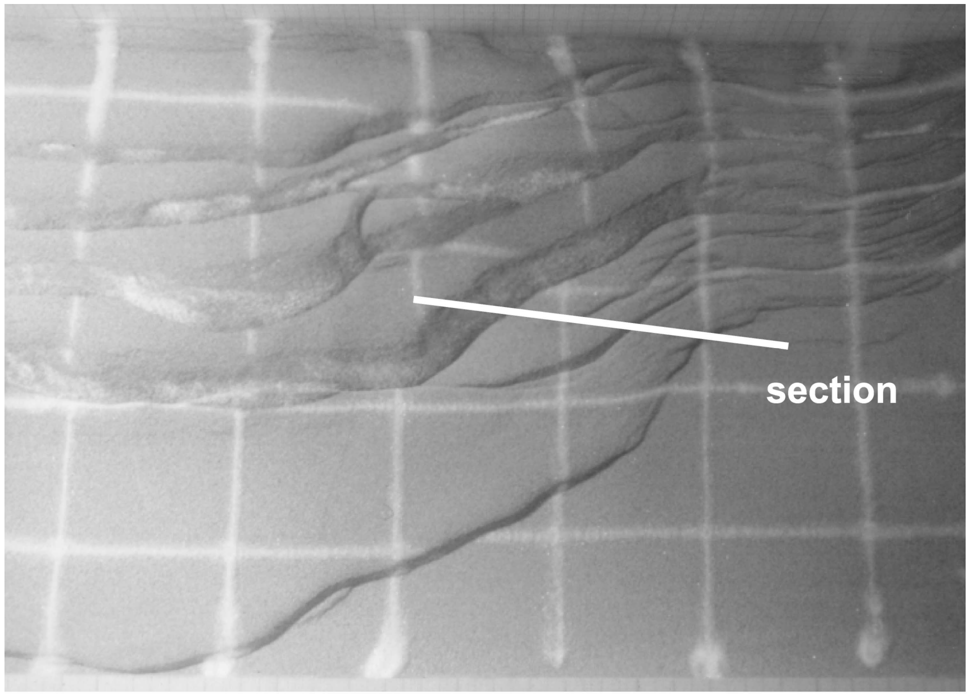
We have already mentioned that the rate of shortening on the 55-km-long section to the south (10°N) would be in the range of 2.5 to 5 mm/yr, with an average horizontal displacement along individual ramps of about 1 mm/yr. This suggests that the frontal thrust structures have absorbed only some of the plate convergence (2 cm/yr) classically proposed for the Quaternary. This difference can be related to shortening processes below the resolution of seismic reflection data and/or additional shortening in the inner parts of the prism. If our age estimations are correct, at least 30 km of imbricate thrusts were added at the front of the accretionary wedge in the Pleistocene. The present rate of accretion could be even more rapid if the upper sedimentary package in the foreland is younger than expected; i.e., if it is largely dominated by Pliocene-Pleistocene influxes from the Orinoco Delta. This rate of accretion can be compared to the 40km-wide accretionary complex inferred from ODP Leg 190 to have been added to the Nankaï Margin in the Pleistocene (Leg 190).

Some ramp anticlines are sealed by the active turbiditic sedimentation of the Orinoco Delta, especially in the core of the prism (Figure 12), whereas the frontal folds clearly are synchronous with syntectonic sedimentation (Figures 8 and 10), which outlines the forward propagation of the compressive deformation toward the front of the prism. This is evidence of propagation of the deformation from the core of the prism to the present front during the very recent Quaternary.

## FLUID MIGRATION AND SUBSURFACE SEDIMENT MOBILIZATION

### *Fluid Migration*

The thermal regime in the Barbados prism area is strongly influenced by fluid migration mechanisms. As a consequence, thermal conditions and fluid dynamics are closely linked. In the Atlantic abyssal plain, at the toe of the prism, heat flows measured show variable values (between 30 and 120 mW/m$^2$; Foucher et al., 1990; Langseth et al., 1990; Henry et al., 1990, 1996). The areas of highest values could



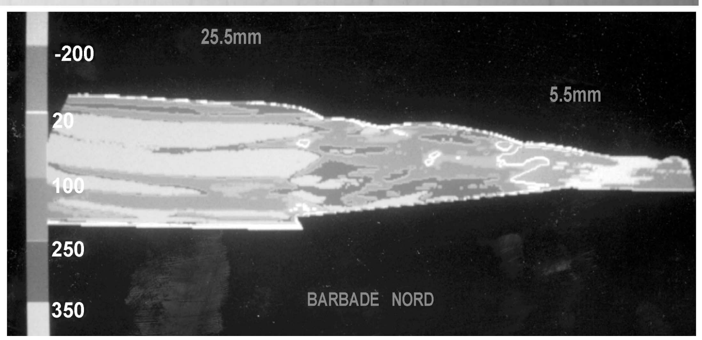




**Figure 7.** Sandbox experiment and X-ray tomography section across the transform zone of the sandbox experiment; compare with the structure of the area shown in Figure 6.

be related to fluid escape issued from below the prism, at the base of the sedimentary pile or in the oceanic crust (Henry, 2000). It is also possible that the oceanic lithosphere is locally abnormally hot with respect to its Late Cretaceous age (the expected heat-flow values would be around 50 mW/m$^2$). Especially in the area of the Barracuda and Tiburon Rises (diffuse boundary between North and South America), this could relate to a possible thermal event during Eocene and Oligocene extensional tectonics (evidenced by several active normal faults during this period). As already mentioned, focused fluid flow has been evidenced along the décollement and related thrusts along the DSDP/ODP transect (Moore et al., 1988). This is best indicated by pore-fluid geochemistry (Blanc et al., 1988; Gieskes et al., 1990; Kastner et al., 1997) and the occurrence of mineral veins (Labaume et al., 1995; Vrolijk and Sheppard, 1991). For instance, the low chlorinity values of fluids in fault planes have been related to dilution by dehydration water of the compacted clays. Similarly, high methane contents have been interpreted as migrating thermogenic methane from the inner part of the prism (Blanc et al., 1988; Gieskes et al., 1990). We cannot exclude the possibility that dilution of ground water and high methane contents could be related partly to destabilization of gas hydrates during the uplift associated with the tectonic thickening of the prism.

The décollement and active ramps branching up act as preferential pathways for fluids expelled during compaction and/or clay diagenesis, in an overall environment in which low-permeability hemipelagic sediments prevent efficient direct flow to the surface. Fluid flows probably originate from the inner, western part of the prism where several kilometers of sediments are tectonically stacked. But in some places, fluids could migrate along strike from the south, as a result of southward deepening of the oceanic basement when crossing oceanic fracture zones (e.g., Tiburon Rise; Screaton and Ge, 1997).

There is growing evidence that enhanced permeability in these faults (as much as $10^{-12}$ to $10^{-13}$ m$^2$ with respect to $10^{-14}$ to $10^{-20}$ m$^2$ in the sediment), and consequently the partitioning of flow between the sediment matrix and the faults, could be spatially heterogeneous and transient (Saffer and Bekins, 1999; Bekins et al., 1995). West-southwest–east-northeast trends of negative seismic amplitude anomalies along the décollement (DSDP/ODP transect) would correspond to kilometer-wide channels of focused fluid flow (Shipley et al., 1994). Quasi-lithostatic pore pressures were inferred at these locations from logging data while drilling (Moore et al., 1998; Moore, Klaus et al., 1998).

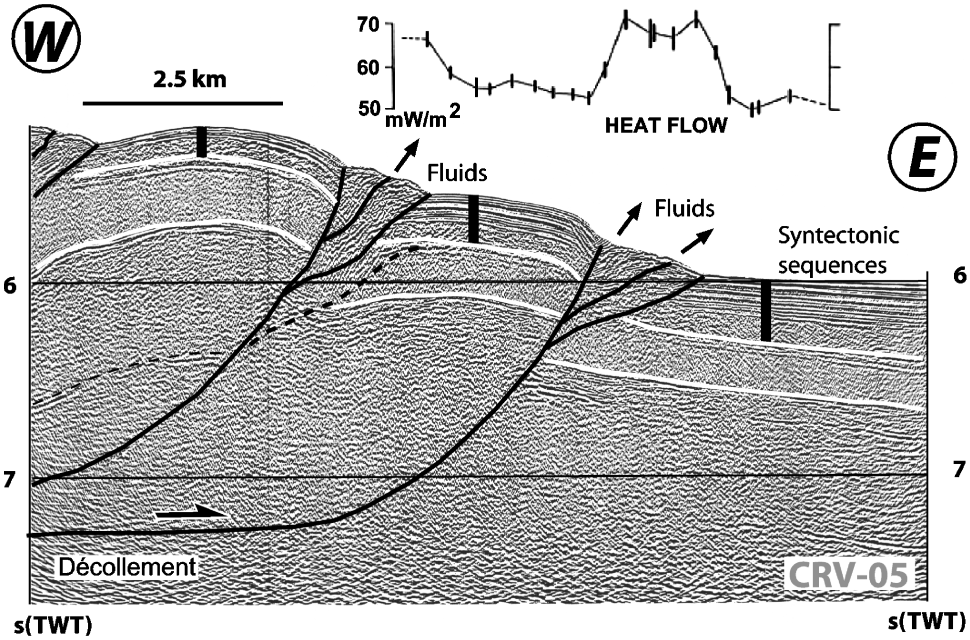



**Figure 8.** Section across the deformation front at 12°20′ N. In the section at the bottom, three sedimentary sequences above the décollement have been assigned tentatively to the Pleistocene, Pliocene, and Miocene, from top to bottom, respectively. The heat-flow measurements at the top are from Foucher et al., 1990.

Temperature measurements made during ODP Leg 110 (Fisher and Hounslow, 1990), as well as sea-floor heat-flow measurements farther south (Foucher et al., 1990; Langseth et al., 1990), show that local thermal anomalies are caused mainly by fluid flow along active faults. However, during ODP Leg 156, two holes were cased in order to provide long-term (1.5-yr) pressure and temperature monitoring. These two holes were located outside the zones of negative polarity along the décollement (cf. supra) where pore pressure and fluid flows are inferred to be the highest (Shipley et al., 1994). In these wells, despite evidence for fluid migration in fault planes mentioned above, the available temperature measurements from ODP Leg 156 wells exhibit almost linear thermal gradients and constant heat flow versus a depth of 82 $mW/m^2$ at site 948 (Foucher et al., 1997) and about 100 $mW/m^2$ at site 949 (Becker et al., 1997). This suggests that the fluid flows in the faults in these two areas have no significant implication in terms of convective heat transport (mainly conductive heat diffusion). The measured overpressures in ODP wells 948 and 949 are relatively moderate, which suggests relatively low fluid-flow rates and is consistent with temperature measurements. But this is not necessarily the case along the west-southwest–east-northeast trends of seismic amplitude anomalies along the décollement, which correspond possibly to kilometer-wide channels of inferred dilatancy and focused fluid flow (Shipley et al., 1994).

Farther south, in the frontal area at 12°20′N, fluid flows have been evidenced along the thrust faults from sea-bottom heat-flow measurements (70 + 5 $mW/m^2$ at the frontal thrust versus 55 + 5 $mW/m^2$ east of the front or within the frontal folds; Foucher et al., 1990). This indicates that the fluid flows are high enough here to generate thermal anomalies along the faults. Similar flows of warm fluids along active faults also have been inferred on the Cascadian active margin from the occurrence of thermally destabilized methane hydrates (Suess et al., 1999).

In the core of the prism, the rare available heat-flow measurements evidence a complex thermal regime, especially in the vicinity of mud volcanoes (cf. infra). Seeps of methane-rich water associated with chemicosynthetic, deep-fauna assemblages are well known in the southern Barbados Ridge (Jollivet et al., 1990), indicating intense fluid dynamics and gas generation in the prism.

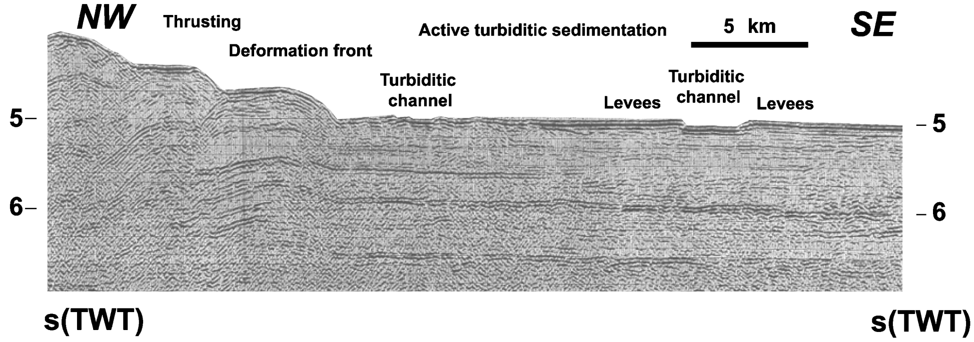



**Figure 9.** Section across the deformation front and the Atlantic abyssal plain at 10°40′ N.

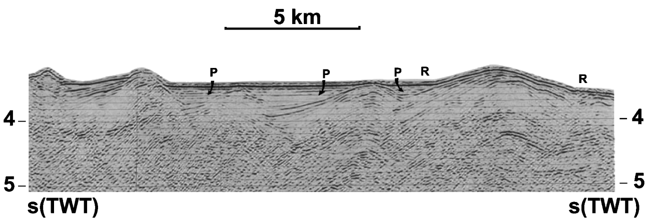



**Figure 10.** Piggyback basin at 10°40′ N. R = frontal ramps; P = piggyback basins.

## *Gas Hydrates*

The occurrence of gas hydrate-bearing sediment is expected in many places from the existence of bottom-simulating reflectors (BSR; Ferguson et al., 1993). BSRs may play an important role in the fluid dynamics of the prism. They are generally interpreted as resulting from a negative acoustic-impedance contrast between solid gas hydrates above, and free gas trapped below bottom-simulating reflectors (BSRs) (Henriet and Mienert, 1998; MacKay et al., 1994). The nature and origin of the gas is not known, but it is worth noting that, although these BSRs are widespread in the southern Barbados Ridge, they generally are not observed at the leading edge of the prism and in the abyssal plain. The BSR occurrence generally is associated with the mud volcano and shale diapir province, which suggests a probable deep origin for the gas (which probably is mostly thermogenic methane). Gas hydrates may thus be absent in these areas, or, alternatively, they may be too diffuse to give rise to significant changes in seismic velocity or density in the host sediments. Conversely, BSR occurrences are widespread in the mud volcanoes and shale diapir province. Had the gas been of shallow and bacterial origin, BSRs would have been present in different tectonic settings. We thus favor a deep and thermogenic origin.

These gases are circulating preferentially through the décollement and overlying ramp network and/or along sandy turbidite intervals. They may not be trappable in the shallow, frontal parts of the prism and foreland, because most of the incoming gas is quickly expelled from the prism along the very active frontal ramps, reaching the sea floor (as they used to do at first stages of frontal folds development), and thus preventing hydrate generation. Open (active) faults reaching the surface are well known to be efficient conduits for rapid oil and gas flows (Moretti, 1998; Sassen et al., 2001). Such a process could severely hamper the plausibility of conventional hydrocarbon trapping in the frontal anticlines. The question remains: has oil or gas been diverted from this flow into potential reservoirs intersected by faults in the past million years or less?

## *Shale Diapirism and Mud Volcanism*

The fluid-pressure regime produces an abundance and diversity of shale diapirs and mud volcanoes (Figure 11). Both are related to an intense fluid expulsion associated with subsurface sediment mobilization. Some mud volcanoes have been locally observed in the Atlantic abyssal plain as far as 23 km east of the deformation front. They are located in basement lows adjacent to oceanic fracture zones (Westbrook and Smith, 1983; Langseth et al., 1988; Henry et al., 1990, 1996; Henry, 2000; Lance et al., 1998; Summer and Westbrook, 2001). In the prism itself, mud volcanism is almost totally absent north of 15°N. In the southern part of the prism, the toe of the tectonic wedge usually is devoid of active mud volcanic activity, with few exceptions. The main province of active mud domes and volcanoes is found only in the core of the tectonic prism (Biju-Duval et al., 1982; Brown and Westbrook, 1987, 1988; Brown, 1990; Gonthier et al., 1994; Griboulard et al., 1991, 1997). Dense occurrences of mud volcanism and shale diapirism are observed along the southernmost transfer zones (Valery et al., 1985). Conversely, the western inner part of the tectonic wedge, an area characterized by extensional structures superimposed on thrust tectonics, is devoid of active mud volcanism. But fossil mud volcanic activity associated with hydrocarbon seepage is known on the island of Barbados (Joe's River Formation; Senn, 1940). The study of nannofossils in cores collected from the south of Barbados indicates that the mobilized sediments in shale domes and mud volcanoes are of Miocene-Pliocene age at the front of the mud volcanism zone (east), whereas Eocene-Pliocene intervals have been mobilized in the inner part (west) (Deville et al., 2003).

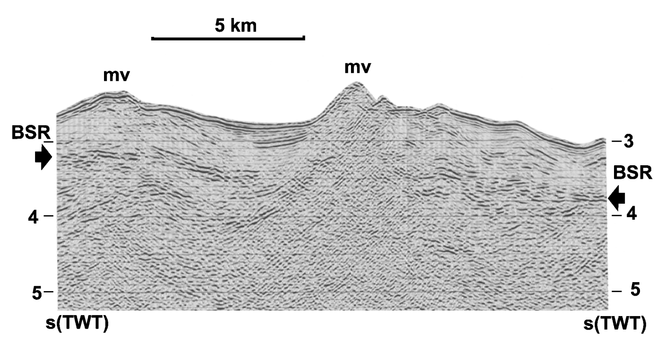



**Figure 11.** Shale diapir associated with bottom-simulating reflectors (BSRs) at 10°40′ N.

Similar mud volcanoes are well known in Trinidad and in many accretionary prisms of many active margins where thick sedimentary wedges are present (Higgins and Saunders, 1974). Mud volcanoes correspond to sedimentary eruption of liquefied material forming cones or mud pies and associated superficial mud flows, whereas shale domes correspond to piercing diapirs of mobilized plastic shales, which probably have never been liquefied. Mud volcanism and shale dome processes both are obviously related to the development of overpressure at depth, contributing to sediment mobilization by reducing strength in the overpressured layer (Hubbert and Rubey, 1959). In this tectonic context, overpressure generation is favored by the conjunction of fast sedimentation rates leading to compaction disequilibria (sedimentary loading) and compressive stress regimes inducing layer-parallel shortening and tectonic overloading. In addition, gas hydrate occurrence in these deep offshore areas is likely to reduce permeability in the superficial levels, thereby slowing fluid expulsion and favoring overpressuring. Moreover, the high deformation rates in accretionary prisms (especially compared to onshore mountain belts) probably have an important role in dynamic development of overpressure (typically nonstatic phenomena), and temperature induces the cracking of hydrocarbons in thick prisms, which is an additional factor for overpressure generation (Hedberg, 1980). Although gas expelled by the mud volcanoes in deep water is most likely to be dissolved, the occurrence of free gas bubbles, especially in the shallowest areas, also is likely to reduce the density of the sediments.

In the front of the tectonic wedge, hydraulic fracturing resulting from excess pore pressure tends to be subhorizontal (least principal stress $\sigma 3$ is vertical). Consequently, water will remain trapped longer in the sediments, but lateral hydraulic connectivity will be enhanced. If such is indeed the case, high pore pressures in the center of piggyback basins, if approaching the lithostatic load, will be transmitted laterally toward the anticline crest where sedimentary thickness and loads are smaller (Figure 13) (Yardley and Swarbrick, 2000). Consequently, pore pressure overcomes the vertical load and catastrophic upward mud extrusion can occur. Low pore-fluid pressure near the surface will favor lateral emplacement as sedimentary sills or chambers from the main vertical mud conduits toward the surrounding formations. This process is well imaged on some seismic sections (Biju-Duval et al. 1985) and has been proved by drilling in Trinidad (Higgins and Saunders, 1974).

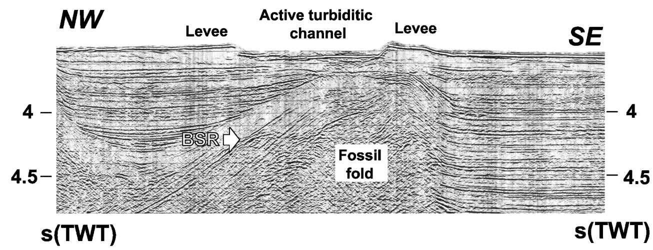



**Figure 12.** **Fossil fold sealed by an active turbiditic channel (see location in Figures 1 and 2).**

If mud volcanoes, as active faults breaching the sea floor, are numerous, they can significantly modify the flow path of water and hydrocarbon migration in the basin. They are very efficient vertical conduits allowing direct escape to the surface, as evidenced by methane-rich cold seeps associated with the development of numerous chemicosynthetic communities (Jollivet et al., 1990; Olu et al., 1997). The occurrence of numerous mud volcanoes certainly will increase the rate of excess pore-pressure dissipation, but it will also increase the rate of sedimentation with input of material from depth in adjacent piggyback basins, in addition to the current turbiditic influxes and hemipelagic sedimentation. This will, in turn, sustain the rate of burial and the generation of high pore pressure from possibly shallower stratigraphic intervals.

Thermal conductivity in shale diapirs, mud volcanoes, and, more generally, in overpressured formations, is usually low because of the high richness of water, which will be enhanced by the adiabatic expansion of thermogenic gas if gas bubbles are present. This significantly modifies the distribution of heat in the vicinity of and in the diapirs with respect to regional values (Bagirov and Lerche, 1999; MacDonald et al., 2000). The temperature regime is dependant on the thermal conductivity distribution, but it also will be highly influenced by the convective regime in mud volcanoes. When mud volcanoes and diapirs are widespread, they seriously disturb the thermal regime in the entire accretionary wedge. This is evidenced by available heat-flow measurements (values higher than 100 $mW/m^2$, as much as 230 $mW/m^2$ in the vicinity of mud volcanoes in a heat-flow background regime lower than 40 $mW/m^2$; see published measurements in Gonthier et al., 1994).

Overpressured zones are known to occur at 3000 m below sea floor in the vicinity of Trinidad, as

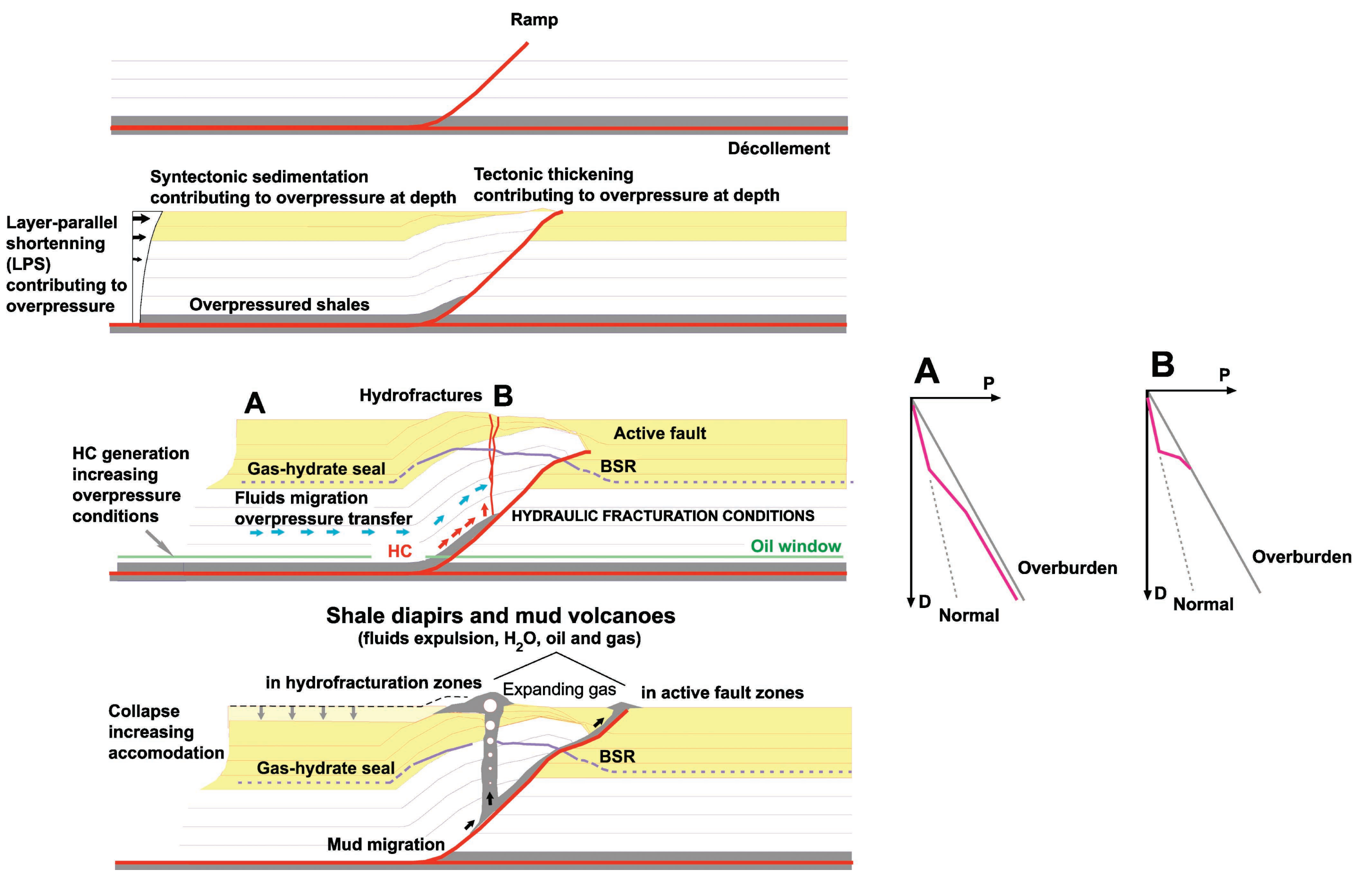



**Figure 13.** Proposed pressure distribution leading to the occurrence of mud volcanoes.

deduced from abnormally low seismic velocities (Heppard et al., 1998). The top of abnormal pressures is located in early Pliocene strata in onshore Trinidad, but becomes younger (up to the early Pleistocene) to the east in the offshore area (Wood, 2000). In addition to compaction disequilibria, processes of hydrocarbon (mainly gas) generation, aquathermal expansion of pore fluids, and clay dehydration also are inferred. A single restricted area of elevated pore pressure at, for instance, 3- to 5-km depth below sea floor will significantly diminish thermal conductivity and, as a consequence, the thermal regime (higher thermal gradient). As a result, oil and gas generation both in and below the overpressure zone are enhanced. On the other hand, hydrocarbon expulsion will be enhanced below the overpressure zone, but would be inhibited within it.

## DISCUSSION AND CONCLUSIONS

Lateral (north to south) changes along the Barbados Ridge deformation front are not progressive but are clearly segmented, with a major change in the thickness of the prism around 13° and 15°N, associated with segmentation of the incoming oceanic lithosphere.

In the northern part of the prism, because the tectonic wedge is relatively thin, organic matter is immature; as a consequence, any hydrocarbon accumulation could only occur assuming long-distance migration from the inner (several-kilometer-thick) part of the prism (i.e., from the west or from the south). The available data on fluid dynamics in this area suggest that such long-distance migration occurs along the sole thrust and frontal faults, but at moderate rates. Sandy potential reservoirs are present in the Eocene, but are totally absent in younger strata. In this northern part of the prism, two types of potential traps can be defined: those preserved below the décollement and resulting from past tectonic events in the abyssal plain (mainly traps associated with normal faulting structures), and those above the décollement and directly associated with the process of accretion. Eocene sandy reservoirs may be found in both of them, as the décollement, hosted initially in early Miocene claystones, deepens farther west to stratigraphic intervals older than Eocene.

To the south, the prism is much thicker, and organic matter is necessarily mature in the lower part of the tectonic wedge, close to the décollement and below. We have no indication whether or not source rocks are present in the Tertiary section. The possible location of Cretaceous formations in the prism or below the décollement remains very poorly constrained. This is a critical point, because the Aptian to Senonian strata generally include excellent, thick, and laterally extensive source-rock intervals in northeastern South America (such intervals correspond to the main source rocks for the origin of oil accumulations in Colombia, Venezuela, and Trinidad). Such levels can be expected also in deep-marine environments where the oceanic lithosphere is as old as the Cretaceous, especially in the southern part of the Barbados prism (west of magnetic anomaly 34, Figure 1).

Potential sandy reservoirs should be present in the late Miocene to Holocene Orinoco deep-sea fan, now partly incorporated in the accretionary prism. Present sediments include both hemipelagic clays and fine- to medium-grained, stacked, thin turbidites, but also individual sand layers as much as 2.3-m thick (Gonthier et al., 1994). Thus, potential reservoirs are mostly deep-marine sandstones originating from the Orinoco River delta and deep-canyons network, deposited in topographic lows on the sea floor. This includes the piggyback basins on the accretionary wedge and the foreland in front of the prism. Piggyback basins are very active (high rate of relative subsidence and tectonic tilting of bounding highs) but possibly short-lived features (less than 2 m.y.). Consequently, they trap a significant volume of turbiditic sandstones only during a short period of time. As these basins are of restricted lateral extent, successive sandy deposits will be vertically stacked. On the contrary, the happenstance of channelled deposits vertically stacked in the abyssal plain should be less, as the axis of the foredeep will migrate as rapidly as the front of deformation propagates. In Trinidad, middle Miocene analogs of such foredeep deposits are the Herrera and Karamat sandstones in the deep-marine Cipero and Lengua shales. They are now reservoirs for oil in complex south-verging anticlinal traps of middle Miocene to late Pliocene age (Penal/Barrackpore field; Dyer and Cosgrove, 1992).

The structures developing at the Barbados Ridge front, as well as at the frontal edge of most accretionary prisms, are very active features. The present front of the southern Barbados Ridge (30-km wide) may have been accreted only in the last 2 m.y. This implies a very rapid volumetric growth of the Barbados Ridge, which cannot simply be extrapolated to the past (accretion is believed to have been active at least since the Eocene). The acceleration of the volumetric growth of the prism could have started only in recent times. This explains the enormous volume of water that has been trapped in the newly accreted sediments in a short period of time, the resulting compaction disequilibria, and, as a consequence, the occurrence of widespread mud volcanoes and, possibly, large heat-flow anomalies.

The frontal folds are well imaged on seismic records. They are located usually in very deep water (in excess of 2000 m), and we can wonder if their mode of growth as previously described can enhance or perturb hydrocarbon trapping processes. These frontal folds develop above upward propagating ramps originating from a basal décollement. These ramps usually reach the sea floor in their earliest stage of development. Developing ramps and folds allow preservation of the accretionary ridge taper. Rates of displacement along fault planes and rates of uplift of anticlinal crests are in the range of a millimeter per year, while displacement over the décollement will be as much as 2 cm/year (plate-convergence rate). These rates will have to be compared with the rates of maturation, expulsion, migration, and entrapment of hydrocarbons in the short interval of the past 1 to 3 m.y. The relative rates and efficiency of fluid migration in faults and mud volcanoes, if present, and in sedimentary drains will be a critical issue for assessing trapping possibilities.

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