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Structure and evolution of the Demerara Plateau, offshore French Guiana: Rifting, tectonic inversion and post-rift tilting at transform–divergent margins intersection

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ABSTRACT

We present the structure and evolution of the eastern part of the Demerara plateau, offshore French Guiana, from the analysis of geophysical data collected during GUYAPLAC cruise. This area is located at the intersection of a transform segment and a divergent segment of a continental margin related to the Early Cretaceous opening of the Equatorial Atlantic. The main structures are NNE–SSW to NNW–SSE trending normal faults on the eastern edge of the plateau, and WNW–ESE to NW–SE trending acoustic basement ridges on its northern edge. When replaced in their Albian position, these structures appear to be parallel to the coeval oceanic accretion axis and transform faults, respectively. The most striking structures are related to a post-rift but syn-transform tectonic inversion, producing E–W to WNW–ESE trending folds, sealed by a regional unconformity. This shortening cannot be related to ridge push, but is probably related to a plate kinematic change 105 My ago, that modified the deformation in the vicinity of the transform fault. Late post-rift evolution also includes a significant Tertiary oceanward tilt of the edge of the Demerara plateau. The driving mechanism of this late tilt is unclear, but may be related to a lithospheric flexure resulting from the loading of the abyssal plain by the Orinoco and Amazon deep-sea fans.

1. Introduction

Formation and evolution of passive continental margins can be explained at regional scale by simple models of lithospheric thinning (McKenzie, 1978; Wernicke, 1985). However, these models do not encompass the numerous segments where the strike of the continental margin is not perpendicular to the plate divergence, as in oblique rifting, transfer zones, or along transform faults. Moreover, the post-rift evolution of numerous passive continental margins is not characterized only by a decreasing subsidence, but often show tectonic inversions (e.g. the Norwegian margin, Vagnes et al., 1998; US West Coast, Withjack et al., 1995; Angola, Hudec and Jackson, 2002).

We present in this paper a new data set of seismic reflection lines on the eastern part of the Demerara Plateau, offshore French Guiana (Fig. 1). The data were collected during GUYAPLAC cruise, performed in 2003 in the framework of the French program EXTRAPLAC to argue on the extension of the Exclusive Economic Zone. The main aims of this study are to discuss the role of obliquity in margin structure and evolution, and to describe the structures associated with the tectonic inversion experienced by this margin.

2. Geodynamic setting

The Demerara Plateau represents a bathymetric extension of the continental shelf offshore Guiana. Water depth increases progressively from South to North, with a steep slope at the northern transition with the abyssal plain, 180 miles (340 km) from the coast line, and a more gentle slope towards NE (Fig. 1).

The Demerara Plateau and its conjugated margin, the Guinean Plateau, are located at the intersection between two domains of the Atlantic Ocean (Fig. 2):

– The Central Atlantic Ocean, that opens since Jurassic times (Klitgord and Schouten, 1986). A very short (only few million years) but widespread magmatic event occurred 200 My ago, and created what has been referred as the Central Atlantic Magmatic Province (CAMP) (Marzoli et al., 1999). This magmatic event came before the oceanic accretion that started during Early Jurassic times in the northern part of the Central Atlantic (Sinemurian; Sahabi et al., 2004), and Middle Jurassic times in its southern part (Bathonian; Klitgord and Schouten, 1986). The Guinean

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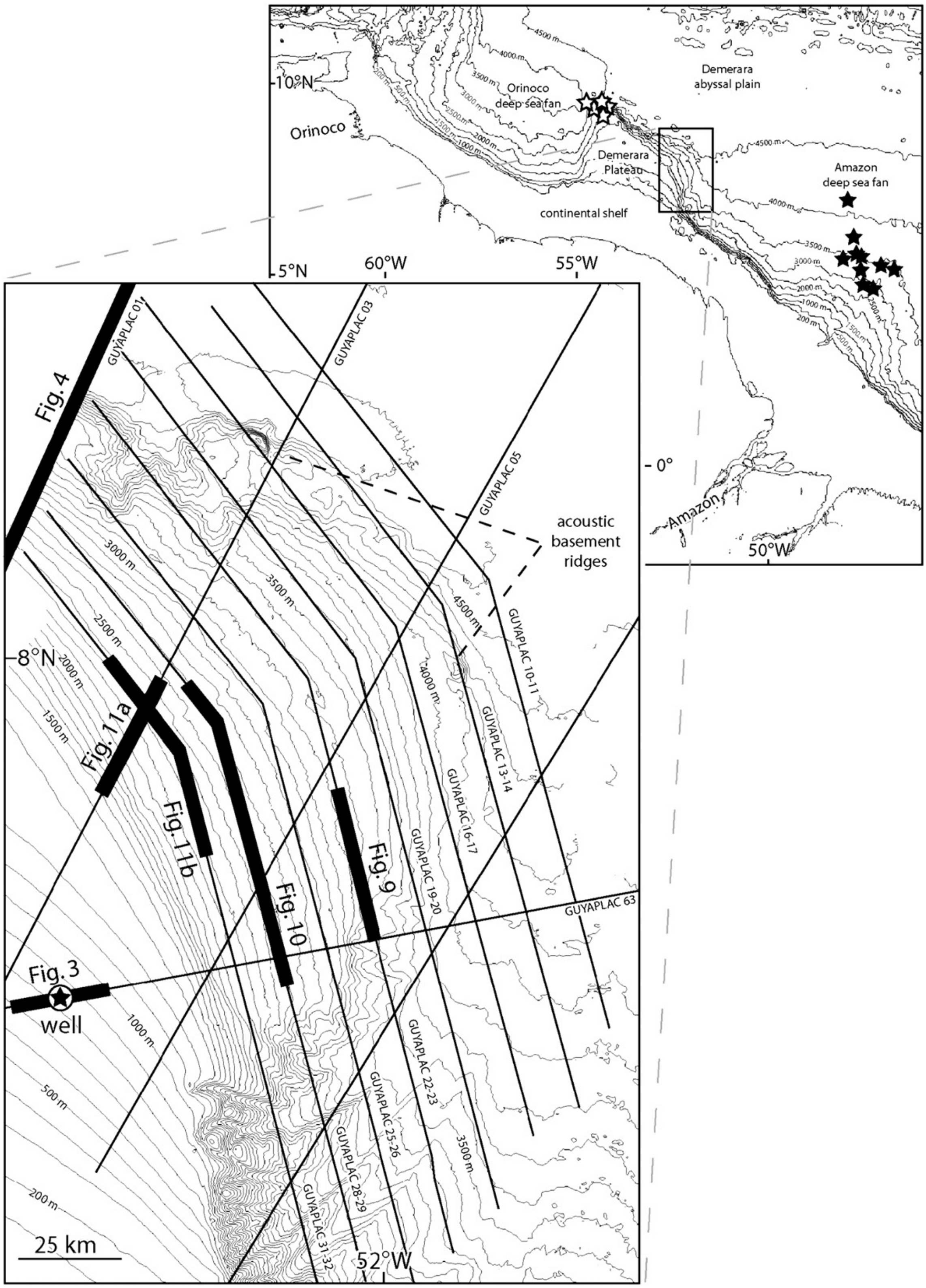

Fig. 1. Location of GUYAPLAC seismic lines in the eastern Demerara plateau. Bathymetry from GUYAPLAC cruise and SHOM data (equidistance 100 m). The upper box locates the Demerara plateau and the Orinoco and Amazon deep sea fans on the northern margin of South America. Stars indicate wells: open stars are sites from Leg ODP 207 (Shipboard Scientific Party, 2004), black stars are Leg ODP 155 sites (Flood et al., 1995), the circled black star locates industrial well G2 described by Gouyet (1988).

and Demerara Plateau were located East of the southernmost termination of the Central Atlantic Ocean, where the divergent plate boundary connected through a transform fault with another spreading axis located to the West in the Caribbean area. South and Southeast of this southern termination, there are only limited indications of Jurassic subsidence and continental sedimentation (Basile et al., 2005).

– The Equatorial Atlantic Ocean opened East of the Demerara Plateau during Early Cretaceous times (Fig. 2). This oceanic domain is characterized by the Chain, Romanche and Saint Paul transform

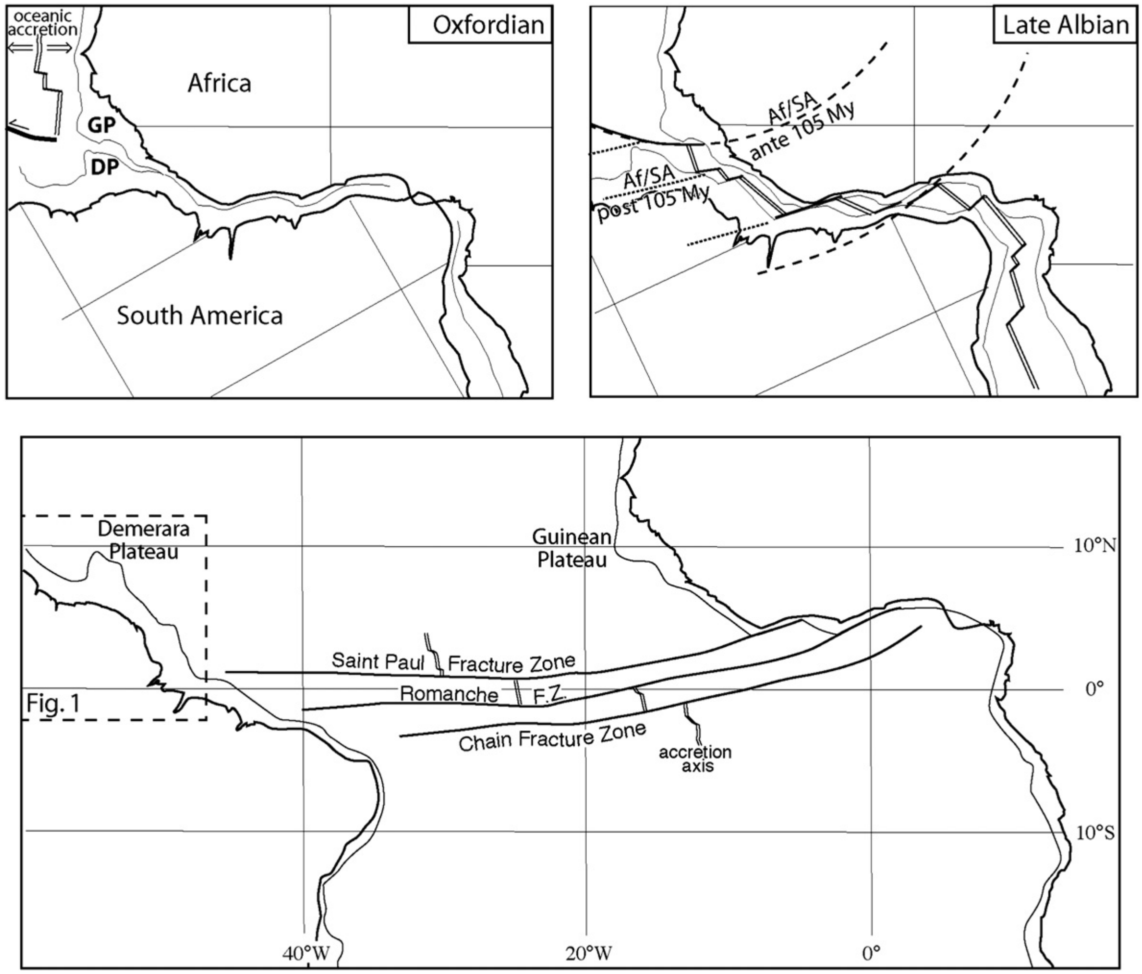

Fig. 2. Present day and reconstructions of Africa–South America positions before (Oxfordian: oceanic spreading in the Central Atlantic) and after the rifting in the Equatorial Atlantic (Late Albian (105 Ma): oceanic spreading in the South and Equatorial Atlantic). Africa is supposed fixed. The present day coastline and 2 km isobath are drawn. DP and GP are Demerara plateau and Guinea plateau, respectively. The dotted curves in the Late Albian reconstruction are the small circles parallel to the relative motion of South America by reference to Africa, and show the plate kinematic reorganization occurring at 105 Ma.
Modified from Basile et al. (2005) and Campan (1995).

faults that offset short spreading axis by several hundreds of kilometers. It resulted from the northward propagation of the South Atlantic Ocean spreading axis, and its connection with the Central Atlantic spreading axis. Rifting in the Equatorial Atlantic occurred during Early Barremian to Aptian (Basile et al., 2005), and the first oceanic crust probably accreted during Late Aptian (Basile et al., 1998). Campan (1995) proposed a kinematic reconstruction with a first Africa/South America displacement parallel to the WNW–ESE Guinea Fracture Zone, and a kinematic change at 105 Ma (Albian) that induced the formation of numerous E–W fracture zones North of Demerara Plateau (Fig. 2). This change was associated with the northward displacement of the triple junction between Africa, North America and South America plates (Müller and Smith, 1993).

As a consequence, the western, northern and eastern borders of the Demerara plateau represent Jurassic divergent, Early Cretaceous transform, and Early Cretaceous divergent margins, respectively. The area investigated in this paper covers the eastern Early Cretaceous divergent margin, and its connection with the northern transform margin.

3. Previous studies

Only few studies were published on the deep structure of the Demerara Plateau. Seismic reflection lines shot and exploration wells drilled by the oil industry during the seventies and early eighties were compiled by Gouyet (1988). More recently, ODP Leg 207 drilled five sites in the northwestern edge of the Plateau (Shipboard Scientific Party, 2004) (Fig. 1). Finally, Greenroyd et al. (2008) presented a crustal-scale study based on wide-angle seismic data. In the northeastern part of the Demerara Plateau, the Moho is 23 to 26 km deep, and rises rapidly below the northern edge of the plateau, at the continent–ocean transition. North and East (Greenroyd et al., 2007) of the Demerara Plateau, the oceanic crust appears to be unusually thin (3.3 to 5.7 km thick).

The oldest sediments recovered in this area are Late Jurassic sandstones, that were dredged at the foot of the northern slope of the Demerara Plateau (Fox et al., 1972). The oldest drilled sediments are Early Cretaceous in age (Valanginian to Aptian limestones), deposited in continental to proximal marine environments (Gouyet, 1988). The sedimentation became more clastic during Barremian–Aptian times, and a 120–125 My-old basaltic lava flow has been drilled in the eastern part of the plateau (Gouyet, 1988).

Along the northern border of the Demerara Plateau, these Early Cretaceous sediments were deformed by SW–NE-trending en échelon folds associated to strike–slip faulting (flower structures) (Gouyet, 1988). These structures are compatible with right-lateral transpression. They were peneplained by a regional unconformity, at least partly in a sub-aerial environment, and covered by late Albian detrital sediments (Gouyet, 1988; ODP Leg 207 Shipboard Scientific Party, 2004). In the eastern part of the plateau, normal faults affected the Early Cretaceous sediments, and also slightly offset the Late Albian regional unconformity (Gouyet, 1988).

Above the Albian unconformity, the Late Cretaceous and Cenozoic sediments are currently less than 1.5 km-thick. Cenomanian–Turonian black shales deposited first in hemipelagic environments. They represent a source rock for oil production in this area, and ODP Leg 207 Shipboard Scientific Party (2004) showed that methanogenesis is still active in those sediments. The more recent sediments are mainly clayey, with some silts and pelagic limestones (Campanian–Maastrichtian, Paleocene, Early Eocene and Middle Eocene chalks: ODP Leg 207 Shipboard Scientific Party, 2004). Their thickness has been restricted by bottom currents (deep western boundary current steered to the SE: Johns et al., 1990) and by slope instabilities (Loncke et al., 2009; ODP Leg 207 Shipboard Scientific Party, 2004). ODP Leg 207 Shipboard Scientific Party (2004) underlined a prominent erosional surface across the entire northwestern plateau. They proposed a Late Oligocene–Early Miocene age for this surface, but this age is not supported by referenced publication nor ODP Leg 207 data.

It is noteworthy that the main regional detrital sources represented by the Orinoco and the Amazon rivers are far from the Demerara Plateau, but built deep sea fans and turbiditic bodies at its northern (Gonthier et al., 2002) and eastern feet (Damuth et al., 1983, 1988; Loncke et al., 2009) (Fig. 1).

4. Data

The seismic acquisition system comprised a 24-channel streamer with a group spacing of 12.5 m, and two Generator Injector air guns with a total volume of 300 in.^{3}. The seismic data were stacked and migrated using water velocity. This provided good seismic images when subseafloor traveltimes were less than 2.5–3 s. The dominant half-period of the seismic wavelet is 3.25 ms which corresponds to the vertical resolution of the seismic profiles (5 to 8 m at sedimentary seismic velocities). The horizontal resolution is 25 m corresponding to the size of the bins while stacking the data.

The total magnetic field was measured using a SeaSPY marine magnetometer, towed 300 m behind the ship. After visual inspection and removal of occasional spikes, the magnetic anomaly was calculated using the IGRF model. Data were subsequently gridded using the GMT minimum curvature method (Wessel and Smith, 1991), using a gridcell size of 5×5 km.

Whereas some limits or artifacts such as side reflections are inherent to 2D seismic acquisition, the most important limitations in the use of the seismic data set comes from the orientation and spacing of the seismic lines, which were designed to allow the best and faster bathymetric coverage. This results in a wide spacing (circa 12 km) between seismic lines roughly parallel to depth contour lines. As a consequence, the seismic lines are frequently oblique with respect to the strike of geological structures, which are not imaged in perpendicular cross sections. Mapping of geological structures is based on interpolations of similar structures cut by several seismic lines, but the confidence in these interpolations obviously depends on the line spacing. The continuity of the structures from one line to the adjacent may be questioned, especially where a given structure is cut by only two lines, and where it orientation changes.

5. Stratigraphy

5.1. Seismic units correlated with well G2 (Gouyet, 1988)

We established the stratigraphy of the studied seismic lines from their intersection with a seismic line presented and correlated by Gouyet (1988) with the well G2 (Fig. 3). From top to bottom, we define the following units:

– The uppermost unit A is poorly reflective, but reflections are continuous laterally. This unit is Neogene in age, and its base fits with the base of Miocene.
– Unit B is characterized by diffractions in the upper part, transparent seismic facies in the middle part, and a reflective lower part. This unit is Paleogene in age, with the base of Eocene located at the boundary between the transparent and reflective facies, and the base of Paleocene at the base of the reflective lower part.
– Unit C is poorly reflective, especially in its lower part where the transparent seismic facies corresponds to the Cenomanian–Turonian black shales. This unit is upper Cretaceous in age.
– The top of unit D is defined by the first reflection below the transparent part of unit C. Unit D is reflective, and is dated Middle to Upper Albian.

Below unit D, we define additional sedimentary units, named E and F, solely based on their seismic character (Section 5.2), that cannot be correlated to the drill hole section because of the poor resolution on the seismic lines at depth in the area of the well G2. Furthermore, these units do not appear to be laterally continuous on the seismic lines.

Also note that this terminology differs from the one used by ODP Leg 207 Shipboard Scientific Party (2004) to describe the seismic lines on the northwestern edge of the Plateau.

5.2. Lateral variations

The seismic units defined near the well G2 present lateral variations of seismic facies, and the lack of lateral continuity makes it difficult to tie together the various facies.

North and Northeast of the well G2, the uppermost seismic unit has a transparent to slightly chaotic seismic facies, lying on sliding surfaces that truncate the lowermost units and rise upslope to the sea floor (Fig. 4). This uppermost unit is postulated to be a lateral equivalent to unit A and probably the top of unit B, but the sliding surfaces does not allow to observe the lateral continuity with the area where these units are clearly defined. In the same northeastern area, the uppermost unit lies on a reflective unit that groups together unit C and the bottom of unit B. The change from transparent to

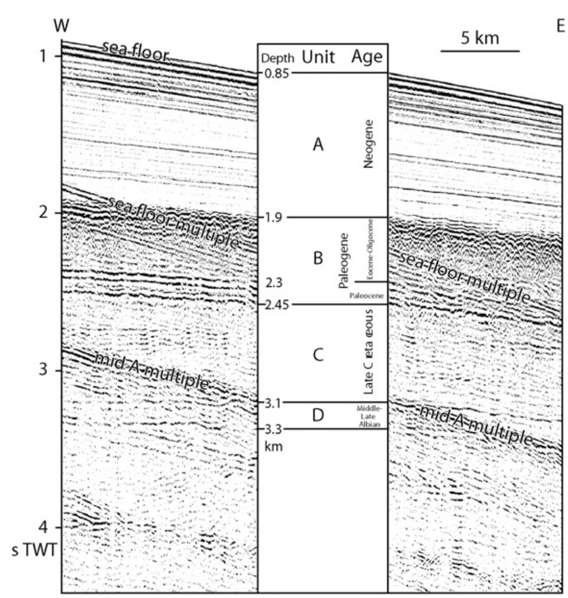

Fig. 3. Seismic units (A to D) from seismic line GUYAPLAC 63, and stratigraphy and approximate depths from well G2 (Gouyet, 1988). The seismic line is located in Figs. 1 and 5. Depth in second double way (sdw). Vertical exaggeration 13 in water.

reflective seismic facies in the lower part of unit C probably reflects a lateral change in Cenomanian–Turonian black shales sedimentary facies. Finally, unit D is laterally continuous, without noticeable changes in seismic facies.

According to the lateral variations of the seismic units, we can define three area in the eastern Demerara plateau (Figs. 5 and 6).

- In the investigated area, the upper plateau is characterized by sub-horizontal units B to D (Fig. 4). The bathymetric slope is related to the northeastward thinning of the more recent unit A. In the studied set of seismic lines, the penetration does not allow to observe the units below unit D, but according to the few seismic lines shown by Gouyet (1988), the underlying sediments are thick and poorly deformed.
- East and Northeast of the upper plateau, the intermediate plateau (Figs. 5 and 6) is characterized by the oceanward tilting of units B to D (Fig. 7a). Below unit D, two additional sedimentary units can be defined from the seismic lines: the poorly reflective unit E overlies the highly reflective unit F (Fig. 4). These two units were deformed and eroded by the regional unconformity onlapped by unit D (Fig. 7a). Above unit D, the seismic units are poorly continuous laterally.
- The lower plateau is defined only along the North and Northeast edges of the studied area. Here appear narrow ridges of acoustic basement, covered by sub-horizontal sediments of units D to A. Some of these ridges outcrops from the sedimentary cover, and appears to be NW–SE trending in the bathymetry (Fig. 1). Below unit D, a transparent acoustic facies infills the trenches between the ridges. The relations between this transparent unit and units E and F cannot be observed on the available seismic lines. The lower plateau is located in the area where the most important thinning of the crust occurs. There, the Moho uplifts from 21 to 14 km deep (Greenroyd et al., 2008).
- Southeast of the lower plateau the sedimentary cover thickens at the foot of the continental lower slope, and the acoustic basement cannot be observed on the available seismic lines. Seismic units E and F can be postulated in the western border of this East Guiana deep basin (Figs. 5 and 6), where strong negative gravimetric anomalies are located (Fig. 8a) suggesting a thick sedimentary section. This area may have a continental basement deeper but similar to the one of the Demerara Plateau. But there is no explanation for the strong positive magnetic anomaly located in the same area (Fig. 8b). Eastward the continent–ocean boundary may strike NNE–SSW, and marks the edge of a smooth oceanic crust which appears homogeneous at large wavelengths on gravimetric anomaly map (Fig. 8a).

North of the Demerara Plateau, the abyssal plain is clearly separated from the plateau by the northern escarpment, underlined by a strong gravimetric anomaly (Fig. 8a) and the limit between positive and negative magnetic anomalies (Fig. 8b). Below the abyssal plain, seismic lines display usual acoustic characteristics of an oceanic crust with basement relief covered by horizontal sediments. East of the plateau, the abyssal plain as it can be defined from the bathymetry (deeper than 4500 m below sea level) covers both a typical oceanic crust northward and a higher area southward which shares the same basement characters (NW–SE trending ridges and negative magnetic anomalies) than the lower plateau (Figs. 5, 6 and 8).

6. Tectonic structures

6.1. Cretaceous normal faults

The oldest structures observed in GUYAPLAC seismic lines are NNE–SSW to NNW–SSE trending normal faults located along the

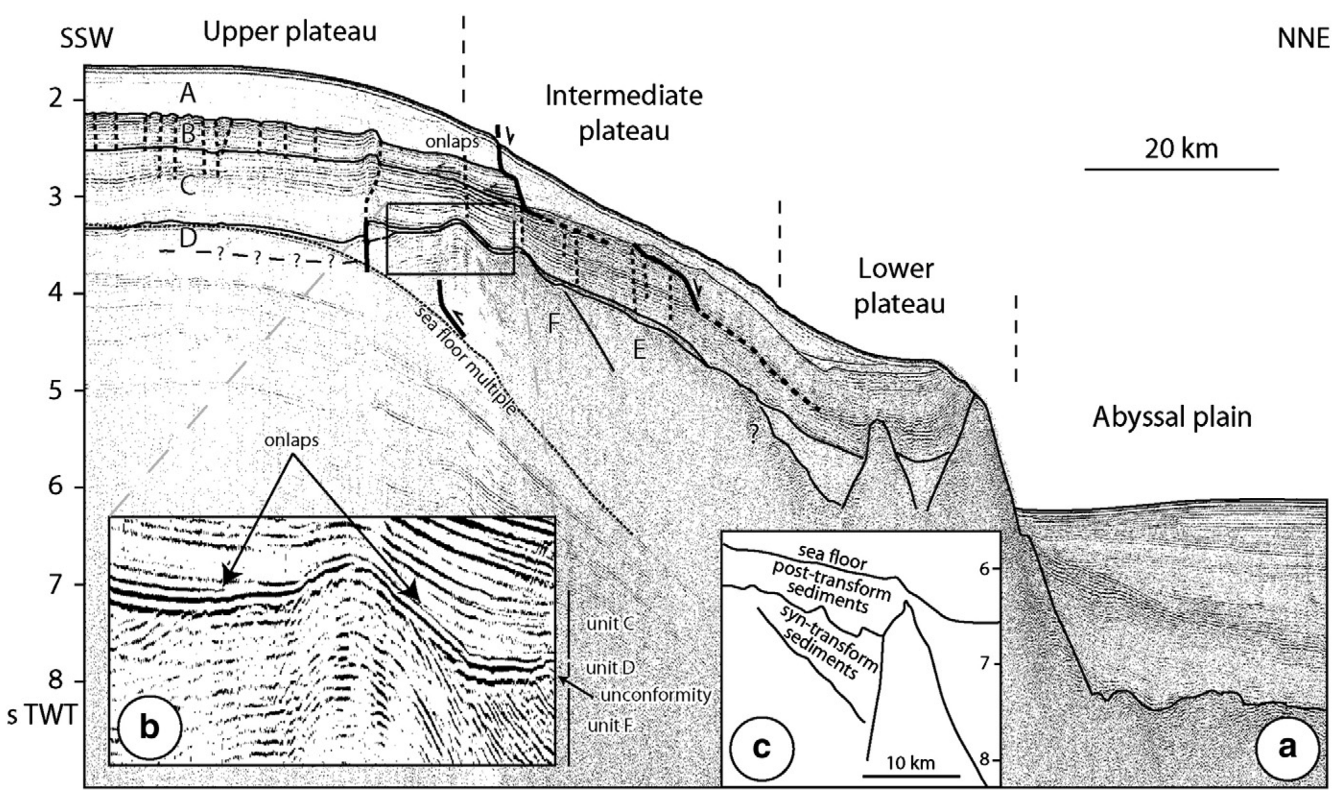

Fig. 4. Seismic line GUYAPLAC 01 (a, with a local zoom: b), located in Figs. 1 and 5 (the southwesternmost part of the line extends out of the western edge of Figs. 1 and 5). 4c: for comparison, line drawing (same scale as Fig. 4a) of a seismic section across the transform margin off Côte d'Ivoire (modified from Basile et al., 1996). A to F are seismic units (see text for details). Thick lines are faults, dotted thick lines are polygonal fault network. Faults are not directly observed in the acoustic basement, but drawn from offsets and tilts of the overlying seismic reflectors. Depth in second double way (sdw). Vertical exaggeration 13 in water.

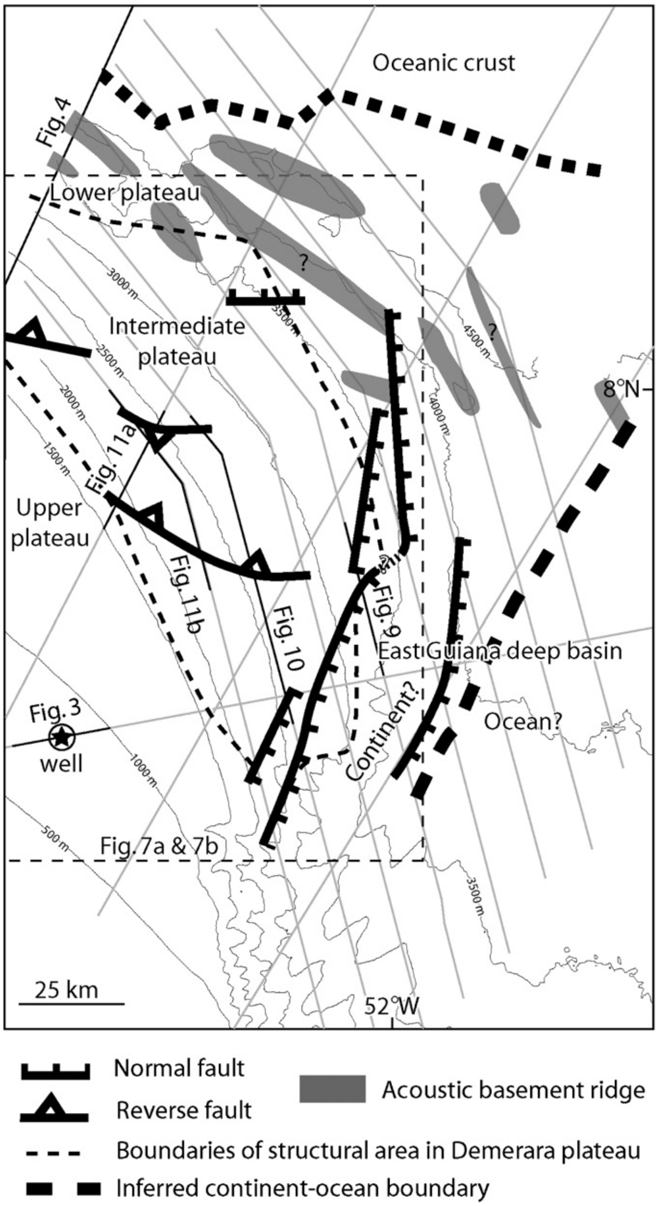

Fig. 5. Structural map of the eastern Demerara plateau. GUYAPLAC seismic lines and bathymetry (equidistance 500 m) are reported. The confidence in the interpolations between seismic lines depends obviously on their wide spacing. The thick dotted lines tentatively limit the oceanic crust northward and eastward from their depths and geophysical properties (see Fig. 8 and text for comments). The thin dotted lines limit the various structural parts of the Demerara plateau. The bathymetric plateau extends down to 4500 m northward, and 4000 m eastward, but the basement characters are similar in the lower plateau and eastward below the abyssal plain. Faults are not directly observed in the acoustic basement, but drawn from offsets and tilts of the overlying seismic reflectors. The map traces of the faults correspond to their intersection with the top of the acoustic basement. Most of fault slip occurred before Unit D deposition. At the given map scale, anticline axes are too close to the thrusts to be shown. They are parallel to the thrusts and located on their hanging wall (Fig. 6).

eastern border of the Demerara Plateau (Figs. 5, 6, 9 and 10). These faults deepen and probably thicken the seismic units E (Fig. 7b) and F East of the plateau, although the restricted seismic penetration does not allow to observe their bases in this area. The dip slip throws increase from NNE to SSW. The top of unit F is clearly offset by the normal faults, and unit E fills in the half-grabens with a slight tilting, thickening towards the border faults, and sometimes onlapping unit F. An angular unconformity overlaid by unit D seals this tilting, but unit D is itself slightly offset by the normal faults (Figs. 9 and 7a). Faulting seems to end inside unit D. This suggests a main extensional phase prior to the Late Albian unit D, but reactivation of the normal faults during Late Albian. It is noteworthy that a bathymetric step persists up to the present day, as the eastern bathymetric limit of the plateau overlies the westernmost normal fault (Figs. 1 and 5).

Northwestward of this area, offsets as well as fan-shaped geometries in units E and F suggest other tilted blocks (e.g. Fig. 10), but the subsequent deformations (see below) and the strike and spacing of the seismic lines do not allow to map these structures.

6.2. Cretaceous shortening

In the southern and southwestern parts of the intermediate plateau, seismic units E and F are folded in asymmetric anticlines (Fig. 11). These folds seem to be related to underlying thrusts. Although the seismic lines are widely spaced, three thrusts and associated anticlines can be tentatively mapped from the available data set (Figs. 5, 6 and 7b). Fold axis trend E–W to WNW–ESE and appear parallel to the northernmost basement ridges. Both northward and southward dipping of fold axial surfaces are observed. The three anticline axis seem to display right-lateral en échelon setting in a NW–SE trending band. Their western terminations are not observed on the available lines, but the decreasing offset of the southernmost thrust from Fig. 11b to a (from East to West) probably indicate that its western termination is closed to the mapped one. A similar decreasing offset is observed towards the eastern termination (compare Figs. 11b and 10). These folds can be interpreted as formed during NNE–SSW shortening, either by reactivation of older en échelon structures, or related to NW–SE right-lateral strike–slip displacement, or both.

Unit F does not show any change of thickness that can be related to this folding, while unit E commonly thins toward the anticline hinges (Figs. 10 and 11). This thinning appears to be related to erosion prior to unit D deposition. However, some onlaps toward the anticline hinges can be observed inside unit E (Figs. 10 and 11). This suggests that folding may have started during unit E sedimentation, with the main shortening phase occurring between deposition of units E and D.

In the northern part of the intermediate plateau, both units E and F are tilted northward (Fig. 7b). In this area, a flat erosional surface cuts units E and F (Figs. 4, 6, 7a, 10, and 11). The overlying unit D lies conformably on this erosional surface suggesting it was a flat and horizontal surface at the time of deposition. The same erosional surface cuts the folded units E and F on the three anticline mapped in Fig. 5, but not on the anticline shown in Fig. 4a. There this surface is slightly irregular, and onlapped by unit D (Fig. 11). In the southernmost intermediate plateau as in the upper plateau, unit D conformably lies on unit E, without any trace of erosion in between. This geometry suggests that unit D sealed a contrasted morphology, with a peneplained flat area in the northern intermediate plateau, residual hills along the anticlines and a basin, probably below sea level, in the southernmost intermediate plateau and in the upper plateau.

6.3. Upper Cretaceous reactivation

Upper Cretaceous tectonic activity is observed only locally: at the southern border of the plateau, normal faults are reactivated during the deposition of the Late Albian unit D (Fig. 9, c.f. Section 6.1); at the northern border of the upper plateau, an anticline fold is reactivated during the deposition of Cenomanian–Turonian unit C, that onlaps each limb (Fig. 4). Southward of this fold, a coeval subvertical fault divides a more subsiding southward area, where seismically transparent black shales probably deposited, from a less subsiding northward area where the coeval sediments exhibit a more reflective seismic facies. The available seismic lines does not allow to orient this structure and to image it in the deeper units.

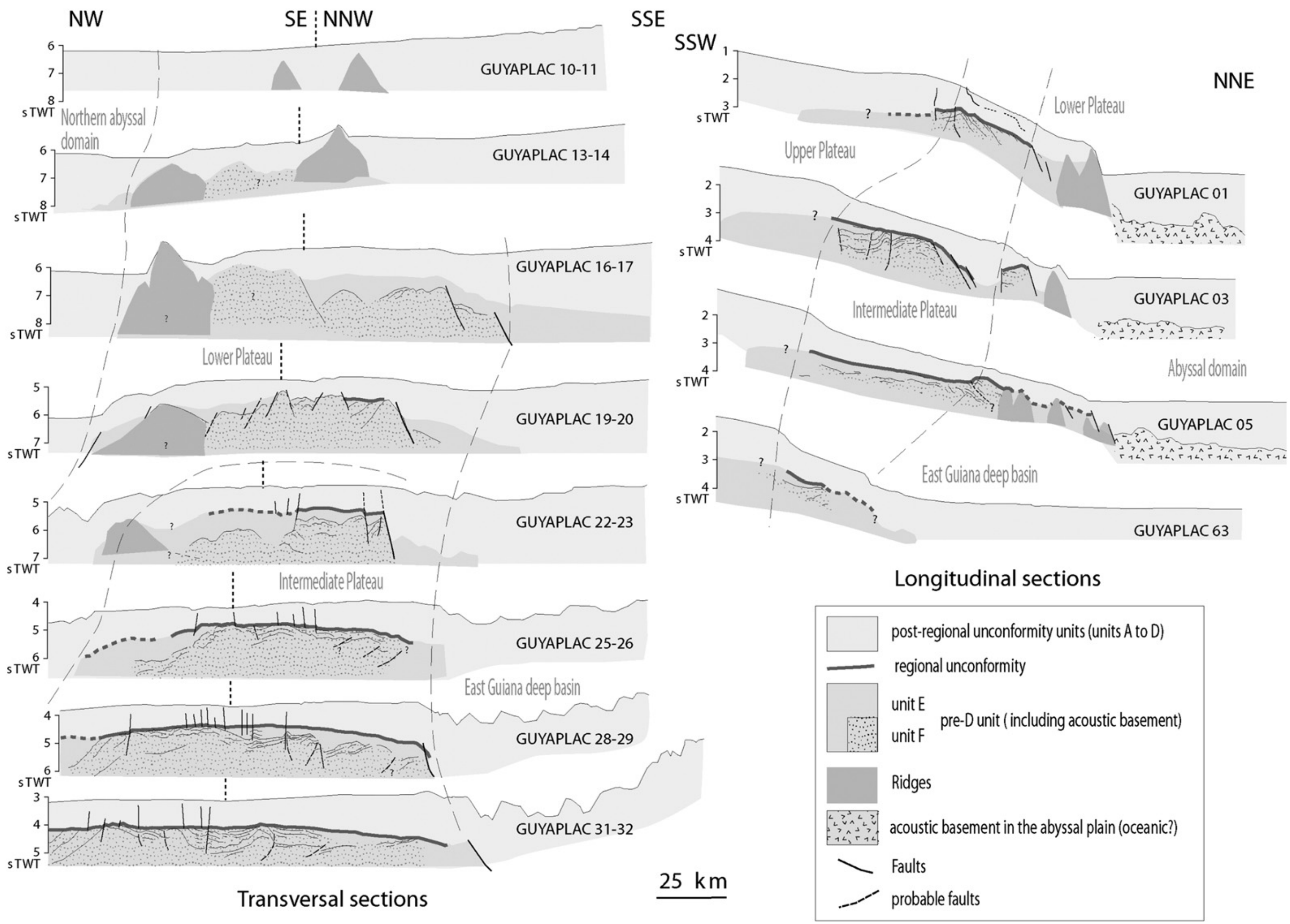

Fig. 6. Line drawings of GUYAPLAC seismic sections, located in Fig. 1. The vertical dotted bar indicate change of line strike on transversal sections. Same vertical and horizontal scale for all sections. Vertical exaggeration 13 in water.

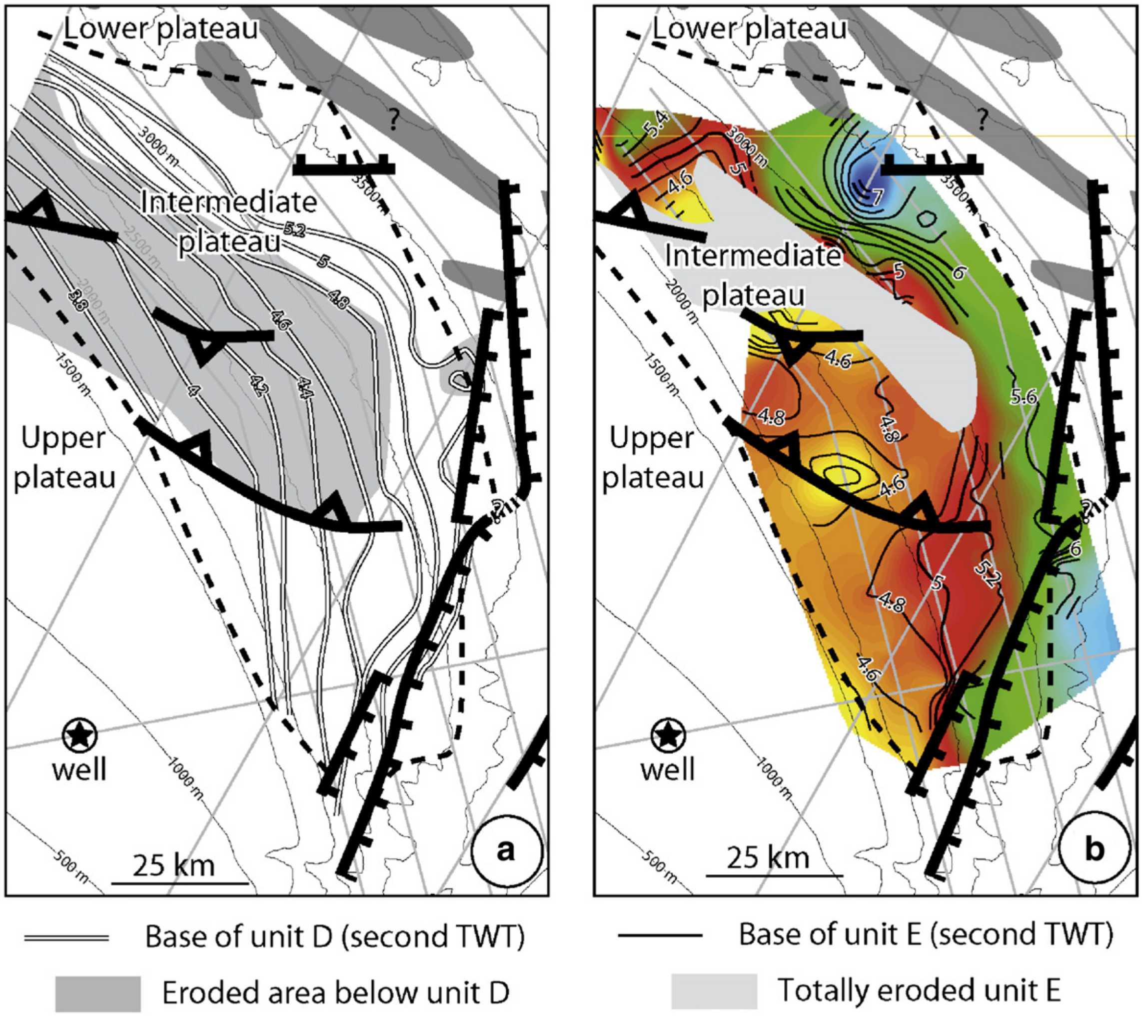

Fig. 7. Depths of the base of unit D (7a) and E (7b) in the intermediate plateau. Structures and bathymetry as in Fig. 5. Due to limited seismic penetration, the base of unit E cannot be observed in the deepest parts, and the color interpolation is only weakly constrained.

6.4. Cenozoic oceanward tilt and gravity-driven slides

As stated above, units B to D are tilted oceanward (towards ENE) in the intermediate plateau (Figs. 6 and 7a). Units A and B are poorly continuous laterally, and appear to be cut by numerous oceanward slides with basal surfaces parallel at depth to the reflectors of unit C (Figs. 4 and 11a). The boundary between the upper and the intermediate plateau is marked by a bathymetric scarp, 100 to 300 m-high (Fig. 1), that represents the uppermost limit of the sliding area. This scarp indicates that gravity-driven sliding is still active.

Unit D deposited during Late Albian times in a roughly flat area (c.f. Section 6.2), close to sea level (Gouyet, 1988; ODP Leg 207 Shipboard Scientific Party, 2004). It is important to note that unit D does not onlap its oceanward dipping base in the intermediate plateau, but lies parallel with it, without any significant change in thickness. There is a slight thickening of unit D in the upper plateau and in the northern and southern parts of the intermediate plateau, on each side of the area where the deformed units E and F were more deeply eroded (e.g. Fig. 4). These changes of thickness of unit D can be simply related to differential compaction of the underlying sediments (e.g. unit E in Figs. 10 and 11). Unit C displays the same characteristics as unit D, i.e. no onlap nor significant change in thickness in the intermediate plateau, and a slight thickening where the underlying sedimentary section is thicker. These patterns suggest that units C and D deposited in a roughly horizontal area, and predate the oceanward tilting of the Demerara plateau, that consequently should be Cenozoic in age. This oceanward tilting may be coeval with the deposition of few onlaps in lower part of unit B along the Upper/Intermediate plateau boundary (Fig. 4).

A peculiar undulated layer has been observed on both eastern lower plateau and in the East Guiana deep basin, in the uppermost seismic unit (unit A and top of unit B). This layer is mostly isopach, with an undulated shape at a pluri-kilometer scale, and a prominent erosional base. Undulations have the same wavelength than the present day bathymetric gullies. A layer exhibiting the same features has also been observed along the western edge of the Demerara Plateau, where it has been interpreted as a regional erosional surface with submarine channels developed during the Late Oligocene–Early Miocene (ODP Leg 207 Shipboard Scientific Party, 2004). Assuming that the erosive base of the undulated layer can represent the gullies formed when the tilting of the intermediate plateau and the gravity-driven slides started, and if this undulated layer could have the same Late Oligocene–Early Miocene age as in the western Demerara plateau, this may indicate that tilting and associated collapse of the superficial sedimentary cover began close to the Paleogene–Neogene boundary.

6.5. Acoustic basement ridges

Several ridges are located in the lower plateau and eastward below the abyssal plain, at the transition between the continental

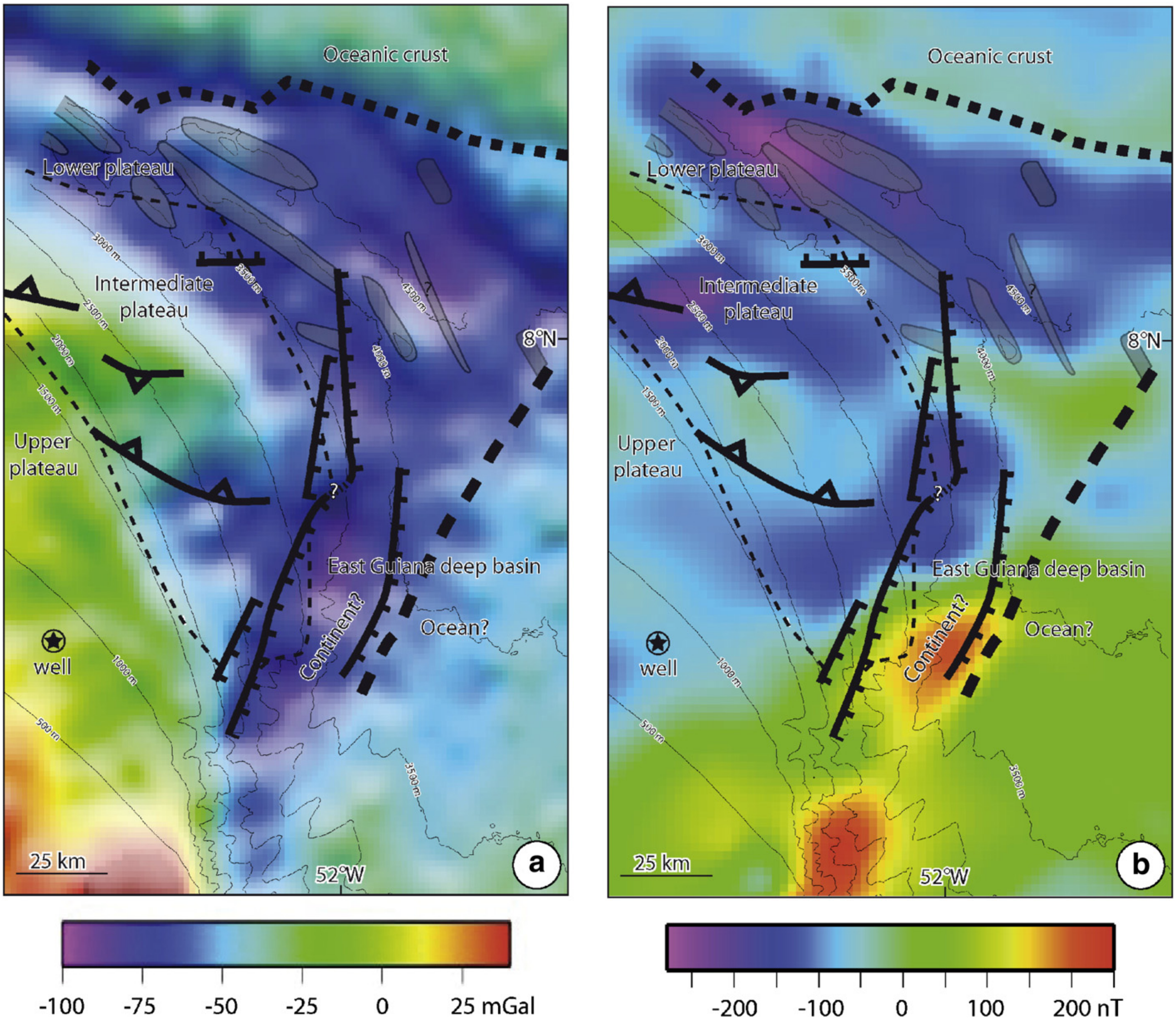

Fig. 8. Geophysical maps of the studied area. Same area and structural features as for Fig. 5. a: Free air gravity anomaly from satellite altimetry (Sandwell and Smith, 2009). b: Magnetic anomaly from gridded GUYAPLAC data.

and the oceanic crusts (Figs. 5, 6 and 8). On the bathymetry of the northern edge of the plateau, these ridges express as N115°-striking steep escarpments (Fig. 1), with a possible left-lateral en échelon setting when compared with the direction of the continent–ocean boundary (Fig. 5). This area presents an ESE–WNW trending negative magnetic anomaly (Fig. 8b), parallel to the anomalies overlying the oceanic fracture zones North of the Demerara Plateau. These ridges prolongate to the eastern border of the plateau, where they strike N150° (Fig. 5) and locally rises from the sedimentary cover (Fig. 1). On seismic lines, these ridges consist in acoustic basement (Fig. 4), but the nature of the rocks remains unknown, and because of the lack of stratigraphic correlations with the upper plateau, the timing of the ridge formation is also unclear. Along the northern edge of the plateau, seismic unit D does not seem deformed at the contact with the ridges, but on the eastern border, the base of the undulated layer is offset by a basement ridge, suggesting this structure may have been emplaced or reactivated during Cenozoic times.

6.6. Polygonal faulting

Seismic units B, C and D are deformed by numerous sub-vertical faults, either with a slight normal slip, or without significant offset (e.g. Figs. 4, 10 and 11). In these units, there are no changes in thickness or seismic facies associated to these faults. As units A and E are poorly reflective, it is not possible to assess if these two units are crossed by the faults. However, some faults clearly reach unit F, and some the sea floor where they are related to pockmarks (Loncke et al., 2009) (Fig. 10). This suggests that at least some of these faults may represent conduits allowing deep fluids to reach the sea floor. The timing of this faulting as not been determined, but should be recent, probably coeval with unit A deposition.

Fault spacing is on the order of few kilometers, possibly less as the quality of the seismic lines probably does not allow to image every fault. Spacing is quite regular, and does not vary with the strike of the seismic lines (Fig. 10), suggesting a polygonal setting of the faults (Cartwright et al., 2003).

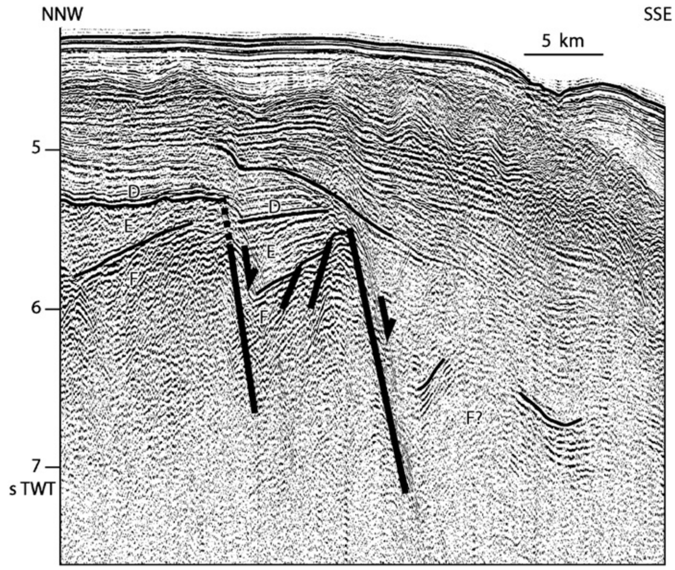

Fig. 9. Seismic line GUYAPLAC 22–23, located in Figs. 1 and 5. Faults are not directly observed in the acoustic basement, but drawn from offsets and tilts of the overlying seismic reflectors. Vertical exaggeration 13 in water.

Polygonal faulting affects the intermediate plateau, but also the northeastern edge of the upper plateau (Fig. 4). This deformation is not observed in the lower plateau or in the abyssal plain.

7. Discussion

7.1. Ridges at the continent–ocean transition

In the absence of indications for the lithological composition of the acoustic basement ridges located near the continent–ocean transition, two hypotheses can be proposed for their formation and setting:

- These ridges may be magmatic intrusions or extrusions at the continent–ocean boundary, which could explain the associated magnetic anomaly (Fig. 8b). Similar narrow and elongated magnetic anomalies have been reported along the toe of the continental slope in the transform margin settings of eastern South Africa and of Lebanon (Schattner and Ben-Avraham, 2007). Gouyet (1988) previously interpreted a dome-shaped basement high as a submarine volcano located at the foot of the northwestern border of the Demerara Plateau. According to Seiler et al. (2009), magmatic intrusions can be postulated along the Gulf of California transform continental margin when oceanic spreading occurred along the margin. The en échelon setting of the ridges on the northern edge of the plateau favors this strike–slip related emplacement, which in turn is not compatible with a Cenozoic intraplate emplacement as suggested by the easternmost ridges (cf. Section 6.5). It is however possible that all the ridges do not share the same nature and timing of formation, even if these ridges appear to be connected.
- These ridges may be built by tectonic processes, as observed along the Côte d'Ivoire–Ghana margin, at the intersection between the transform and divergent segments. There, several 25 km-long acoustic basement ridges are en échelon-disposed along the Romanche fracture zone (Basile et al., 1992, 1993). One of these ridges has been drilled during ODP Leg 159, and consists in synrift (Late Albian) sediments deformed inside the transform fault (Basile et al., 1998). Towards the divergent basin, this ridge bounds a 15 km-wide oceanward tilted block, narrower but similar in shape to the northward intermediate plateau where seismic units E and F are tilted oceanward (Fig. 4) (Basile et al., 1996). In both cases, the tilting is sealed by a regional unconformity. In the Côte d'Ivoire–Ghana margin, these ridges have been interpreted as remnants of a horse-tail termination of strike–slip faults (Christie-Blick and Biddle, 1985) at the intersection with the divergent segment (Basile et al., 1992). It is also noteworthy that most of the thinning of the crust occurs in both cases in the ridges area (Greenroyd et al., 2008; Sage et al., 1997), and that the northern ridges in the Demerara plateau were parallel to the transform fault (Fig. 12a) prior to plate kinematic reorganization that occurred 105 My ago (Fig. 2).

In any case, as some of the ridges are cropping out, direct sampling is most wanted to discriminate the two hypothesis.

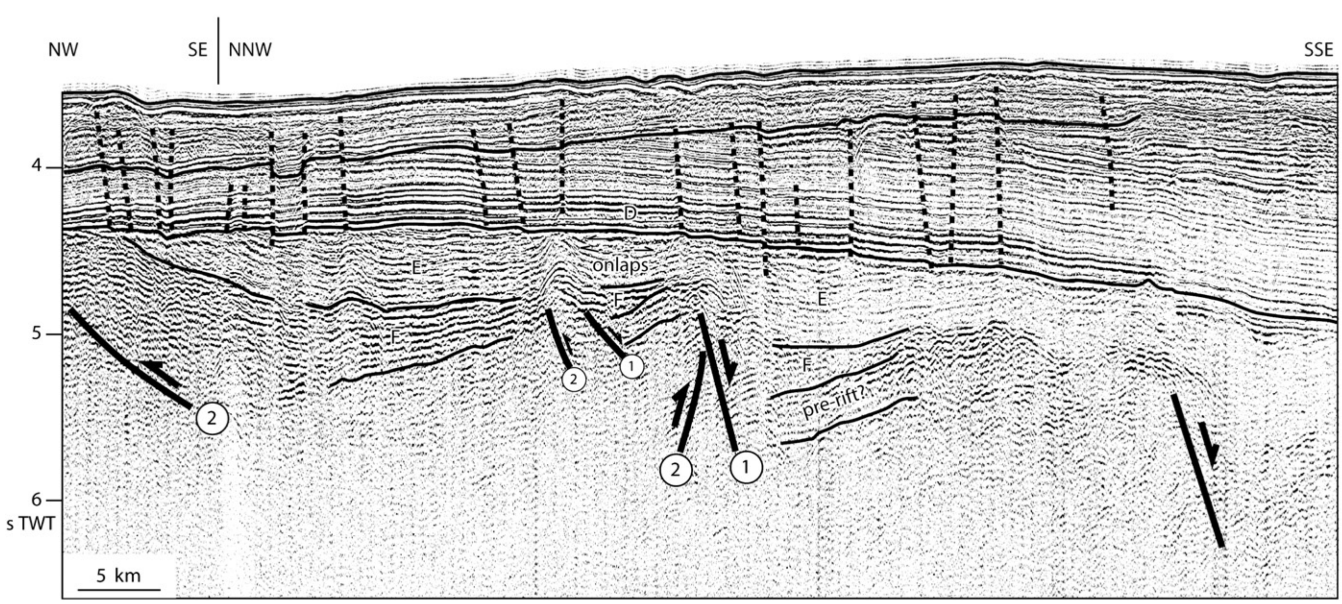

Fig. 10. Seismic line GUYAPLAC 28–29, located in Figs. 1 and 5. The circled numbers refer to the first and second stage of deformations, respectively. 1: normal faults coeval with seismic unit F. 2: reverse faults reactivating the same structures, coeval with the upper seismic unit E, and older than unit D. Faults are not directly observed in the acoustic basement, but drawn from offsets and tilts of the overlying seismic reflectors. Dotted thick lines indicate faults with polygonal network. Vertical exaggeration 13 in water.

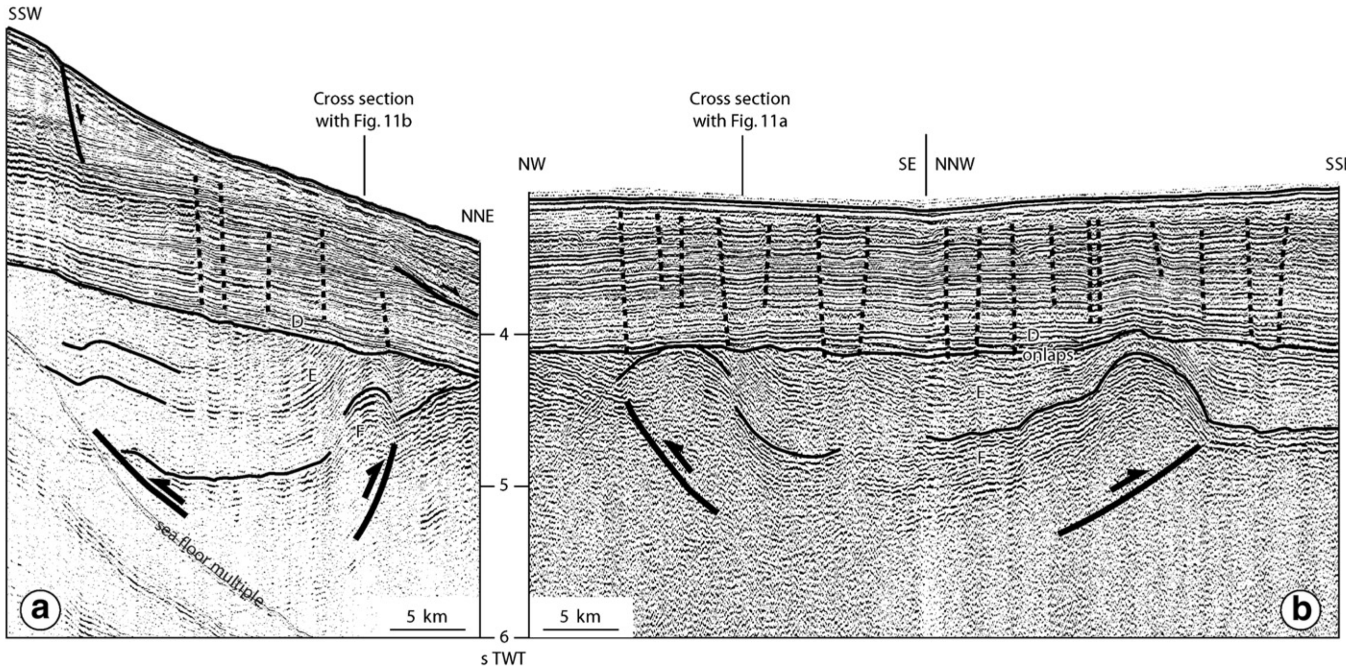

Fig. 11. Seismic lines GUYAPLAC 3 (11a) and 31–32 (11b), located in Figs. 1 and 5. Faults are not directly observed in the acoustic basement, but drawn from offsets and tilts of the overlying seismic reflectors. Vertical exaggeration 13 in water.

7.2. Tectonic inversion

It has been a frequent observation that passive continental margins, far from lithospheric plate boundaries, experienced limited shortening during their post-rift evolution. The main mechanism proposed to explain this shortening was the ridge push (e.g. Hudec and Jackson, 2002; Withjack et al., 1995), producing an intraplate effect of far field stresses. However, some post-rift shortening were also related to plate kinematic reorganizations, inducing deformation localized on the main preexisting structures such as the ocean–continent boundary (e.g. Iberia and Gulf of Biscay: Masson et al., 1994). Shortening has also been identified during rifting (Turkana rift: Le Gall et al., 2005), where it is related to oblique rifting on preexisting crustal weakness zones. Increasing the obliquity of rifting leads from divergent to transform continental margins, where folding is commonly observed, and associated to strike–slip tectonics (Christie-Blick and Biddle, 1985). Shortening can be observed in the sedimentary cover as positive flower structures parallel to strike–slip faults (e.g. in Ghana: Antobreh et al., 2009), oblique en échelon folds in sheared bands above or on the edge of the main strike–slip displacement zone (e.g. in Côte d'Ivoire–Ghana: Basile et al., 1993; Lamarche et al., 1997). Block rotation around vertical axis can also induce local shortening and folding in sheared zones (Basile et al., 1993). Finally, shortening clearly increases with transpression, and partitioning of strike–slip and convergent displacements can form a fold and thrust belt along a transform continental margin, as exampled by the Cenozoic West Spitzbergen margin (Braathen and Bergh, 1995; Lowell, 1972).

While the studied area is located at the intersection of transform and divergent margins, both segments appear to have been folded during the formation of the margin. Folding and associated thrusting are not restricted to the transform margin, but extend over the intermediate plateau, that covers most of the divergent margin. Seismic units F and E provide guidelines for the timing of tectonic shortening: they deposited during the main extensional phase, as they exhibit fan-shaped geometry bounded by normal faults, but the uppermost part of unit E is also coeval with shortening, as it onlaps the anticlines and is itself folded. The main regional unconformity, sealed by the middle to Late Albian unit D, is itself locally deformed both by normal faults (southward) and by folding (northward). These observations question both the mechanisms of deformation on this margin and the timing of rift to passive margin transition.

The southern border of the Guinean Plateau represents the conjugated transform margin of the northern border of the Demerara Plateau. As the Demerara Plateau, the Guinean Plateau is characterized by a post-rift tectonic inversion older than Cenomanian times (Benkhelil et al., 1995). WNW–ESE-trending folds are observed on the shallow edge of the Guinean Plateau (Tricart et al., 1991), and NW–SE-trending folds on the slope West of 17°30′W (Benkhelil et al., 1995) (Fig. 12a). Both deformation age and tectonic style are similar to the shortening deformations observed on the Demerara Plateau. However, no shortening deformations were described on the eastern side of the Guinean Plateau, which is the conjugated margin of the studied eastern Demerara Plateau.

When the two conjugated margins are replaced on their Late Albian fit (Fig. 12a), the Cretaceous fold axes on the Demerara Plateau appear to be roughly perpendicular to the coeval oceanic accretion axis, and consequently cannot be related to a ridge push mechanism. In this reconstruction, the two folded areas face each other on each side of the transform fault, in the last area where the American and African continental lithospheres are still in contact. However, the shortening directions inferred from the trends of folds axis differ by 60° and cannot be related to a homogeneous regional stress field (Fig. 12a). The folds located in the investigated area of the Demerara Plateau can be interpreted as resulting of a right-lateral displacement sub-parallel to the main transform fault (Fig. 12a). Southwest of the Guinean Plateau, folding can be related to N–S right-lateral strike–slip, parallel to a regional dextral fault described by Benkhelil et al. (1995), whose relationship with the E–W transform fault is unclear. Anyway in both Guinean and Demerara plateaux fold axes appear to be parallel to and localized on preexisting structures.

The Cretaceous shortening in the southwestern part of the Guinean Plateau is supposed to reactivate Jurassic extensional structures, related to the rifting of the central Atlantic (Benkhelil et al., 1995), whereas the inverted structures of the eastern part of the Demerara Plateau are localized by a Cretaceous syn-rift structure, related to the rifting and transform faulting in the equatorial Atlantic.

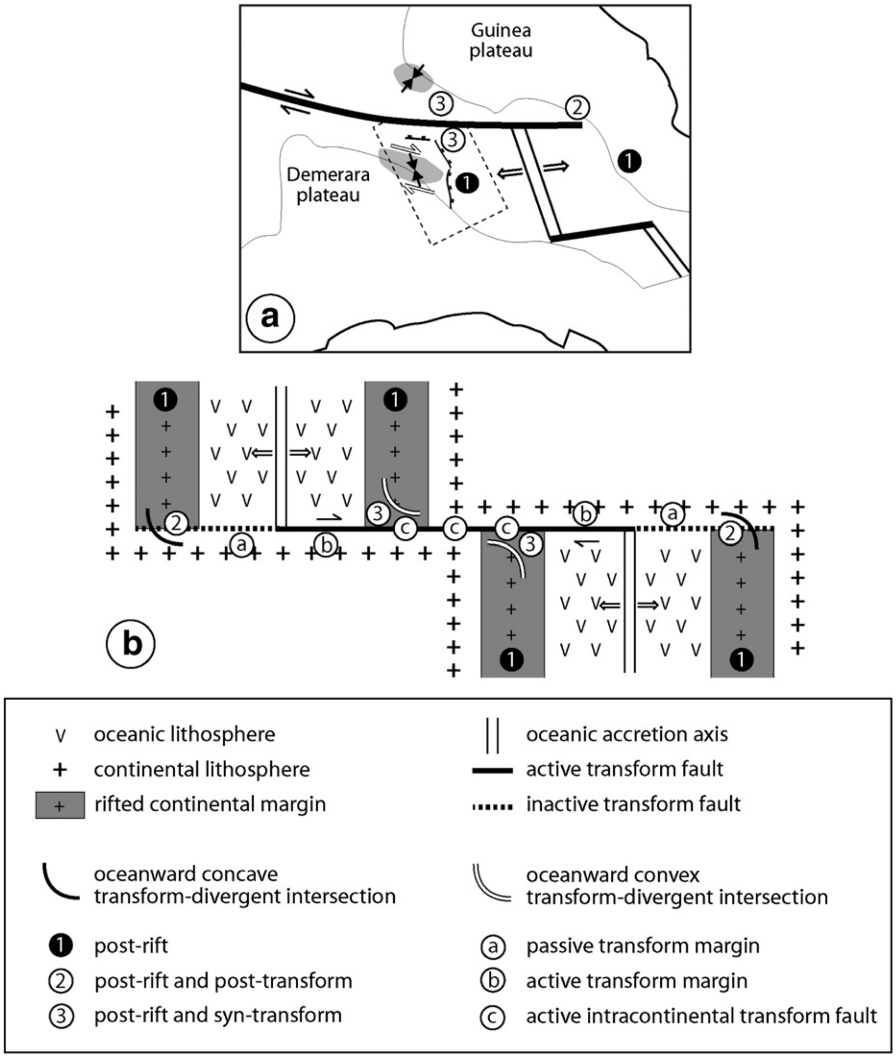

Fig. 12. a: Reconstruction of the Demerara and Guinea conjugated margins in their Late Albian (105 Ma) position, with Africa supposed fixed (modified from Basile et al. (2005) and Campan (1995)). The present day coastline and 2 km isobath are drawn. The dotted box indicates the studied area (Fig. 5). Post rift shortening occurred in the shaded area, and the black arrows indicate shortening directions. Deformations in the Guinea plateau from Benkhelil et al. (1995). b: Asymmetry and diachronism along transform margins and at transform–divergent margins intersections.
Concave/convex terminology according to P. De Clarens (personal communication).

However, because of the limited seismic penetration, the deeper structures cannot be observed in the available data from the Demerara Plateau, and it is not possible to decipher if the thrusts were located on older structures. In the southwestern part of the thrusted area, previous tilted blocks are inverted (Section 6.1 and Fig. 10), but there the extensional structures cannot be mapped, and it is not possible to determine if they trended NNE–SSW as the adjacent (and not inverted) normal faults or WNW–ESE as the later thrusts.

During Cretaceous times, both area first localize E–W extension, then folding that can be explained by right-lateral displacement along almost perpendicular structures. It is noteworthy that this Cretaceous inversion is coeval with a geodynamic reorganization that seems to be related to the start of the oceanic accretion in the studied area: the displacement between Africa and South America changes from WNW–ESE during rifting (parallel to the transform margin segments) to W–E during drifting (parallel to the present day oceanic fracture zones) (Fig. 2) (Benkhelil et al., 1995; Campan, 1995). A possible explanation of the right-lateral reactivation of the edges of the two plateaux is a slight convergence component along the transform margin just before the plate kinematic reorganization. However, the reactivation did not end when the two continents lost their last contact in this area, as attested by the strongly attenuated late deformation in the Cenomanian–Turonian sediments of both Demerara (c.f. Section 6.3) and Guinean plateaux (Benkhelil et al., 1995) (Fig. 12a). This suggests that the shortening is related to the active transform fault (intracontinental or active transform margin, Fig. 12b) rather than to the intracontinental setting.

7.3. The significance of the post-rift unconformity

In continental divergent margins, the post-rift unconformity generally refers to a regional erosional unconformity that erases the crest of some tilted fault blocks, coeval with the deposition of sediments sealing normal faults, and with the formation of the first oceanic crust (Falvey, 1974). Therefore this post-rift unconformity classically dates the transition from the rift (intracontinental) to the passive margin (continent–ocean boundary) stages.

For continental transform margins, an additional stage of active transform margin separates the intracontinental and passive margin stages (Basile et al., 1993; Mascle and Blarez, 1987). Moreover, the passive margin stage does not start with the oceanic accretion, but later when the oceanic accretion axis drifts along the transform margin. Consequently, the post-transform unconformity that seals the strike–slip faults is more recent than the post-rift unconformity in the adjacent divergent margins. More, this post-transform unconformity is diachronic along a transform margin, from older ages in the concave to younger ages in the convex transform–divergent intersections (Fig. 12b).

In the studied area of the Demerara Plateau, the seismic unit D covers the main regional unconformity, observed both on transform and divergent segments, and across their convex intersection. However, this unconformity seals both extensional and compressional structures, and is itself slightly deformed. This unconformity results from a subaerial erosion, that probably postdates the peak of deformation, but predates the end of the deformation.

From the changes of thickness of the various seismic units, one can propose that unit F may represent the syn-rift unit, and that the boundary between units E and F corresponds to the post-rift unconformity. Both ages and kinematic constrains imply that units E, D and the lower part of unit C are post-rift, but syn-transform. The youngest deformations occurred during the deposition of the lower unit C, but we did not observe any associated post-transform unconformity, probably because the whole Demerara Plateau was below sea level at this time. On the contrary, the main unconformity overlain by unit D occurred during the syn-transform stage, when the transform margin is still active and deformed, but before the post-rift thermal subsidence drown of the margin. We suggest that the regional unconformity observed over the Demerara Plateau may be related to a change in plate kinematic, but not to the end of rifting or transform deformation. A plate kinematic reorganization is expected to affect only the active transform segments, and consequently should be restricted to the convex transform–divergent intersections (western Guinean Plateau and northeastern Demerara Plateau), and should not be observed in the concave transform–divergent intersections (eastern Guinean Plateau (Benkhelil et al., 1995) and southeastern Demerara Plateau).

7.4. Tertiary tilt of the edges of the Plateau

One of the main characteristics of the Demerara Plateau is the Cenozoic oceanward tilt of its northern edge. In the studied area, the upper and lower plateaux remain horizontal, and this tilt affects only the intermediate plateau, approximately 60 km wide, with a slight increase of the slope downward. West of the studied area, the oceanward tilt affects the edges of the plateau whatever the nature and structure of the basement (Gouyet, 1988; ODP Leg 207 Shipboard Scientific Party, 2004). This oceanward tilt is not observed on the conjugated Guinean margin, where the Cretaceous unconformity stayed close to the horizontal.

From the lack of onlaps on the basal surface and the constant thickness of the sedimentary layers, this tilt is dated Cenozoic in age, possibly Late Oligocene–Early Miocene from an extrapolation with the dating of a regional erosional surface in the western Demerara Plateau (c.f. Section 6.4). Anyway, this tilt occurs at least few tens of million years after the end of rifting or the end of the transform faulting. This delay, as well as the restricted width of the tilted area, excludes thermal subsidence alone as an explanation, as this subsidence is expected to reach a maximum at the beginning of the post-rift stage, and to affect the whole divergent margin.

This vertical displacement may be driven by mantle dynamics. For example, Praeg et al. (2005) interpreted successive Cenozoic stages of tilting of the NW European continental margin as a result of changes in underneath mantle flow. Post-rift tilting along the continental edge may also be explained by an edge-driven convection cell emplaced at the continent–ocean boundary (King and Anderson, 1998). But there is no clear reasons to restrict this mechanism to the Demerara Plateau (and not on the adjacent margins) and to such a short wavelength of tilting. Not to mention its Neogene timing which does not have any clear explanation neither.

On the other hand, the Demerara Plateau tilting may be coeval with the Late Miocene uplift of the Eastern Cordillera (Colombia) and Venezuelan Andes, that changed the course of the Orinoco from the Caribbean Sea to the Central Atlantic, and connected the Amazon to the Atlantic (Hoorn et al., 1995). These changes in the drainage patterns of northern South America induced the development of two major deep-sea fans on each side of the Demerara Plateau. These thick sedimentary bodies in the abyssal plain loaded the oceanic lithosphere. If the oceanic and continental lithospheres are coupled across the continent–ocean boundary, loading the oceanic lithosphere should induce an oceanward flexure of the continental margin. As a thin lithospheric elastic thickness can be expected for the previously thinned and deformed continental crust of the intermediate plateau, this flexure should occur on a short wavelength and with a relatively steep slope. Watts et al. (2009) proposed such a mechanism to explain the uplift of the Guyana shield as a flexural consequence of the loading of the margin by the Amazon fan. They computed a 750 m-deflection induced in the Demerara area by the load of the Amazon fan alone, but did not encompassed the loading by the Orinoco deep-sea fan. This flexural mechanism also implies a positive feed-back, as a flexural tilt of the margin should induce gravity-driven slides of its sedimentary cover, unloading the margin but increasing the load in the abyssal plain, which would increase the flexure. However, this hypothesis remains purely conjectural, and needs to be discussed with better constrains on the geometry, age, and deposition environments of both margin and abyssal plain sediments, which is beyond the scope of this paper.

8. Conclusions

The new data set described in this paper allows us to describe the structures at the transform–divergent margins intersection East of the Demerara plateau. These tectonic structures consist mainly in normal faults in the eastern divergent segment, and perpendicular ridges on the northern transform segment. Both segments were later deformed by a post-rift but syn-transform shortening, during which deformations were localized on preexisting structures. This shortening is not compatible with a ridge push mechanism, but can be explained by strike–slip motion parallel to and in the vicinity of the coeval transform fault. Shortening is postulated to be related to a plate kinematic reorganization occurring 105 Ma ago at the time of the end of intracontinental transform faulting. The post-transform evolution is dominated by an unusual Cenozoic seaward tilting of the border of the plateau, whose origin is unclear, but that may be related to the loading of the adjacent oceanic lithosphere by the Orinoco and Amazon deep sea fans.

Two points have to be underlined:

- The divergent and transform structures of the margin are not parallel to the present day accretion axis and oceanic transform fault, respectively, as a plate kinematic reorganization occurred 105 Ma ago at the time of the end of intracontinental transform faulting. This suggests that the common assumption that drift kinematics prolongs rift kinematics may be questionable, especially in transform margin setting.
- The post-rift unconformity is not the main unconformity at the transform–divergent intersection in the studied area. This main unconformity is associated to the syn-transform shortening, and is well expressed because of the shallow depth (near sea level) at this time. It is related to a plate kinematic reorganization and not to the incipient oceanic accretion. This suggests that the post-rift (coeval with the formation of the first oceanic crust) and post-transform (coeval with the drifting of the oceanic axis along the margin) unconformities are not systematically the main stratigraphic unconformities at divergent–transform margin intersections.

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