Evolution of the Southern Guinea Plateau: Implications on Guinea-Demerara Plateau formation using insights from seismic, subsidence, and gravity data

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Abstract

The Guinea Plateau (offshore Guinea) and its conjugate, the Demerara Plateau (offshore French Guiana), comprise two of the most prominent passive continental margins in the Atlantic Ocean. The conjugate plateaus formed as a result of two periods of rifting, the Jurassic opening of the Central Atlantic Ocean and the northward-propagating Cretaceous opening of the Southern Atlantic Ocean. Although several studies are published on the Demerara Plateau that explain the evolution of its multi-rift history and the effect of rifting on its distinct geometry, the Guinea Plateau, and in particular its south-eastern margin, remain relatively unexplored in the literature. Here we present interpretations of the structure and evolution of the Guinea Plateau using recent 2-D and 3-D seismic-reflection data collected at the intersection of the southern and eastern margins. We substantiate our study with calculated subsidence curves at four locations along the southern margin, as well as two 2-D gravity forward models along regional seismic-reflection profiles to estimate stretching factors (β) and crustal thicknesses. We combine our results with previous studies concerning the south-western Guinea margin, and compare them to published interpretations regarding the conjugate margins of the Demerara Plateau. The resolved amounts of rift-related volcanism, listric-style normal faults, and moderate stretching factors suggest that a component of upper-crustal asymmetry (simple shear) and depth-dependent stretching may have persisted at the Demerara-Guinea conjugate margins during Cretaceous rifting of the equatorial segment of the Southern Atlantic Ocean.

1. Introduction

Passive continental margins are broadly classified using tectonic models based on a combination of structural, stratigraphic, and magmatic processes. This classification scheme has led to passive continental margins typically being distinguished by their divergent or transform tectonics, (non)presence of salt structures, and the amount of volcanism produced during continental rifting. The formation and evolution of such passive continental margins are often

explained by regional-scale models that focus on the relative contributions of pure and/or simple shear (McKenzie, 1978; Wernicke 1985; Barbier et al., 1986). These models have implications for the types of structural deformation observed along conjugate passive margins. Under conditions of pure shear, magmatic activity should be centered on the rift axis where considerable MORBs may be extruded, while similar structural deformation on both sides of the rift axis persists. In contrast, under simple-shear conditions, one margin of the rift axis displays listric-style deformation, while the conjugate retains planar-style deformation (Ziegler and Cloetingh, 2004).

In the Southern Atlantic Ocean, conjugate volcanic-rifted margins are found south of the Walvis Ridge, offshore Namibia and Brazil (Orange Basin – de Vera et al., 2010; Pelotas Basin – Stica et al., 2014) and display significant packages of extrusive volcanics manifested as seaward-dipping reflectors (SDRs). Further north, in the central segment, are the well-documented salt basins of the Santos-Campos Basin and offshore Angola that have become known for their significant oil reserves (Davison, 2007; Marton et al., 2000). Further north still, is the equatorial segment, defined as the region bounded in the south by the Chain and Romanche fracture zones (FZs) and in the north by the Marathon and Guinea FZs (Fig. 1). The equatorial segment contains one of the best-preserved examples of an extinct transform passive margin (the Ivory Coast-Ghana transform margin), which gave rise to an oblique spreading model for the evolution of transform passive margins (Mascle and Blarez, 1987; Mascle et al., 1988; Mascle et al., 1997). This model has subsequently been used to characterize observations of upper-crustal deformation and crustal thicknesses of the equatorial Demerara Plateau using seismic-refraction and reflection data sets (Greenroyd et al., 2007, 2008; Basile et al., 2013).

Although there have been significant recent academic geophysical studies conducted on the evolution of the Demerara Plateau, chiefly the Amazon Cone Experiment (Greenroyd et al., 2007), GUYAPLAC cruise (Basile et al., 2013), and GuyanaSPAN survey (Reuber et al., 2016),

relatively little work has been done to unravel the evolutionary history of the conjugate Guinea Plateau. We present in this paper an interpretation of newly acquired 2-D and 3-D seismic reflection data across the southern and eastern margins of the Guinea Plateau (Fig. 1). The seismic horizons are regionally tied to two exploration wells, GU-2B-1 and SABU-1. In addition, we calculate subsidence curves at both well locations, and two additional pseudo-wells. Lastly, 2-D crustal thickness models are computed using satellite-derived free-air gravity anomalies (Sandwell et al., 2014). Our results are integrated with previous studies to delineate the evolution of the Southern Guinea Plateau, with comparisons made to the well-studied conjugate Demerara Plateau.

2. Regional setting

The Guinea Plateau is located off the coast of Guinea, West Africa, in the equatorial segment of the Southern Atlantic Ocean (Fig. 1). The plateau itself is a salient bathymetric feature that extends nearly 300 km from the continental coast and is visible from space. For this reason, the Guinea Plateau, and its conjugate, the Demerara Plateau offshore French Guiana, have served as attractive markers for tectonic plate reconstructions. Global reconstructions, based on magnetic lineations of the seafloor, have shown that the southern-most extent of the Central Atlantic Ocean, located north of the Marathon FZ (Fig. 1), opened during the Middle-Late Jurassic (Klitgord and Schouten, 1986). At this time, Africa and South America were still connected, and the western margin of the Demerara-Guinea Plateau complex was adjacent to the Central Atlantic Ocean. Moulin et al. (2010) infer that the first oceanic crust in the Southern Atlantic occurred in the Early Cretaceous between 132–134 Ma near the southernmost tip of Africa and South America. Subsequent stepwise, northward-propagating rifting continued to separate the two continents until 105–107 Ma, when a through-going ridge-transform-ridge system developed between the Demerara and Guinea Plateaus (Moulin et al., 2010; Edge, 2014).

3. Previous Studies

No new studies with specific focus on the Guinea Plateau have been published in the last 2–3 decades. Sheridan et al. (1969) presented a ~600-km-long seismic-refraction profile extending from Freetown, Sierra Leone, out into the Sierra Leone abyssal plain. Jones and Mgbatogu (1982) used coincident gravity, magnetic, and seismic observations collected across the Guinea-Bissau, Guinea, and Sierra Leone continental margins to develop the first published crustal models and geologic history of the Guinea Plateau. Their contributions were critical for creating an awareness of the Guinea Plateau's western and southern margin evolutions as the distinction of two episodes of rifting: the Jurassic opening of the Central Atlantic and the Cretaceous opening of the Equatorial Atlantic. The authors also created the first highly-detailed bathymetric map of the region, and highlighted the pronounced features of the Guinea Fracture Zone.

Masclé et al. (1986) collected and interpreted a suite of continuous, single-channel seismic reflection lines, in addition to six dredging/coring sites, and established a preliminary structural and stratigraphic sketch of the Guinea Marginal Plateau (i.e. southern margin). Following up on this study, Benkhelil et al. (1995) presented interpretations based on a denser grid of single-channel 2-D seismic lines collected during the Equamarge I & II cruises. The authors resolved two main tectonic episodes affecting the Southern Guinea Plateau: (1) rift-related normal faulting, and (2) reverse faulting (via inversion), folding, and transcurrent faulting. However, the location of their data and study focused on the western tip of the Guinea Marginal Plateau. At this point in time, no comprehensive geophysical study exists of the southern E-W Guinea Plateau margin as it approaches the intersection with the N-S margin located offshore Sierra Leone. As for the Demerara Plateau, a thorough discussion of previous studies can be found in Basile et al. (2013).

4. Seismic and Well Data

Extensive 2-D and 3-D seismic reflection data were collected by Hyperdynamics Corporation and were made available to the University of Arizona (Fig. 1). Two depth-migrated 3-D datasets, Surveys A & C, cover approximately 7800 km² of the Southern Guinea Plateau. The line spacing is 25 m x 25 m for both surveys, resulting in excellent resolution and confidence in seismic interpretations. An additional 84 depth-converted 2-D lines provide coverage in and beyond the 3-D surveys for regional interpretation. Lastly, two exploration wells, GU-2B-1 and SABU-1, provide well-ties for seismic interpretation.

The seismic stratigraphy that defines the basis for all interpretations presented in this study is shown in Figure 2, and is summarized here from bottom to top:

- The deepest portions of the seismic data are interpreted as Jurassic and older basement. Neither of the exploration wells penetrate deep enough to sample Jurassic stratigraphic units. Many of the discernable reflectors display prominent convex-upwards geometries that bound chaotic seismic data (see Figs. 2 and 3 for examples). We interpret these as related to the Freetown Complex intrusives near the coast of Sierra Leone, which are major gabbroic intrusions dated at 193 ± 3 Ma and interpreted to be a consequence of the stress regime associated with the initial stages of Atlantic rifting (Beckinsale et al., 1977; Mgbatogu et al., 1988). No Mesozoic strata have been reported on the coasts of Guinea-Bissau, Guinea, and Sierra Leone; therefore, clastic Jurassic sedimentary rocks are inferred to be continentally-derived from Paleozoic and Pre-Cambrian basement rocks of the Bove Basin and Leo Uplift, respectively. The top of the Jurassic (145 Ma) is

marked by an angular unconformity evident in the seismic data. The unconformity is interpreted to be related to the break-up unconformity of the Central Atlantic Ocean rift event (~175 Ma; Pindell, 1985) and/or a prominent local decline in sea-level.

- Above the Jurassic unconformity are pre-rift Neocomian sedimentary rocks. The Neocomian-aged units are regionally characterized by high-impedance, “rail-road-track” reflections interpreted to be constituents of carbonate platforms (Nibbelink, pers. comm). This sequence has been drilled and identified on the Guinea Plateau shelf in the GU-2B-1 well and the A2-1 well located on the Demerara Plateau (Reuber et al., 2016).
- Aptian strata overly the Neocomian carbonates and consist of sandstones, mudstones, and rift-onset basalts. Sandstones and mudstones logged in the GU-2B-1 well consist of fine-to-medium, subangular, well-sorted grains implying near-sea-level conditions distal from the source. Basalts drilled on the shelf exhibit low densities, ranging from 2350–2600 kg/m³ and velocities in the range of 4.0–5.5 km/s.
- Albian strata mark the transition into syn-rift sedimentation. Thicknesses range from 0.2–0.3 km on the shelf, to a maximum of 4.0 km in the deepest sections of the rift basins. On the shelf, sandstones, mudstones, minor evaporites, and volcanoclastics were drilled (GU-2B-1). The volcanoclastic rocks appear as high-impedance reflections in the seismic with logged densities around 2200 kg/m³ and velocities of ~4.0 km/s. Only the uppermost Albian units have been drilled above the large syn-rift half grabens, where volcanic trachytes with densities of 2600 kg/m³ and velocities ~5.0 km/s were found. Argon-Argon dating on core samples resolved an emplacement date of ~106 Ma for the trachytes. These trachytes are interpreted to be genetically related to the silica-under-saturated alkaline-nepheline-syenite suite known as the Los Archipelago ring structure emplaced 104.3 ± 1.7 Ma and located off the south-western shore of Conakry, Guinea (Moreau et al., 1996). In the half-grabens, kilometers-thick clastic syn-rift fill is inferred based on

layered reflection geometries and low amplitudes. The top of the Albian is marked by an unconformity in the seismic, which marks the termination of syn-rift kinematics.

- Several sequences of sandstones and shales on-lap onto the Albian unconformity. Late Cretaceous inferred volcanic cones are found locally on the plateau (McMillan, 2012; McDougall et al., 2013). Thick sequences of carbonates are bounded in-between the Cretaceous and Oligocene unconformities, deposited during times of post-rift margin subsidence. Deposition of sands and muds via submarine channels characterize the top of the stratigraphy and continue to be deposited on the present-day seafloor.

4.1 3-D Seismic Interpretation

Four interpreted seismic lines from the 3-D datasets are shown in Fig. 3. Lines 3D-1 and 3D-2 are oriented mostly parallel to the margin slope, while Lines 3D-3 and 3D-4 are perpendicular to these and extend from the shelf out into the deep water. All four lines emphasize the deformation associated with Cretaceous rifting, i.e. faulting, sedimentation, and volcanism. Line 3D-1 runs sub-parallel to the shelf edge and shows an oblique view of the dominant rift-graben in the region, the Baraka graben, informally named after the basin-bounding Baraka fault. The Baraka fault is a very large listric fault interpreted to sole out into basement rocks at a depth of about 8.0 – 9.0 km, while offsetting pre-rift sediments up to a maximum of ~4.0 km and allowing large volumes of clastic syn-rift deposits to accumulate. Although the Baraka fault displays considerable displacement at its center, interpretation of 3-D data shows the displacement diminishes within ~50 km to the east and west of its center (see discrepancy in faulted Aptian depths from lines 3D-3 and 3D-4, Fig. 3). Additional half-graben basins and their bounding normal faults persist beyond the Baraka graben towards deeper water. Lines 3D-3 and 3D-4 (Fig. 3) emphasize the complexity of these basin and fault geometries. In line 3D-3, it is evident that the volume of syn-rift fill decreases toward deeper water in successive half-grabens. Meanwhile, in

line 3D-4, we interpret only two listric faults that cut through the Aptian and sole out into basement rocks. Instead, toward the deeper water, large-scale faulting gives way to smaller-scale semi-planar faulting that soles out into the top of the Aptian. Similar changes in fault geometries are found to the west of Line 3D-3 (not shown) and may be a result of the transtensional tectonics of the region.

The base of the syn-rift fill is defined by the rift-onset unconformity (ROU). At rifted margins, pre- or syn-rift doming and erosion creates an ROU, which is typically recognized in the seismic data as an angular unconformity with top-lap truncations of seismic reflectors from below (Franke, 2013). In Lines 3D-1 and 3D-4, we see clear examples of a prominent top-lap angular unconformity, which we associate with rifting initiated in Aptian time based on the presence of Aptian-aged basalts drilled on the shelf. We used this relationship to map an Aptian horizon in areas of deformation associated with rifting. Conversely, the top of the syn-rift basins are beveled by the break-up unconformity (BU), a common marker at rifted margins that represents the change from rift-to-drift and initiation of oceanic spreading (Falvey, 1974; Franke 2013). The BU on the Southern Guinea Plateau is associated with a late-Albian age and is prominent in our 3-D seismic data as the boundary at which the majority of syn-rift faults are truncated. The Albian BU is also associated with the tops of several paleo-seamounts in the deeper water (Fig. 2), volcanoclastics in the shallow water, and on-lapping clastic strata across the continental slope (Fig. 2 & 3). The development of the Albian BU in the deepwater may be related to hiatuses in pelagic sedimentation at passive continental margins caused by bottom-water currents (Hubbard, 1988). Consequently, the several submarine channels dissecting the continental slope-rise margin of the present-day Southern Guinea Plateau seafloor are interpreted to have persisted since the Cretaceous rifting event.

Albian volcanoclastics on the shelf are interpreted to be lava flows likely deposited in shallow-water conditions. Their low densities and low seismic velocities may imply that the volcanoclastics were a product of significant back-and-forth wave-tide action that broke down the extrusive flows and mixed them in with sands and shales (Nibbelink, pers comm). This contrasts, however, with the Albian paleo-seamounts that are found in the deeper water (Fig. 2 & 4). Their relatively higher densities and seismic velocities may suggest that they were subjected to less mechanical erosion than their counterparts deposited on the shelf. Regardless, both expressions of the Albian volcanics are the result of a period of syn-rift volcanism near the end of rifting (106 Ma).

Horizon maps of the Aptian ROU and Albian BU are shown in map and perspective views in Fig. 4. The plan-view maps for both unconformities show interpreted fault traces that resulted from the Cretaceous rifting event. Note that many additional minor faults may exist, particularly within the half-graben basins; however, faults with traces less than 3.0 km in length are not included in this regional study. The fault traces show that the extent of large-scale listric faulting is localized to a relatively small area ~70 km in length and ~40 km in width. Beyond this region to the east-northeast, the large listric faults diminish and give way to volcanoes, volcanic flows, and small planar faults. To the west-southwest, offsets along the Baraka fault become much less pronounced as it begins to mark the hinge-zone at the end of the shelf, as found on other rifted margins of the Atlantic Ocean (e.g. Steckler and Watts, 1982).

In addition to rift-related faulting, both map views also show the locations and extents of volcanic features associated with their respective geologic times. Notice the amount of Aptian (rift-onset) volcanism apparently is less than the amount of Albian volcanism. However, it is important to note that the amount of mapped Aptian rift-onset basalts shown here is likely a minimum due to difficulties in differentiating the basalts from other possible rock types in the rest

of the section. For example, in line 3D-2 (Fig. 3) we highlight an example of interpreted Aptian basalt flows that appear to have accumulated in an old paleo-topographic low. However, in many of the rift grabens, it is difficult to distinguish the amount (if any) of basalts located on the Aptian ROU (e.g. Lines 3D-3 & 3D-4). Kardell (2016) analyzed the volumes of rift-related volcanism in the Southern Guinea Plateau and reported values far below the average values found on typical volcanic-rifted margins in the Atlantic Ocean (Franke, 2013). Albian volcanics, however, are dispersed spatially throughout the datasets, and do not appear confined only to faults.

4.2 Regional 2-D Seismic Interpretation

We extended our interpretations beyond the area imaged by the 3-D surveys via 84 depth-converted 2-D seismic profiles (Fig. 1). All horizon and fault picks on these profiles were constrained by interpretations made in the 3-D data. The seismic resolution of the 2-D data is not as high as the 3-D data; however, regional-scale faulting and volcanic features are still conspicuous. Investigation of the regional 2-D lines shows a transition from the listric-style deformation that dominates the Southern Guinea Plateau from 14.5°W–15.5°W, into interpreted transtensional faulting west of 15.5°W.

Line 2D-1, a ~230-km-long seismic profile that spans the Guinea Plateau shelf out onto the abyssal plain, is shown in Fig. 5. This seismic profile is unique within this study in that it images a significant segment (~80 km) of oceanic crust, and consequently the narrow transition from continental shelf to oceanic crust (≤ 50 km). Onset of oceanic crust in the seismic data is interpreted as the juxtaposition of layered seismic reflectors of continental origin, with the chaotic and highly-fractured reflectors of oceanic crust. In addition, the sea-floor bathymetry smooths significantly beyond the stretched continental crust as it transitions into deeper water underlain by oceanic crust. Line 2D-1 also shows the reduced offset of syn-rift units across the Baraka fault.

269 Instead, we observe significant offset across the gently dipping ($\sim 20^\circ$) fault located south of the
270 Baraka. This fault, herein referred to as the transtensional Demerara-Guinea Transform fault,
271 appears to offset the Aptian ROU and the Albian BU before the transition into oceanic crust
272 occurs. The Aptian ROU ultimately truncates into the oceanic crust, while the Albian BU is
273 interpreted to persist above the first 30–40 km of oceanic crust, resulting in syn-rift strata being
274 deposited on top of oceanic crust (Fig. 5).

275
276 Farther to the west, and up on the shelf, Line 2D-2 shows a dramatic difference in
277 sedimentary thicknesses relative to those observed in the 3-D datasets (Fig. 6). In particular, pre-
278 rift strata thicken towards the west, while syn-rift strata decrease in thickness until they are nearly
279 indiscernible. Faulting is observed in the upper-Neocomian section, which increases to over 3 km
280 in thickness. Aptian-aged units appear fractured with disrupted reflectors; however, their
281 geometries are not able to be accurately resolved. Above this is a thin veneer of Albian syn-rift
282 units that disappear towards the west. Additionally, post-rift strata also display decreasing
283 thicknesses towards the west. In contrast, the entire section has water depths of approximately
284 one km before slightly increasing at the western edge. Increasing Aptian and Neocomian unit
285 thicknesses towards the west implies that this region experienced increased accommodation
286 space and subsidence prior to Cretaceous rifting, likely a result of the Jurassic rifting of the Central
287 Atlantic.

288
289 Line 2D-3, shown in Fig. 7, is located near Line 2D-2 except oriented NE-SW rather than
290 E-W (Fig. 1). The persistence and resolution of the faulting, however, is more pronounced than
291 that imaged in Line 2D-2. Faulting of the Aptian units appears to continue down through the
292 Neocomian and, at times, the Jurassic. We interpret a negative (i.e. normal-slip) flower structure
293 (Harding, 1985) based on the fault-plane geometries and oppositely tilted reflectors in the Aptian
294 section. Similarly increasing Aptian and Neocomian unit thicknesses towards the west support

the influence of Jurassic rifting also inferred on Line 2D-2. These pre-existing normal faults, created during the Jurassic, were likely transtensionally reactivated during the subsequent oblique rifting of the Equatorial Atlantic, creating the observed flower-structure geometry.

Regional maps of the Aptian and Albian horizons (Fig. 8) show, particularly on the shelf, a minimal amount of rift-related volcanic rocks beyond the boundaries of the 3-D surveys. Instead, the existence of volcanics appears to be limited to volcanism related to the onset of sea-floor spreading, manifesting as paleo-highs of the oceanic crust (e.g. Fig. 5). In addition, the maps show that there is almost no discrepancy in the depths of the horizons towards the west. In other words, a computed isopach of the two horizons would resolve little-to-no syn-rift sedimentation towards the western boundaries of the 2-D seismic profiles. Therefore, the maps in Fig. 8 show us the limited influence of Cretaceous rifting towards the western portion of the Guinea Plateau.

An improved structural outline of the Southern Guinea Plateau that integrates interpretations presented in this report with previous interpretations discussed in Benkhelil et al. (1995) is shown in Fig. 9. This structural map indicates the transition of rift-related tectonics across the entire southern margin, which includes: (1) from 14.5°W–15.5°W, a localized zone of listric faulting that locally cuts down into Jurassic-aged crust, is coincident with, and surrounded by, several volcanoes and volcanic flows; displacement along the listric faults terminates and/or diminishes into a series of smaller, more-planar normal faults that sole out into Aptian strata. (2) From 15.5°W–17.3°W, faulting is dominated regionally by the E-W-oriented transtensional Demerara-Guinea Transform fault, with a reduced amount of rift-related volcanism and offset across the Baraka fault. Near 17.0°W, faulting is interpreted to have a NW–SE orientation and displays a negative flower-structure geometry that offsets Aptian and older units. (3) West of 17.3°W, Benkhelil et al. (1995) interpret a NNW–SSE-trending transcurrent fault zone, surrounded by several normal faults, inverted normal faults, and associated folds oriented roughly

NW–SE. These authors interpreted their findings to be the result of polyphase tectonic activity; first, N-S trending synsedimentary normal faults created during Jurassic divergent rifting, followed by folding, structural inversion, and transcurrent faulting created during Cretaceous transform motion between the Demerara-Guinea conjugate plateaus.

5. Backstripping and Tectonic Subsidence

The syn-rift and post-rift subsidence of passive continental margins can be explained, to a first order, as the result of mechanically stretching the continental lithosphere, producing thinning and passive upwelling of hot asthenosphere (McKenzie, 1978; Le Pichon and Sibuet, 1981). In order to quantify the rifting and subsidence history of the Guinea Plateau, we have applied the backstripping technique of Steckler and Watts (1978) to two exploration wells and two pseudo-wells created from seismic stratigraphy (Fig. 1). The backstripping technique allows us to remove the effects of sedimentary loading and isolate the component of tectonic subsidence, i.e. subsidence due to thermal cooling and contraction.

Decompaction of sediment layers was accomplished using the method outlined in Sclater and Christie (1980). Empirical functions for porosity-depth changes for each lithology were extracted from exploration well data. This approach simulates what would happen if younger strata were progressively removed allowing underlying strata to decompact and return to their former depths. During backstripping, paleo-water depths were inferred based on sediment lithology as microfaunal and palynological assemblages were not available. For this reason, multiple backstripping curves with varying paleo-water depths were computed for each well location in order to quantify uncertainties in the tectonic subsidence. Paleo-sea level values were taken from Haq et al. (1987). Tectonic subsidence values were calculated as point loads and the flexural strength of the lithosphere was not considered (Steckler and Watts, 1978).

The first column of Fig. 10 shows the backstripping results for the Guinea Plateau. In each plot, the decompacted sediment and tectonic subsidence curves show the convex-upward shape characteristic of rifted margins. Superimposed on the tectonic curves are exponential curves computed using the McKenzie (1978) uniform stretching model (USM). The parameters used for the USM and backstripping analyses are presented in Table 1. By fitting USM thermal subsidence curves to our tectonic subsidence curves, we estimate the amount of stretching that has occurred at different locations of the Guinea Plateau (i.e. β -values). Note that, where applicable, USM curves were computed for both the Jurassic Central Atlantic thermal subsidence and the Cretaceous Equatorial Atlantic rifting event. For the GU-2B-1 well, the tectonic subsidence signal due to faulting during the Cretaceous rift event is indiscernible. Therefore, we believe this region of the Guinea Plateau has been experiencing exponentially decaying thermal subsidence since the Late Jurassic. In contrast, at SABU-1 located on the continental slope, the two rift events are well distinguished; thermal subsidence from the Late Jurassic is abruptly disturbed by extension and faulting during the opening of the Equatorial Atlantic (e.g. Baraka basin, Fig. 2). Even further into the deepwater, calculations of tectonic subsidence for pseudo-well 1 (PW-1) produces a subsidence curve that displays similar results to those of the SABU-1 well, but, with less thermal subsidence from Jurassic rifting and more extension and thermal subsidence from Cretaceous rifting. The subsidence results for pseudo-well 2 (PW-2) are quite different than the others, with the signal being dominated by Jurassic thermal subsidence. We identify no discernable subsidence due to faulting in the Cretaceous; however, the PW-2 location is ~200 km west of GU-2B-1, but only 50 – 100 km east of the continental slope of the Central Atlantic (Figs. 1 and 6), therefore this result is not too surprising.

In the second column of Fig. 10, we show the results of our backstripping and USM calculations for four locations on and around the Demerara Plateau. It is important to note that

the calculations are based solely on pseudo-wells inferred from published seismic lines and velocities found in Greenroyd et al. (2007 and 2008) and Basile et al. (2013); therefore, large uncertainties due to our depth conversions are inherent. Although our calculations are first-order estimates, there is a strong consistency in subsidence amounts and stretching factors between the conjugate plateaus. In addition, both plateaus show the combination of both rifting events in their subsidence patterns. Extension factors computed from Jurassic thermal subsidence data on the Demerara Plateau are less than those derived from the Guinea Plateau data. However, this may be more a function of the locations on the Demerara Plateau used for backstripping analysis, none of which are less than ~150 – 200 km from the western edge of the Demerara Plateau.

Table 1. Parameters used in Subsidence and Mechanical modeling

Parameter	Value
Density of water	1000 kg/m ³
Density of sediment grains	2350 kg/m ³
Density of crust	2800 kg/m ³
Density of mantle	3330 kg/m ³
Zero elevation crustal thickness	33 km
Thermal thickness of the lithosphere	115 km
Temperature of the mantle	1333 °C
Thermal diffusivity	8.0 x 10 ⁻⁷ m ² /s
Thermal expansivity	3.28 x 10 ⁻⁵ / °C

6. 2-D gravity modeling

Differences in crustal thickness across passive continental margins provides an alternative measure of the amount of extension experienced as a result of rifting (referred to as β_c in this paper). In this study, we used free-air-anomaly (FAA) gravity data from Sandwell et al. (2014) in order to generate 2-D crustal models. The FAA gravity data are shown as an inset map in Fig. 11, from which two profiles were extracted for crustal-model computation. Both the FAA gravity

map and profiles show the characteristic continental margin “edge effect” in which a FAA high is aligned with the shelf-slope break, accompanied by a FAA low at the shelf rise (Rabinowitz, 1982; Watts and Marr, 1995). Somewhat unusual is the consistently positive FAAs across the vast majority of the Guinea Plateau shelf. The inset map shows a very strong SE–NW-trending anomaly high from Sierra Leone in the south to offshore Guinea Bissau in the north. These consistent anomaly highs may be associated with gabbroic intrusions dated at 193 ± 3 Ma interpreted to be related to the opening of the Central Atlantic Ocean (Beckinsale et al., 1977; Mgbatogu et al., 1988). For crustal modeling, section geometries and densities within the upper 10 km were constrained by seismic reflection and well data. At crustal depths greater than 10 km (i.e. lower continental crust, oceanic crust, and mantle), densities were inferred from nearby crustal modeling studies offshore Sierra Leone (Jones and Mgbatogu, 1982) where seismic-refraction data was available (Sheridan et al., 1969). Moho geometries were determined through a combination of modeled best-fits to the gravity data as well as reasonable agreement ($\leq 10\%$) of crustal thickness variations determined for the nearby Sierra Leone passive margin.

In general, our calculated results closely match the observed gravity values, with average errors being ± 4 mgal for both profiles. We find the greatest discrepancies occur at gravity highs oscillating across the continental shelf, where dense intrusive rocks may exist as parts of the basement complex. In profile A, a steep gravity increase occurs oceanward of the gravity low (Fig. 11). We interpret this as an effect of sharp changes in water volumes caused by submarine channels on the ocean floor, which have relatively large effects, as well as a ~20-km segment of relatively higher density transitional crust. The transitional crust may be interpreted in the seismic reflection data (Fig. 5) as an area of highly fractured/disrupted reflectors at the boundary between interpreted oceanic and continental crust. Meanwhile in profile B, a smaller magnitude FAA low is observed at the continental rise and is followed by a less dramatic increase in FAA towards zero above isostatically compensated oceanic crust. Another difference between the two profiles

is the thickness of the oceanic crust and its overlying sedimentary sections. In profile A, the oceanic crust is thick (~7 km), with a rugose topography, and a ~2–3-km-thick draping of sedimentary units. Conversely, Profile B shows thin oceanic crust (~5 km), with smooth topography, and ~5-km-thick overlying sedimentary section. Although it is difficult to explain the difference in oceanic crustal thicknesses, the rugose topography in Profile A may be related to its position within the transtensional regime of the Guinea Plateau margin that crosses the Guinea Fracture Zone, a lineament along which volcanic rocks and seamounts continue to occupy the present-day seafloor (Fig. 1). Meanwhile, the thicker post-rift strata seen in Profile B are likely a function of the profile's position, which spans the divergently rifted section of the Guinea Plateau, a region characterized by considerably more syn- and post-rift sedimentation.

Compiled magnetic anomaly data from Maus et al. (2009) are also presented in Fig. 11. Jones and Mgbatogu (1982) hypothesized that a small magnetic peak coinciding with the gravity low at the continental rise defined the location of the ocean-continent transition (OCT) offshore Sierra Leone. Along the Southern Guinea Plateau margin, however, we see no such correspondence. In Profile A, there is a broad magnetic high seaward of the gravity low at the continental rise. The magnetic anomalies of Profile B are even more ambiguous, with one broad magnetic high located above the syn-rift grabens, and another seaward of the OCT. Similarly, previous studies examining magnetic anomaly signatures across magma-poor continental margins have also illustrated inconsistencies between magnetic peaks and the transition from continental crust into oceanic crust. At offshore Newfoundland, Funck et al. (2003) found seafloor spreading anomaly M3 to be clearly above thinned continental crust rather than oceanic crust (Minshull, 2009). At the Southern Guinea Plateau margin, additional problems in identifying the OCT arise due to almost 40 Myr of no magnetic reversals (121 – 83 Ma; Granot et al., 2012), known as the Cretaceous Normal period (a.k.a. Cretaceous Superchron or C34), which creates additional uncertainty in resolving accurate ages of initial oceanic crust formation via plate-

reconstruction techniques. For this reason, we believe the OCT of the Southern Guinea Plateau margin is best determined using the combination of seismic reflection interpretation and gravity modeling presented in this paper.

7. Discussion

7.1 Oblique Tectonics of the Southern Guinea Plateau

Interpretation of high-quality 2-D and 3-D seismic reflection data presented in this paper provides insight into a region of the previously mostly unexplored Southern Guinea Plateau. In particular, the 3-D data image distinctive rift-related listric-style faults and syn- and post-rift volcanism found along the Guinea Plateau's south-eastern margin (Fig. 3). To the west of the 3-D surveys, 2-D seismic reflection data image the transtensional tectonics defined by the pronounced E–W trend of the continental shelf-slope margin. When combined with interpretations presented on the westernmost tip of the Guinea Plateau by Benkhelil et al. (1995), it is possible to develop a more complete understanding of the rift tectonics of the Southern Guinea Plateau (Fig. 9).

Oceanic crustal ages provide strong evidence that the Western Guinea Plateau margin of the pre-Cretaceous Guinea-Demerara conjugate plateau formed as a result of the Jurassic rift event (Jones and Mgbatogu, 1982; Klitgord and Schouten, 1986; Davison, 2005; Reuber et al., 2016); however, its effect on the Southern Guinea Plateau margin has only been the subject of speculation due to a lack of seismic resolution of the structure within Jurassic strata. As such, Benkhelil et al. (1995) presumed a prominent N10°W-trending fault zone (Fig. 9) as likely being a reactivated structure related to Jurassic rifting. Likewise, in our 2-D data, it is also difficult, if not impossible, to accurately resolve consistent patterns of faulting beneath the Jurassic unconformity. However, in the 3-D data, mapping of Jurassic and Neocomian faults via seismic

coherency and curvature attributes has resolved orientations of $0^\circ \pm 5^\circ$ for geologic sections of both ages. This suggests that the prevailing tensional stress regime of the Jurassic penetrated, to some degree, to the location of our 3-D surveys. Therefore, the mapped NNW–SSE-oriented faults punctuating the Neocomian section of Line 2D-2 (Fig. 6) are also likely related to crustal weaknesses generated by the opening of the Central Atlantic Ocean. It is probably through long-lived subsidence along these faults that the pre-rift (i.e. pre-Albian) rock units westward of 17.5°W were able to accumulate such pronounced thicknesses (Figs. 6 and 10).

During Aptian time, as the Austral and Central Segments of the South Atlantic had already begun to open via stepwise northward-propagation, the Guinea-Demerara conjugate plateau began to split apart in two distinct motions: (1) from 14.5°W – 15.5°W , the plateaus rifted divergently, with only minor rift-onset volcanism, extending in a WSW–ENE direction. Listric faults and half-grabens formed in this narrow zone, and significant thicknesses of clastic strata accumulated (up to thicknesses of 4 km) within the Baraka basin. (2) West of 15.5°W , the plateaus slid past one another along the Demerara-Guinea Transform, the continental continuation of the Guinea Fracture Zone, with its location probably along a zone of pre-existing Pan-African weakness (Benkhelil et al., 1995). The motion was transtensional in nature, as evidenced by the synthetic and antithetic dip-directions of the listric faults to the east, the normal offsets of syn-rift units across the Demerara-Guinea Transform (Fig. 5), and the negative flower structure orientations observed in line 2D-3 (Fig. 7). Volumes of syn-rift deposits decrease towards the west, as the offset of pre-rift Aptian strata diminishes across faults (Fig. 8). In Albian time, oceanic accretion began between the two plateaus, and a through-going ridge-transform-ridge system was fully developed by 105–107 Ma (Basile et al., 2013; Edge, 2014). Consequently, several Albian volcanics with an emplacement date of around 106 Ma were deposited on the shelf-edge and slope of the divergently rifted section of the Guinea Plateau. Along the transtensional segment of the Guinea Plateau, we interpret that the oceanic accretion axis progressively drifted

westwards along the Demerara-Guinea Transform, causing a post-rift unconformity that post-dates formation of oceanic crust along the transform margin (Basile et al., 2013). Therefore, Albian-aged units persist above oceanic crust across the transform margin up until the oceanic accretion axis had drifted beyond the continental margin (Fig. 5).

7.2 Asymmetrical rifting

A direct comparison of upper-crustal deformation between the Demerara and Guinea conjugate plateaus reveals several differences. Fig. 12 shows a summary of Cretaceous rift-related deformations restored to their positions during the late-Albian (100 Ma). Faulting on the eastern margin of the Demerara Plateau is from Basile et al. (2013). Neither the extent nor the amount of Aptian-Albian volcanism found on the Guinea Plateau has been reported to date on the conjugate Demerara Plateau. Basile et al. (2013) report roughly N-S-trending normal faults along the eastern edge of the Demerara Plateau, backed by inverted normal faults with a WNW–ESE orientation. Conversely, on the southeastern edge of the Guinea Plateau, we find massive NW–SE-oriented listric faults with no evidence of inversion tectonics. However, inversion structures were found to the northwest by Benkheilil et al. (1995), and both groups of authors attribute local compressional episodes to a plate kinematic reorganization in the late Albian.

The reasons for the apparent asymmetry between the conjugate margins are unclear. Even if some of the observed differences in faulting and volcanism are a matter related to discrepancies in data resolution and penetration depths, it is unlikely such profound contrasts would exist artificially. In fact, varying degrees of asymmetry between conjugate rift margins are not entirely unusual. Favre and Stampfli (1992) documented asymmetries at the Gulf of Suez, the Red Sea, and the Central Atlantic rift system along Morocco and the North American coast. The authors found disparities in crustal deformation in each rift setting, from pre-oceanic rifts (Gulf of

Suez) to rift zones with fully developed oceanic accretion axes (Red Sea and Central Atlantic). For each set of conjugate margins, Favre and Stampfli (1992) attributed the asymmetry as a function of simple shear models with upper and lower plate settings (Wernicke, 1985). According to Wernicke (1985), lower plate margins are characterized by large tilted blocks that sole out onto a mid-to-lower crustal detachment surface, while conjugate upper plate margins are dominantly flexural and consists of planar-style extensional faults and large-scale plateau-basalt volcanism. As such, the rift models of Favre and Stampfli (1992) imply uniform simple-shear-type extension in the upper crust contrasted with dominantly pure-shear extension at deeper crustal and mantle-lithosphere levels.

Asymmetrical rifting between the Demerara-Guinea conjugate plateaus is an attractive model based on our observations; however, there are inconsistencies between the Demerara-Guinea rift system and those described by Favre and Stampfli (1992). The listric fault blocks observed on the Guinea margin imply its configuration is a lower plate margin, with the planar-style extension observed on the Demerara margin implying upper-plate-margin deformation. However, the issue of observed volcanism on the Guinea margin, and an apparent lack of significant volcanism on the Demerara margin, is contrary to Wernicke's (1985) simple-shear rifting models. In general, the upper plate is expected to experience volcanism due to greater attenuation of the lower lithosphere (i.e. thinner lower crust). However, interpreted volumes of volcanics along the Guinea margin (Kardell, 2016) are volumetrically small compared to magma-rich margins (Franke, 2013), as the predicted amounts due to rift-onset volcanism are insignificant, and instead much of the volcanism occurred near the time of crustal separation. In addition, crustal strength along the Guinea margin may already have been compromised by documented dike swarms during the early Jurassic (Beckinsale et al., 1977).

Additional questions arise from the narrow width of stretched continental crust along the divergently rifted Guinea margin (Fig. 11 - Profile B). Listric-style extension appears to occur over a ~40 km zone from the continental shelf edge to oceanic crust. Meanwhile, the eastern Demerara Plateau margin appears to also contain a margin-width of ~40–50 km (see Basile et al. (2013)). The aforementioned asymmetrical margins of the Red Sea, Gulf of Suez, and Central Atlantic consist of margin widths ranging from 100–200 km. Likewise, the extensively studied non-volcanic Galicia margin also contains ≥ 100 km of stretched continental and/or transitional crust (Reston and McDermott, 2014). Model experiments show that the width of a rift zone depends on crustal thickness and the depth at which the brittle/ductile transition zone resides at the onset of lithospheric extension (Ziegler and Cloetingh, 2004). Therefore, the narrow width of extension along the divergent segments of the Demerara-Guinea plateau margins may have been a function of below-average crustal thickness and a shallow brittle/ductile transition, perhaps influenced from thermal perturbations resulting from Jurassic rifting of the Central Atlantic. At any rate, without proper seismic refraction and reflection lines that illuminate the differences in upper- and lower-crustal structure across the two divergently rifted margins, we can only speculate at the degree of asymmetry between the margins. Although upper-crustal asymmetries are apparent, further examinations of their crust-mantle boundaries will be decisive in generating an appropriate cross-sectional rift model.

7.3 Extension discrepancy

The amount of extension measured from offsets along upper-crustal faults (i.e. β_f) at many passive margins is less than the amount of extension measured from crustal thinning (β_c) and subsidence (β) (Kusznir and Karner, 2007; Reston and McDermott, 2014). Edge (2014) measured the amount of Cretaceous extension along 54 2-D seismic profiles that traverse the Southern Guinea Plateau margin. His results produced an average extension factor of $\beta_f = 1.11$ throughout

the margin, with a maximum value of $\beta_f = 1.39$ occurring across the Baraka half graben. His average β_f seems quite low for a margin that has extended to the point of continental separation and formation of an oceanic accretion axis. We believe this is in part due to the Guinea margin having been influenced by at least two stages of extension, both of which need to be considered when analyzing the total amount of stretching. In our subsidence results (Fig. 10), we see that thermal relaxation from the Jurassic rift event is imprinted on all of the tectonic subsidence curves. Therefore, by only measuring the offset of the Aptian rift-onset horizon, the subsidence due to Jurassic extension is mistakenly ignored.

Other proposed reasons for underestimates of extension measured from upper-crustal faulting include insufficient seismic resolution at depth causing significant subseismic deformation to be excluded (Reston and McDermott, 2014). These authors found that subseismic distributed deformation can account for $30 \pm 17\%$ of the total extension. In other words, up to almost 50% of the margin extension can be accommodated through faults too small to be resolved seismically. If we apply the end-member case of 47% subseismic deformation to the β_f -factors of the Guinea Plateau, the maximum extension measured across the continental slope becomes $\beta_f = 2.62$, resulting in a final crustal thickness ranging from 11.9 – 13.4 km (assuming an initial crustal thickness of 31.2 – 35 km). This amount of extension is more than enough to compensate for the discrepancies between β_f -factors computed from faults and β -factors computed from subsidence analysis (Fig. 10); however, in the deepwater segment of the margin, near the OCT, neither of the β_f - nor β -factors are large enough to account for the actual crustal thickness resolved through gravity modeling (see Fig. 11; Profile B). At the PW-1 pseudo-well location, the crust is stretched by $\beta_c = 3.21 - 3.60$, and up to $\beta_c = 4.66 - 5.23$ at the OCT. This suggests that rifting at the Southern Guinea Plateau may have had some component of depth-dependent stretching (DDS; Royden and Keen, 1980) in which the extension of the whole continental crust and lithospheric mantle greatly exceeds that of the upper crust (Kusznir and Gardner, 2007). Kusznir and Gardner

(2007) have documented that extension and thinning of a large number of non-volcanic rifted margins is partitioned with depth; however, it is unclear how this partitioning with depth affects asymmetrical margins because it predicts ‘upper plate’ behaviour on both conjugate margins. This requirement is inconsistent with the distinction of both upper and lower plate margins for the rift models outlined in Wernicke (1985) and Favre and Stampfli (1992). Therefore, due to conflicting implications between DDS and asymmetrical rift models, we hypothesize that DDS accounts for only a fraction of the observed extension discrepancy of the Southern Guinea Plateau rift margin. Instead, the extension discrepancy is likely the result of a combination of DDS, subseismic faulting, and polyphase (multi-rift) faulting.

8. Conclusions

Interpretation of new, high-quality 2-D and 3-D seismic reflection data located across the southern margin of the Guinea Plateau provides insight into the plateau’s structural evolution. Within the 3-D data, we resolve minimal rift-onset volcanism, large-scale listric normal faulting, a pulse of volcanism at the end of rifting during onset of oceanic-crust formation, and varied amounts of post-rift subsidence dependent on location along the margin. In addition to interpreted vertical profiles, we present 3-D maps of the Aptian rift-onset unconformity and Albian breakup unconformity that emphasize their spatiotemporal volcanic features as well as the notably narrow width of the divergently-rifted section of the Southern Guinea Plateau margin. Interpretations made in the 3-D data are carried over to regional 2-D profiles that image the E–W-striking section of the margin. Here, a shift in tectonics is inferred from divergent rifting in the east, to transtensional separation along the Demerara-Guinea Transform in the west. We integrate our results with previous seismic exploration efforts and produce a structural map that shows that the Southern Guinea Plateau margin is defined by oblique tectonics, with a dramatic decrease in Cretaceous syn-rift sedimentation west of the 3-D surveys.

624

625 We supplement our seismic interpretations with computed subsidence curves from four
626 locations along the plateau, as well as two crustal-scale gravity models that traverse from the
627 continental shelf out across the ocean-continent transition. We draw comparisons between our
628 results for the Guinea Plateau and published results on the conjugate rifted Demerara Plateau
629 and make at least three main points:

- 630 1. Backstripping and subsidence analyses of the Southern Guinea Plateau reveal the
631 influence of a multi-rift history on its evolution. Subsidence through thermal relaxation of
632 the Jurassic rift event that opened the adjacent Central Atlantic is greatest towards the
633 western margin of the Guinea Plateau, but its signal is also seen in pre-Cretaceous rift
634 strata of the Equatorial Atlantic margin, probably along buried N–S Jurassic faults created
635 during the E–W extension of the Jurassic rift event. Both events should be considered
636 when tracking the evolution of the Southern Guinea Plateau margin.
- 637 2. When restored to late-Albian times, apparent upper crustal asymmetry is found between
638 the conjugate rift margins. The divergently rifted segment of the Southern Guinea Plateau
639 is defined by large-scale listric faulting interpreted to sole out into basement rocks, as well
640 as conspicuous extrusive volcanism emplaced during late-Albian separation of the
641 margins. This is in contrast to planar-style extension, limited volcanism, and inversion
642 episodes reported on the conjugate divergently rifted Demerara Plateau. Therefore,
643 asymmetrical rifting may have dominated during early crustal separation between the
644 conjugate margins.
- 645 3. Apparent discrepancies exist between the amounts of extension measured across upper-
646 crustal faults in the seismic reflection data, extension factors for the crust and mantle
647 lithosphere resolved from subsidence data, and crustal thickness variations determined
648 from gravity modeling. The upper crustal extension discrepancy along the continental shelf
649 and upper slope can be explained through a combination of subseismic deformation and

polyphase (multi-rift) faulting; however, beneath the deepwater continental rise and slope, near the ocean-continent transition, these factors are still not enough to account for extended crustal thicknesses resolved in the gravity models. In addition, extension factors computed by fitting a uniform stretching model to the backstripping data acceptably match crustal thicknesses up to the deepwater regions. Near the ocean-continent transition, extension of the whole continental crust and lithospheric mantle greatly exceeds that of the upper crust; therefore, the Southern Guinea Plateau margin may provide evidence of some degree of depth-dependent stretching during rifting.

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FIGURES

Fig. 1 (Top): Present-day geodynamic setting of equatorial segment of the Southern Atlantic Ocean. The Guinea Plateau (G.P.) and Demerara Plateau (D.P.) are outlined along West Africa and South America, respectively. Fracture zones (FZs; light-gray lines) and the Mid-Atlantic Ridge (red line), modified from Edge (2014), show the paths along which the two conjugate plateaus have been separating since the Cretaceous. The Marathon FZ represents the dividing line between the (Jurassic) Central Atlantic Ocean to the north and the (Cretaceous) Southern Atlantic Ocean to the south. (Bottom) Contoured bathymetric map of the Guinea Plateau showing locations of seismic data used in this study. Red boxes outline the two 3-D surveys, A & C, while red lines highlight the 2-D seismic lines shown in this paper. Purple circles show the two exploration wells, GU-2B-1 and SABU-1, the yellow circles, PW-1 and PW-2, are pseudo-well locations used in subsidence analysis discussed in the text (5. Backstripping and Tectonic Subsidence). Inset map in bottom-left shows the named sections/segments of the South Atlantic Ocean. The segments are divided by prominent FZs. Map shown at plate-reconstruction time of

845 Chron-34 (83 Ma) as presented in Moulin et al. (2010) and emphasizes the north-ward
846 propagation of the Southern Atlantic Ocean rifting event.
847

Fig.2 (Right): Generalized stratigraphic section of the Southern Guinea Plateau based on seismic and well data. (Left) Arbitrary seismic profile taken through both 3-D surveys. Naming convention of rift-timing in relation to stratigraphic section is shown to the left of the stratigraphic column. Unless explicitly stated elsewhere, pre-rift strata refer to all strata deposited prior to the Cretaceous rifting event. The Jurassic, Aptian, Albian, Cretaceous, and Oligocene horizons are typically manifested as unconformities, while the Neocomian (brown) is a conformable horizon representing the top of Neocomian carbonates.

Fig. 3: Four interpreted seismic profiles from the 3-D surveys, A & C. Inset map of line locations shown in bottom-right corner of Line 3D-1. Lines 3D-1 and 3D-2 are inlines oriented roughly parallel to the strike of the shelf margin. Lines 3D-3 and 3D-4 are crosslines that provide a dip-direction perspective of the continental shelf-slope-rise margin. Horizon/unconformity colors are consistent with the stratigraphy presented in Fig. 2. The semi-transparent green lines beneath the Aptian ROU in Lines 3D-1 and 3D-4 highlight the top-lapping angular unconformity characteristic of rift-onset unconformities (Franke, 2013). The red-shaded polygons represent interpreted Albian volcanics.

Fig 4. Horizon maps and perspective views (V.E. = 3:1) of the Aptian (bottom) and Albian (top) unconformities interpreted in the 3-D datasets and colored by depth beneath sea level. Interpreted volcanics differentiated as either Aptian or Albian shown as colored polygons on their respective horizon map. A massive volcano is outlined in yellow on both maps, as it is interpreted as having been active throughout many successive periods of geologic time on the basis of its seismic, gravity, and magnetic signatures (see Fig. 3, Line 3D-1 for seismic example). The 3-D perspective of the Aptian ROU emphasizes the geometry of faulting caused by the Cretaceous rift event; meanwhile, the 3-D perspective of the Albian BU emphasizes several paleo-seamounts emplaced at the end of rifting and subsequent onset of oceanic spreading (~106 Ma).

Fig. 5: (Top) Uninterpreted seismic profile 2D-1 (Fig. 1). The interpreted section (Bottom) uses the same color conventions as presented in Fig. 2, with the addition of oceanic crust (chaotic reflectors; dark green). Note the narrow extent of extension-related faulting. Also notice the reduced offset of pre- and syn-rift strata across the Baraka fault. Significant offset of the Aptian and Albian horizons is seen across the Demerara-Guinea Transform fault, showing the transition from a divergent regime (listric faults) into an oblique regime (transtensional faults) along this region of the Southern Guinea Plateau margin. Albian syn-rift fill is interpreted to persist from continental crust onto highly-fractured nascent oceanic crust. This is a result of BU complexities associated with continental transform margins, where newly formed oceanic crust drifts along the transform margin during the initial stages of oceanic accretion (Masle and Blarez, 1987; Basile et al., 2013). COT: Continent-Ocean-Transition (V.E. = 4:1).

Fig. 6: Seismic profile 2D-2. Location of this E–W profile is ~70 km west of the 3-D surveys. Notice the extreme increase in unit thicknesses between the Neocomian and Jurassic horizons, contrasted by very thin “syn-rift” fill between the Albian and Aptian horizons. This suggests that this region of the Guinea Plateau was more affected by the Jurassic rifting event of the Central Atlantic, causing significant subsidence of Jurassic (and older) sequences, ultimately allowing large accumulations of carbonates to develop. In contrast, the Cretaceous rift event resulted in little-to-no syn-rift deposition at this location on the Guinea Plateau. Instead, it may have caused reactivation and/or fracturing and transform faulting to occur in Aptian – Neocomian units (dashed fault lines). Note that the western edge of the seismic data is not interpreted due to limited seismic fold in addition to edge effects resulting from its location near the boundary of the seismic velocity model used to convert 2-D seismic lines from time to depth.

Fig. 7: Seismic profile 2D-3. Note the profile is a time section. Relative horizon thicknesses appear similar to line 2D-2, however the Aptian and Albian unconformities are roughly coincident, with no Cretaceous syn-rift fill. The faulting patterns appear to be similar to “flower structures” found preserved at other transform margins and boundaries (e.g. Mascle et al., 1997). The faulting is slightly transtensional as evidence of a concave-upwards geometry of the Jurassic and Neocomian horizons.

Fig. 8: Regional horizon maps for the Aptian and Albian unconformities along the Southern Guinea Plateau. Horizon picks were extended beyond the 3-D Survey boundaries (gray rectangles) through regional mapping on 84 depth-converted 2-D profiles (Fig. 1). The Baraka fault is interpreted to persist ~40 km to the WNW of Survey C (dashed black line); however, offset of the Aptian unconformity dramatically diminishes. Conversely, the nearly east-west transform fault (i.e. Demerara-Guinea Transform) south of the Baraka fault, interpreted to be a marker of intracontinental rifting between the Demerara and Guinea Plateaus, continues to offset and down-drop the Aptian and Albian unconformities for ~100 km to the west.

Fig. 9: Structural map combining interpretations using our data (black lines, tan and gray polygons) and interpretations presented in Benkheilil et al. (1995), shown in blue. Our interpretations focus on the divergent, listric-style deformation and volcanism that dominates the south-eastern margin of the Guinea Plateau. West of the listric faulting, transtensional faulting and a decrease in rift-related deposition and volcanism transition into a large transcurrent fault zone and inversion as interpreted by Benkheilil et al. (1995). This evident transition in deformational structures emphasizes the progression of the obliquely rifted Southern Guinea Plateau margin that manifested during the mid-Cretaceous.

Fig. 10: Comparison of backstripping curves computed at four locations along the Southern Guinea Plateau in the first column (two wells and two pseudo-wells; Fig. 1) and four locations along the Eastern and Northern Demerara Plateau (locations shown in inset maps and

approximate coordinates given after pseudo-well names). The subsidence data for the Demerara Plateau was computed using unit ages and thicknesses inferred from interpreted seismic sections presented in Greenroyd et al. (2007, 2008) and Basile et al. (2013); seismic methodology also used for PW-1 and PW-2 for the Guinea Plateau. The dotted green line represents the total decompacted sediment thickness, while the solid black line shows the tectonic subsidence computed from backstripping. The red (Guinea Plateau) and blue (Demerara Plateau) error bars represent the amount of uncertainty in the tectonic subsidence at each time interval (see text). The Cretaceous rifting event of the South Atlantic is highlighted by gray boxes. Best-fit curves of a uniform stretching model with crust and mantle extension, β , for each period of rifting are shown as dashed black lines. Note for DP4, that the stretching-model curve for the Cretaceous rift event does not include data points from the Miocene – present, as an influx of sediment from the Amazon basin appears to be generating accelerated subsidence. Numbers in parentheses after well name indicate water-bottom depth below sea level.

Fig. 11: Forward-modeled 2-D gravity profiles computed using the Talwani et al. (1959) method. Free-air gravity data (top-left map) derived from Sandwell et al. (2014), magnetic data (top-right map) derived from Maus et al. (2009). Both profiles A & B are taken along 2-D seismic reflection profiles; upper crustal (≤ 10 km) geometries and densities are constrained by seismic interpretation. Lower crust and mantle densities are average values based on crustal-scale velocities and densities. Lower crust/mantle boundary depths are determined through gravity modeling, with the constraint that they cannot deviate dramatically from a crustal model offshore nearby Sierra Leone computed using seismic refraction constraints and presented in Jones and Mgbatogu (1982). On profile B, crustal thicknesses derived from stretching factors (β) are shown assuming an initial crustal thickness ranging from 31.2 km – 35.0 km. Note the mismatch between the gravity-derived results at PW-1, perhaps implying a uniform stretching model is not a perfectly valid assumption for the rifted Southern Guinea Plateau. GU-1= GU-2B-1 and SB-1 = SABU-1. Profile A is from Line 2D-1 (Fig. 5). Profile B 2-D seismic line not shown because it is nearly coincident with line 3D-3 shown in Fig. 3. Numbers in brackets after well names represent distance from well location.

Fig. 12: Late Albian (100 Ma) plate-reconstruction diagram showing interpreted deformation associated with Cretaceous rifting of the Demerara-Guinea Plateau. Plate-reconstruction diagram modified from Edge (2014) and Reuber et al. (2016). Demerara Plateau interpretations are from Basile et al. (2013), and are representative of the conjugate segment of the Demerara Plateau to the 3-D seismic data of the Guinea Plateau. Notice the distinct asymmetry in upper-crustal faulting and volcanism between the conjugate margins. No large-scale listric faults and volcanics were interpreted by Basile et al. (2013), while those characteristics are what define the south-eastern Guinea Plateau. This degree of asymmetry is suggestive of simple-shear extension during Cretaceous rifting in order to describe the discrepancy in faulting and volcanism between the two conjugate margins.