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Flexural unloading and uplift along the Côte d'Ivoire-Ghana Transform Margin, equatorial Atlantic

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Abstract. Recent Ocean Drilling Program sampling of the Côte d'Ivoire-Ghana margin of West Africa provides for the first time the opportunity to study the development of a marginal ridge that formed along a sheared passive margin adjacent to the continent-ocean transition after the end of intracontinental wrenching. We model its evolution using a two dimensional flexural backstripping technique. The model is constrained by existing seismic refraction and reflection data on the crustal structure and stratigraphy and paleobathymetric evidence from the cores. Following rifting at 120 Ma (Aptian), intracontinental wrenching continued until ~105 Ma (mid Albian), when South America and Africa separated along this transform. While the possible presence of thicker crust under the Marginal Ridge compared to the Deep Ivorian Basin may explain some of the ridge's topography, the entire uplift can be readily modeled as a flexural response to unloading along a shallow-dipping (~25°) fault at the time of continental separation. Forward subsidence modeling of the Marginal Ridge suggests that an effective elastic thickness of 2.5 km at 105 Ma is most appropriate to match the observed structure. Conduction of heat from the oceanic plate across the continent-ocean transition drove temporary uplift of at least 1200 m peaking just before 89 Ma, when sedimentary data show deposition within the photic zone (0–50 m). This study shows that significant, transient thermal uplift can be found in sheared continental passive margin settings even where magmatism is insignificant.

1. Introduction

During continental breakup, divergent margins are affected by extension and shearing. Extensional passive margins are often typified by broad zones of thinned crust (100–200 km) close to the continent-ocean transition, at least in the nonvolcanic case [e.g., *Montadert et al.*, 1979; *Boillot et al.*, 1988], whereas sheared passive margins show sharp (<30 km) continent-ocean transitions, with deformation focused close to the former plate boundary [e.g., *Scrutton*, 1979; *Mascle et al.*, 1995]. Although divergent passive margins have been studied extensively using both geophysical and geological methods, notably drilling by the Deep Sea Drilling Project (DSDP) and the Ocean Drilling Program (ODP) [e.g., *Montadert et al.*, 1979; *Sawyer et al.*, 1994], the database for sheared passive margins is far less comprehensive. Drilling by ODP Leg 159 [*Mascle et al.*, 1996] on the Côte d'Ivoire-Ghana margin of West Africa (Figure 1) now provides the first opportunity to study directly the tectonic evolution of a classic transform passive margin.

The Côte d'Ivoire-Ghana margin lies at the eastern end of the Romanche Fracture Zone, which accommodates ~950 km of offset, making this one of the largest such boundaries on Earth. The margin is marked by a substantial basement ridge standing ~1900 m above the adjacent rifted continental crust along the continental (northern) side of the Côte d'Ivoire-Ghana margin. This ridge has been interpreted as a relict of the thermal rejuvenation of the plate margin associated with passage of a seafloor spreading center, a hypothesis principally supported by multichannel seismic data showing that the Marginal Ridge was created close to the time of spreading ridge intersection [*Basile et al.*, 1993]. Alternatively, uplift along the margin could be related to magmatic underplating during ridge passage (see Exmouth Plateau [*Lorenzo and Vera*, 1992]). *Clift et al.* [1997] noted that the wavelength of the uplift was compatible with a flexural origin and pointed out that since the spreading ridge passage occurred during the Early Cretaceous, any thermal anomaly would have long decayed.

In this paper we use new drilling-derived constraints on the vertical motions of the Marginal Ridge, in conjunction with seismic and gravity data, to examine the influence that flexural unloading across the continent-ocean transition may have on the form and magnitude of the Marginal Ridge during its formation, prior to coupling to the oceanic crust of the Gulf of Guinea. We also assess the degree of thermal rejuvenation during ridge-transform intersection. We use a flexural cantilever model for continental extension to examine the development of the Marginal Ridge. Application of this type of model to ancient rift systems, such as the North Sea, Jeanne d'Arc Basin and East Africa, has achieved good matches of the rift architecture and sedimentary fill and allowed the lateral variability in extension to be quantified [e.g., *Hendrie et al.*, 1991; *Kusznir et al.*, 1995]. This study builds and substantially improves upon an earlier study of the Côte d'Ivoire-Ghana margin in which *Clift et al.* [1997] estimated

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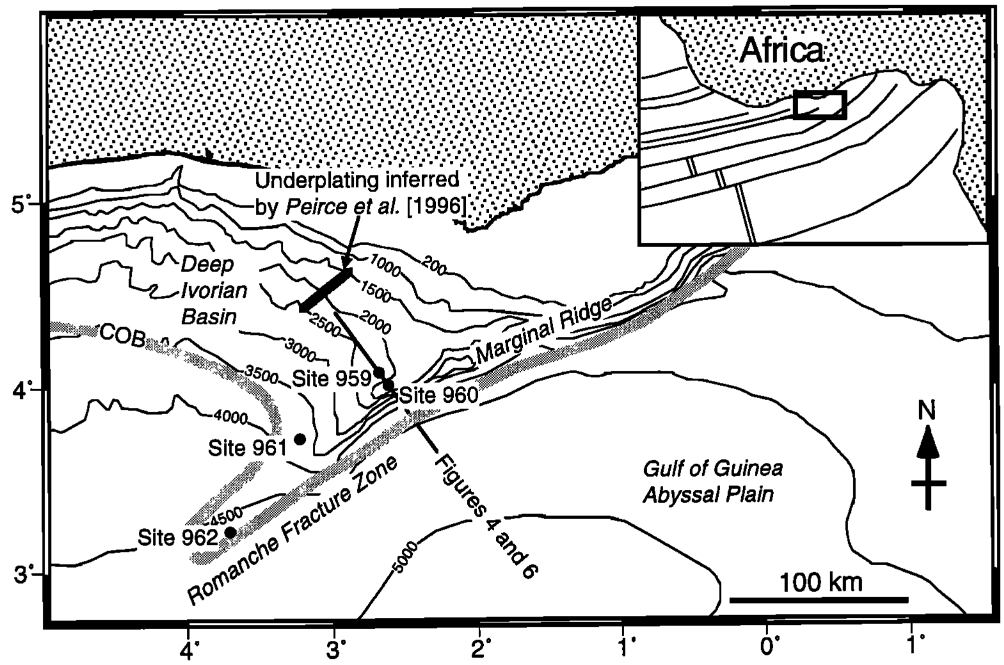

Figure 1. Location map showing the Côte d'Ivoire-Ghana margin and the position of Ocean Drilling (ODP) Sites 959–962 along the crest of the marginal ridge. COB is continent-ocean boundary. Gulf of Guinea is presumed oceanic crust. COB is poorly constrained in the Deep Ivorian Basin but is probably east of the COB along the ridge, where it is constrained by the drill sites, because of the west directed shearing of material along the transform.

the impact of the flexural strength of the continental lithosphere along this margin.

2. Transform Margin Tectonics

Sheared passive continental margins differ from rifted passive continental margins in the style and duration of brittle faulting, as well as in their thermal evolution during continental separation. High-standing continental marginal ridges, 50–100 km wide and bounding deep sedimentary basins are often formed in such settings. Kinematic models for the evolution of sheared continental margins discern three phases that distinctly influence their tectonic and sedimentary development (Figure 2) [*Mascle and Blarez*, 1987]. During the first phase (Figure 2) intracontinental wrenching dominates along the future transform zone. Cross structures known as accommodation zones (areas of complex shearing and folding) [*Rosendahl*, 1987; *Bosworth*, 1987] and large-scale transfer zones [*Gibbs*, 1984; *Etheridge et al.*, 1985; *Tankard and Welsink*, 1987] may represent the features that develop into marginal ridges if they experience relatively lower crustal extension than the surrounding areas or are preferential sites of magmatism and hence underplating. It is noteworthy that the geometry and kinematics of the accommodation zones evolve with fault propagation and continued extension. *Hayward and Ebinger* [1996] demonstrated that the along-axis segmentation of the Afar rift is controlled by the degree of extension and heatflow. Individual segments shorten with increasing extension.

Wrenching may be accompanied by either extension, which generates pull-apart basins [*Burchfiel and Stewart*, 1966], or compression, which creates push-up ranges (e.g., San Andreas) [*Crowell*, 1974] and thickens the transform boundary. In the rifted margin setting only extensional faulting occurs in the rifted margins adjacent to the transform boundary. Both extensional and transform deformation are important at the intersection between the transform offsets and thinned continental crust.

During the second phase of margin development, transform motion between the continental and oceanic plates (Figure 2) culminates in the passage of an oceanic spreading ridge along the margin. Flexural unloading of a seaward-dipping transform fault, in a manner analogous to footwall uplift in extensional settings, provides an alternative method for producing permanent uplift close to the continent-ocean transition. *Clift et al.* [1997] note that the wavelength of the Côte d'Ivoire-Ghana Marginal Ridge is consistent with an unloading origin and suggest that this accounts for ~75% of the total height of the Marginal Ridge above the floor of the adjacent Deep Ivorian Basin. The effectiveness of this mechanism depends on the magnitude of the seaward dip of the fault that unloads the margin and the flexural rigidity of the continental plate, which may be low shortly after rifting [*Karner and Watts*, 1982; *Holt and Stern*, 1991; *Kooi et al.*, 1992]. Recently, *Watts and Stewart* [1998] used three-dimensional models to calculate low effective elastic thicknesses (Te <10 km) for the Gabon margin during continental breakup. Relatively high flexural rigidity and steep angles of dip of the fault will tend to increase the amount of uplift during unloading, provided that the heave (apparent horizontal displacement) on the fault remains constant. Local subsidence is predicted when the fault dips landward.

The juxtaposition of increasingly younger oceanic lithosphere against the sheared continental margin results in

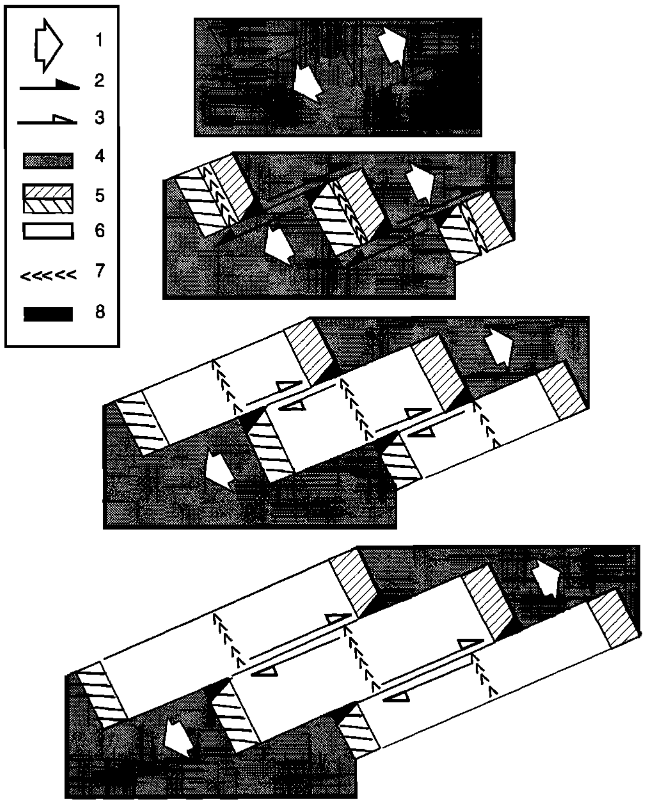

Figure 2. Schematic evolution of a sheared passive margin after *Mascle et al.* [1996], showing how crust in the position of the ODP drill sites experiences both extension related to the adjacent rift, and shearing related to the transform offset: 1, divergence; 2 and 3, transform motion between continental and oceanic crust respectively; 4, continental crust; 5, thinned continental crust; 6, oceanic crust; 7, ridge axis; and 8, Marginal Ridge.

temporary uplift due to lateral heat conduction from the oceanic crust lithosphere [e.g., *Todd and Keen*, 1989]. Thermal expansion of the crust and thermal erosion of the mantle lithosphere effectively reset the tectonic age of the plate. Uplift can be as much as ~2500 m at the time of ridge-margin juxtaposition when the transform offset is large (900 km) [*Todd and Keen*, 1989]. Such uplift is suggested as causing the large erosional unconformities seen along transform margins in several places (e.g., Newfoundland) [*Todd et al.*, 1988] and the Barents Sea [*Faleide et al.*, 1993; *Vågnes*, 1997]. *Todd and Keen* [1989] predict only a modest uplift (<150 m) during intracontinental wrenching due to frictional heating along the fault zone, although this process is poorly understood.

The intersection of spreading ridge and transform margin marks the start of the final, passive phase of evolution in which both oceanic and continental plates subside because of cooling and thickening of their respective lithospheres (Figure 2). Flexural coupling between the plates at this stage can result in enhanced subsidence of the continental margin, because of the more rapid subsidence of the oceanic plate [*Lorenzo and Wessel*, 1997; *Sandwell and Schubert*, 1982]. This process will tend to reduce the magnitude of a marginal ridge since the downward flexure is strongest closest to the continent-ocean boundary. *Clift et al.* [1997] estimate the modern flexural downwarping of the Côte d'Ivoire-Ghana margin could be as much as 650 m in the vicinity of the ODP drill sites.

3. Geologic Background

The Côte d'Ivoire-Ghana sheared passive margin was formed in an area previously affected by Late Proterozoic Pan-African deformation, although the cratonic blocks separated by Pan-African greenstone belts may be Archean [*Shackleton*, 1986]. Pan-African deformation is generally considered to predate 550 Ma [*Castaing et al.*, 1993], and since the area was tectonically inactive for a long period of time prior to breakup, this means that the continental lithosphere would have been thermally equilibrated by this time. The age of continental breakup itself and the initiation of seafloor spreading are not well constrained because of the lack of marine magnetic anomalies on the adjacent seafloor [*Rabinowitz and LaBrecque*, 1979]. Regional plate reconstructions using magnetic anomalies in the southern and central Atlantic, together with continental paleomagnetic constraints show that breakup began in the Early Cretaceous (post 140 Ma) [*LePichon and Hayes*, 1971; *Rabinowitz and LaBrecque*, 1979], although more recently an age of ~120 Ma (early Aptian) has been proposed for breakup in this region [*Klitgord and Schouten*, 1986]. Biostratigraphic data from the onshore Benue Trough of Nigeria and Mayo Oulo Lere Basin of Cameroon [*Brunet et al.*, 1988] confirm that the Barremian (121–127 Ma) [*Gradstein et al.*, 1995] was the time at which rift-related sedimentation started. For the purpose of modeling the vertical motions of the margin we choose an age of 120 Ma (early Aptian) for the start of rifting, in agreement with *Wilson and Guiraud's* [1992] estimate for the start of rifting and strike-slip motion in the Benue Trough.

North of the Côte d'Ivoire-Ghana margin, continental crust thins progressively toward the west [*Sage*, 1994] with a passage to fully oceanic crust just west of the ODP Leg 159 sites. A transect from north to south through the region of the ODP drill sites runs from rifted continental crust of the Deep Ivorian Basin (Figure 1) into fully oceanic crust of the Gulf of Guinea. It is important to note that the oceanic crust in the Gulf of Guinea dates from the time of ridge-transform intersection, whereas the rift-related deformation dates to the earlier phase of continental breakup.

The Côte d'Ivoire-Ghana margin, like many transform continental boundaries, is marked by the presence of a prominent basement ridge, termed the Marginal Ridge [*Le Pichon and Hayes*, 1971], running parallel to the continent-ocean transition (Figure 1). Seismic evidence for transpression in the form of positive flower structures is known from the Côte d'Ivoire-Ghana margin, although the degree of crustal shortening in this particular area does not appear to be great [*Basile et al.*, 1996].

Conductive thermal models predict maximum transient uplift when the spreading ridge passed along the sheared margin. Biostratigraphic data suggest that ridge passage probably occurred before the end of the Turonian (90.5–89 Ma) [*Mascle et al*, 1996]. Nannofossil fauna at ODP Site 962 at the westernmost edge of the marginal ridge date deformed uppermost Albian sediments (planktonic foraminifer *Rotalipora Appenninica* zone; 97–99.5 Ma) that were deposited prior to the end of intracontinental wrenching and so constrain the ridge-transform intersection there to be younger than 97 Ma [*Mascle et al.*, 1996]. This assumes that deformation predates final passage of the spreading ridge.

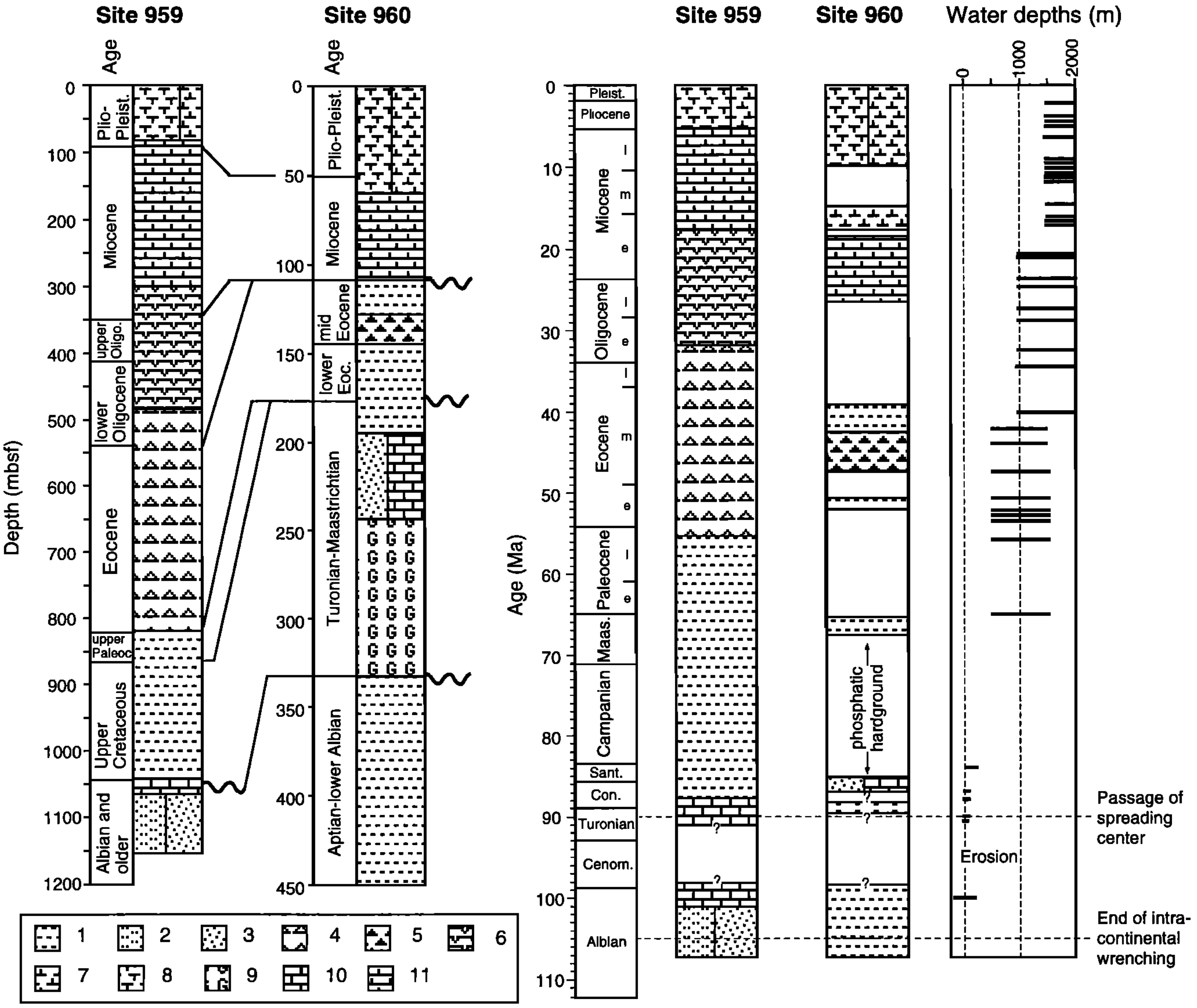

Figure 3. Schematic sedimentary logs of ODP Sites 959 and 960 showing the basic lithology and age data. 1, claystone; 2, siltstone; 3, sandstone; 4, porcellanite; 5, chert; 6, diatomite; 7, nannofossil ooze; 8, foraminiferal ooze; 9, grainstone; 10, limestone; and 11, nannofossil chalk. Wavy lines represent unconformity. Chronostratigraphic diagram showing the size of the sedimentary hiatus on the Marginal Ridge in relation to the major tectonic events to have affected the region. The range of possible paleowater depths are shown as horizontal bars and are derived from core data at both Sites 959 and 960 [*Mascle et al.*, 1996]. On the scale of the entire ridge the slight elevation difference between Sites 959 and 960 is insignificant.

Similarly, at Site 960 (Figure 3), Albian clastic sediments are separated by angular unconformity from overlying, untectonized shallow marine limestones of Turonian-Coniacian age (nannofossil zone CC13, 89 Ma) [*Gradstein et al.*, 1995], which places an upper limit of 89 Ma on the time of intersection. Calcite veins penetrating the Turonian carbonates have anomalously negative $\delta^{13}C$ and $\delta^{18}O$ isotope characteristics [*Marcano et al.*, 1998], and indicate paleotemperatures 20°–30°C above normal seawater (~2°C), postdating the deposition of the sediments at 90.5–89 Ma. As the limestones are no older than 89 Ma, the age of hot fluid flow must also be equal to or younger than 89 Ma. However, this does not mean that ridge-transform intersection postdated 89 Ma because of the lag time between ridge-transform intersection and hot fluid migration. The thermal data are compatible with ~90 Ma being the time of ridge passage at the location of Site 960. For example, assuming a thermal diffusivity of 1 mm^2s^{-1}, a pulse of heat applied at the continent-ocean transition would take 12 m.y. to penetrate 20 km into the plate or 3 m.y. to advance 10 km. The time estimate of 90 Ma for ridge-transform intersection is consistent with large scale plate reconstructions [*Klitgord and Schouten*, 1986] and the conclusion of *Bellier* [1998], who uses planktonic foraminifers to show that an open deep marine connection was not available between the Tethys and the South Atlantic until the Turonian. As the Romanche Fracture

Zone is the largest in the central Atlantic, the separation of South America and Africa along this feature would be the last major obstacle to deep water exchange.

We estimate intracontinental wrenching ceased at ~105 Ma (mid Albian) in the vicinity of ODP Sites 959 and 960, half way between rift initiation (120 Ma) and ridge passage (90 Ma). This date is also consistent with the marine sedimentation seen at Site 959. The deep water, bathyal water depths estimated at Site 962 in CC9 times (*R. Appenninica* subzone; 99–97 Ma) make it unlikely that these sediments were deposited in an intracontinental pull-apart wrench setting [*Mascle et al.*, 1996]. By assuming a constant rate of plate separation it is possible to estimate the time at any given point on the margin at which an active ocean-continent system replaces the intracontinental transform regime, being halfway between the start of rifting and spreading ridge-margin juxtaposition (Figure 2). The change from rifting to spreading will clearly affect thermal interactions across the transform by changing the relative amounts of rifted continental versus oceanic crust that passes any particular point on the margin. However, the timing of ridge passage will be unaffected.

4. Stratigraphy and Structure of the Marginal Ridge

Existing multichannel seismic data (Figure 4) from the Marginal Ridge show that postrift sediments thin with a toplap relationship on to the acoustic basement of the Marginal Ridge. Sedimentation close to the ridge crest, as sampled at ODP Site 960, was disrupted by several hiatuses and condensed sections marked by hardground development (Figure 3). In contrast, west of the ridge crest at Site 959 sedimentation has been more continuous. The interpreted seismic section (Figure 4) was converted to a depth rather than time section using the seismic velocity determinations of *Sage* [1994]. Biostratigraphic constraints from the ODP wells allow

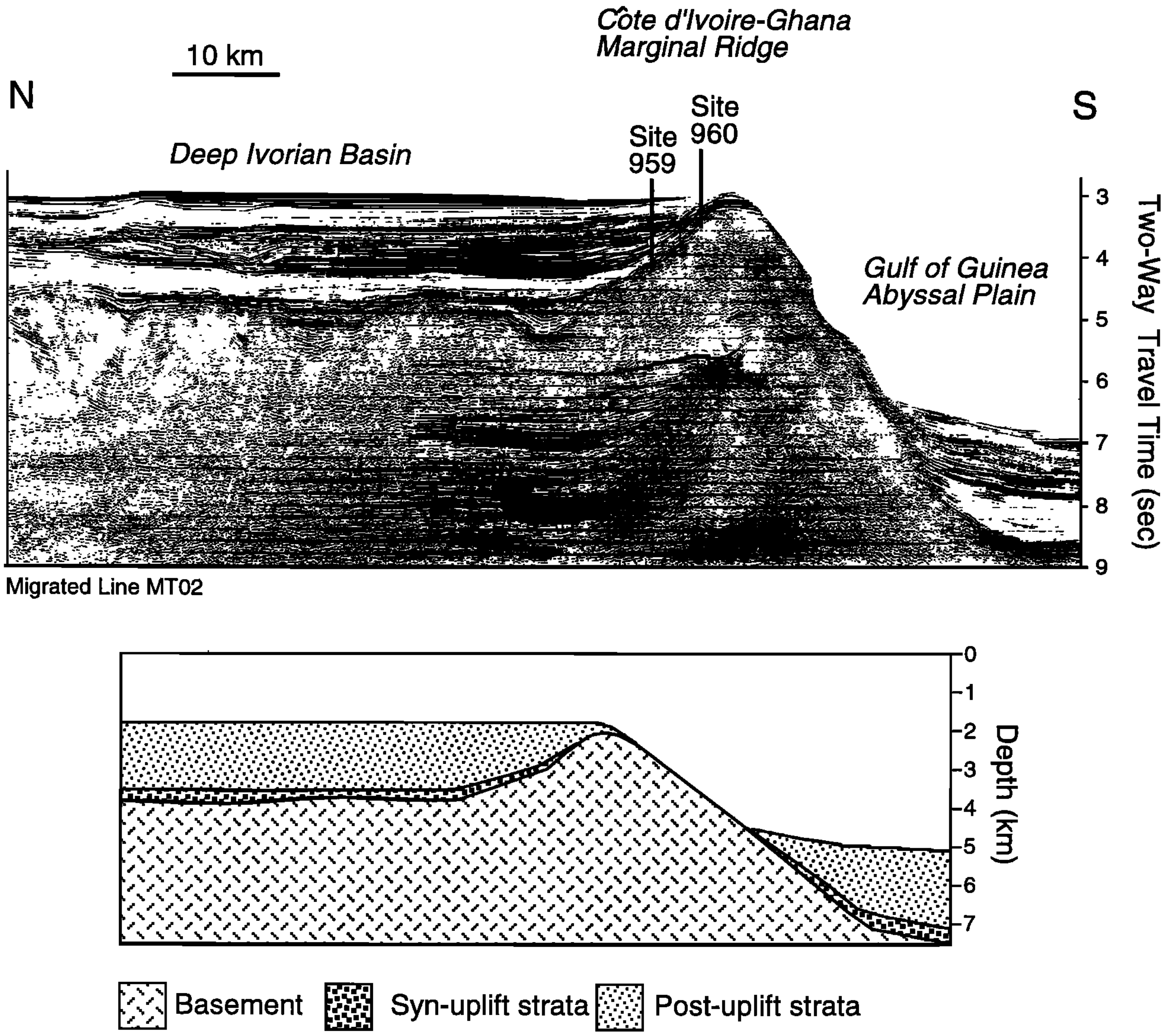

Figure 4. Seismic profile of Marginal Ridge in the vicinity of Site 959 and 960 from seismic line MT02 [*Basile et al.*, 1993, 1996]. Model stratigraphy taken from line MT02 and converted to depth using the average interval velocities of *Sage* [1994].

the interpreted reflectors to be dated. Critically, the stratigraphy was then divided into a "basement" i.e., rocks formed prior to unloading of the margin (although some of these are deformed sediments that were deposited during intracontinental wrenching), a sequence formed during ridge uplift at 105 Ma, and a postuplift sequence. In this area the pre-uplift rocks are the tectonized Aptian-Albian sediments of the intracontinental wrench phase and their basement. The identification of synuplift sediments from the seismic profile is difficult because their thickness is small, and they do not appear to be preserved on the crest of the ridge where it has been drilled. The duration of the hiatus seen at Sites 959 and 960 is constrained to being pre-early Turonian and post uppermost Albian and spans the estimated time of continental separation and thus the time at which flexural uplift of the Marginal Ridge would have occurred. We infer that the crest of the ridge experienced subaerial exposure and erosion prior to transgression at 89 Ma. As the predicted time of ridge passage and thus maximum possible thermal uplift is ~90 Ma at Sites 959 and 960, any sediments deposited on the ridge between 105 and 90 Ma must have been eroded just prior to renewed carbonate sedimentation in the wake of spreading ridge passage.

The Cenozoic history of the ridge, as reconstructed from the sediment facies and accompanying benthic foraminifer assemblage of the recovered cores [*Mascle et al.*, 1996], is one of pelagic sedimentation in significant water depths, summarized on Figure 3. The change from siliceous-dominated to carbonate-dominated sedimentation in the early Miocene reflects global changes in the planktonic microfauna and is not indicative of water depth changes.

5. Subsidence Modeling

If flexural backstripping significantly improves upon the one dimensional backstripping method of *Sclater and Christie* [1980] as applied to the Marginal Ridge by *Clift et al.* [1997], it must also represent a simplification of the actual three-dimensional system. Uncertainties may be introduced into the analysis if significant loads are applied to the plate out of the section chosen or if the sediment loads considered are actually supported to a large extent outside the line of the section; for example, the Marginal Ridge might be depressed because of being better predictors of the extensional deformation of the continental crust. In contrast, *Roberts and Kusznir* [1998] point out that necking models do not incorporate upper crustal brittle faulting, and while useful on a large scale, they are inappropriate when considering the tilting and stratigraphic development of a single fault block, as we attempt in this case. Because necking models do not incorporate faulting, they cannot account for the footwall uplift of individual faults and, consequently, tend to underpredict rift flank topography. Detachment models may be appropriate for the Côte d'Ivoire-Ghana margin, but there is no particular reason to favor these over the flexural cantilever model used here.

The flexural cantilever model provides a reasonable simulation of the rapid drop in mechanical strength predicted for a quartz-dominated crust when the increase in temperature with depth causes a change in the style of crustal deformation from brittle failure above to creep below [e.g., *Carter*, 1976]. However, as noted by *Van der Beek* [1998], the model does not account for upper mantle strength and so is least applicable to rifts in cold cratonic areas (e.g., Baikal, East Africa) and works best in areas where the crust and mantle are more decoupled, i.e., hot (e.g., Aegean Sea and western United States). In the case of the Côte d'Ivoire-Ghana margin we are modeling the flexural unloading of a rifted margin at 105 Ma shortly after the rift phase at 120 Ma, and consequently, we anticipate a hotter than normal lithosphere appropriate to the flexural cantilever model.

The flexural cantilever model assumes the same amount of extension in the lower crust and mantle lithosphere as in the upper crust and distributes this in a sinusoidal fashion over any given wavelength, typically 100 km. Such a model implies a depth of necking in the middle of the crust at <15 km, although necking is not specifically incorporated into the flexural cantilever model. This necking depth is shallower than the values of 15–20 km used in rifted margin settings to explain the seismically determined shape of the Moho and the size of the rift flank uplift [e.g., *Kooi et al.*, 1992; *Janssen et al.*, 1993] but is in accord with the 7.2 km predicted by *Watts and Stewart* [1998] for the Gabon margin. A relatively shallow depth of necking seems appropriate for modeling the unloading of a previously rifted margin given that the plate is expected to be weak compared to the prerift condition. If this assumption is incorrect and necking is significantly deeper in the crust, then our forward models will tend to underpredict the uplift of the Marginal Ridge for any given elastic thickness (Te). Consequently, when a forward model provides a match for the modern and restored Cretaceous ridge and overlying stratigraphy, the Te used for this model would be an overestimate of the real figure if necking is deeper than 15 km.

In the models developed here we chose to extend brittle faulting to a depth of 15 km because for continental lithosphere with a quartz-dominated rheology and a geothermal gradient of 15–18°C km^{-1} the brittle-ductile transition is usually estimated as lying between 10 and 15 km [e.g., *Zuber et al.*, 1986]. This range of values is supported by the observed depth of seismic activity in noncratonic lithosphere [*Bufe et al.*, 1977; *Kusznir and Park*, 1984], although seismicity is large sediment accumulations farther west in the Deep Ivorian Basin that could be interpreted as tectonically induced postrift subsidence. Three dimensional studies of the Gabon margin [*Watts and Stewart*, 1998] suggest that because flexual rigidity in rifted margins during and shortly after breakup is low, the first-order behavior of the margin can be successfully captured by two-dimensional models. Furthermore, seismic reflection data from *Peirce et al.* [1996] and *Sage* [1994] indicate that the sediments of the Gulf of Guinea Abyssal Plain and Deep Ivorian Basin do not thin or thicken rapidly for some distance on either side of the studied section, suggesting that a two-dimensional modeling approach will capture many of the most important features of the margin.

We assess the feasibility of a flexural uplift origin for the Marginal Ridge at the time of separation between Brazil and Africa (i.e., ~105 Ma) by forward modeling. The vertical tectonic reconstruction of *Clift et al.* [1997] applied the one-dimensional backstripping technique of *Sclater and Christie* [1980] to the ODP wells on the Marginal Ridge. This approach has the intrinsic problem of having to assume local isostatic compensation of the crust during and after stretching. Although this is commonly considered a reasonable approximation in extensional margin settings, at least during the synrift period, *Clift et al.* [1997] noted that the wavelength of the Marginal Ridge uplift is consistent with a flexural deformation of the margin, which would require a finite flexural

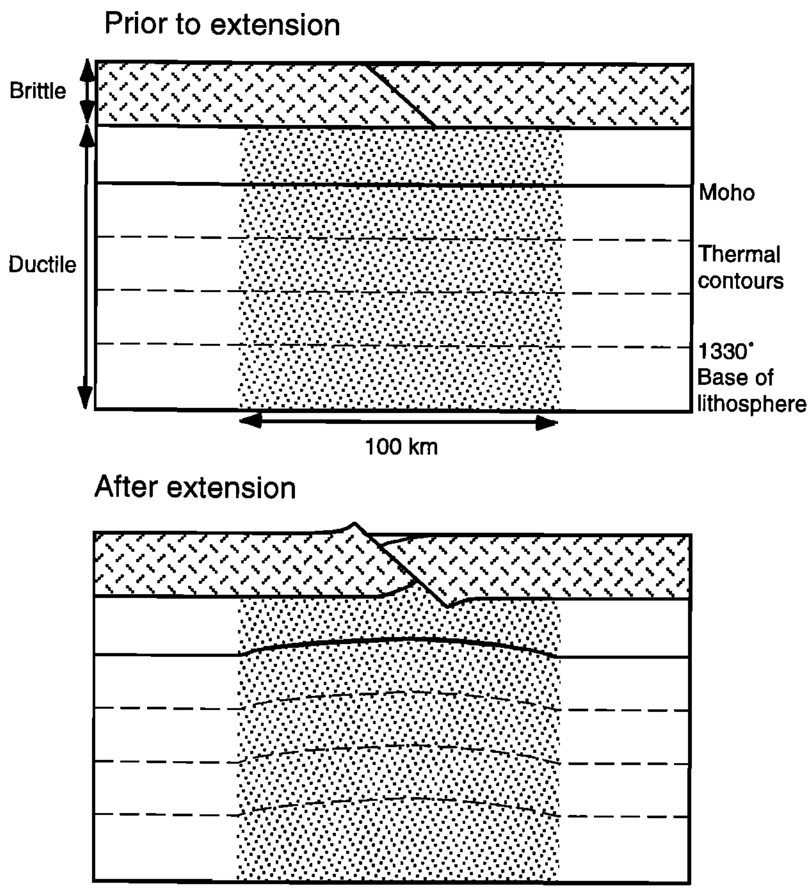

Figure 5. A schematic representation of the flexural-cantilever model for lithospheric deformation showing simple shear faulting in the brittle upper crust and pure shear in the lower crustal and mantle (redrawn from *Kusznir et al.* [1991]).

rigidity to the plate at the time of formation. While the flexural strength of the continental crust may be low (Te ≈ 5 km) during and immediately after rifting [e.g., *Karner and Watts*, 1982; *Holt and Stern*, 1991; *Kooi et al.*, 1992; *Watts and Stewart*, 1998], forward modeling the architecture and sedimentary fill of some active rift zones (e.g., East Africa [*Ebinger et al.*, 1991] and Late Baikal [*Van der Beek*, 1997]) shows that elastic thicknesses can be significant even at this stage (Te = 20–30 km). As a result, the most appropriate way to model the vertical motions of the sheared margin is through a method in which the flexural rigidity of the plate can be accounted for in addition to the thermal subsidence due to cooling and thickening of the mantle lithosphere. This type of analysis has not been possible before along the Ghana margin or other transform passive margins because of the lack of paleowater depth information with which to constrain the model.

5.1. Flexural Cantilever Model

We employ a flexural cantilever model for continental deformation [*Kusznir and Egan*, 1989; *Kusznir et al.*, 1995] to model the evolution of the Marginal Ridge. The flexural cantilever model considers deformation in the upper crust to be brittle and that in the lower crust and lithospheric mantle to be ductile (Figure 5). The model has been criticized by some workers for underpredicting the elastic thickness of the continental crust compared to estimates based on gravity and seismic data and for overestimating the amount of rift flank uplift, most notably in the Baikal Rift [*Van der Beek*, 1997]. *Van der Beek* [1997] instead argues in favor of either crustal necking models, such as that employed on the Gabon margin by *Watts and Stewart* [1998], or detachment faulting models as noted at >25 km in rifted cratons [*Foster and Jackson*, 1998; *Jackson and Blenkinsop*, 1993].

5.2. Forward Modeling the Marginal Ridge

In producing a forward model we aim to match the present-day shape and depth of the Marginal Ridge crest as well as the modern stratigraphy. We then use the mantle lithosphere extension predicted by the preferred forward model to flexurally backstrip the basement in order to estimate the vertical history of the sheared continental margin in response to extension across the Marginal Ridge at the end of the intracontinental wrenching phase between Brazil and Africa and the start of ocean-continent shearing (i.e., ~105 Ma).

The flexural cantilever model is only applicable to the transform margin situation assuming that there is some degree of flexural coupling across the continent-ocean transition. The models of *Sandwell and Schubert* [1982] assume that the younger, hotter oceanic side of the boundary becomes locked mechanically against the colder continental side after the transform becomes tectonically quiescent. As a result, the

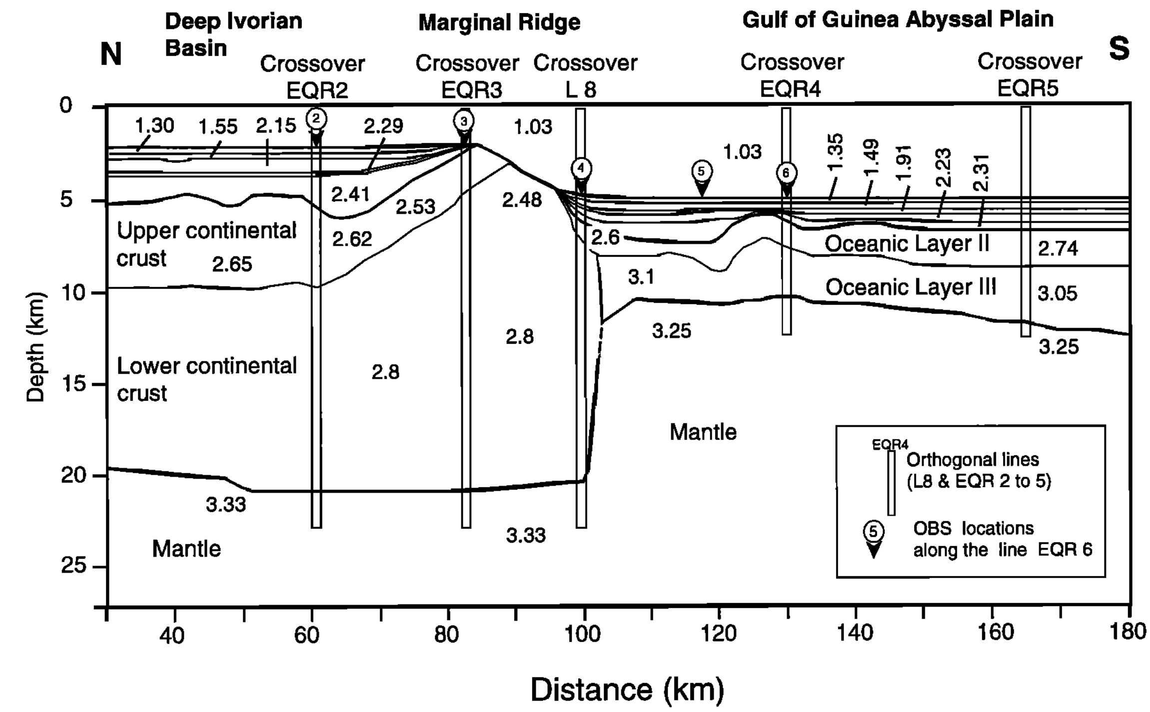

Figure 6. Density model for the Côte d'Ivoire-Ghana margin in the vicinity of ODP Sites 959 and 960 from *Sage* [1994]. See Figure 2 for location of transect.

seafloor on the continental side may bend down toward the fracture zone and on the oceanic side bow up by several hundred meters relative to the overall thermal subsidence [*Lorenzo and Wessel*, 1997]. Gravity modeling of transform plate boundaries by *Christeson and McNutt* [1992] provided further evidence that coupling across the boundary is commonly observed. A more detailed study would be required to quantify the degree of coupling in this setting, but for the purpose of this modeling study we presume that this margin behaves in a similar fashion to those studied elsewhere.

Evidence for uncoupled flexure being important in the margin's development prior to the passage of the spreading ridge comes from the combined gravity and seismic study of *Sage* [1994]. The basement and upper-lower crustal boundary are seen to shallow toward the continent-ocean transition, suggestive of uncoupled flexure of the plate, but the Moho does not mirror this pattern (Figure 6). Between the crossover point of seismic lines EQR3 and L8 the continental crust appears to thin by almost 3 km whereas the continental Moho remains relatively flat (Figure 6). The pressures required to maintain the Moho at this depth are almost an order of magnitude greater than those required by a flexure model. Most likely, this reflects greater uncertainties with increasing depth in the seismic interpretation. Indeed, no Moho reflections were recorded by *Sage* [1984] under the Marginal Ridge in the region of the studied transect. Although an upward flexed Moho would disrupt the fit between the crustal model and the observed gravity, this may represent greater degrees of serpentinization along the continent-ocean boundary than anticipated by *Sage* [1994]. Thus, at face value the crustal cross section of *Sage et al.* [1997] indicates flexural deformation of the margin compatible with the weakly coupled conclusion of the gravity model.

In the case of the Côte d'Ivoire-Ghana margin the slope of the southern edge of the Marginal Ridge represents the main fault that performs the unloading. There are no water depth controls for the ridge immediately following unloading and uplift at 105 Ma because of an unconformity spanning that period on the ridge crest. We anticipate that the Marginal Ridge may have experienced transient uplift, possibly as much as 2.5 km, because of juxtaposition of the spreading ridge at 90 Ma. This process is not included in the flexural cantilever model, and so, comparison of the forward model with the observed paleobathymetry at this time is not possible as a test for a forward model. As a result, the suitability of a forward model is principally assessed by comparing the present depth to the ridge crest and adjacent Deep Ivorian Basin as well as to the overall geometry of the ridge. Although extension may not have been as strong under the Marginal Ridge itself, this is difficult to quantify from existing seismic or gravity data, as mentioned above. Tectonized Aptian-Albian sediments at the base of Holes 959 and 960 were deposited during intracontinental wrenching, following the onset of rifting (phase 1), but predate the uplift of the Marginal Ridge (phases 1 and 2; Figure 4).

We assume that lithospheric extension during initial continental rifting (~120 Ma) was uniform across the entire width of the section studied because the basement of the Deep Ivorian Basin is relatively flat for several kilometers away from the Marginal Ridge (Figure 4). Using seismic refraction and gravity-derived crustal thicknesses of 35 km for unrifted African crust and 16.5 km for the Deep Ivorian Basin landward of the Marginal Ridge [*Sage*, 1994], the extension at 120 Ma is taken to have a β of 2.1 (pre-rift crustal thickness/post-rift crustal thickness). Estimating the earlier extension is important because in order to model flexural unloading during

Table 1. Physical Parameters Used in the Preferred Forward Model

Parameter	Value
Te	2.5 km
Heave	27 km
Initial angle of fault	25°
Age of extension	105 Ma
Pure shear width	100 km
Initial crustal thickness	35 km
Brittle layer thickness	15 km
Density of crust	2700 kg m^{-3}
Density of mantle	3300 kg m^{-3}
β at 120 Ma	2.1

continental separation, we need to have an estimate of the inherited thermal subsidence from that earlier rifting event. We assume that mantle lithosphere extension at 120 Ma equals the crustal extension determined by seismic means [*Sage*, 1994], which is reasonable given the observed thermal subsidence in the Deep Ivorian Basin. For the purpose of trying to show that the Marginal Ridge could be generated solely by a process of flexural unloading and in view of the lack of reliable crustal thickness estimates for the ridge, we assume that this crust was extended to the same degree as that under the adjacent Deep Ivorian Basin at 120 Ma.

In attempting to generate a forward model that duplicates the modern Marginal Ridge we vary the amount of extension β across the sheared margin, the effective elastic thickness (Te), and the angle and heave of the faults. We consider both extension parallel to the Marginal Ridge (i.e., perpendicular to the section seen in Figure 4), which generated the thinned crust of the Deep Ivorian Basin at 120 Ma, and the subsequent extension perpendicular to the Marginal Ridge in order to simulate the unloading experienced during continental separation at 105 Ma. The extension causing unloading of the Marginal Ridge at 105 Ma increases rapidly southward from the crest of the ridge to the oceanic crust, where by definition, $\beta = \infty$. In order to model successfully vertical motions of the margin we increase the extension south of the margin at 105 Ma to create a β of 6, at which point the thermal subsidence of the basement closely approaches that of oceanic crust [*McKenzie*, 1978]. The precise lateral variation in β between the ridge crest and the oceanic crust is controlled by the dip of the detachment fault, the amount of heave, and the width over which the ductile extension is distributed in the mantle [e.g., *Weissel and Karner*, 1989]. The depth of necking is fixed at >15 km for the flexural cantilever model.

In modeling the extension across the transform margin at 105 Ma we choose the physical parameters shown in Table 1 to apply to the flexural cantilever model. Density values for the crust and mantle, 2700 and 3300 kg m^{-3}, are taken from *Sclater and Christie* [1980]. In order to model the loading and compaction of the sediment column during the forward modeling, as well as the decompaction of sediment during the flexural backstripping we used petrophysical data from the sediment cores from the ODP wells. We assume that porosity can be described as an exponential function of depth, where porosity Ø, at depth Z below sea level, is given by

$$Ø = Ø_0 e^{-cz},$$

where $Ø_0$ is the porosity at the sediment surface and C is the controlling constant [*Athy*, 1930]. For the shale, sand, and limestone we use the values 63, 49 and 70% for $Ø_0$, 0.51, 0.27, and 0.71 km^{-1} for C, respectively. These values provide reasonable agreement with the values measured by the shipboard measurements on core samples taken during ODP Leg 159 [*Mascle et al.*, 1996]. However, shaley sediments from the tectonized Aptian-Albian units, designated basement on the cross section (Figure 4), often show lower porosities than expected. This reflects compaction due to the pressures of deformation in the transform zone rather than regular burial-related compaction and dewatering [*Mascle et al.*, 1996]. Since we treat these rocks as basement their lack of further compaction following their deformation prevents this from becoming a source of error in our analysis. In order to calculate the loading effect of the sediment we use sediment matrix densities of 2.72, 2.65 and 2.71 g/cm^3 for shale, sand and limestone. Again, these are in close agreement with actual measurements from the core taken at each of the drill sites.

5.3. Flexural Backstripping

Flexural backstripping is a reverse-modeling process that removes, in step-wise fashion, the thermal subsidence predicted from the forward extension model and progressively unloads and decompacts the postrift sedimentary section deposited since the time of unloading and uplift. Correction is made for eustatic sea level variations using the model of *Haq et al.* [1987], which shows sea level in the mid Cretaceous being ~200–250 m higher than today. Provided that the model and observations can be made to match at the present day, differences between backstripped model and observation at 90 Ma can be used to estimate uplift due to the lateral heat conduction because this process is not included in the flexural cantilever model. This influence would not have affected basement depths prior to 105 Ma (onset of ocean-continent transform motion) and should be largely dissipated today. Assuming that the flexural forward model is appropriate the backstripping method allows the depth of the ridge crest at any given time in the past to be estimated.

6. Results

6.1. Flexural Forward Models

Figure 7 shows our preferred forward model, which reproduces the overall shape and depth of the Marginal Ridge. The parameters used in this model are provided in Table 1. A detachment dipping at ~25° provides the best fit given the seismically determined constraints on heave and ridge shape. A purely flexural unloading model can generate the main features of the permanent elevation and geometry of the Marginal Ridge observed today, although as we now document, the forward model is quite sensitive to changes in the key parameters.

The heave on the fault is quite well constrained by the seismic profile (Figure 4) at ~27 km. Degradation of the fault surface after exposure will tend to shallow the angle and increase the apparent heave. Figure 8 shows the predicted forward model for a 60° fault, first with the same 27 km heave and then with a reduced heave of only 6 km. In both cases the uplift of the ridge significantly exceeds the backstripped model prediction by placing the ridge well above sea level at the time of uplift. In the case of the 27 km heave the ridge is

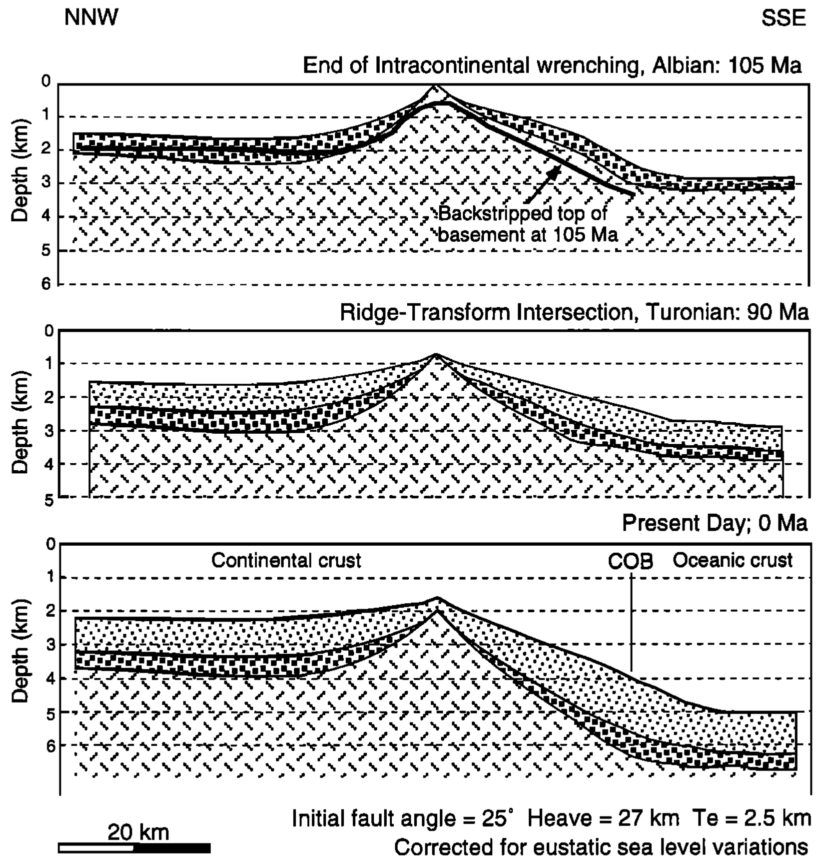

Figure 7. Forward model of the Marginal Ridge showing the predicted geometry of the basement assuming unloading of the margin at 105 Ma, along a 25°, seaward dipping detachment. The solid line shows the profile predicted from the backstripped, unloaded model at this time. Forward model of the Marginal Ridge at 90 Ma assuming normal thermal subsidence since unloading of the margin at 105 Ma.. Forward model of the Marginal Ridge at the present day.

still subaerial even at 90 Ma, even if there is no thermal rejuvenation at that time. While erosion could be invoked to lower the height of the ridge, the models, especially the 6 km heave model, also have problems in reproducing the gentle southern slope to the Marginal Ridge. That this is composed of in situ basement material has been confirmed by submersible diving [*Mascle and Equanaute Scientific Party*, 1994]; so, it is not feasible to shallow the slope by infilling through mass wasting since the ridge uplift. A low heave model also predicts a horizontal distance between the Marginal Ridge crest and the base of the slope that is substantially smaller than observed.

Matching the heave of 27 km and the limiting marginal uplift at the time of unloading to the present elevation of the basement above the Deep Ivorian Basin seen on the seismic profile requires a low effective elastic thickness (Te) and a shallow depth of necking (<15 km). A Te of 2.5 km at the time of continent separation provides the best match. Slight changes in this variable of only 1–2 km produce models with poor fits to the observed data. For example, Te = 1 km (Figure 8) reduces the relative height of the ridge to ~1 km, whereas a Te of 5 km results in uplift that far exceeds the observed water depths at 105 Ma even without accounting for the effect of spreading ridge migration (Figure 8). Even after accounting for subaerial erosion and thermal subsidence such initial conditions are incapable of reproducing the modern form of the Marginal Ridge and onlapping stratigraphy. Consequently, we predict that models with Te >3 or <1.5 km will not satisfy the observed ridge geometry, assuming necking at <15 km. A necking depth of 25 km, such as is estimated for cratonic rifts, would provide a much higher ridge than is observed, so that Te would have to be <2.5 km in order to provide a match. As noted above, such necking depths are not considered appropriate to this setting.

Because we interpret the current southern slope of the Marginal Ridge to be a degraded paleodetachment surface, the modern slope places a lower limit on the final angle at 12°. In turn, the 12° modern dip of the southern slope to the Marginal Ridge limits the dip of the initial seaward dipping fault to <~25° given the constraints on heave and ridge shape. The low initial 25° angle of the seaward dipping fault detachment in the best fit model is unusual because transform zones are typically associated with subvertical faults [e.g., *Harding et al.*, 1985]. In practice, it is difficult to produce a model that compares well with the modern Marginal Ridge with any detachment >30°, as the initial uplift becomes too large despite the low Te and shallow depth of necking. Although reducing the value of the favored model, Te (<2.5 km) allows higher-angled faults to generate less uplift, these values are incompatible with the shape of the ridge and its overlying stratigraphy. As for Te, the forward model is sensitive to changes in the slope angle.

Differences between the preferred forward model and the

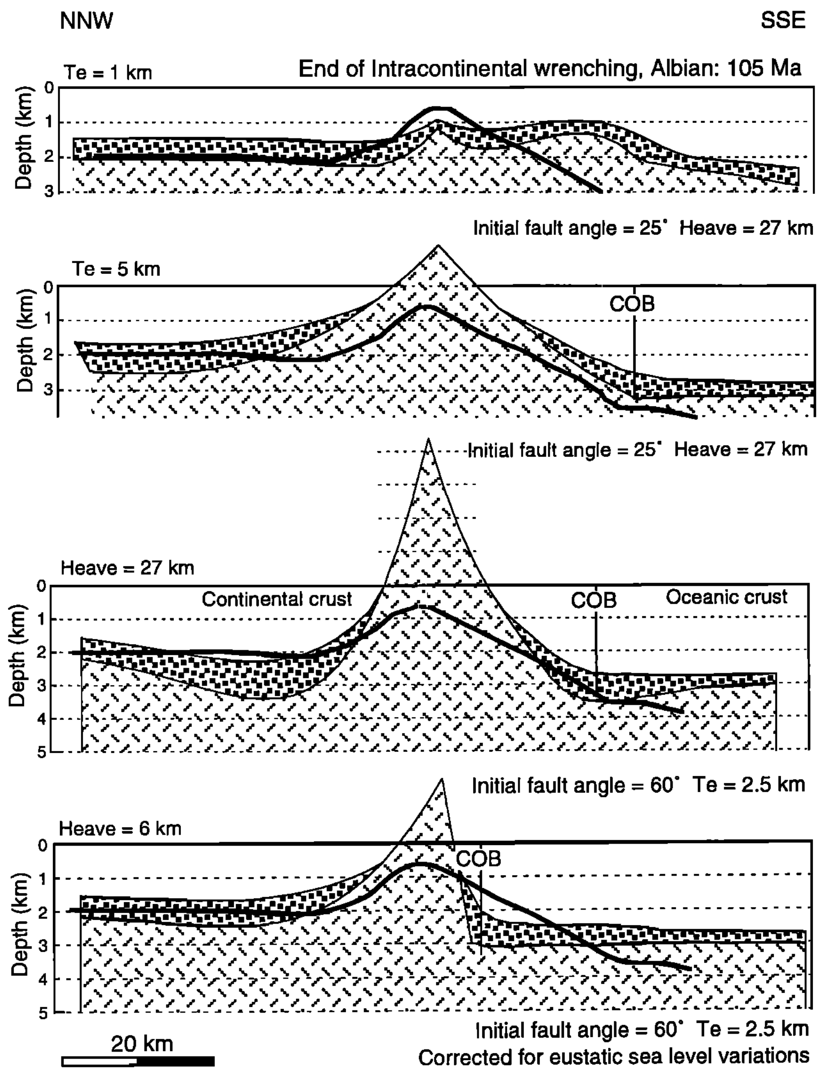

Figure 8. Forward model of the Marginal Ridge at 105 Ma showing the different sizes of uplift and postrift stratigraphic geometries generated by changing key variables in the model Te = 1 km, Te = 5 km, initial angle of fault is 60°, and initial angle of fault = 60° and heave = 6 km.

observed data may be explained by subaerial erosion of the ridge crest induced by the thermal uplift during the passage of the spreading center at 90 Ma, a process not simulated in the flexural cantilever model. The simple unloading model produces a sharp, rather than the observed rounded, ridge shape. The model also predicts submarine conditions since unloading at the time of continental separation (105 Ma), whereas ODP cores indicate a subaerially eroded unconformity between the Albian tectonized units and the Turonian shallow water carbonates. Thermal uplift and ensuing erosion should not affect the predictions for the form of the Marginal Ridge during its initial uplift, prior to its juxtaposition with the hot oceanic crust, or for the present-day, by which time the additional heat will have dissipated. Finally, although the model is in good agreement with the total sediment thickness in the Deep Ivorian Basin, a limitation in the modeling routine that sediment flux be constant across the section results in overprediction of the thickness of sediment on the southern slope of the ridge itself. This does not significantly affect other predictions of the forward model.

6.2. Flexural Backstripping

Flexural backstripping provides important information concerning the thermal subsidence and thus the nature of thermal rejuvenation at the time of spreading ridge passage, although it does not provide information on the flexural origin of the Marginal Ridge. Figure 9 shows the backstripped model calculated using our preferred β profile derived from the forward model. Although β does increase rapidly seaward from the Marginal Ridge crest, the mantle-driven thermal subsidence predicted from the inferred crustal extension is insufficient to restore the crest of the ridge to sea level at 90 or 105 Ma. Changing the mantle extension values, while leaving the crustal thickness constant, shows that in order to restore the top of the ridge to sea level at 90 Ma (when shallow water limestones were being deposited; Figure 3) a net mantle β of almost 4 would be required. This is almost twice as much as the crustal thinning measured by the previous seismic refraction study (Figure 6) and may be used as evidence that thinning of the mantle lithosphere was greater than that of the crust under the Marginal Ridge, at least at 90 Ma.

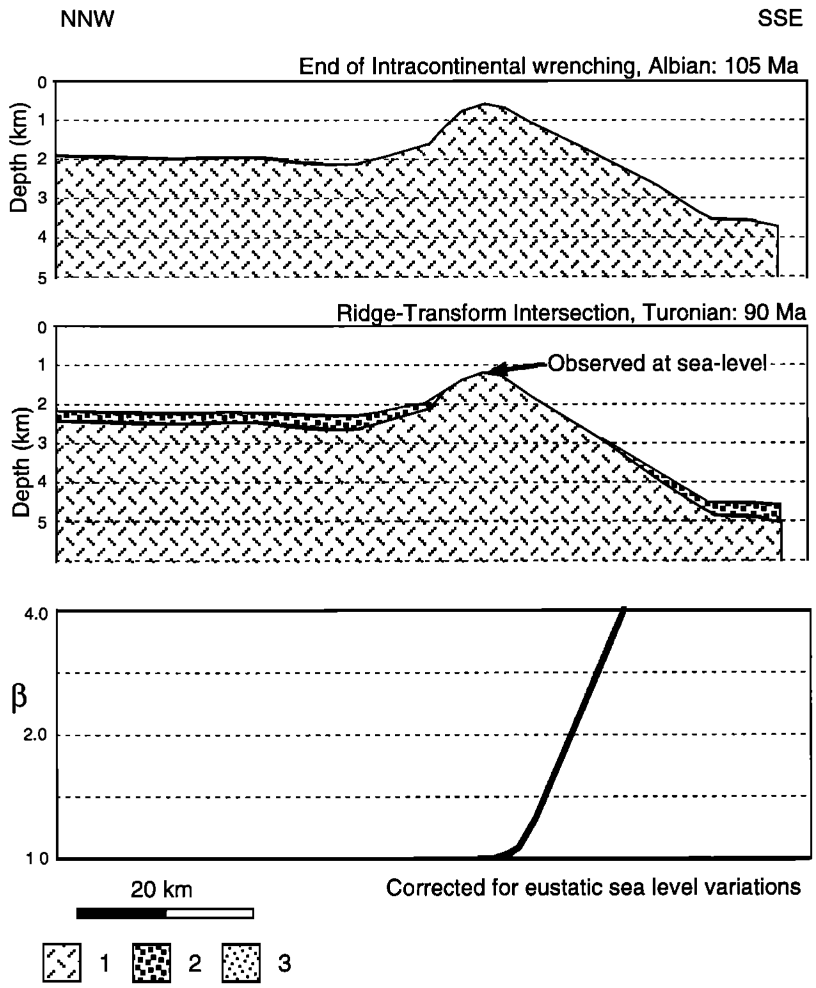

Figure 9. Backstripped profile across the Marginal Ridge at 105 Ma (end of intracontinental wrenching) and at 90 Ma (ridge-transform intersection). The lateral variation in extension across the margin during continental separation at 105 Ma. The model uses an elastic thickness of 2.5 km and assumes extension started at 120 Ma ($\beta = 2.1$) to the left of the transform margin. The backstripped model does not account for thermal rejuvenation of the lithosphere during ridge-transform intersection. Note the mismatch between predicted depth at 90 Ma and the known presence of the ridge within the photic zone.

Shoaling of the Marginal Ridge and subaerial erosion immediately prior to limestone deposition at 89 Ma could result from differential tectonic extension of the mantle at 105 Ma (end of intracontinental wrenching) or from thermal erosion of the mantle during spreading ridge passage at 90 Ma. Neither of these processes are reproduced by the flexural cantilever forward model and simple unloading across the Marginal Ridge at 105 Ma, which is why the backstripped model does not predict the observed shallow water conditions at 90 Ma. The seaward dipping geometry of the detachment makes differential tectonic thinning of mantle under the Marginal Ridge unlikely as the thinning would have to be substantially offset to the north away from the region where deformation would be expected in the mantle. A south dipping detachment would be expected to focus mantle extension south of the ridge crest.

Differential extension would predict elevation of the Marginal Ridge crest far above sea level at 105 Ma prior to undergoing thermal subsidence. In contrast, uplift due to passage of the ridge would delay peak uplift until 90 Ma, so that the ridge could have been submarine for much of the time 105–90 Ma. The erosion that caused the unconformity at 90 Ma would then be predicted to occur immediately prior to the subsequent transgression. Since the period between 90 and 105 Ma at Sites 959 and 960 is marked by unconformity (Figure 3) the paleowater depth during this interval is unknown, and both options are compatible with the data. However, given the unusual nature of the extension required for a tectonic thinning of the mantle and the known presence of the oceanic spreading center, we suggest that the shoaling of the ridge at 90 Ma is driven by thermal erosion of the mantle lithosphere due to conduction during continent-ocean shearing, peaking at or just after the time of final spreading ridge passage.

7. Sensitivity of the Flexural Cantilever Model

Simulation of Marginal Ridge uplift at 105 Ma by the flexural cantilever model is sensitive to variation in key modeling parameters (see Table 1), some of which are better constrained (crustal and mantle densities, heave, depth of

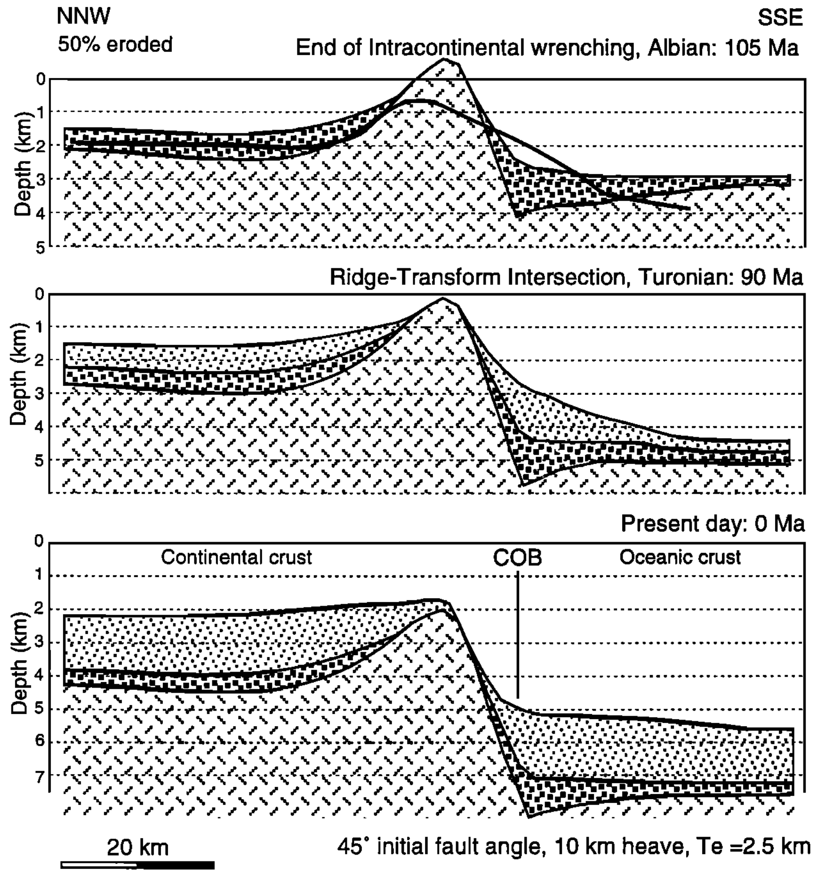

Figure 10. Forward models for evolution of the Marginal Ridge with an initial faulting angle of 45°. Model shows good agreement with the observed paleowater depths at 90 Ma and in the present day. Erosion of 50% of the subaerial topography provides the best fit. However, the model poorly mimics the modern southern slope of the Marginal Ridge and does not match the backstripped model at 105 Ma.

brittle faulting, and depth-porosity trends) than others (Te, crustal thickness, and angle of faulting). Uncertainties due to variations in the compaction of the sedimentary cover are at least an order of magnitude lower than those associated with water depth estimates, especially as the flexural cantilever model incorporated actual porosity and density data from the well sites. Differences in the initial crustal thickness of the unloading continental plate may change the buoyancy of the continental crust that is driving the uplift. For example, if a crustal thickness of 20 km is used instead of 16.5 km, then this would increase the amount of uplift by ~20% or 400 m. However, crustal thickness estimates for the Deep Ivorian Basin have smaller uncertainties of 1–2 km, determined by a combined ray-tracing and ray theory waveform modeling of ocean bottom seismometer wide-angle reflection and refraction data [*Sage*, 1994].

Consequently, we focus on the Te, angle of faulting, and heave as the principle variables that can significantly change the model results. The forward models produced results substantially different from the observations: Te changesonly slightly, but faulting angle and heave, which need to change significantly to make a larger difference to the mode, are more robust. To demonstrate this we run alternative models to quantify the effect of changing these variables on ridge height, shape, and subsidence history. Of these factors the heave is the best constrained because it can be measured directly from the seismic profile and because submersible observations confirm the continental composition of the lower ridge slope down to at least 4500 m water depth [*Mascle et al.*, 1994], implying a heave of at least 12 km. The lack of seismic evidence for a major unconformity below that level and the relative lack of magmatism along the continent-ocean transition suggests that much of the lower slope also comprises continental material. This horizontal distance would have increased because of slope degradation, although the pinch-out of older sedimentary sequences toward the top of the modern ridge indicates that the degree of slope degradation since uplift has not been dramatic. Assessing the effect of Te variations is justified in view of the high values (Te = 20–30 km) that have been estimated in modeling several active rift zones [e.g., *Van der Beek*, 1997], although as noted above, even a low value of 5 km produces incompatible forward models.

8. Feasibility of High-Angle Fault Models

Low angle detachments (~25°) reproduce the shape of the ridge accurately but have water depth predictions that require additional thermally driven uplift in order to match the observation of a hiatus from 105 to 90 Ma at the ridge crest followed by shallow water carbonate sedimentation. In

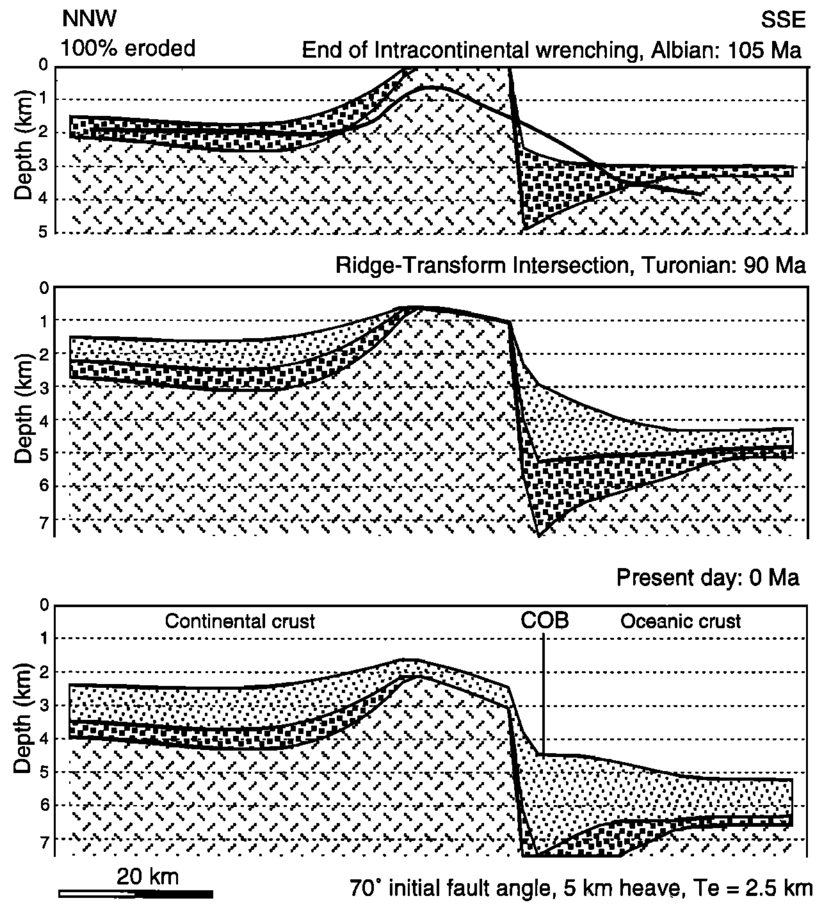

Figure 11. Forward models for evolution of the Marginal Ridge with an initial faulting angle of 70°. In this case, 100% erosion of subaerial topography is required to match the modern basement depth. The model shows a poor agreement with the reverse model or the shape of the modern slope of the Marginal Ridge.

contrast, moderate-to-high-angle fault models with dip values greater than the best fit 25° described above are advantageous in some regards but require small heaves of only 5 km, much less than that seen on the seismic profiles, and also significantly overpredict the gradient of the modern southern slope to the ridge. A larger initial uplift at 105 Ma along a high-angle fault can generate an unconformity and result in a delay in the start of marine sedimentation without having to invoke additional thermal rejuvenation of the margin to drive uplift peaking at 90 Ma. High-angle faults are also more typically associated with transform fault zones [e.g., *Scrutton*, 1982].

We examine the influence of applying significant subaerial erosion to the large subaerial ridges that accompany high angle forward models to see if high-angle fault models can be reconciled with the modern observed stratigraphy and the backstripping predictions. In the first instance we choose a fault dipping at 45° and vary the heave and erosion rate to provide the best possible fit to the present-day depth to the ridge top as well as to the observed time of shallow marine sedimentation at 90 Ma (Figure 10). The forward model, shown at 105, 90, and 0 Ma, results in several important matches with the recorded ridge history. By choosing to erode 50% of any subaerial topography the model successfully predicts not only the unconformity seen at Site 960, the start of shallow marine sedimentation, but also the rounded form of the Marginal Ridge crest. This reinforces the conclusion derived from the unconformity seen in cores that the Marginal Ridge suffered subaerial erosion prior to 90 Ma limestone sedimentation. The match with the modern ridge depth and thickness variations in the sedimentary cover is also good in this model, although the overall ridge shape is much too steep.

If we use an initial fault angle to 70°, 100% of the subaerial topography at 105 Ma must be eroded in order to keep the topographic difference between the Deep Ivorian Basin and the Marginal Ridge consistent with the present day (Figure 11). Unfortunately, this also results in a highly inaccurate predicted shape to the ridge, notably the prediction of tilted wave-cut terrace, and a steep lower slope to the ridge with little resemblance to modern seismic and bathymetric observations. Attempts to amend the 70° model to fit the observations more closely can result in better fits for some aspects but causes additional problems. For example, a reduction in the amount of subaerial erosion produces a more realistic rounded top to the ridge crest but makes the modern ridge crest too high above the Deep Ivorian Basin. Similarly, by reducing the heave on the fault the ridge-basin relief can be preserved, but like all high-angle models, the southern slope of the ridge remains too steep, and the horizontal distance between the crest of the Marginal Ridge and the base of the slope is too small. Where erosion is set to sea level, the crest of the ridge is predicted at the correct horizontal distance from the base of the ridge but necessitates the generation of a wave cut terrace (Figure 11). We conclude that flexural unloading across high-angle faults cannot explain the form of the Marginal Ridge observed on the Côte d'Ivoire-Ghana margin.

9. Discussion

9.1. Transient Thermal Uplift

Transient uplift during the passage of a spreading ridge south of the margin following breakup provides a mechanism by which lateral heat conduction can drive uplift between 105 Ma and the start of marine sedimentation at Site 960 at 89 Ma. The amount of uplift required to raise the backstripped Marginal Ridge at 90 Ma from ~1200 m water depth (Figure 9) to sea level is within the limits of the 2000+ m estimated by simple heat conduction models [e.g., *Todd and Keen*, 1989] or the 1300–1400 m estimated by thermomechanical models [e.g., *Gadd and Scrutton*, 1997], although both these models assume local isostatic compensation. Indeed, *Gadd and Scrutton* [1997] note that with regional isostatic compensation their estimate falls to 335–470 m using a Te of 10 km. Our estimation of a low Te (2.5 km) would suggest uplift toward the higher end of *Gadd and Scrutton's* [1997] range.

The rounded form of the Marginal Ridge crest suggests subaerial erosion, but how much is unknown. A minimum estimate of uplift may be gained by assuming that elevation above sea level and erosion were small. In this case uplift at 90 Ma would be at least ~1200 m (Figure 9). Estimating an upper limit to the uplift is more difficult. However, even if we assume that the ridge-transform intersection removed almost the entire mantle lithosphere ($\beta = 10$) and this occurred as much as 4 m.y. prior to the resumption of marine sedimentation at 89 Ma, then for a ridge in local isostatic equilibrium, no more than ~310 m of thermal subsidence would be expected [*McKenzie*, 1978], thus limiting uplift to ~1510 m. A finite flexural rigidity would reduce this value significantly by compensating for the uplifting force over a wide area of the margin away from the most strongly affected southern edge. However, if the mantle lithosphere was effectively thermally eroded away, it is unlikely that the plate's rigidity could be significant, and so, the isostatic estimate provides a reasonable upper limit to the thermally driven uplift. Our estimate of a low Te (2.5 km) suggests that uplift approaching the maximum values is feasible in this example.

In practice, the net total uplift generated would depend on the degree of erosion. For example, by assuming that 50% of the exposed relief was eroded the total uplift would increase to ~1540 m, which is close to the thermomechanical model estimates of *Gadd and Scrutton* [1997] for a locally compensated margin and much more than their regional isostatic predictions. All of these estimates exceed the estimate of 390 m made by *Clift et al.* [1997] using the same drill data but a local isostatic assumption.

9.2. Underplating and Uplift

Magmatic underplating of the continental margin during passage of the spreading axis has been proposed as a method of generating permanent uplift along some volcanic sheared margins, most notably the southern Exmouth Plateau [*Lorenzo et al.*, 1991]. However, seismic reflection and wide-angle refraction studies close to the Marginal Ridge have suggested that the rift architecture and crustal structure of the Deep Ivorian Basin are most similar to nonvolcanic margins, which argues against magmatic underplating in this case [*Sage*, 1994]. *Peirce et al.* [1996] used seismic refraction evidence located farther north (Figure 1) to show up to 8 km of high velocity (7.3 km s^{-1}) lower crust close to the extensional continent-ocean transition, which tapers away to zero over a distance of ~100 km. This material was interpreted as gabbro underplated to the extended continental crust during breakup, in a manner similar to that noted in the northeast Atlantic [*Mutter et al.*, 1988]. While this implies a more volcanic-type margin, strangely lacking in seaward dipping reflectors, it does not appear to affect the Côte d'Ivoire-Ghana Marginal Ridge itself. The gabbros proposed by *Peirce et al.* [1996] would represent a rather brief excess melting episode early in the breakup process, somewhat predating the generation of the Marginal Ridge. In contrast, the thinner than normal oceanic crust in the Gulf of Guinea [*Sage*, 1994] suggests that magmatism at the time of ridge passage past the ODP drill sites was, if anything, less than normal and thus unlikely to have added significant volumes of melt to the continental margin.

9.3. A Flexural Origin?

If the Côte d'Ivoire-Ghana Marginal Ridge has been unloaded during continental separation, then it might be anticipated that the conjugate sheared margin in northeast Brazil would be depressed. The transform continental boundary north of the Potiguar Basin in northeast Brazil shows an oceanward dipping basement [*Matos*, 1992], which may, in part, be due to loading of the "upper plate" to the detachment. However, the presence of a major oceanward-dipping unconformity, removing an increasing amount of synrift sediment toward the continent-ocean transition, suggests that much of this tilting is related to uplift, erosion, and subsequent subsidence due to passage of the spreading ridge along this margin. The predicted loading of the conjugate margin assumes a perfect mass balance between the two margins, whereas there is evidence for continental fragments along the length of the Romanche Fracture Zone [*Honnorez et al.*, 1994], which could conceivably unload both margins. The evidence from the conjugate margin is equivocal, but is not in contradiction with a flexural origin for Marginal Ridge formation in West Africa.

Forward models using a flexural cantilever approach and a Te of 2.5 km provide the best fit to the modern observed stratigraphy and reconstructed subsidence profile but may underpredict the initial Te and/or the initial angle of south dipping detachment. Te may increase with time following rifting [e.g., *Karner and Watts*, 1982] but will not have a large effect on the size and width of the Marginal Ridge, which is principally generated at the point of continental separation (~105 Ma). An increasing flexural rigidity on the continental side of the margin would result in less subsidence of the Deep Ivorian Basin relative to the Marginal Ridge than if Te remained low. Thus a forward model that does not account of this effect will tend to predict a Marginal Ridge at ~105 Ma that is smaller than it would have been, as the model unrealistically allows the ridge to grow relative to the basin due to differential loading.

It may be argued that lithospheric heating during the original rifting process, prior to ocean-continent transform activity, may mask the thermal effect of ridge passage on the Marginal Ridge. However, the degree of thermal rejuvenation during ridge passage depends on how vigorous the conduction process between continental and oceanic plates is and how much of the original rift-related thermal event remains by the time of ridge passage. Rift-related heating is a function of extension and is well constrained by the observed crustal thicknesses [*Sage*, 1994]. Masking of the spreading ridge-related uplift is most likely to be a factor if the transform offset is short and/or spreading is very rapid because little time has elapsed during which the rifted continental crust can recover before receiving the spreading ridge-derived thermal pulse. A short transform offset or rapid spreading also reduces the amount of time for heat conduction from oceanic to continental plates. Neither of these possibilities are applicable to the Côte d'Ivoire-Ghana margin. On this margin at the location of the ODP drill sites the original extension is modest ($\beta \approx 2.1$) and results in a peak lithospheric geothermal gradient of ~23°C km^{-1}, compared to as much as 87°C km^{-1} predicted by *Todd and Keen* [1989] for ridge passage and 10–11°C km^{-1} for normal unrifted continental lithosphere [*Vitorello and Pollack*, 1980]. A pulse of heat from a juxtaposed spreading ridge would dwarf the residual heat from the earlier rift event. In addition, because rifting finished between 15 and 30 m.y. prior to the juxtaposition of the spreading ridge, this allows time for much of the rift-related thermal pulse to decay. As the rate of postrift thermal re-equilibration is related to the square root of age, with a thermal decay period of 80 m.y. for an equilibrium lithosphere thickness of 125 km [*McKenzie*, 1978], we estimate that only 38–56% of the original rift-related thermal pulse remained by the time of ridge passage, i.e., the background geothermal gradient for thermal rejuvenation measured only ~15–18° km^{-1}.

Late stage thermal rejuvenation of the Côte d'Ivoire-Ghana Marginal Ridge does not seem to be significant. Hotspot activity on the Cameroon Line during the Cenozoic-Recent [*Halliday et al.*, 1990] is >1000 km distant and given that even the largest modern hotspots appear to have effective thermal radii of ~1000 km (e.g., Iceland) [*White and McKenzie*, 1989], this feature is not considered important to the thermal evolution of the Marginal Ridge. This assumption is reinforced by the lack of any evidence from seismic or drill data for uplift of the ridge at the time of initiation of the Cameroon Line at ~30 Ma [*Fitton and Dunlop*, 1985].

10. Conclusions

Forward modeling the extension across a sheared margin at the time of continental separation shows that basement ridge features of the size and wavelength of the Côte d'Ivoire-Ghana Marginal Ridge can be generated by flexural unloading along oceanward dipping detachments during the final separation of continental blocks. Such ridges are not related to the passage of spreading ridges along the margin, unless this results in magmatic underplating, which does not appear to be the case in West Africa. If flexural unloading along a low angle (25°) detachment is appropriate to explain the modern ridge, then significant thermal rejuvenation of the plate margin is predicted to account for the shallow paleowater depths observed at the time of spreading ridge passage at 90 Ma. The low dip to the inferred detachment, 25°, is much less than might be expected for a transform setting but is required by flexural models in order to match the modern day basement shape, stratigraphy and inferred subsidence history.

Models involving moderate- (45°) and high-angle (60°) detachment faults to generate uplift of the Marginal Ridge successfully predict an erosional hiatus between the tectonized units and transgressive cover but produce basement geometries that do not agree with those observed. Moderate- to high-angle fault systems uniformly overpredict the gradient on the southern slope of the Marginal Ridge. The initial height of the ridge above the floor of the Deep Ivorian Basin is always too high with moderate- and high-angle detachments unless the heave is reduced to unreasonably small values and/or erosion is strongly increased, which results in wave cut terraces not observed on seismic data. Similarly, only very low values of flexural rigidity (Te = 1.5–3.5) are able to account for the form of the ridge. Te of 5 km or greater result in ridges that are too big and steep.

Thermal rejuvenation of the margin during ridge-transform intersection is predicted to have generated at least 1200 m of uplift, possibly as much as 1540 m, depending on the exact timing of ridge passage and the degree of subaerial erosion suffered. These estimates are based on the difference between the recorded water depths at 90 Ma and the depths predicted from the flexural backstripping of the interpreted seismic data, assuming simple thermal subsidence after extension. Uplift estimates are substantially more than those derived by backstripping assuming local isostatic compensation [*Clift et al.*, 1997] and somewhat more than thermomechanical models that assume regional isostatic compensation [*Gadd and Scrutton*, 1997].

We question the reliability of the higher crustal thicknesses imaged under the Marginal Ridge by refraction surveys [*Sage*, 1994], as the flat Moho is difficult to reconcile with the apparent flexural deformation of the upper crust. Nonetheless, the presence of thicker crust under the Marginal Ridge compared to the adjacent Deep Ivorian Basin could also generate much of the observed feature. Definitive testing of this hypothesis must be undertaken where the crustal thicknesses close to the continent-ocean transition are well defined.

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