Crustal structure of the continent–ocean transition off the Côte d’Ivoire–Ghana transform margin: implications for thermal exchanges across the palaeotransform boundary

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Accepted 2000 June 6. Received 2000 May 9; in original form 1999 May 17

SUMMARY

Crustal structure of the continent–ocean transition off the Côte d’Ivoire–Ghana transform margin is determined by 2-D modelling of wide-angle seismic data and gravity data. The resulting models, together with geological data recently collected by drilling on the margin, are compared with available thermomechanical models of transform margins. In particular, we test the predicted effects on the transform margin crust of continental heating by the adjacent oceanic lithosphere. These effects include: (i) ductile flow of the lower continental crust either parallel to the transform direction or perpendicular to it, from the continent towards the ocean; and (ii) thermal uplift of the transform margin.

In the region of the Côte d’Ivoire–Ghana transform margin, the seismic and gravity data indicate that the continental crust remains uniform in thickness as the continent-ocean boundary is approached. The interpreted models lack a transitional domain between continental and oceanic crusts, and abnormally thin oceanic crust is found along the margin. These results exclude large-scale ductile flow of lower continental crust during the margin formation. The geological data are, moreover, not consistent with thermal uplift of the margin during the period of contact between the transform margin and the spreading centre. Our geophysical and geological data thus infer only moderate heating of the transform continental margin by the adjacent oceanic lithosphere. Moderate heating of the Côte d’Ivoire–Ghana transform margin during margin formation implies that the thermal exchange between the transform margin and the adjacent spreading centre were lower than predicted by available thermal models. Two main factors may have contributed to reduce thermal exchanges across the continent-ocean boundary. The first factor is that the oceanic edge may have been colder than assumed by thermal models. An abnormally cold spreading axis along the transform margin when the oceanic crust was emplaced is supported by the atypical structure of the oceanic crust along the Côte d’Ivoire–Ghana transform margin. The oceanic crustal structure is similar to that found along oceanic fracture zones, where oceanic crust is also emplaced near a ridge-transform intersection. In both cases, abnormally low thermal-gradients at the spreading centre along the transform margin may be explained by the cumulative effect of: (i) a cold edge effect of the adjacent continental lithosphere; (ii) accretion at the end of a ridge segment, away from the main magma supply point along the spreading axis; and (iii) an efficient upward cooling of the oceanic lithosphere owing to deep hydrothermal circulation within the faulted lithosphere. The second factor causing moderate thermal exchanges is the possibility of large amounts of water in oceanic lithospheric faults, which reduces the lateral thermal conduction in the lithosphere, since water is much less conductive than rock.

During the margin formation, the oceanic edge thus seems to have acted as a thermal buffer between the cold continental margin and the hot spreading centre. We
propose that this thermal buffer hypothesis may be generalized to most transform margins, implying a lower amount of continental heating than assumed by the available thermomechanical models.

Key words: continental margins, crustal structure, gravity, seismic structure, thermal conductivity, transform faults.

1 INTRODUCTION

Transform margins bring into contact continental and oceanic lithospheres with strongly different thermal and mechanical properties. The thermal and mechanical contrasts across the transform boundary are particularly important when the continental margin faces the spreading centre of the adjacent plate. This situation implies thermal and mechanical exchanges across the transform boundary. Such changes were widely modelled during the last few years, mostly in order to constrain the consequences of thermal exchanges on the crustal structure and the crustal deformation better.

The purpose of this paper is to compare transform margin structure predicted by available thermomechanical models with the observed structure inferred from new geophysical and geological data available at the Côte d'Ivoire-Ghana transform margin. This comparison will enable us to evaluate the thermal exchanges across the continent–ocean boundary in the Côte d'Ivoire case, and specifically the degree of continental heating produced by an oceanic spreading centre passing along the transform margin.

2 THERMOMECHANICAL EVOLUTION OF TRANSFORM CONTINENTAL MARGINS

Transform margin evolution can be divided into three main stages (Maasche & Blaise 1987) (Fig. 1). During stage 1, active shearing occurs between two continental lithospheres which are similar in composition and thickness (Fig. 1a). The period of continent–continent contact terminates when a normal thickness continental lithosphere comes into contact with extended continental lithosphere located on the facing plate (Fig. 1b). This configuration marks the beginning of a progressive asymmetry of the thermal and mechanical properties between the two adjacent lithospheres, which are similar in nature but different in thickness. During stage 2, transform motion occurs between oceanic and continental lithospheres, and the spreading centre sweeps along the transform margin (Fig. 1c). At this time, the continent–ocean boundary is a sharp transition between two lithospheres with strongly different mechanical and thermal properties. Stage 3 begins once the spreading centre has completed its pass along the transform margin; then, shearing ceases along the continent–ocean boundary, and the continental and oceanic lithospheres are part of a single plate (Fig. 1d).

Following this schematic evolution, the thermomechanical evolution of transform margins has been modelled widely (Todd & Keen 1988; Lorenzo & Vera 1992; Gadd & Scrutton 1997). These studies used a thermal model based on time-dependent cooling, in which oceanic lithosphere heats continental lithosphere by lateral conduction across a vertical boundary (Qc in Figs 1b–d). The continental heating increases during stage 2, as the oceanic spreading centre approaches the observation point M (Fig. 1c). At the end of stage 2, the temperature of the transform margin in contact with the spreading centre could reach 700 °C at 8 km depth, and 1200 °C at about 12 km depth (Todd & Keen 1989; Lorenzo & Vera 1992). The temperature gradually decreases during stage 3, as the spreading centre moves away (Fig. 1d). Frictional heating generated by tectonic activity at the transform fault (Qf in Figs 1a–c) appears to be negligible relative to conductive heating (Todd & Keen 1989).

The thermomechanical models suggest that the continental heating alters the mechanical behaviour, and thus the crustal structure, of transform margins. First, heating greatly reduces the viscosity of continental material at deep crustal and lithospheric levels. Thermal models show that the viscosity ratio (predicted viscosities of the thermal model/viscosities at thermal equilibrium) at the continental edge is decreased by a factor of 100 within a 50–80 km wide area lying along the transform boundary (Reid 1989; Todd & Keen 1989). Depending on the stage of margin evolution, two kinds of stress can induce ductile flow of the low-viscosity continental material. (i) During strike-slip motion (stages 1 and 2), shear stress and viscosity coupling across the transform boundary can induce a ductile flow of lower continental crust parallel to the transform motion. The rate of such ductile flow depends on a balance between the viscous coupling across the transform boundary and the crustal viscosities. During stage 2, as the spreading centre approaches, both the continental viscosity and the viscosity coupling across the transform boundary are reduced by the temperature increase at the continental edge. The reduced viscosity coupling across the transform boundary becomes too low to allow ductile flow parallel to the transform boundary (Reid 1989; Vagnes 1997). The gradual decrease in flow rate induces thinning of deep continental crust, resulting in isostatic subsidence (Reid 1989; Vagnes 1997). (ii) After stage 1, the difference in crustal thickness of the adjacent crusts implies large lateral density variations across the transform fault, and consequently a strong lateral lithostatic gradient (P in Figs 1c and d). The resulting horizontal stress, which could reach 100 MPa at 12 km depth, may induce a ductile flow perpendicular to the continent–ocean boundary, directed away from the lower continental crust towards the adjacent oceanic lithosphere.

Second, thermal expansion of the continental lithosphere induces uplift of the transform margin, which is consequently exposed to subaerial erosion (Todd & Keen 1989; Lorenzo & Vera 1992; Gadd & Scrutton 1997; Vagnes 1997; Fig. 1c). Once the oceanic spreading centre has passed by, the continental edge thermally subsides but, for a while, the uplift still progresses into the continent due to lateral conduction (Fig. 1d).

The continental heating predicted by thermomechanical models implies permanent changes in the crustal structure...
Figure 1. Schematic model of the 3D evolution of transform margins, based on available thermomechanical models. Medium grey corresponds to continental lithosphere, light grey to extended continental lithosphere, and dark grey to oceanic lithosphere. Point M, located on the continental edge (plate A) successively faces: (a) a continent; (b) an extensional margin; (c) an oceanic lithosphere that gradually thins with the decreasing distance to the adjacent spreading centre; (d) an oceanic lithosphere that gradually thickens with the increasing distance from the spreading centre. During stages 1 and 2, M experiences active shearing (a to c). By the end of stage 2, active shearing ceases once the spreading centre has passed; during stage 3, plates A and B belong to a single plate (d). Qf indicates the moderate frictional heating generated by tectonic activity along the transform fault (a to c). Qc indicates the conductive heating from the hot oceanic lithosphere towards the continental lithosphere (b to d). The heating is supposed to induce thermal uplift of the continental edge, with maximum effect when adjacent to the spreading centre. P indicates the lithostatic pressure due to the sharp crustal thinning across the lithospheric boundary. Lithostatic pressure is maximum during continent-ocean contact (c and d). Theoretical location of the Côte d’Ivoire-Ghana example is indicated by a rectangle on the insets showing the 2-D transform margin evolution.

which should be detectable by geological and geophysical investigations. First, crustal thinning either by erosion (upper crust) or by ductile flow (lower crust) can be determined by geophysical modelling of wide-angle seismic data and gravity data. Second, the amount and age of uplift related to thermal expansion can be deduced from seismic reflection data interpreted together with geological samples collected on transform margins.

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3 THE CÔTE D’IVOIRE-GHANA TRANSFORM MARGIN EXAMPLE

3.1 Geodynamical setting

In this paper, we focus on the Côte d’Ivoire-Ghana transform margin. Because this margin results from a major transform motion (horizontal slip now reaches 890 km; Mascle 1976), we can expect that the consequences of the continental heating on the crustal structure are particularly well-expressed. In the study area, the transform margin forms the southern edge of an extensional margin (Figs 1b and 2b). This position close to an extensional/transform margin intersection implies that the three main stages of evolution of the transform margin were recorded by the thick sedimentary cover deposited on the subsiding extensional margin. Owing to this particular position, it was possible to deduce the timing of the main tectonic events related to the transform margin evolution from a large set of seismic reflection data available on the continental margin, interpreted together with samples provided by deep drilling (ODP Leg 159) and diving cruises (Mascle et al. 1997; Basile et al. 1998). The deep structure of the continental margin and the adjacent oceanic domain is, moreover, well-constrained by wide-angle and multichannel seismic data (Edwards et al. 1997a, b; Peirce et al. 1996; Sage et al. 1997a, b). Finally, geological and geophysical data provide estimates consistent with kinematic reconstructions (Klitgord & Schouten 1986; Nürnberg & Muller 1991) for the duration of each evolutionary stage of the Côte d’Ivoire-Ghana transform margin: continent-continent shearing (stage 1) lasted about 20 Ma, continent-ocean shearing (stage 2) lasted 10 Ma, and the transform margin has been passive (stage 3) for 80 Ma (Basile et al. 1998). These values are approximately the same as values used for computations of thermal models, so that a direct comparison between the Côte d’Ivoire-Ghana transform margin data and the available thermomechanical models can be made.

3.2 Crustal structure of the Côte d’Ivoire–Ghana transform margin and the adjacent oceanic crust

In the northern part of the Gulf of Guinea, the Côte d’Ivoire–Ghana transform margin is located between a deep continental basin (Deep Ivorian Basin) to the north, and the Gulf of Guinea abyssal plain to the south (Figs 2 and 3). In the Deep Ivorian Basin, velocity versus depth models have been deduced from wide-angle data (Sage 1994; Peirce et al. 1996) together with multichannel seismic data (Sage 1994). The models indicate velocities and crustal thicknesses indicative of an E–W extensional margin. The structural relations between the extensional margin and its southern and northern transform borders suggest that the basin can be considered as a mega pull-apart basin (Mascle & Blarez 1987).

The southern border of the extensional margin is a well-expressed marginal ridge (Figs 2 and 3), about 50 km wide and 130 km long, which marks the southwest end of the Côte d’Ivoire–Ghana transform margin. The deep structure of the marginal ridge is given by wide-angle seismic modelling and shows the same crustal structure as the adjacent extensional margin, with tilted blocks and a westward gradual thinning of the continental crust (Sage et al. 1997b). Such a crustal structure strongly suggests that the marginal ridge is part of the Côte d’Ivoire pull-apart basin.

The detailed structure of the adjacent oceanic crust is deduced from wide-angle seismic modelling and multichannel seismic profiles (Sage et al. 1997a). The oceanic crust is abnormally thin (3–4 km) and is characterized by the absence of velocities indicative of oceanic layer 3 (6.6–7.2 km s⁻¹), and by a high velocity gradient between crust and mantle velocities (Edwards et al. 1997a; Sage et al. 1997a). On the other hand, the multichannel seismic profiles show lateral changes in the seismic facies and attenuation below the top of basement, suggesting a heterogeneous oceanic basement (Sage et al. 1997a). The atypical velocities, together with the heterogeneity displayed by the multichannel seismic profiles, can be explained by an oceanic basement mostly composed of mantle-derived ultramafics, but associated with discontinuous gabbroic and basaltic bodies (Sage et al. 1997a), similar to oceanic basement emplaced along intraoceanic fracture zones or at slow spreading ridges (Cannat 1993, 1996; Niu & Batiza 1994).

Additional wide-angle seismic and gravity data are presented below, in order to constrain more accurately the crustal structure across the continent-ocean transform boundary. With these new constraints, we are better able to evaluate potential continental ductile flow parallel or transverse to the transform margin during stages 2 and 3 of its formation.

4 CRUSTAL STRUCTURE OF THE CONTINENT–OCEAN BOUNDARY ALONG THE CÔTE D’IVOIRE-GHANA TRANSFORM MARGIN

4.1 Crustal structure at the Côte d’Ivoire–Ghana continental slope from 2-D wide-angle seismic modelling

Wide-angle seismic line L8, 84 km in length and parallel to the transform direction, was located at the bottom of the continental slope (Fig. 2a). Data were recorded during RRS Charles Darwin cruise 55 by two digital ocean bottom seismographs (DOBSSs), separated by 75 km; the DOBSSs were operated by the Institute of Oceanographic Sciences Deacon Laboratory (Kirk et al. 1982; Kirk & Whitmarsh 1992). The line was shot at a 300-m shot spacing, using a 7064 m² (116 L) air gun array. Profile L8 was crossed by 25 near-perpendicular single channel seismic reflection lines, spaced at a distance of 2.5 km (Fig. 2a); the reflection lines were recorded during the EQUAMARGE II cruise (1988). The profile was also crossed by two multichannel seismic lines recorded during the EQUASIS cruise (1990, 1991).

Wide-angle seismic data recorded by DOBSS W (at the west end of the line) and DOBSS E (at the east end) are shown on Figs 4(a) and 5(a), respectively. 2-D modelling of traveltimes was first carried out by ray tracing for both refracted and reflected arrivals. The synthetic seismograms were then compared to model the observed amplitudes (Figs 4b and 5b), using the asymptotic zero-order ray theory (Cerveny et al. 1977; Zelt & Smith 1992). Ray trace diagrams and the corresponding traveltimes computed for our final model are displayed in Figs 4(c and d) and 5(c and d); the final 2-D velocity–depth model is shown on Fig. 6, and the velocities and thicknesses computed for the model units are given in Table 1. Velocity uncertainties, based on χ²-tests, were computed by inversion.

The data show three high amplitude reflections (Figs 4a and 5a) that were correlated with the main reflectors displayed on the reflection profiles recorded on the continental slope (Fig. 3).
The corresponding refracted arrivals define three sedimentary layers (units I-III; Figs 1 and 6, Table 1) which overlay the acoustic basement identified on the reflection lines (Fig. 3).

The top of acoustic basement is marked on both the reflection and wide-angle seismic profiles by a high amplitude reflection (SO) (Figs 3 and 4a). The geometry of this interface was given by the reflection profiles crossing line L6.

The acoustic basement was divided into several units (Fig 6) defined by refracted and reflected arrivals clearly identified on both DOBS record sections. A high velocity contrast between units IV and V was needed to reproduce the amplitude of reflection P2P (Figs 4b and 5b). Unit V is characterized by a P2 refracted arrival with lower amplitudes, and two main internal reflections (Figs 4a and 5a), and was divided into three units for modelling. However, only small velocity contrasts between units Va, Vb and Vc were needed to model the amplitude of the waves reflected within unit V (0.1 km s⁻¹ for Va/Vb interface, and 0.15 km s⁻¹ for Vb/Vc interface). The vertical velocity

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Table 1. For each unit of one best velocity-depth model computed for wide-angle seismic data line L8 (see Fig. 6), are indicated the velocity ranges at the top and at the bottom of the unit and the thickness range along the line. The uncertainties given for velocities and thicknesses are based on \( \chi^2 \) tests and were computed by inversion.

<table>
<thead>
<tr>
<th>Unit number</th>
<th>Velocities at top of the unit (km s(^{-1}))</th>
<th>Velocities at bottom of the unit (km s(^{-1}))</th>
<th>Unit thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>1.6±0.01</td>
<td>1.71±0.01</td>
<td>410±50±15</td>
</tr>
<tr>
<td>II</td>
<td>2.05±0.05</td>
<td>2.23±0.05</td>
<td>470±380±30</td>
</tr>
<tr>
<td>III</td>
<td>2.35±0.05</td>
<td>2.51±0.05</td>
<td>5±80±30</td>
</tr>
<tr>
<td>IV</td>
<td>4.8±0.05</td>
<td>4.9±5.15±0.05</td>
<td>20±2700±50</td>
</tr>
<tr>
<td>V(a)</td>
<td>5.8±6.0±0.05</td>
<td>6.3±6.33±0.05</td>
<td>1800±3820±100</td>
</tr>
<tr>
<td>V(b)</td>
<td>6.4±0.05</td>
<td>6.5±6.53±0.35</td>
<td>3000±3010±500</td>
</tr>
<tr>
<td>V(c)</td>
<td>6.6±6.65±0.1</td>
<td>6.67±6.85±0.1</td>
<td>2900±8200±500</td>
</tr>
<tr>
<td>V(v)</td>
<td>5.67±7.0±0.1</td>
<td>6.7±7.1±0.1</td>
<td>1300±2000±500</td>
</tr>
<tr>
<td>VI</td>
<td>8.1±0.15</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 4. Ray tracing of the best fit model for DOBS W (Line L8, see location in Fig. 3). (a) Display of the observed air gun data. Vertical geophone traces plotted with a reduction velocity of 6.0 km s^{-1}, and band-pass filtered from 4 to 15 Hz. Amplitudes are scaled linearly with range. SO corresponds to reflection at top of the acoustic basement displayed on multichannel seismic data (see Fig. 3). PmP and PmP corresponds to intracrustal reflections, and PmP corresponds to reflection at the bottom of the crust. SO and PmP mark refracted arrivals within the acoustic basement defined on multichannel seismic data. (b) Synthetic seismograms plotted with reduction velocity of 6.0 km s^{-1} and computed for velocity-depth model displayed on (d) and in Fig. 6. Amplitudes are scaled linearly with range. (c) Computed travel-time solution for the 2-D model (pluses) shown on (d) is superimposed on the observed data plotted with the same characteristics as (a), (d) 2-D velocity-depth model and rays.

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Figure 5. Ray tracing of the best fit model for DOBS E (Line L8, see location in Fig. 2). Same legend as Fig. 4.
Figure 6. Final velocity–depth 2-D model computed for wide-angle seismic line L8. Solid lines indicate velocity contrasts; dashed lines indicate change in velocity gradient. Velocities are in km s⁻¹. For unit I to unit III, the bracketed values correspond to averaged velocities computed for the units. For unit IV to unit VII, velocities are given at the top and at the bottom of each unit. Arrows mark the DOBS positions. G marks the intersection with the gravity line discussed below.

computed for the Deep Ivorian Basin (Peirce et al. 1996) and for the marginal ridge (Sage et al. 1997b) (Figs 7b and 8). The only exception is the occurrence of the reflective unit (unit V') at the base of the continental crust, which is not observed beneath the Deep Ivorian Basin and the marginal ridge (Fig. 8). The velocities modelled for this unit are almost the same as velocities for unit V and are similar to velocities computed for the base of the continental crust beneath the Deep Ivorian Basin and the marginal ridge (Figs 7b and 8), suggesting that unit V' is part of the continental crust. The complexity of the reflections observed within the lower continental crust close to the palaeotransform boundary may be explained by shear motion effect at ductile continental levels. But we can however, not exclude the possibility that the complexity of the PmP reflection may be due to out-of-plane structure, since the continental crust rapidly thins towards the transform boundary, as indicated by previous gravity models (Sage et al. 1997b; Pontoise et al. 1990).

Finally, the wide-angle seismic model L8 shows that, at the base of the continental slope, the crustal structure is typical of an extensional margin, with a gradual westward thinning, and is similar to the structure of the adjacent marginal ridge (Fig. 8). The velocity–depth models computed for the Ivorian Basin, the marginal ridge and the continental slope, suggest that perpendicular to the transform margin, the continental crust maintains a nearly constant thickness towards the transform boundary, as far as the base of the continental slope (Fig. 8); no thinning of continental crust related to the transform margin is observed. On the oceanic side, multichannel seismic data display intracrustal reflectors which are continuous from EQR4 to the basement slope-break; this suggests that the oceanic crust extends up to the continental slope (Sage et al. 1999a). These results support a continent–ocean transition less than 5 km wide and a steep continental Moho.

In order to check crustal models deduced from wide-angle seismic data with an independent set of geophysical data, and in order to determine the likely crustal structure and geometry of the continent–ocean transition, we present below a gravity model across the transform margin and the adjacent oceanic crust.

4.2 Gravity modelling of the continent–ocean transition

The gravity profile discussed in this section is 200 km long, extending from the Deep Ivorian Basin to the adjacent oceanic domain (G in Fig. 2a). The model is however, calculated from a longer profile, which extends 500 km into the African continent and 500 km into the Atlantic ocean, in order to include the gravimetric effect of distant oceanic and continental structures on the gravity anomaly observed close to the ocean–continent transition.

Marine gravity data were recorded during ATLANTIS 1973 and EQUAMARINE 1988 cruises (G1 and G2 in Fig. 2a). Onshore gravity data were taken from IRD and DMA–ORSTOM databases. Data were projected along a 155° direction, normal to the general trend of the transform margin (Fig. 2a), in order to avoid 3D effects related to the sharp thinning of the crust across the continent–ocean transition.

The starting 2-D gravity model was deduced from both the wide-angle seismic data and the reflection seismic data available along the gravity profile. We obtained the geometry of the sedimentary layers by converting the two-way travel times of the available seismic reflection profiles into depths, using the velocities from wide-angle seismic modelling (Sage 1994; Peirce et al. 1996; Sage et al. 1997a,b). Depths to deeper interfaces were given by the 2-D velocity–depth models. The averaged velocities modelled for the crustal layers were converted into densities, using the linear relationship given by Carlson & Raskin.
(1984) for igneous oceanic crust, and by Barton (1986) for sediments and igneous continental crust. The density of mantle was assumed to be constant (3.33 g cm\(^{-3}\)) along the profile, because no velocity change was identified from one transverse velocity-depth model to another. A density of 1.027 g cm\(^{-3}\) was assumed for the sea water.

The length and orientation of the gravity profile used for modelling, as well as constraints imposed by the reflection data on the units' geometry, especially at the oceanic edge, improve the previously published gravity models established across the Côte d'Ivoire-Ghana transform boundary (Pontoise et al. 1990; Sage et al. 1997b).

The computed gravity anomaly for the 2-D model was compared to the observed anomaly, and the densities and interface geometries were adjusted to achieve a satisfactory fit. In order to test the validity of the velocity-depth models, we achieved a forward modelling fit by moving the interfaces between the transverse refraction lines only, and by changing the densities in the ranges given by the velocity–density laws defined above. Because the averaged velocities of the crustal layers were constant from one wide-angle seismic model to another, within 50 km of the palaeotransform boundary (see EQR2, EQR3, L8, Fig. 8), we chose to keep the densities laterally constant. L9, which lies 75 km away from the palaeotransform boundary, presents a slightly different velocity-depth profile within continental crust (Peirce et al. 1996; Figs 7 and 8). For this line, a main crustal layer, with velocities between 3.5 km s\(^{-1}\) and 7.3 km s\(^{-1}\) replace units III to V modelled for refraction lines EQR2, EQR3, and L8 (Fig. 8). For this reason, and because we found no evidence for lateral change within the upper crustal structure from multichannel seismic data between EQR2 and L9, the density values given for EQR2, EQR3 and L8 were extrapolated towards L9, within bounds compatible with the average velocities computed for line L9 (Fig. 7b).

The final density model is displayed on Fig. 9. The misfit between the computed and observed gravity anomalies is less than 1.5 mGals along the profile (Fig. 9a). The proposed model respects the check points provided by wide-angle seismic models (Fig. 9b), and the modelled densities are still within the bounds given by Barton (1986) and Carlson & Raskin (1984).
On the oceanic side, the averaged amplitude of the gravity anomaly is related to the crustal thickness; shorter wavelengths of the gravity anomaly are related to the top of basement geometry, which is constrained by seismic reflection data. Close to the continent–ocean boundary, the computed gravity anomaly fits the data well if the crust–mantle transitional layer observed on the EQR4 wide-angle seismic model is extended up to the slope-break observed at the top of the acoustic basement (Fig. 9b). This transitional layer must be thickened by a factor of two towards the continent to fit the observed gravity anomaly.
anomaly. This is in good agreement with multichannel seismic data which show such a thickening of the reflective zone attributed to the crust-mantle transitional layer (Sage et al. 1997a).

The marginal rise is characterized by a positive gravity anomaly of 35 mGal (Fig. 9a). The gravity anomaly is correctly modelled if the crustal interfaces given by the wide-angle seismic models for EOR3 and L8 rise between the two lines. A deeper source for the anomaly, at the depth of the Moho for example, would induce a gravity anomaly wider than observed.

The continent-ocean boundary is characterized by a -50 mGal anomaly located 2 km south of the wide-angle seismic line L8, at the basement slope break (Fig. 9a). Close to the continent-ocean boundary, the amplitude and the wavelength of the gravity anomaly are mainly controlled by the Moho depth. The computed gravity anomaly fits the data well for a 21.4 ± 0.2 km thick Moho beneath the marginal rise and the continental slope, and a sharp Moho rise to 10 km depth beneath the oceanic crust. The steeper the Moho rise, the greater the amplitude of the difference between the positive and negative peaks of the gravity anomalies. For a smoother rise occurring over 4 km, the difference between the positive and the negative peaks decreases to 60 mGal. The Moho depth beneath the continental edge mainly controls the amplitude of the positive anomaly. A 300 m variation in Moho depth induces a 10 mGal misfit on the gravity anomaly amplitude.

Despite the fact that solutions deduced from gravity modelling are non-unique, our final 2-D gravity model is consistent with the main results of wide-angle seismic data. Gravity modelling, moreover, gives the crustal structure perpendicular to the transform margin, between the refraction lines.

Suggested by the along-strike velocity-depth models (Fig. 5), the density-depth model is compatible with identical crustal (unit V) structures from the Deep Ivorian Basin (L9) to the foot of the continental slope (L8). This result suggests that the deep crustal structure of the extensional margin was not altered by the transform motion even adjacent to the transform margin itself. On the proposed gravity model, the continental Moho does not change depth, whereas the top of the continental basement does change substantially. Despite the fact that small variations in Moho topography (< 1 km) cannot be resolved, this locally implies a state of isostatic imbalance. At 20 km and 80-90 km along the gravity model, the top of basement corresponds to the top of tilted blocks (Sage 1994).
Altimetric studies conducted on other extensional margins show that the tilted blocks are characterized by short wavelengths (<50 km) which are supported by lithospheric rigidity (Diamant et al. 1986; Verhoef & Jackson 1991) so that the tilted blocks are not balanced at depth by Moho depth variations. At the continental slope, the mass deficit suggested by the gravity models implies upward vertical forces at the continental edge. There, the state of isostatic imbalance could be explained by the gradual erosion of the continental slope.

The gravity model is compatible with oceanic crust thinning towards the ocean–continent boundary, as previously suggested by the wide-angle seismic models EQR4 and EQR5 (located in Fig. 2; Sage et al. 1997a). The gravity model shows that the anomalously thin oceanic crust may extend over a 50-km wide area southeast of the continental–ocean boundary (Fig. 9), supporting an abnormally low magmatic budget at the north end of the spreading centre during accretion. The gravity model is compatible with a thin oceanic crust underlain by a transitional layer between crust and mantle, as previously suggested by the wide-angle seismic data and the multichannel seismic data (Sage et al. 1997a). Velocity and density modelled for this transitional layer (7.0–8.0 km s$^{-1}$ and 2.99 g cm$^{-3}$, respectively) can be interpreted either by underplating (LASE Study Group 1986; White et al. 1987) or by partial serpentinization of upper mantle peridotites (Spudich & Orcutt 1980; Christensen 1986).

With a thin oceanic crust, hydrothermal circulation, which is expected down to 7–8 km below top of basement, can allow partial alteration of mantle peridotites into serpentinite (Cannat 1993). Partial alteration of the peridotites can explain the density and velocity of the transitional layer, and supports the hypothesis of a cold accretory process when the crust was emplaced along the transform margin. On the other hand, underplating is observed in particularly hot environments (such as near a hot spot), implying thick magmatic oceanic crust and, in some circumstances, intrasediment volcanism (Lorenzo et al. 1991) which are not compatible with the seismic data available on the transform margin and the adjacent oceanic edge off Côte d'Ivoire–Ghana. These observations rule out an underplating hypothesis.

Our gravity model, constrained by the wide-angle seismic data, provides strong support for a sharp continent–ocean transition at the transform margin, which is modelled as a subvertical interface between continental crust to the north and anomalous oceanic crust to the south. There is no transitional domain (other than atypical oceanic basement) between continental and oceanic lithospheres. Perpendicular to the transform margin, the continental Moho shallows rapidly at the basement slope-break from 21.5 km to 10 km depth over a distance as short as 2.5 km (Fig. 9).

The gravity line orientation, normal to the general trend of the transform margin, minimizes the 3-D effect of the sharp crustal thinning towards the transform boundary. However, we cannot exclude a 3-D effect of the gradual crustal thinning related to the extensional margin. On the eastern side of the gravity line, the crustal thickness seems to remain constant over a few tens of kilometres (Fig. 8). On the other hand, continental Moho rapidly shallows westward, implying a mass excess relative to the location of gravity line G (Fig. 2), which may have influenced the gravity modelling. In that case, the proposed model possibly presents slightly underestimated continental Moho depths, but this possible 3-D effect would not alter the steepness of the Moho at the continent–ocean boundary.

4.3 Transform margin uplift and maximum continental heating dated from geological data

The numerous seismic reflection lines shot over the Côte d'Ivoire–Ghana marginal ridge (Fig. 2) display a well-defined unconformity (Fig. 3) which marks the last uplift above sea level recorded by the transform margin. This unconformity was previously attributed to thermal uplift coinciding with the spreading centre passing along the margin (Basile et al. 1993), which is expected to have occurred about 80 Ma (Campanian), from kinematic reconstructions of the South Atlantic (Klitgord & Schouten 1986; Hünrb erg & Müller 1991). However, drilling of the unconformity at Sites 959 and 960 of ODP leg 139 revealed that the erosion was older than the late Albian (97–100 Ma) sediments lying on the unconformity at Site 959, and that the marginal ridge was already tilted northward at that time (at the top of the ridge, Site 966 was emergent whilst on the northern slope, Site 959 was at about 100 m below sea level; Basile et al. 1998). In this area, no more recent uplift has been recorded by the sedimentary section (Basile et al. 1998), indicating that the main margin uplift occurred during the earlier intracontinental shearing stage of the margin formation. Furthermore, the geological samples show only moderate heating (120–180 °C), as indicated by low-grade metamorphism (Bekheli et al. 1996; Benkhellil et al. 1997; Bouillin et al. 1997), clay transformations (Holmes 1998) and fluid inclusions (Lespinasse et al. 1989). The maximum palaeotemperatures were recorded by geological samples located below the late Albian unconformity, and consequently pre-date the end of the intracontinental transform motion.

Rocks younger than late Albian present no evidence of subsequent heating contemporaneous with passing of the spreading centre. The transform margin uplift is, moreover, followed during stage 2 by subsidence of the continental edge at Early Coniacian time (88 Ma) (Oboh-Ikuenobe et al. 1995), that is 10 Ma before the expected time of the spreading centre passing along the transform margin.

The only geological event that can be related to passing of the oceanic spreading centre is an acceleration of the continental slope erosion. Apatite fission tracks measured on samples recovered either at the top of the marginal ridge or along the continental slope show that the rocks rapidly cooled down to 60 °C at different times, depending on the sample location along the margin (Bouillin et al. 1998). Along the continental slope, a first set of ages at 90 Ma suggests that the slope erosion started some 10 Ma before the spreading centre passed. The sediments delivered by this erosion event were probably deposited within the active transform fault zone, as the rocks observed and sampled in the Romanche fracture zone, 400 km west of the Côte d'Ivoire–Ghana continental margin (Honeoe et al. 1994). A second set of fission tracks ages between 80 and 70 Ma provides evidence for an acceleration of denudation processes immediately after the spreading centre passed. The products of this second erosional phase were deposited at the foot of the continental slope, in the depressions of the newly emplaced oceanic crust. This second phase, clearly less important than the first one, could be related to local reactivation of the transform margin in the vicinity of the spreading centre, in good agreement with the kinematic reconstructions (Klitgord & Schouten 1986; Nürnberg & Müller 1991).

In summary, the uplift and maximum palaeotemperature recorded by the marginal ridge sediments can both be attributed...
to the intracontinental shearing stage of the transform margin formation (stage 1). During continent-ocean shearing (stage 2), the uplift was followed by subsidence of the marginal ridge, and by the formation of the southern continental slope that probably became steeper close to the adjacent spreading centre.

5 DISCUSSION

5.1 Evidence for limited heating of the Côte d'Ivoire–Ghana continental transform margin

Perpendicular to the transform margin, the wide-angle seismic and gravity data provide evidence of a constant thickness of the continental crust within 100 km of the continent-ocean boundary, but of an abnormally thin oceanic crust lying along the transform margin. This result, corroborated by the gravity model displayed in Fig. 9, rules out the hypothesis of crustal thinning at the transform margin by large scale ductile flow at deep crustal levels during stages 2 or 3 of the transform margin formation.

On the other hand, geological data from the Côte d'Ivoire–Ghana transform margin imply that at main thermal event (Holmes 1998) occurred before the end of the Aptian (Basile et al. 1998; Bouillin et al. 1998) and the maximum uplift before late Albian (Basile et al. 1998), coeval with intracontinental transform motion (stage 1). Subsequently, both subsidence and cooling occurred at the continental edge (stages 2 and 3; Basile et al. 1998). These results are not compatible with available thermomechanical models which indicate a maximum thermal uplift when the spreading centre passes, and continued uplift for millions of years thereafter due to lateral conduction towards the adjacent continent (Todd & Keen 1989).

We conclude that during the contact between the Côte d'Ivoire–Ghana transform margin and the adjacent spreading centre, the temperature along the transform margin was not high enough either to allow ductile flow, or to induce thermal uplift of the continental edge. The thermal exchange across the continent-ocean boundary was thus lower than predicted by available thermomechanical models.

Two possible hypotheses may explain the low amount of thermal exchange across the continent-ocean boundary: (i) the thermal gradient at the adjacent spreading centre was lower than expected by models; (ii) the structure of the lithosphere close to the continent-ocean boundary prevented efficient thermal conduction between the hot oceanic lithosphere and the cold continental margin.

5.2 Restricted heat supply from oceanic lithosphere toward the Côte d'Ivoire–Ghana transform margin

During the direct contact between a transform margin and the adjacent spreading centre, the available thermal models postulate that the amount of heat lost by lateral conduction towards the adjacent continent is balanced by heat supply from the ridge axis. Thermal models thus assume a constant thermal gradient, typical of spreading centres, throughout the adjacent oceanic lithosphere. The atypical structure of the oceanic crust emplaced along the Côte d'Ivoire–Ghana transform margin is, however, common in low magma budget accretionary processes (Edwards et al. 1997a; Sage et al. 1997a). This result implies that the thermal gradient at the spreading centre decreased towards the adjacent continent when the oceanic crust was accreted.

A similar atypical structure of oceanic crust due to alteration of accretionary processes is commonly described at slow spreading centres (Bown & White 1994; Niu & Batiza 1994; Cannat 1996), along non-volcanic rifted margins (Boillot et al. 1989), and along oceanic fracture zones (Detrick & Purdy 1980; Cormier et al. 1984; Minshull et al. 1991; White et al. 1992).

The velocity and gravity models proposed for the oceanic crust 70 km away from the Côte d'Ivoire–Ghana continent-ocean boundary are typical of Atlantic oceanic crust (White et al. 1992; Edwards et al. 1997a; Sage et al. 1997a), and thus support a spreading rate that was fast enough to produce oceanic crust of normal thickness (Edwards et al. 1997a).

On the other hand, the progressive thinning of the oceanic crust towards the continent-ocean boundary suggests that the atypical crustal structure is more likely due to the influence of the transform margin.

Along oceanic fracture zones, where oceanic crust also is emplaced at a ridge-transform intersection adjacent to colder lithosphere, two mechanisms, which can also be applied to transform margins, are put forward to explain the reduction in the magma budget. First, the cold edge effect of the adjacent lithosphere may reduce the thermal gradient at the spreading centre (Cormier et al. 1984; Fox & Gallo 1984; White et al. 1992). However, the geological and geophysical studies presents above support a low amount of thermal exchange across the continent-ocean boundary along the Côte d'Ivoire–Ghana transform margin. Moreover, abnormally thin and heterogeneous oceanic crust is also known along second order discontinuities, where the offset is too small for a significant edge effect (Minshull et al. 1991). Second, a sharp reduction in the magma budget is inferred at the ends of ridge segments bounded by fracture zones, due to increasing distance from the main magma supply points which are centred along individual spreading cells of the ridge segments (Detrick & Purdy 1980; Minshull et al. 1991; White et al. 1992).

An expected consequence of a low magma budget accretionary process is that magmatic accretion is partially replaced by tectonic accretion at ridge segment ends (Lagabrielle et al. 1998; Show & Lin 1986). This leads to intense faulting at the end of the ridge segment, including the passive side of the spreading centre where no active shear occurs. The intense faulting, together with the overthickening of the brittle oceanic lithosphere at the cold segment ends (Cannat 1996), is expected to favour deep hydrothermal circulation down to 10 km depth (Calvert & Potts 1985; Cannat 1993). More efficient hydrothermal circulation is known to greatly increase the cooling rate of the oceanic lithosphere (Lin & Parmentier 1989), and thus contributes to diminishing the oceanic thermal gradient along fracture zones.

Near the Côte d'Ivoire–Ghana transform margin, the seismic and gravity models show that the oceanic crust is abnormal over a distance of 50–70 km away from the ocean-continent boundary. This is more than twice as large as the abnormal portion of oceanic crust near oceanic fracture zones (e.g. Louden et al. 1986; Prince & Forsyth 1988). The unusual extent of atypical oceanic lithosphere may result from the remaining cold edge effect of the continental lithosphere on initial seafloor spreading processes just after continental drifting. The abnormally wide cold belt of oceanic lithosphere along the transform margin could also be partly explained by a regional thermal minimum, as observed along some of the Mid-Atlantic Ridge segments (Thibaud et al. 1998). In particular, a thermal
minimum is described in the upper mantle of the present day equatorial Atlantic, along the Romanche fracture zone (Bonatti et al. 1993). Such a thermal anomaly, explained by downwelling mantle flow occurring in the equatorial belt of the Atlantic (Bonatti et al. 1993, 1996) could also have existed in the past, when the oceanic crust was emplaced along the transform margin. Another explanation is that an additional fracture zone runs a few tens of kilometres seaward of the continental slope (Edwards et al. 1997a), leading to broadening of the zone of altered accretionary processes, and of the resulting atypical crust, oceanward.

To consider another theory, lateral conduction from the oceanic lithosphere towards the continental margin may have been less efficient than expected from thermal models. The large amount of sea water expected within the thick faulted lithosphere may have reduced the thermal conductivity of oceanic lithosphere, since water is less conductive than rocks. In summary, the structure of the oceanic crust along the Côte d'Ivoire–Ghana transform margin rules out the assumption of a typical oceanic thermal gradient at the spreading centres when the oceanic crust was accreted, which was inferred by the available thermomechanical models. The low thermal gradient spreading centre, which probably extended up to 50–70 km away from the Côte d'Ivoire–Ghana continent–ocean boundary, seemed to act as a 'thermal buffer' between the hot ascending asthenospheric upwelling and the continental margin, such a 'thermal buffer' would be characterized by reduced thermal conductivity and by efficient upward cooling due to active hydrothermal circulation.

We suggest that the abnormal thermal structure of the end of the spreading centre, which we infer to exist off Ivory Coast, should be a general feature of transform margins.

5.3 Comparison with other transform margins

Only a few geophysical models are available which account for the structure of the oceanic lithosphere adjacent to transform margins. However, some data tend to support the hypothesis that generally abnormal oceanic crust bounds transform margins. Along the Southwest Newfoundland transform margin, wide-angle seismic modelling and multichannel seismic data support an abnormal oceanic crust, in which a typical layer 3 is missing (Reid et al. 1988; Todd et al. 1988; Keen et al. 1990), suggesting a cold accretionary environment (Reid & Jackson 1997). The lack of a typical layer 3 and a gradual downward transition to typical mantle velocities are also documented along the North Baffin Bay transform margin (Jackson et al. 1990; Reid & Jackson 1997), and heterogeneous crust typical of fracture zones is described near the Falkland transform margin from gravity results (Lorenzo & Wessel 1997). These data suggest a relatively reduced magmatic activity associated with a cold thermal state along transform margins when the oceanic crust was accreted.

A relatively cold thermal regime is also supported by some geological studies conducted on transform margins. The amount of continental uplift is in some cases lower than expected from the thermomechanical models (Reid & Jackson 1997; Vågnes 1997). Moreover, when stratigraphic data are available, they indicate that uplift of the transform margin clearly begins during stage 1 (Exmouth transform margin, Lorenzo & Vera 1992; Senja fracture zone, Vågnes 1997) and ends before the spreading centre has passed along the margin (Senja fracture zone, Vågnes 1997).

However, the Exmouth transform margin seems to be an exception. Both geological and geophysical studies indicate an abnormally hot environment during the transform margin formation, due to the vicinity of a hot spot. From seismic and gravity data, high rates of volcanism and underplating are inferred at the transform margin and the adjacent continental domain (Mutter et al. 1989; Lorenzo et al. 1991). Adjacent to the transform margin, the oceanic crust is 6 km thick (Lorenzo et al. 1991), indicating a normal thermal gradient at the spreading centre. For this particular case, the crustal structure of the transform margin seems to be altered by continental heating, as expected from available thermomechanical models. The geological data indicate a 1350-m uplift of the continental edge at the continent–ocean boundary, lasting 40 Myr once the spreading centre had passed by (Lorenzo & Vera 1992). Furthermore, crustal models show that the oceanic crust emplaced along the continent–ocean boundary is underlain by a wedge with density 2.95 g cm$^{-3}$ (Lorenzo et al. 1991) and average P-wave velocities 7.5 km$^{-1}$ (Lorenzo & Vera 1992), compatible with underplated material provided by continental ductile flow. We thus infer that the southern edge of the Exmouth Plateau can be regarded as an example where sufficient temperature at depth related to an abnormally hot environment allowed, at the continental edge, both thermal uplift and lower crust ductile flow towards the adjacent ocean.

6 CONCLUSIONS

Available thermomechanical models of transform margins suggest that significant heating of the continental lithosphere occurs by conduction from the adjacent oceanic lithosphere. Such heating is expected to produce uplift of the transform margin, and thinning of the continental crust toward the continent–ocean boundary, by ductile flow of the lower crust. Data from the Côte d'Ivoire–Ghana transform margin are not consistent with these models, suggesting only moderate heat transfer by conduction from oceanic lithosphere towards the adjacent transform continental margin.

Based on the crustal structure of the adjacent oceanic crust, we propose that the reason for the discrepancy is the low temperature of the oceanic lithosphere when it was accreted along the Côte d'Ivoire–Ghana transform margin, due to: (i) the cold edge effect of the adjacent continental lithosphere; (ii) accretion at the end of a ridge segment, far from the main magma supply point; (iii) upward heat loss provided by efficient hydrothermal circulation; and (iv) reduced thermal conductivity of the hydrated lithosphere.

We propose that the 50–70-km wide belt of atypical oceanic crust and lithosphere bounding the transform margin acted as a 'thermal buffer' between the ascending asthenospheric upwelling and the transform margin. This 'thermal buffer', by causing temperature to decrease gradually along the spreading centre towards the continent, may have prevented efficient heating of the adjacent continent.

These conclusions are based on the Côte d'Ivoire–Ghana example, where the abnormally cold oceanic belt may be partly explained by a regional thermal anomaly. However, some geophysical and geological data lead to support the hypothesis of cold oceanic lithosphere and moderate thermal exchange across the continent–ocean boundary along most transform margins. The concept of moderate thermal exchange between continental and oceanic lithospheres along transform margins.

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has now to be tested by computing new thermomechanical models, with more realistic values of the oceanic thermal gradient and conductivity at the oceanic edge.

ACKNOWLEDGMENTS

The gravity data were processed and modelled in the Laboratoire de géodynamique et gravimétrie (IPG Paris, France) with the generous help of M. Diamant, Ch. Deplus, and S. Bonvallet. The wide-angle seismic data presented above were collected under NERC Grant GR3/701 to the University of Edinburgh. We thank R. Scrutton, R. Edwards and Marvin Saunders for their assistance in collecting and processing the wide-angle data presented above. We also sincerely thank Ph. Charvis and G. Boillot for their constructive criticisms of earlier versions of this manuscript, as well as George Spence for suggestions which helped us to improve the English text. We thank J.M. Lorenzo, W.S. Holbrook and an anonymous reviewer for their reviews which helped us to improve the paper. This work was supported by grants from the 'Ministère des Sciences et Techniques', and the INSU and TRD financial support. Contribution Geosciences Azur (CNRS-UNSA-UPMC-IRD) n°326.

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