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Temporal and spatial correlations between changes in plate motions and the evolution of rifted basins in Africa

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ABSTRACT

We present the results of a backstripping study, incorporating over 200 wells in 14 different basins or groups of basins located on the African interior or continental margins. These results, presented as mean tectonic subsidence rates averaged over appropriate periods, are compared with the opening histories and changes in plate motions of the Atlantic and Indian Oceans and the Tethys-Mediterranean area for seven time-steps from the Late Jurassic until present. The results indicate a temporal and spatial correlation between phases of rapid subsidence in African basins and changes in plate motions and, therefore, are promising in terms of applicability of large-scale plate motions in the studies of paleo-stresses in sedimentary basins, especially at times of initial rifting and formation of subduction zones. Reactivation of the basins appears to be primarily controlled by the orientation of the basin and underlying basement structures with respect to the stress direction. The basins are preferably formed within weaker zones, mostly Pan-African shear zones, and the distance from the active plate boundary to the weak zone plays a minor role in whether or not it will be reactivated. The presence of three important hot spots, which were active during the early stages of break-up of Gondwana, may have been important in preweakening the lithosphere at their sites. This may have played an additional role in the localization of the basin deformation.

INTRODUCTION

The present-day African continent consists of the remnant heart of Gondwanaland, the initial break-up of which started as early as the Late Carboniferous (Fig. 1). Continental rift systems, developed within Africa

at that time, are indicative of the Karoo rifting period. The Karoo lasted until Middle Jurassic and led eventually to the opening of the Indian Ocean and the drifting of Madagascar, India, Australia, and Antarctica away from Africa (Lambiase, 1989; Binks and Fairhead, 1992). During the same period, rifting occurred along the present northwestern margin of Africa, from Middle Triassic onward. Crustal separation commenced in the Early Jurassic, creating the Central Atlantic between Africa and North America (Klitgord and Schouten, 1986; Favre and Stampfli, 1992). The opening of the Central Atlantic caused the development of sinistral transtensional movements between Africa and Eurasia and the opening of the Alboran-Ligurian-Penninic Ocean as an oceanic pull-apart basin (Laubscher and Bernoulli, 1977; Dercourt et al., 1986; Ziegler, 1988). This ocean was located (paleogeographically) in the present-day western Mediterranean area. A second phase of rifting, from Late Jurassic until early Aptian, resulted in the opening of the South Atlantic and the separation of South America from Africa (Fig. 1). At the same time, the West and Central African rift system, which extends from Nigeria and Niger to Kenya, developed as a series of intracontinental rifts (Browne and Fairhead, 1983; Fairhead, 1988a, 1988b; Binks and Fairhead, 1992; Guiraud and Maurin, 1992). Sea-floor spreading in the South Atlantic domain started in the south, in the Cape Basin, at ca. 130 Ma (Rabinowitz and Labrecque, 1979) and propagated to the north, reaching the Gabon margin by ca. 119 Ma (Masce et al., 1986). From early Aptian until late Albian rifting took place in the Equatorial Atlantic domain, and the final breakup, resulting in a complete separation of Africa and South America, occurred in the late Albian (Guiraud et al., 1992). The most recent rifting and opening phase in the

history of Gondwana started in the late Eocene and led to the development of the Red Sea, Gulf of Aden, and East African Rift System (Fig. 1) (Rosendahl et al., 1992).

The Mesozoic and younger tectonic history of the African plate, predominantly that of these three major rifting phases (e.g., Binks and Fairhead, 1992) but also reflecting a number of reorganizations of nearby plate boundaries, is recorded in a series of sedimentary basins distributed throughout the African interior and along the continental margins. Although the locations and orientations of these basins are in most cases controlled by the trends of older Pan-African shear zones (Daly et al., 1989; Maurin and Guiraud, 1993), their origin and further development are intrinsically related to the break-up of Gondwana and the opening histories of the Atlantic and Indian oceans, as well as to the tectonic evolution of the Tethys-Mediterranean regime. Former studies have shown that tectonic processes taking place at plate boundaries influence the state of stress of large parts of the adjoining plates and, by this, affect the evolution of the sedimentary basins located on these plates (Cloetingh, 1988; Ziegler, 1988; Bally et al., 1989). The area affected by the stress field can be as large as a lithospheric plate and is not necessarily limited by the transition of oceanic to continental lithosphere (Cloetingh, 1988; Ziegler, 1988; Leighton and Kolata, 1990). Part of a continent and the adjacent ocean can share a regional stress field as shown by Cloetingh (1988) for the compressional stress field of western and central Australia together with the adjoining Indian Ocean, and the extensional stress field of eastern Australia and the Tasman Sea. This can lead to the reactivation of sedimentary basins located far from the active plate boundary as shown by Ziegler (1988, 1990) for the European plate, where many basins were reactivated during

Figure 1. Paleogeographic reconstructions of the southern continents showing the three major rifting phases in the breakup history of Gondwana. (A) Karoo rifting and early opening of the Central Atlantic Ocean. (B) South Atlantic rifting and beginning of the West and Central African Rift System. (C) Opening of the Red Sea–Gulf of Aden and rifting in the East African Rift System. Dark shaded areas mark the newly formed oceans; light shaded areas are the extensional basins. Dotted areas are the cratons, with W = West African craton; C = Congo craton; K = Kalahari craton. (Modified from Binks and Fairhead, 1992; and Maurin and Guiraud, 1993.)

plate (Fig. 2, Table 1). The studied basins form a representative selection of the total number of basins located on the African plate: along the Atlantic and Tethys-Mediterranean margins (Fig. 3A), along the Indian Ocean margin (Fig. 3B), and from the West and Central African rift system. We will also briefly discuss the influence of the Jurassic Karoo plume (Fig. 3B) and the Cretaceous Tristan da Cunha and St. Helena hot spots (Fig. 3A) on the evolution of the basins. We backstripped over 200 wells from the fourteen African basins or basin groups in order to highlight the timing and intensity of the tectonic events involved in their development from the Late Jurassic until the present.

TECTONIC SUBSIDENCE ANALYSIS

We discuss the subsidence history of the basins labeled in Figure 2 from Late Jurassic until present for seven time slices: Late Jurassic, Neocomian and Barremian, Albian, Cenomanian through Coniacian, Campanian and Maastrichtian, Paleocene through mid-Eocene, and Oligocene-Miocene. These periods cover the main phases of rifting related to the opening of the South and Equatorial Atlantic Ocean (Late Jurassic, Neocomian-Barremian, and Albian) and to the more recent development of the Red Sea, Gulf of Aden, and East African rift system (Paleocene–mid-Eocene and Oligocene-Miocene). Furthermore, we will discuss a time slice that covers the period just after the breakup of Africa and South America (Cenomanian-Coniacian) and one that covers the period just after the Santonian compressional event (Campanian-Maastrichtian). This compressional event marks an important change in the relative movement of Africa and Eurasia. For each of these time slices we have determined the mean tectonic subsidence rate of the individual basins, based on backstripped well data. The water-loaded tectonic subsidence has been calculated for a set of over 200 wells, distributed among the 14 study areas (see Table 2), following procedures described by, for example, Steckler and Watts (1978), Sclater and Christie (1980), and Bond and Kominz (1984). We have adopted the Harland et al. (1990) time scale. The age-control of the wells is primarily based on biostratigraphy. Compaction effects have been corrected for, using standard depth-porosity relations for the different lithologic types observed (cf. Sclater and Christie, 1980), and, where available, depositional water depths have

the Alpine collision. An additional role in the localization of deformation is probably played by the activity of mantle plumes (Bailey, 1992). They tend to preweaken the lithosphere, especially in areas where the lithosphere is already thinned, as underneath most Pan-African shear zones crosscutting the African plate. These shear zones were affected by the Pan-African tectono-thermal event (650–550 Ma), which caused a thin-

ning of the lithosphere (Black and Girod, 1970; Thorpe and Smith, 1974; Bailey, 1992). Most of the younger volcanism appears to be restricted to these thinned areas (Bailey, 1992). In the present study we investigate the relationship between the Late Jurassic and younger plate reorganizations in the Atlantic, Indian, and Tethys-Mediterranean domain and the subsidence pattern in 14 study areas throughout the African

Figure 2. Location of the basins used for the backstripping study. 1, Senegal Basin; 2, Gabon Coastal Basin; 3, Kwanza Basin; 4, Outeniqua Basin; 5, Mozambique and Rovuma Basins; 6, Morondava Basin; 7, Lamu Embayment; 8, Somali Basin; 9, Sagaleh Basin; 10, Abu Gharadig Basin; 11, Sirte Basin; 12, Termit Trough; 13, Southern Chad Basins; 14, Sudan Rifts.

been incorporated. Error estimates for minimum and maximum limits of compaction effects are on the order of tens of meters. Error estimates for paleo-water depths are

<50 m for shallow water deposits but increase to 100–500 m for deeper water deposits. Eustatic sea-level fluctuations, which are of the order of <200 m (Vail et al.,

1977), have not been taken into account. A local isostatic compensation mechanism was adopted in the backstripping calculations.

Most tectonic subsidence curves display a stepwise pattern, with periods of more rapid subsidence interrupted by hiatuses and/or phases of slower subsidence, indicating periods of tectonic activity alternating with periods of relative tectonic quiescence. Changes in subsidence rate can, in most cases, be correlated over the entire basin, as illustrated for the Doba basin in Chad (Fig. 4). Furthermore, changes in subsidence rate can very often be correlated between different basins located at different parts of the plate, as will be discussed later. This correlation between different basins throughout the plate indicates a regional cause for the changes in subsidence rate. The average duration of periods of rapid subsidence (i.e., vigorous tectonic activity) is on the order of 10 m.y. for most studied basins, and the maximum observed tectonic subsidence does not exceed 2.5 km for any of the wells, although many of the wells do not reach basement.

Seismological data indicate that the lithospheric thickness across the Central African shear zone varies between 150 and 200 km in

TABLE 1. LOCATION AND INITIATION OF THE STUDIED BASINS

Basin name	Location	Initiation	Related tectonic process
<u>Atlantic Margin</u>			
Senegal Basin [1]	Mauritania - Senegal - Gambia - Guinea Bissau	Late Triassic-Early Jurassic	Rifting and opening Central Atlantic
Gabon Coastal Basin [2]	Gabon	Neocomian	Rifting and opening South Atlantic
Kwanza Basin [3]	Angola	Neocomian	Rifting and opening South Atlantic
Outeniqua Basin [4]	South Africa	Middle Jurassic	Opening Mozambique Basin
<u>Indian Ocean Margin</u>			
Mozambique and Rovuma Basins [5]	Mozambique	Permo-Triassic	Karoo rifting
Morondava Basin [6]	Madagascar	Permo-Triassic	Karoo rifting
Lamu Embayment [7]	Kenya	Permo-Triassic	Karoo rifting
Somali Basin [8]	Somalia - Ethiopia	Permo-Triassic	Karoo rifting
Sagaleh Basin [9]	Somalia	Permo-Triassic	Karoo rifting
<u>Tethys - Mediterranean area</u>			
Abu Gharadig Basin [10]	Egypt	Middle to Late Jurassic	Left-lateral opening of the Tethys
Sirte Basin [11]	Libya	Late Cretaceous	Possibly, subduction in the Tethys domain
<u>West and Central African rift system</u>			
Termit Trough [12]	Niger - Chad	Neocomian	Rifting and opening South Atlantic
Southern Chad Basins [13]	Chad - C.A.R.	Neocomian	Rifting and opening South Atlantic
Sudan Rifts [14]	Sudan	Late Jurassic - Neocomian	Rifting and opening South Atlantic

Note: [1] Templeton, 1971; Latil Brun et al., 1986; Bellion, 1991; [2] Teisserenc and Villemin, 1990; [3] Brognon and Verrier, 1966; De Klasz, 1978; Lunde et al., 1992; [4] Dingle et al., 1983; Bate and Malan, 1992; [5] Mougnot et al., 1986; Coster et al., 1989; Nairn et al., 1991; [6] Boast and Nairn, 1982; Coffin and Rabinowitz, 1992; [7] Kent, 1982; Coffin and Rabinowitz, 1992; [8] Piccoli et al., 1986; Bosellini, 1989, 1992; [9] Bosellini, 1989, 1992; [10] Sestini, 1984; Bayoumi and Loft, 1989; [11] Gumati and Nairn, 1991; Guiraud and Maurin, 1992; [13] Guiraud et al., 1987; Genik, 1992, 1993; Guiraud and Maurin, 1992; [14] Schull, 1988; Wycisk et al., 1990; Bosworth, 1992; McHargue et al., 1992.

Figure 3. A. Present-day map of the Atlantic Ocean, showing the locations of the main topographic features and the studied basins along the Atlantic margins of Africa (modified from Dziewonski et al., 1992). TC, Tristan da Cunha hot spot; SH, St. Helena hot spot. B. Present-day map of the Indian Ocean, showing the locations of the main topographic features and the study areas along the Indian Ocean margin of Africa (modified from Schlich, 1982). K, Karoo hot spot. B., Basin; C. B., Coastal Basin; R., Ridge; Pl., Plateau; Emb., Embayment.

the undeformed regions, and the crustal thickness in these areas is on the order of 35 km (Plomerová et al., 1993). For these crustal and lithospheric thicknesses a thinning factor of 1.8, which is the original thickness of the lithosphere or crust divided by its thickness after extension, can be regarded as an upper limit, given the observed tectonic subsidence and basin ages. For an isostatic model with the given crustal and lithospheric thicknesses, a thinning factor of 1.8, and a rifting duration of 10 m.y., the maximum thermal subsidence rate is 20 m/m.y., immediately after rifting. Smaller thinning factors or larger rifting durations would lead to slower postrift subsidence. We therefore deduce that for the basins located in the continental interiors a subsidence rate of 20 m/m.y. is an upper limit for thermal subsidence. This does not imply that low subsidence rates (<20 m/m.y.) cannot be tectonically controlled but simply implies that we

TABLE 2. WELL DATA OF THE FOURTEEN STUDY AREAS

Number	Basin name	Number of wells	Maximum penetration (m)	Oldest horizon
1	Senegal Basin	20	4270	M Jurassic
2	Gabon Coastal Basin	21	3942	Neocomian
3	Kwanza Basin	20	4111	U Juras./Neoc.
4	Outeniqua Basin	8	4900	M-U Jurassic
5	Mozambique + Rovuma Basins	17	5957	U Jurassic
6	Morondava Basin	2	5470	Triassic
7	Lamu Embayment	11	4310	M Jurassic
8	Somali Basin	9	4196	Permo-Triassic
9	Sagaleh Basin	3	3361	M Jurassic
10	Abu Gharadig Basin	11	4725	M-U Jurassic
11	Sirte Basin*	36	3970	M Jurassic
12	Termit Trough	8	4272	Barremian
13	Southern Chad Basins	22	4098	Barremian
14	Sudan Rifts	15	4580	U Jurassic

*Subsidence study carried out by Van der Meer and Cloetingh, 1993

Figure 4. Subsidence pattern of the Doba Basin, Chad. Note that changes in subsidence rate are coeval throughout the basin.

cannot discriminate between these two subsidence mechanisms. For subsidence rates larger than 20 m/m.y. we conclude that the driving mechanism is tectonic. We have arbitrarily discriminated between moderate (20–40 m/m.y.) and high (>40 m/m.y.) mean subsidence rates in order to investigate the dependence of subsidence rates on the location and orientation of a basin through time.

For continental margins, thermal subsidence rates are normally much higher. In most cases the thinning factor exceeds 1.8, leading to a substantially larger thermal anomaly and higher values for thermal subsidence, immediately after rifting. Thus, the discrimination between a thermal or tectonic driving factor, for low mean subsidence rates, and unambiguously tectonic, for moderate or high subsidence rates, does not hold for marginal basins. Rapid increases in tectonic subsidence rate usually have a tectonic cause, although it does not have to be regional. Local processes, like salt diapirism and marginal collapse, can also cause phases of rapid subsidence, which show up in the subsidence curves. In the discussion part we will deal with this in further detail.

In the following part we will discuss the results for the seven time slices, determined by calculating the mean subsidence rate of each basin, using all available subsidence curves. For each time slice the mean subsidence rates of the individual basins and the configuration of the southern continents are displayed in the right panels of Figures 5–11. The left panels of these figures show a se-

lection of the subsidence curves that are used to calculate the mean subsidence rate, to illustrate the subsidence pattern in a few basins where important changes occurred during the specific time slices. Since subsidence rates higher than 20 m/m.y., in the African continental basins studied here, have a tectonic cause and thus are probably related to a tensional stress field, we will discuss the possible origin of this tensional stress field during the time slices in question.

CORRELATION OF SUBSIDENCE RESULTS WITH PLATE MOTION CHANGES

Figure 12 summarizes the tectonic subsidence results, together with the occurrences of observed unconformities described in the geological literature. Changes in subsidence rates in the studied basins and the observed unconformities can be correlated over the entire African plate (see, for instance, the Santonian unconformity, evident along the Atlantic margin, the South African margin, and in the Sirte and Southern Chad basins, which is coeval with a change in subsidence rate in most basins). These changes in subsidence rates and observed unconformities can be compared with documented changes in plate motions of the Atlantic and Indian Oceans and the Tethys-Mediterranean, summarized in Figure 13 and hereafter discussed in turn for the seven given time slices.

Late Jurassic

During the Late Jurassic the central Atlantic Ocean was opening after a rifting period that lasted from Middle Triassic until Middle Jurassic times (Klitgord and Schouten, 1986; Favre and Stampfli, 1992). Intimately linked with the opening is the early history of the Tethys-Mediterranean area, where spreading occurred in the Alboran-Ligurian-Penninic Ocean from 155 Ma onward (Laubscher and Bernoulli, 1977; Dercourt et al., 1986; Ziegler, 1988). In the Indian domain, north-south-directed opening of the Somali and Mozambique Basins started in the late Middle Jurassic, at 160 Ma (Rabinowitz et al., 1983). The area of the Mozambique basin was probably pre-weakened by the Karoo plume. This pre-weakening contributed to the rifting and the final breakup between eastern and western Gondwana (Fig. 3B; Storey et al., 1992). Substantial eruptive activity occurred from 198 to 173 Ma in southeastern Africa in a 1500-km-long zone extending from Mozambique to Natal (South Africa; Cox, 1988, 1992). The notion that the final breakup was 20 m.y. later suggests a causal relation with a change in stress field, possibly related to subduction in the Proto-Pacific (Storey et al., 1992). Coeval with the onset of opening in the Indian Ocean, a change in rotation pole occurred in the central Atlantic (Klitgord and Schouten, 1986). The high subsidence rate along the northwest African margin can be linked with the opening of the Central Atlantic and the observed subsidence in the Abu Gharadig basin and, more widespread in northeast Africa, is related to the left lateral opening of the Tethys (Fig. 5; Bayoumi and Loft, 1989; Klitzsch and Squyres, 1990). Similarly, rapid subsidence of the basins along the Indian margins can be related to the opening of the Indian Ocean. For the basins in Mozambique, no data on the deepest parts were available, so no estimate can be given of the Late Jurassic subsidence rate for these basins. According to Mascle et al. (1987), Coster et al. (1989), and Nairn et al. (1991), the Mozambique and Rovuma basins were in a thermal sag phase during this period of time. Effects of the extension in the Indian domain are noticeable farther inland, as shown by the rapid subsidence in the easternmost Sudan Basin. This basin is linked with the Indian margin via Pan-African shear zones that were reactivated during the Karoo rifting period (Daly et al., 1989).

Figure 5. Left panel: Tectonic subsidence curves of selected wells from the Senegal, Outeniqua, Morondava, and Somali Basins and the Blue Nile Rift. The shaded bar indicates the period for which the mean tectonic subsidence rates are shown in the right panel. For clarity of display, no error bars are shown. C, Cenomanian-Turonian; p, Pliocene and younger. Right panel: Paleogeographic reconstruction of the southern continents during the Late Jurassic (modified from Scotese and Golonka, 1992), showing the mean tectonic subsidence rate of the studied basins. Basins shown in white subside with rates between 0 and 20 m/m.y.; striped basins have intermediate subsidence rates between 20 and 40 m/m.y.; and the dark-colored basins are subsiding rapidly with rates of 40–100 m/m.y. In white, with thin outlines, are shown the continental masses; light shading indicates the continental margins, and dark shading marks the oceanic areas. Note the high subsidence rates along the Central Atlantic and Indian Ocean margins.

In the latest Jurassic, when rifting between South Africa and South America began (Rabinowitz and LaBrecque, 1979), the direction of maximum tensional stress changed in southeastern Africa from north-northwest–south-southeast to northeast–southwest (Castaing, 1991), possibly leading to unconformities observed in the basins along the Indian Ocean margin (Piccoli et al., 1986; Coster et al., 1989; Dualeh et al., 1990; Greene et al., 1991). Subsidence in the Outeniqua Basin was renewed, after a period of strong erosion (Dingle, 1978; Bate and Malan, 1992; Van der Merwe and Fouché, 1992), and unconformities of latest Jurassic age are observed not only in the Indian domain, but also in the Tethys-Mediterranean domain (Bayoumi and Lofty, 1989; Rossi et al., 1991). Basaltic Late Jurassic volcanic activity in the Orange Basin of Namibia must be related to the Tristan Da Cunha plume, which was located beneath the southern part of the South Atlantic rift (Fig. 3A; Light et al., 1992).

Early Cretaceous

The Early Cretaceous is the period of the opening of the South and Equatorial Atlantic during the Barremian and Albian, respectively. Between 132 and 126 Ma (Barremian), when sea-floor spreading in the South Atlantic began south of the Walvis-Rio Grande Ridge, a reorganization of the Central Atlantic ridge occurred (Klitgord and Schouten, 1986; Savostin et al., 1986). Rifting and early opening in the South Atlantic led to rapid subsidence along the Atlantic margin, as shown in the Kwanza and Gabon Coastal Basins, and caused a north-south–directed stress field in West and Central Africa (Fig. 6). This resulted in a reactivation of the east-west–trending, Pan-African–aged, Central African shear zone (Daly et al., 1989; Maurin and Guiraud, 1993) on top of which the West and Central African Rift System developed as a system of east-west–trending rifts interconnected by northwest-southeast–trending strike-slip faults (Guiraud and Maurin,

1991, 1992). Subsidence rates in these basins are moderate during the Neocomian-Barremian. The area of the South Atlantic rift and the western part of the West and Central African Rift System have been preweakened by the presence of two important hot spots: the St. Helena and Tristan da Cunha hot spots (Fig. 3A; Wilson and Guiraud, 1992; Wilson, 1993). They were both active during the latest Jurassic and early Cretaceous. The Tristan da Cunha hot spot generated massive flood basalt volcanism in Brazil (Parana province) and minor basalt volcanism in Namibia, both onshore (Etendeka province) and offshore (Hawkesworth et al., 1992; Light et al., 1992; Wilson, 1993). The St. Helena hot spot was cooler and did not cause any massive flood basalt volcanism but was responsible for the development of major dike swarms in the Brazilian Parnaíba and Amazone Basins (de Matos, 1992). It probably did preweaken the lithosphere, at the site of the Equatorial Atlantic rift and the Benue Trough in the westernmost part of the West and Central African Rift Sys-

A

B

Figure 6. A. Tectonic subsidence curves of selected wells from the Kwanza, Mozambique, and Morondava Basins, the Lamu Embayment, and the Abu Gabra Rift, and paleogeographic reconstruction, showing the mean tectonic subsidence rates, for the Neocomian-Barremian. Figure conventions are same as in Figure 5. Note the high subsidence rates along the South Atlantic margin and the increase in subsidence rate in the Morondava Basin, suggesting renewed extension at the Indian Ocean margin. B. Paleogeographic reconstruction of the Tethys-Mediterranean area, showing the location of the main continental blocks and oceans during the Aptian (modified from Dercourt et al., 1986). Stars show future location of Turkey and Oman ophiolites.

tem. The location of the final breakup of the South Atlantic rift is very likely determined by the presence of these two hot spots (Bott, 1992). Hot spots will lead to a deceleration in basin subsidence during the period the hot spot is active and a more rapid thermal subsidence afterwards (Wilson, 1993). The Gabon Coastal and Kwanza basins do not appear to be influenced very much by either of these hot spots. Volcanic rocks with an approximate age of 140 Ma have only been dated in the Kwanza basin (Schlumberger, 1991), and both basins do not show a deceleration of subsidence during the syn-rift phase, nor a rapid postrift subsidence.

The Lamu Embayment and the Mozambique Basins, along the Indian margin, are linked with the South Atlantic Ocean via a complex set of Pan-African mobile belts, trending north-south from Mozambique and Kenya to Sudan, where they link with the Central African Shear Zone. The reactivation of these shear zones and the subsidence in the Lamu Embayment and Mozambique Basins is probably related with the opening of the South Atlantic (see also Genik, 1993). This would imply that the distance from the active plate boundary to the basin is not the controlling factor; more likely, it is the rheology of the underlying

basement. During the same time (Barremian), spreading in the Somali Basin stopped (Segoufin and Patriat, 1980; Rabinowitz et al., 1983) and was taken over by spreading between India-Madagascar, at one side, and Australia-Antarctica, at the other (Norton and Sclater, 1979; Segoufin and Patriat, 1981; Cande and Mutter, 1982; Tarling, 1988; Powell et al., 1988). Subsequently, the Indian-African plate started to be subducted underneath Eurasia, in the late Barremian (Patriat et al., 1982). (According to Scotese et al. [1988] there was a Neo-Tethyan plate to the north of the Indian-African plate, during this time, separated

Figure 7. Tectonic subsidence curves of selected wells from the Senegal and Gabon Coastal Basins, the Lamu Embayment, and the Doba Basin, and paleogeographic reconstruction, showing the mean tectonic subsidence rates, for the Albian. Figure conventions are same as in Figure 5. Note the increase in subsidence along the Atlantic margin and the Central African Shear Zone.

from the latter by the Tethyan rift. During the Early Cretaceous, this Neo-Tethyan plate was being subducted underneath Eurasia.) The cessation of sea-floor spreading in the Somali Basin and the opening of an oceanic basin between India-Madagascar, now forming part of the African plate, and Australia-Antarctica may have played an important role as well in the subsidence along the Indian Ocean margin of Africa. This is shown by the reactivation of the Morondava Basin, which is not connected with the South Atlantic Ocean, ruling out a causal relationship between the increased subsidence in this basin and the opening of the Atlantic. In the Tethys-Mediterranean area, the collision of Apulia, an African promontory, and Eurasia began at 130 Ma (Barremian), which resulted in a counterclockwise rotation of Apulia and the possible opening of the Mesogea oceanic basin (Dercourt et al., 1986; Ricou et al., 1986; Ziegler, 1988; Le Pichon et al., 1988). The Barremian unconformity that has been observed in the southern Sirte basin (Rossi et al., 1991) can be correlated with this opening.

During the Early Aptian, when the South Atlantic Ocean opened and the subsidence of transtensional basins began in the Equatorial Atlantic domain, the direction of max-

imum tensional stress in West and Central Africa changed from north-south to northeast-southwest (Guiraud et al., 1992). Unconformities of early Aptian age are observed in almost every studied basin, with the exceptions of the Sudan rifts and the Abu Gharadig basin (Brognon and Verrier, 1966; De Klasz, 1978; Dingle, 1978; Bellion and Guiraud, 1984; Guiraud et al., 1987; Bosellini, 1989; Coster et al., 1989; Dualeh et al., 1990; Teisserenc and Villemin, 1990; Rossi et al., 1991; Bate and Malan, 1992; Bosellini, 1992; Guiraud et al., 1992). Coeval with the opening of the Equatorial Atlantic (ca. 110 Ma, Albian), a ridge reorganization took place in the South Atlantic (Rabinowitz and LaBrecque, 1979; O'Connor and Duncan, 1990), and a change in spreading direction in the Central Atlantic (Olivet et al., 1984). Possibly from this time on subduction occurred in the Tethyan Alboran-Ligurian-Penninic Ocean due to rotation of the Apulian block and sinistral translation between Eurasia and Africa (Fig. 6B; Dercourt et al., 1986; Savostin et al., 1986; Ziegler, 1988). Accelerated subsidence during the Albian along the Central and South Atlantic margins, probably caused by the rifting and opening in the Equatorial Domain, can be observed in the

Senegal and Gabon basins, respectively (Fig. 7; see also Latil-Brun and Flicoteaux, 1986; Latil Brun and Lucazeau, 1988). Accelerated subsidence of the Lamu Embayment, of the Mozambique basin, and of the basins of the West and Central African Rift System during the Albian affirms the assumption that the history of all these basins was linked with the opening of the Atlantic Ocean. The change in extension direction in West and Central Africa lead to the opening of northwest-southeast- and north-south-trending basins in the West and Central African Rift System, while northeast-southwest-trending faults were reactivated as strike-slip faults (Guiraud and Maurin, 1992). This implies that the weakness of the underlying basement in itself is not the only controlling factor in basin formation and reactivation. The orientation of these weaker zones with respect to the stress direction is at least as, if not more, important.

Late Cretaceous

At the beginning of the Late Cretaceous, Africa was separated from all surrounding continents and, apart from at its northeastern margin, surrounded by active sea-floor spreading axes. These spreading axes, how-

Figure 8. Tectonic subsidence curves of selected wells from the Senegal, Mozambique, Abu Gharadig, and Doba Basins and the Abu Gabra Rift, and paleogeographic reconstruction, showing the mean tectonic subsidence rates, for the Cenomanian through Coniacian. Figure conventions are same as in Figure 5. Note the high subsidence rates over the whole continent while both the Atlantic and the Indian Ocean are open.

ever, were not stable, leading to changes in the state of stress in the African plate. Between 96 and 90 Ma (Cenomanian) there was a critical change in the spreading direction in the Indian Ocean, and India started to move rapidly toward the north with respect to Australia and Antarctica (Powell et al., 1988). Subsequently, India and the Seychelles separated from Madagascar, forming the Mascarene Basin (Norton and Sclater, 1979; Schlich, 1982). According to Scotese et al. (1988) the rapid northward movement of India was caused by the subduction of the Tethyan spreading axis and the initial subduction of the oceanic part of the Indian-African plate. Rapid subsidence along the Indian Ocean margin was probably linked with the rifting and rapid opening of the Mascarene Basin (Fig. 8). Because both Madagascar and India formed part of the African plate at this time, it seems explicable that rifting in the Mascarene Basin caused an acceleration of subsidence in the nearby basins of Madagascar and Mozambique (see also Mougenot et al., 1986; Nairn et al., 1991; Young, 1992). No important changes in the entire Atlantic have been documented for Cenomanian through Coniacian times. A differential spreading rate

between the Central, Equatorial, and Southern sectors of the Atlantic Ocean, as described earlier by Fairhead (1988a, 1988b) and Daly et al. (1989), caused a sinistral displacement along the equatorial fracture zones and the Central African shear zone. Whether this is responsible for the increase in subsidence rate along the West and Central African rift system and in the Senegal and Gabon Coastal Basins remains an open question. The high subsidence rates of the Senegal Basin can be partly related as well to the relaxation of the thermal perturbation related to the Equatorial opening phase or to salt tectonics. In the Gabon Coastal and Kwanza basins, most of the Late Cretaceous and younger deformation is related to salt movement, but in some cases this may have been initiated by deeper structural movements (Teisserenc and Villemin, 1990; Lunde et al., 1992). In the Neo-Tethys domain, between 100 and 95 Ma (late Albian–Cenomanian) spreading stopped and subduction commenced close to the abandoned spreading ridge (Boudier et al., 1985; Dercourt et al., 1986). Oceanic obduction of the Turkey and Oman ophiolites started, and between 90 and 80 Ma they were obducted on the northeastern edge of the Arabo-

African continent (McCulloch et al., 1981; Boudier et al., 1985). It is very likely that the transition from spreading to subduction played an important role in the evolution of the Egyptian Abu Gharadig Basin. It is less clear, however, whether this also played a role in the evolution of the basins along the Atlantic Margin and Central African shear zone. We do not have a convincing explanation ready for the increase in subsidence rate along the Atlantic Margin and Central African shear zone yet (see also Bosworth, 1992; Genik, 1993). The Late Cretaceous is also known for a new peak in widespread volcanic activity in Africa (Bailey, 1992). Kimberlite ages in the African cratons show a peak between 95 and 85 Ma (Smith et al., 1985; Nixon, 1987). In many of the studied sedimentary basins there is evidence for Late Cretaceous volcanic activity (Wilson and Guiraud, 1992). Dolerite sills in the Sudanese Muglad Basin and the Termit Trough in Nigeria reveal ages of 82 ± 8 Ma and 95–85 Ma, respectively (McHargue, 1992; Genik, 1992). In the Lower Benue Trough a peak in magmatic activity occurred between 95 and 83 Ma (Wilson and Guiraud, 1992), and Turonian extrusion of basalts was widespread throughout Mada-

Figure 9. Tectonic subsidence curves of selected wells from the Lamu Embayment, the Sirte Basin, the Termit Trough, and the Abu Gabra Rift, and paleogeographic reconstruction, showing the mean tectonic subsidence rates, for the Campanian and Maastrichtian. Figure conventions are same as in Figure 5. Note the increase in subsidence rate in the Sirte Basin and Termit Trough after the Santonian compressional event.

gascar (Boast and Nairn, 1982). The relationship between this enhanced volcanic activity and plate motion changes or basin evolution is not clear. The reactivation of deep-seated lithospheric fracture zones may be providing pathways for the ascent of magmas. The reactivation of these fracture zones in itself is related to changes along plate boundaries and plate interactions, thus pointing to a possible relationship between the magmatism and plate motion changes, as has been suggested previously by Bellion and Crevola (1991).

At ca. 85 Ma (Santonian), a major change in the relative movement between Africa and Eurasia occurred (Laubscher and Bernoulli, 1977; Savostin et al., 1986; Westphal et al., 1986). Up till then the mega-movement between Africa-Arabia and Eurasia had been sinistral and extensional; at 85 Ma it changed to a scissor-like movement along the African-Eurasian boundary, resulting in a head-on collision between Africa and Eurasia (Dercourt et al., 1986; Ziegler, 1988). Sea-floor spreading in the Alboran-Ligurian-Penninic Ocean and in the Mesogea oceanic basin stopped (Tapponnier, 1977; Dercourt et al., 1986). At the same time the direction of spreading in the Central, South,

and Equatorial Atlantic Oceans changed again (Rabinowitz and LaBrecque, 1979; Klitgord and Schouten, 1986). After this time spreading half-rates started to decrease in the South Atlantic (Cande et al., 1988; Nürnberg and Müller, 1991). The compressional event, related with the Alpine collision, caused a period of strong tectonic activity in North Africa (Reyre, 1984) and Europe (Ziegler, 1990) and put an end to the rapid subsidence in most African basins. It is marked by erosional unconformities in the Senegal and Gabon Coastal basins (De Klasz, 1978; Guiraud et al., 1987; Teisserenc and Villemain, 1990). Unconformities of the same age have been dated in the Outeniqua basin in South Africa, in the Southern Chad basins, and to the northeast of the Sirte basin, in the Al Jabal al Akhdar (Dingle, 1978; Röhlich, 1991; Bate and Malan, 1992; Guiraud et al., 1992; Van der Merwe and Fouché, 1992). During the Campanian and Maastrichtian, following the inversion of most African basins and the associated erosion, subsidence was renewed in many basins (Fig. 9), predominantly the ones striking northwest-southeast, under a east-northeast-west-southwest to east-west maximum tensional stress direction (Benkhelil, 1989;

Guiraud, 1993), showing again the importance of the orientation of a basin and the underlying basement structures, with respect to the stress direction, in basin reactivation. The only plausible explanation for this stress regime and the renewal of subsidence is the ongoing, possibly accelerated, subduction of the Neo-Tethys at the northeastern African margin (see also Zonenshain et al., 1990; Genik, 1993).

Tertiary

During the earliest Tertiary, between 65 and 62 Ma, the oceanic ridge in the Mascarene Basin, southeast of Madagascar, lengthened to the east, creating the Madagascar Basin (Norton and Sclater, 1979), and then jumped toward the northeast to a location between the Seychelles and India (Schlich, 1982; Patriat and Segoufin, 1988). The Seychelles became part of the African plate again. At the same time the Southwest Indian Ridge developed, separating the Madagascar Basin from the Crozet Basin (Patriat, 1979; Schlich, 1982). Coeval with the ridge jump in the Mascarene Basin, subsidence along the margins of Kenya and Somalia increased and sediments were depos-

Figure 10. Tectonic subsidence curves of selected wells from the Lamu Embayment, the Somali and Sirte Basins, and the White Nile Rift, and paleogeographic reconstruction, showing the mean tectonic subsidence rates, for the Paleocene through mid-Eocene. Figure conventions are same as in Figure 5. Note the high subsidence rate in Somalia and Kenya coeval with the ridge jump in the Mascarene Basin.

Figure 11. Tectonic subsidence curves of selected wells from the Senegal, Kwanza, and Mozambique Basins and the White Nile Rift, and paleogeographic reconstruction, showing the mean tectonic subsidence rates, for the Oligocene-Miocene. Figure conventions are same as in Figure 5. Note the increase in subsidence in eastern Africa and Kwanza Basin.

Figure 12. Ages of the unconformities in the studied basins as described in the literature and mean tectonic subsidence rates as calculated from the backstripped subsidence curves. Dark shading marks periods of rapid subsidence (40–100 m/m.y.), light shading marks periods of moderate subsidence (20–40 m/m.y.), dashed outlined boxes mark periods of slow subsidence (0–20 m/m.y.) for part of the basin. Wavy line, unconformity; G, unconformity only dated in Gabon Coastal Basin; K, unconformity only dated in Kwanza Basin; S, unconformity only dated in southern Sirte Basin. (For references see Table 1 and text.) Note that the Aptian, middle-late Eocene, and Pliocene-Quaternary subsidence histories are not incorporated in the backstrip study.

Figure 13. Points of time referring to major changes in the opening histories of the Atlantic and Indian Oceans, the Red Sea, and the Tethys-Mediterranean area (for references see text). Am., America; Afr., Africa; obd., obduction.

ited unconformably on top of the older sections (Fig. 10; Bosellini, 1989, 1992; Dualeh et al., 1990). At, or just after, the Cretaceous-Tertiary boundary event, the rotation pole for South America and Africa changed abruptly (Cande et al., 1988) and the spreading-axis south of the Falkland-Agulhas fracture zone jumped westward, reducing the original offset along this fracture zone by 1000 km (LaBrecque and Hayes, 1979; Hartnady and le Roux, 1985). This was paralleled by the development of an unconformity in the Outeniqua basin (Dingle, 1978; Bate and Malan, 1992; Van der Merwe and Fouché, 1992). A coeval change in spreading direction is seen in the Central Atlantic, marked by an unconformity in the Senegal Basin (De Klasz, 1978). From 65 to

49 Ma, spreading rates were very low in the South Atlantic (Cande et al., 1988; Nürnberg and Müller, 1990). This is in good agreement with the low subsidence rates along the Atlantic margin for the Paleocene-Eocene period. According to Dewey et al. (1989) the convergence between Africa and Europe slowed down about the time of the Cretaceous-Tertiary boundary. Several depositional discontinuities are mentioned for the earliest Paleocene in the Southern European regions (Schwan, 1980, and references therein) and observed in northern Africa, as shown in the Abu Gharadig basin (Bayoumi and Loft, 1987). Sedimentation in the Sirte basin was continuous, but an abrupt deepening of the basin occurred at the Cretaceous-Paleocene boundary (Gu-

mati and Kanes, 1985). Subsidence along the West and Central African Rift System only occurred, as in the previous period, in basins that were oriented northwest-southeast, like the White Nile Rift and the Termit Trough.

At ca. 50 Ma (Eocene) the Indian plate velocity decreased dramatically, probably due to its collision with the Eurasian plate (Patriat and Achache, 1984; Patriat and Segoufin, 1988). In the mid-Eocene, when subduction of the Indian plate beneath Tibet had slowed down, a new plate reorganization occurred. At this stage India and Australia became part of the same plate, and motion along the Ninety east transform stopped (Patriat et al., 1982; Patriat and Achache, 1984; Patriat and Segoufin, 1988;

Veevers et al., 1991). The direction of spreading along the Carlsberg and Central and Southeast Indian Ridges changed from north-south to northeast-southwest, causing a separation of the Chagos and Mascarene plateaus, and spreading rates along the Carlsberg and Central Indian Ridges decreased to almost zero (Schlich, 1982; Talling, 1988). This last system of spreading ridges is still active in the present-day Indian Ocean. The early Eocene unconformities observed in the Mozambique and Rovuma basins are probably related to the collision of India with Asia (Coster et al., 1989). Between 45 and 42 Ma the rotation pole for South America and Africa again changed abruptly (Cande et al., 1988; Nürnberg and Müller, 1991), marking the end of the period of slow spreading. Similar changes of the motion of the rotation pole and direction of sea-floor spreading in the mid-Eocene are seen in the Central Atlantic (Klitgord and Schouten, 1986). Eocene unconformities are evident in the Senegal Basin (Bellion and Guiraud, 1984), in the Gabon Coastal basin (Teisserenc and Villemin, 1990), and in the Kwanza basin (Brognon and Verrier, 1966). From mid-Eocene until present no major changes occurred in spreading rate and direction in the entire Atlantic, although some minor slow-downs and increases in spreading rate are mentioned (Klitgord and Schouten, 1986; Cande et al., 1988). A major north-south compression phase occurred in the Alpine-Pyrenean region in the middle to late Eocene (Bergerat, 1987). This compressional phase had effects in northwestern Africa, like in Mali and in the Benue Trough in Nigeria (Guiraud et al., 1987; Bellion and Guiraud, 1988), and correlates with the observed (late) Eocene unconformities in the White Nile Rift and the Termit Trough (De Klasz, 1978; Bellion et al., 1983; Bosworth, 1992; Guiraud et al., 1987).

A decrease in the spreading rate, at 35 Ma, between North America and Africa, but not yet between North America and Europe, caused a large displacement of the relative rotation pole of Africa and Europe and a change in convergence direction between Europe and Africa (Savostin et al., 1986; Ziegler, 1988). At 20 Ma the sea-floor spreading between North America and Europe slowed down as well, and the rotation pole moved back to approximately its former location. The Neogene history of the Mediterranean is characterized by the evolution of small extensional basins, as in the Western Mediterranean (e.g., Horvath et

al., 1981; Cloetingh et al., 1992; Kooi et al., 1992; Janssen et al., 1993), and contemporary subduction of the remaining parts of oceanic crust. The Oligocene-Miocene is the period of the rifting and opening of the Red Sea, Gulf of Aden, and the East African Rift System (Rosendahl et al., 1992). The Lamu Embayment and the Mozambique and Rovuma Basins were affected by the development of the East African Rift System, and it is not very surprising that subsidence is renewed in these basins (Fig. 11). The Oligocene unconformities observed in these basins are either associated with stepwise rifting activity or with the India-Asia collision (Bosellini, 1989; Coster et al., 1989; Dualeh et al., 1990; Bosellini, 1992). The increase in subsidence in the White Nile Rift and the Termit Trough (see also Guiraud et al., 1987, 1992) are probably also correlated with this rifting phase, and the orientation of the main stress directions probably differed little from the Paleocene-Eocene stress directions as, again, only northwest-southeast-trending basins were reactivated. The opening of the Red Sea and Gulf of Aden, and the rifting in East Africa involved major volcanic activity. This volcanism is not restricted to East Africa but was much more widespread. In northern and western Africa, volcanic fields are restricted to the Pan-African tectonothermally reactivated areas and occur preferably at intersection zones of two or more differently oriented lineaments or fault zones. Many of the volcanic fields (e.g., Hoggar, Tibesti, Cameroon) are associated with domal uplifts, suggesting that they overlie shallow hot spots (Wilson and Guiraud, 1992). Only in some cases (like the Benue Trough) the volcanic fields coincide with the zones of Cretaceous and younger rifts.

The present elevation of the African continent, and especially that of Cretaceous, Eocene, and Miocene shoreline deposits, indicates that large parts of the continent have been uplifted, probably during the late Tertiary (Bond, 1978; Sahagian, 1988). This epeirogenic uplift is accompanied by tilting and rapid subsidence of the continental margins and deepening of the offshore basins, as described for the Angolan Kwanza Basin and Cabinda (Brice et al., 1982; Lunde et al., 1992). Data of the Shell Oil Co. suggest a similar late Tertiary tilting for the Gabon Coastal Basin, with erosion of the landward part of the basin and rapid deposition in the offshore part. Large-scale tectonic movements during the late Miocene have been described for both the Walvis

Ridge and the Angolan Basin (Sibuet et al., 1984). Anomalously thin lithosphere beneath large parts of southern Africa and associated negative Bouguer gravity anomalies suggest a thermal cause for the epeirogenic uplift and a general state of tension for the African plate (Brown and Girdler, 1980; Pavoni, 1992). The zone of largest negative gravity anomalies and maximum thinning, associated with the East African Rift System, extends to the west reaching the Atlantic margin off Angola. This zone possibly also relates to the observed increase in subsidence and the early Miocene unconformity in the Gabon Coastal and Kwanza basins (Brognon and Verrier, 1966; Teisserenc and Villemin, 1990). For the increase in subsidence in the northern Senegal Basin, offshore Mauritania, we do not have a satisfying explanation, but it is most probably related to halokinetic movements.

CONCLUSIONS

A comparison between the subsidence history of 14 basins or groups of basins and the breakup history of Gondwana shows that a strong correlation can be established between changes in plate motions and reactivation of extensional basins in the African plate. This observation can be helpful in the studies of paleo-stresses in sedimentary basins, especially during initial rifting and initiation of subduction zones. Documented rifting phases, like the opening of the South Atlantic Ocean, appear to have affected not only the subsidence rate of the basins located close to this zone of active rifting (in this case along the South Atlantic margin), but also basins located at the other ends of this plate, such as the basins along the Indian Ocean Margin. These latter basins are connected with the South Atlantic margin via a set of Pan-African shear zones, which were reactivated as a consequence of the rifting and thus caused changes in the regional stress field. The rheology of the underlying basement appears to have played a more important role in the formation and reactivation of the studied sedimentary basins than the distance of the basin from the active plate boundary. This can be concluded from the observation that the Late Jurassic and younger rifts in the African plate are preferably formed within Pan-African shear zones, rather than within old cratonic areas located nearer to the active plate boundary. The preweakening effect of the lithosphere by hot spots such as the Karoo, Tristan da Cunha, and St. Helena may have

played an additional role in the localization of deformation. Not only the location of the basin is important in whether or not it will be reactivated, but also the orientation of the basin and underlying basement structures with respect to the direction of the prevailing stress field. This is nicely demonstrated in the post-Santonian basin reactivation restricted to northwest-southeast-trending basins, possibly related with the northeast-directed subduction of the Afro-Arabian plate and the Tertiary west-east extension in East Africa.

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