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Tectonics and stratigraphy of the East Brazil Rift system: an overview

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ABSTRACT

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The East Brazilian Rift system (EBRIS) constitutes the northern segment of the South Atlantic rift system which developed during the Mesozoic breakup of South America and Africa. Following crustal separation in the Late Aptian, it evolved into a passive continental margin.

Along the continental margin six basins are recognized, while three onshore basins form part of an aborted rift. Three continental syn-rift stratigraphic sequences are recognized, spanning Jurassic to Barremian times. The Jurassic (Syn-rift I) and Neocomian (Syn-rift II) phases were most active in the interior rift basins. During the Barremian (Syn-rift III), rift subsidence rates were twice as large as during the Neocomian (Syn-rift II), both in the interior rift and in the marginal rift segments, indicating that rift axis did not migrate from the interior to the marginal setting.

Rift magmatism was centered on the southern EBRIS and peaked between 130 and 120 Ma during syn-rift phase II. Rift phase III was followed by a transitional marine, evaporitic megasequence of Aptian age, which directly overlies the rift unconformity and a marine drift megasequence which spans Albian to Recent times.

During the Late Cretaceous, sedimentation rates responded to first-order eustatic sea-level fluctuations. Tertiary accelerated sedimentation rates can be related to local clastic supply which filled in spaces inherited from previous starved conditions. Between 60 and 40 Ma, post-rift magmatism, centered on the Abrolhos and Royal Charlotte banks, is probably related to development of a hot spot associated with the Vitória–Trindade Seamount Chain.

Although crossing three distinct Precambrian tectono-thermal provinces, ranging from Archean through Late Proterozoic, rift structures follow a general NE trend, subparallel to the principal basement fabric. A NW–SE oriented stress field appears to be compatible with both Neocomian and Barremian phases of crustal extension.

Profiles transverse to the rift axis indicate crustal stretching factors ranging between $\beta = 2.16$ and 2.88. In the shallow portions of the rift, surface extension and crustal thinning seem to be compatible; however, in the deep portions of the basins, this relationship could not be tested.

Reinterpretation of refraction profiles, north and south of the Walvis–São Paulo Ridge transform, indicates that seafloor spreading, from M3 anomaly to Aptian off Pelotas Basin, was taken up by crustal extension in the São Paulo Plateau. Differences in stretching rates may have been accommodated by extension across the Ponta Grossa Arch.

The Early Aptian syn-rift/post-rift transition in the EBRIS marginal basins does not coincide with the onset of the drift phase during the Early Albian. This apparent discrepancy may be explained by a change from distributed margin-wide extension to a focused mode of extension near the future continent/ocean boundary.

1. Introduction

The East Brazil Rift system (EBRIS) consists of six rifted continental margin basins, referred to from south to north as the Pelotas, Santos, Campos, Espirito Santo, Bahia Sul and Sergipe–Alagoas basins. The aborted intracontinental rifts, formed by the Recôncavo, Tucano and Jatobá grabens, form an integral part of this rift system (Fig. 1).

More than 10,000 wells were drilled and several hundred thousand kilometers of seismic were

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recorded by PETROBRÁS and provide control on the stratigraphic and structural framework of these basins, amongst which particularly the Campos, Recôncavo and Sergipe–Alagoas contain major hydrocarbon accumulations.

Classic overviews of the Brazilian marginal basins can be found in Ponte and Asmus (1978), Asmus and Porto (1980), Asmus and Guazelli (1981) and Ojeda (1982). Recently, the East Brazilian marginal basins have been discussed by Chang et al. (1988) and Guardado et al. (1989). In accordance with the objectives of the Geodynamics of Rifting Symposium, the synthesis presented here emphasizes the structural and crustal framework of the Mesozoic EBRIS basins.

2. Megatectonic setting

The plate tectonic setting, which governed the development of the Brazilian continental margin

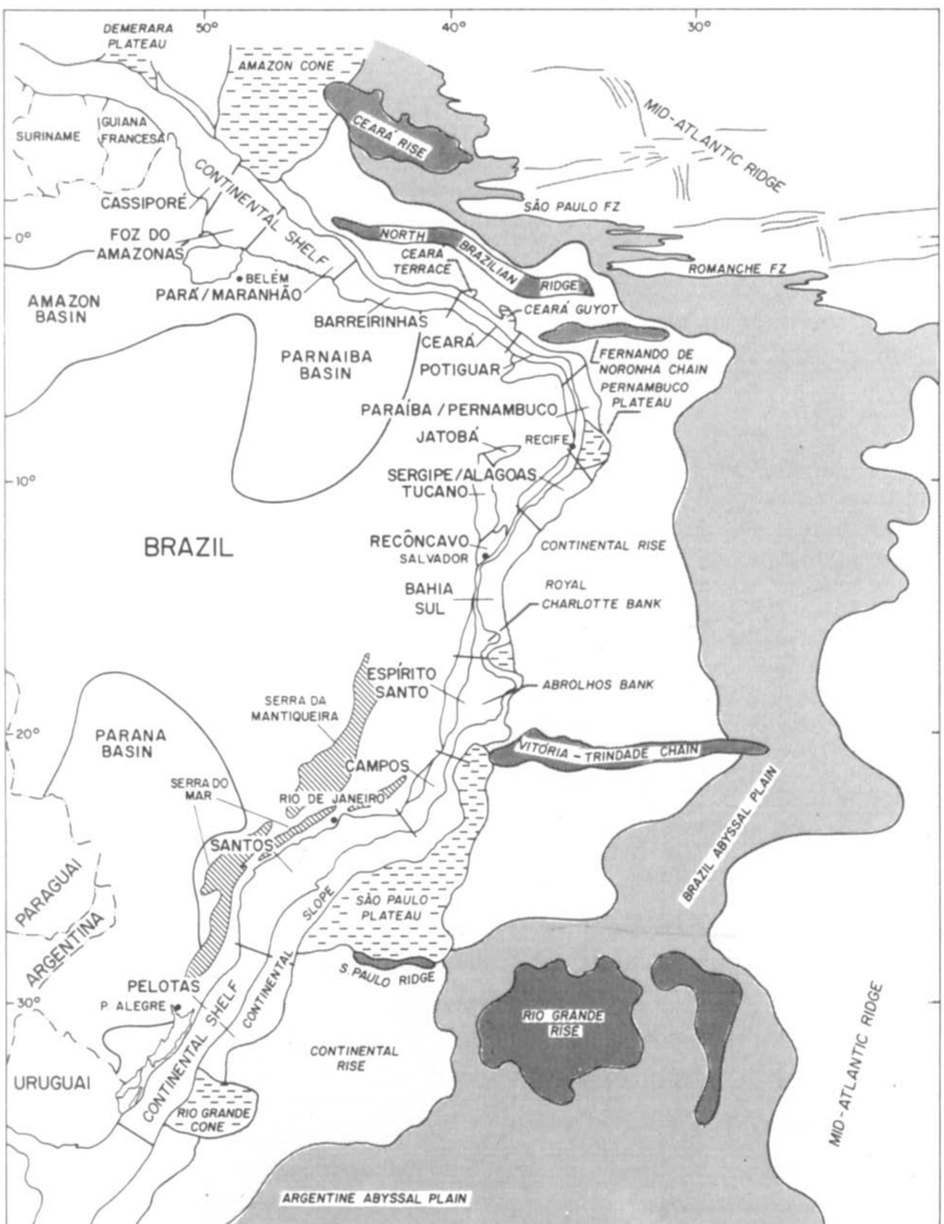

Fig. 1. Index map of Brazilian continental margin basins. Physiographic features after Palma et al. (1979).

is, in comparison to other parts of the World, geologically straightforward. Mesozoic rifting culminates in the splitting of the very large and fairly rigid (South America and Africa) continental blocks. These continental masses belonged to the Gondwana Palaeozoic supercontinent which included South America, Africa, India, Australia and Antarctic.

During the break-up of Africa and South America, two very different margins developed around Brazil (Mascle, 1976); the North Brazilian Equatorial margin evolved in response to strike-slip motion between Brazil and Africa, causing development of complex shear-dominated basins (Gorini, 1977; Zalan, 1984; Azevedo, 1986; Szatmari et al., 1987; Mascle et al., 1988; Guiraud and Maurin, 1992, this volume). In contrast, the East Brazilian margin, evolved out of the EBRIS, into a passive margin, as a consequence of orthogonal crustal extension.

In spite of this simple rifting geometry, reconstruction of the plate motion during the early history of seafloor spreading is still controversial and even more so during the rifting stage. This is due to the presence of a wide magnetic quiet zone spanning earliest Aptian (anomaly M0, 118 Ma) to Campanian times (anomaly 34, 84 Ma), which prevents mapping of seafloor anomalies immediately adjacent to the continental margin and thus impedes tracking of plate motion during the early drift stage. Therefore, plate kinematics were traditionally derived from flow lines along fracture zones (Le Pichon, 1968; Le Pichon and Hayes, 1971; Francheteau and Le Pichon, 1972; Sibuet and Mascle, 1978). Understandably, models based on such data are not able to resolve details of motions between the continental margins of the South Atlantic ocean.

Based on the mapping of Mesozoic magnetic anomalies to the south of the Walvis–São Paulo ridges (Fig. 2), Rabinowitz and La Brecque (1979) developed a more powerful model of plate motion, according to which the continents moved apart as single rigid blocks. This model predicted a Neocomian E–W motion of the continents south of the Walvis–São Paulo ridges, progressively more NE–SW motion to the north of these ridges, shear-extension in the Sergipe–Alagoas/Niger rift system, transtension along the eastern and compression in the western Equatorial margin. Initiation of seafloor spreading in the South Atlantic was thought to have commenced during the Late Berriasian–Early Valanginian (M12 anomaly, 138 Ma).

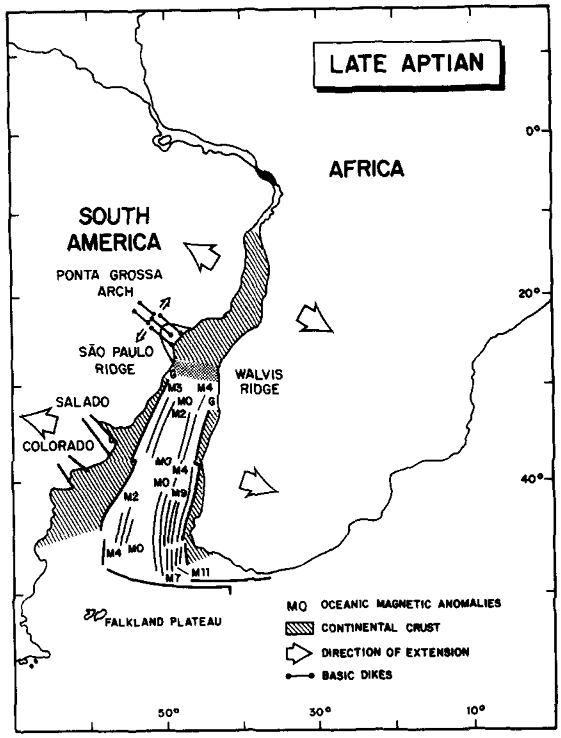

Fig. 2. Reconstruction of the South Atlantic in the Late Aptian with Mesozoic magnetic lineations (Rabinowitz and LaBrecque, 1979). Salado and Colorado basins and Ponta Grossa dike swarm are intraplate extensional features accommodating differences in rates of plate divergence across Walvis/São Paulo Ridge (south: seafloor spreading; north: rifting).

Based on reflection seismic data, Austin and Uchupi (1982) realized that most of the Mesozoic magnetic anomalies mapped by Rabinowitz and La Brecque (1979) occurred on thinned continental crust. This implied that seafloor spreading in the South Atlantic was initiated much later than previously believed, around anomaly M9 time (Late Valanginian–Early Hauterivian, 131 Ma). North of 33°S, seafloor spreading commenced around anomaly M4 time (Mid-Hauterivian, 127 Ma), whereas north of Walvis–São Paulo ridges, first oceanic crust was emplaced near the Aptian/Albian boundary (113 Ma).

Kinematic models, considering intraplate deformation, were proposed to accommodate differential strain along the various segments of the continental margins. Leyden (1976) invoked a clockwise rotation of southern South America, in order to accommodate differential spreading north and south of the Walvis–São Paulo ridges. As a consequence, the Cretaceous Salado and Colorado basins hinged open towards the Atlantic Ocean and the Paraná basalts were extruded along the Torres–Posadas lineament. Pindell and Dewey (1982) proposed that this differential motion in the South Atlantic should be accommodated by rifting in the Benue Trough. Curie (1984) added, without any field evidence, a right-lateral transcontinental strike-slip movement across the Paraná Basin in continuation to the Walvis–São Paulo ridges. Conceição et al. (1988) proposed that this discontinuity ran along the NW trending Curitiba–Maringá shear zone along the Ponta Grossa Arch. This shear zone is inferred from magnetic lineaments, which coincides with the Ponta Grossa dike swarms. However, the proposed shear zone models are incompatible with the geological significance of this dike swarm. Dikes are good paleostress indicators and are emplaced always normal to the minimum principal stress (Suppe, 1985). Therefore, our preferred model is one involving strain accommodation by rotation of South America relative to Africa, thus producing extension in a NE–SW direction, and injection of NW–SE oriented dikes (Fig. 2).

However, intraplate deformation is not necessary if seafloor spreading south of São Paulo–Walvis Ridge is accompanied by continental crustal stretching north of it. Although the nature of the crust underlying the São Paulo Plateau is still in debate, evidence is mounting in favor of a highly stretched continental crust (Kowsmann et al., 1982; Guimarães et al., 1982; Van der Ven, 1983). The principal event of crustal stretching occurred in EBRIS between 130 and 120 Ma (Chang et al., 1988; Conceição et al., 1988); this is compatible with the above interpretation, because it coincides with the age of the seafloor spreading magnetic anomalies M4 to M0, which are well developed south of Walvis Ridge (Fig. 2). If this is the case, then intraplate deformation would need only to account for differences between the rate of seafloor spreading to the south and continental crustal extension to the north. This difference should be quite small and could be realistically accounted for by the small amount

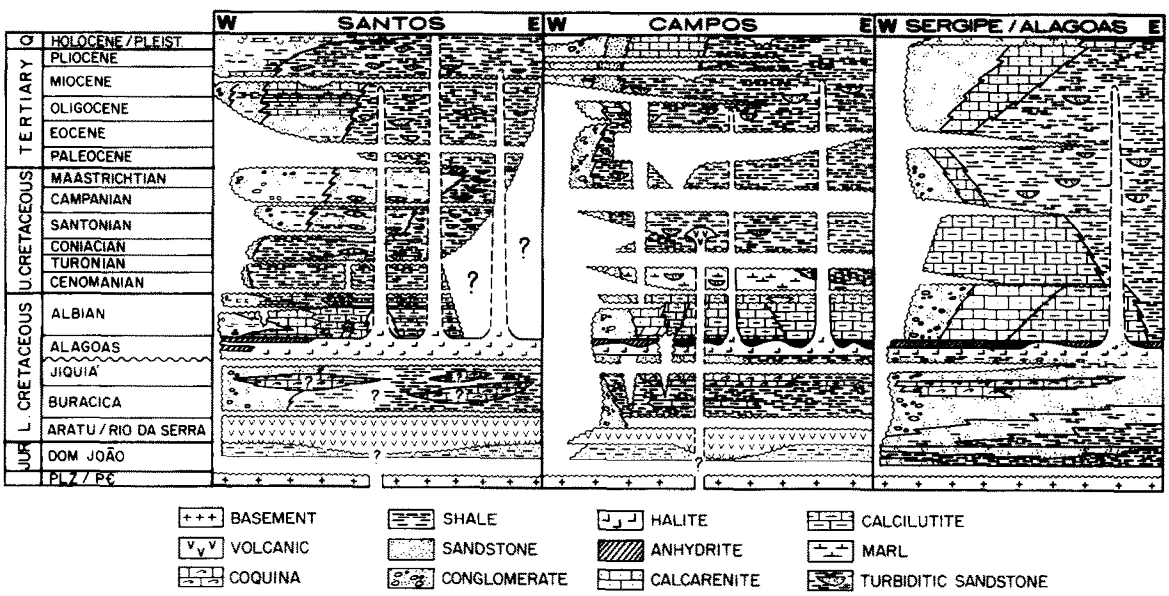

Fig. 3. Santos, Campos and Sergipe–Alagoas basins stratigraphic chart. Early Cretaceous local stages are related to international stages in Fig. 6.

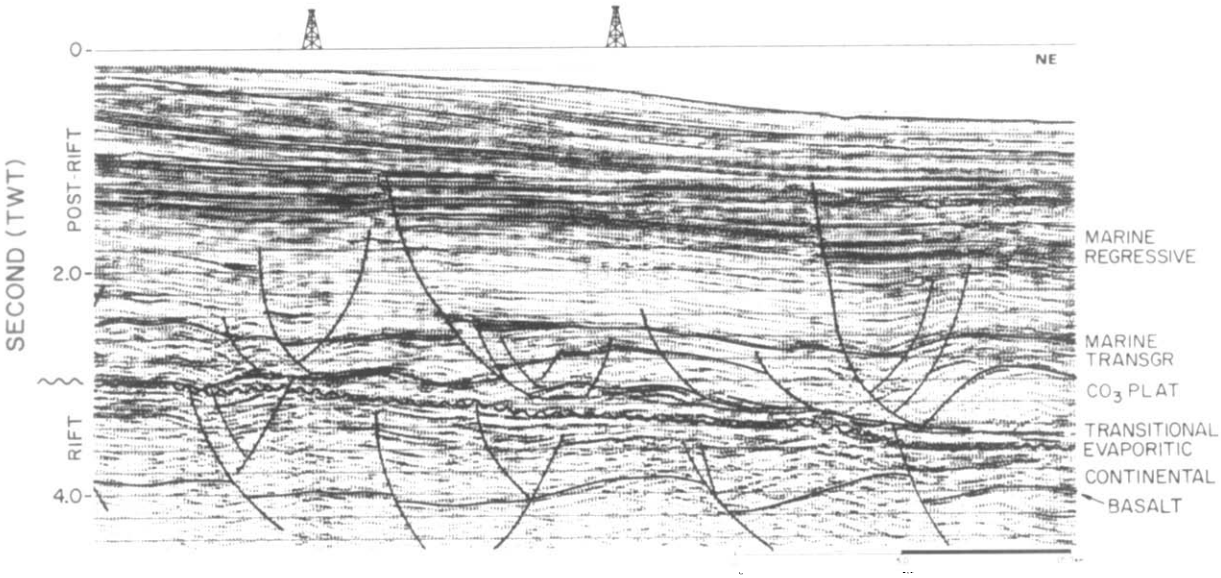

Fig. 4. Seismic section through the Campos Basin, showing the megasequences, horst and graben rift structures, the rift/post-rift unconformity. Post-rift growth faults and associated roll-over structures are due to salt movement.

of extension present in the region of the Ponta Grossa Arch dike swarm. In a 40 km composite profile across the area of densest dike intrusion, representing a small portion of the dike swarm, a cumulative width of 11 km of basaltic dikes was measured (Ussami et al., in press).

3. Stratigraphical development

Sediments occurring in the rifted basins and along the continental margin of EBRIS range in age from Late Jurassic to Recent. The evolution of these basins is best described if the sedimenta-

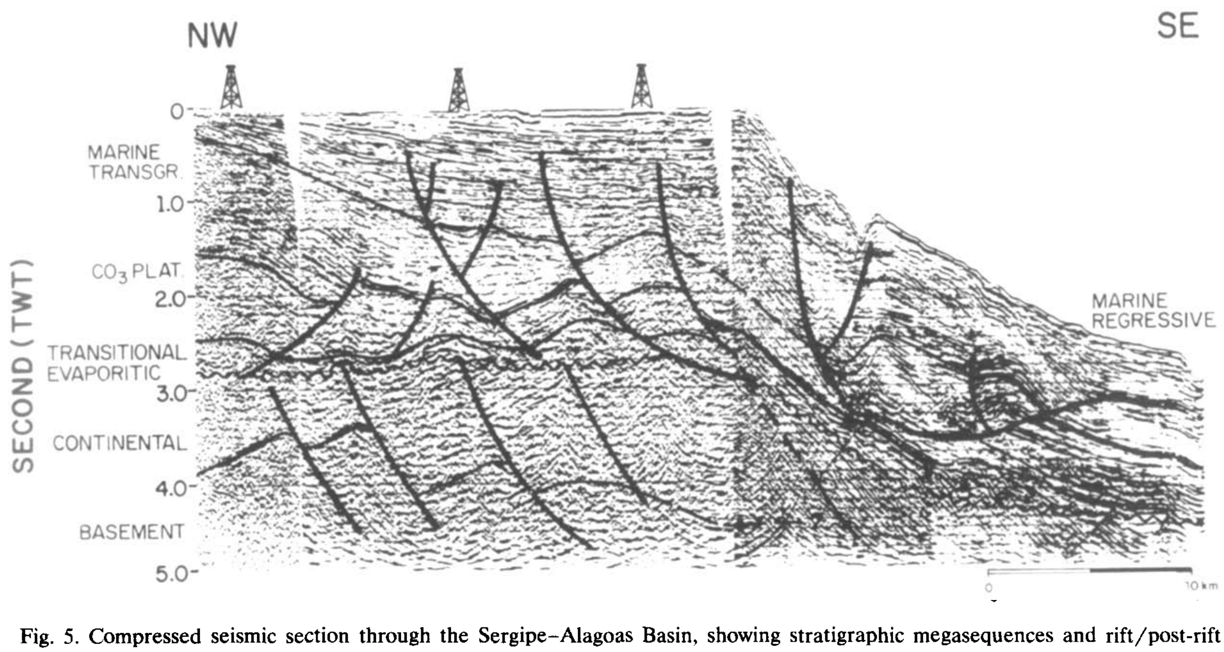

Fig. 5. Compressed seismic section through the Sergipe–Alagoas Basin, showing stratigraphic megasequences and rift/post-rift regional unconformity. Note synthetic rift faults.

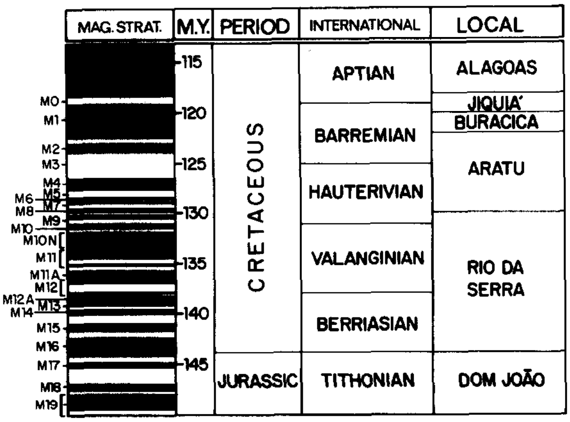

Fig. 6. Chronostratigraphic chart of the Late Jurassic and Early Cretaceous relating local stages, based on ostracod zonation, to international ages. Time scale according to Harland et al. (1982) (Simplified from Beurlen and Figueiredo, 1988).

tion record is subdivided into five megasequences (Figs. 3, 4 and 5), namely the continental syn-rift, the transitional evaporitic, the shallow marine carbonate platform of the early drift stage, and the open marine-transgressive, and regressive cycles of the passive margin stage (Ponte et al., 1978; Chang et al., 1988).

In this paper emphasis will be given to the continental syn-rift alluvial, fluvial and lacustrine deposits which accumulated during Late Jurassic to Aptian times. A tentative correlation chart of the local stages, based on ostracod zonation, with international scales is given in Fig. 6; due to the endemic nature of continental faunas, uncertainties in correlations should be taken into account.

3.1 Continental megasequence

The continental or syn-rift megasequence is best known in three—Campos, Recôncavo and Sergipe–Alagoas—basins located along the eastern continental margin of Brazil.

The continental megasequence is composed of three depositional sequences which are characterized by different facies associations and structural styles (Figueiredo, 1981). The basal sequence (Syn-rift-I) is little affected by faulting and more widespread than the overlying sequences. For this reason, it was originally considered as a pre-rift sequence (Ponte and Asmus, 1976; Ponte et al., 1978; Asmus and Porto, 1980); however, the apparent wider distribution of this basal sequence, while still confined to the rift, is due to the erosion of the younger sequences. Furthermore, its stratigraphic continuity with the overlying taphrogenic sequences and the occasional identification of growth faulting within this basal sequence corroborate the present interpretation that it represents the first syn-rift sediments. The other two sequences (Syn-rift-II and III) were affected by intense faulting and were unquestionably deposited in a tectonically active rift basin.

Magnavita and Cupertino (1988) equated the lithostratigraphic units of the Recôncavo, Tucano and Jatobá rifts with examples from the East African rift system presented by Le Fournier et al. (1985); according to such a comparison, the Brazilian Syn-rift sequence-I could be equated to stages 1 and 2 of Le Fournier et al. (1985), Syn-rift-II with stage 3a and Syn-rift-III with stage 3b.

Syn-rift sequence-I

The Syn-rift sequence-I consists of Late Jurassic sediments deposited in a large depression, known as African–Brazilian Depression (Ponte et al., 1971; Estrella, 1972), which succeeded the development of very large Palaeozoic intracratonic basins. These huge, some 1500 km long, N–S trending lows were rapidly filled with a complex package of coarse-grained fluvial and alluvial fan deposits and minor evaporites accumulating in local playa-lakes. Eolian sands are also common in this sequence. Erosional remnants of these sediments are preserved in the Bahia Sul, Recôncavo–Tucano, Jatobá, and Sergipe–Alagoas basins.

To the south of Bahia Sul Basin, in the rift basins along the Atlantic Ocean, time-equivalent sedimentary rocks are replaced by basaltic lava flows which are overlain by the flows and sediments of the Syn-rift sequence-II.

The following discussion of the Syn-rift sequence-I is based on outcrops, well and seismic data for the Recôncavo Basin, where it attains a thickness of up to 600 m (Fig. 7). In this basin, the Syn-rift sequence-I is subdivided into three formal lithostratigraphic units: Aliança, Sergi and Itaparica Formations. A transitional section between the continental Itaparica and the lacustrine sediments of the overlying Candeias Formation is represented by its basal Tauá Member.

The Aliança Formation is characterized by continental red shales and sandstones containing subordinate evaporitic beds. From bottom to top it can be subdivided into the Afligidos, Boipeba, and Capianga Members (Fig. 8).

The Afligidos Member consists of basal coarse clastics and evaporites, that were deposited under arid alluvial–fluvial conditions, whereby in local evaporitic pans stratified shales, carbonates and evaporites (anhydrite and halite) accumulated. The upper section, composed of red, white and purple shales, reflects a gradual transition to a lacustrine environment.

The Boipeba Member corresponds to an eastward prograding clastic wedge of conglomerates, sandstones and minor red shales, that covered most of the earlier lacustrine basin; as such it reflects a reactivation of the clastic source area.

The Capianga Member consists mainly of lacustrine red shales and white and yellow siltstones and fine-grained sandstones, containing occasional thin beds of ostracod limestones; the Capianga Member is in part a lateral equivalent of the Boipeba clastics.

In the Sergipe–Alagoas Basin the Candeeiro

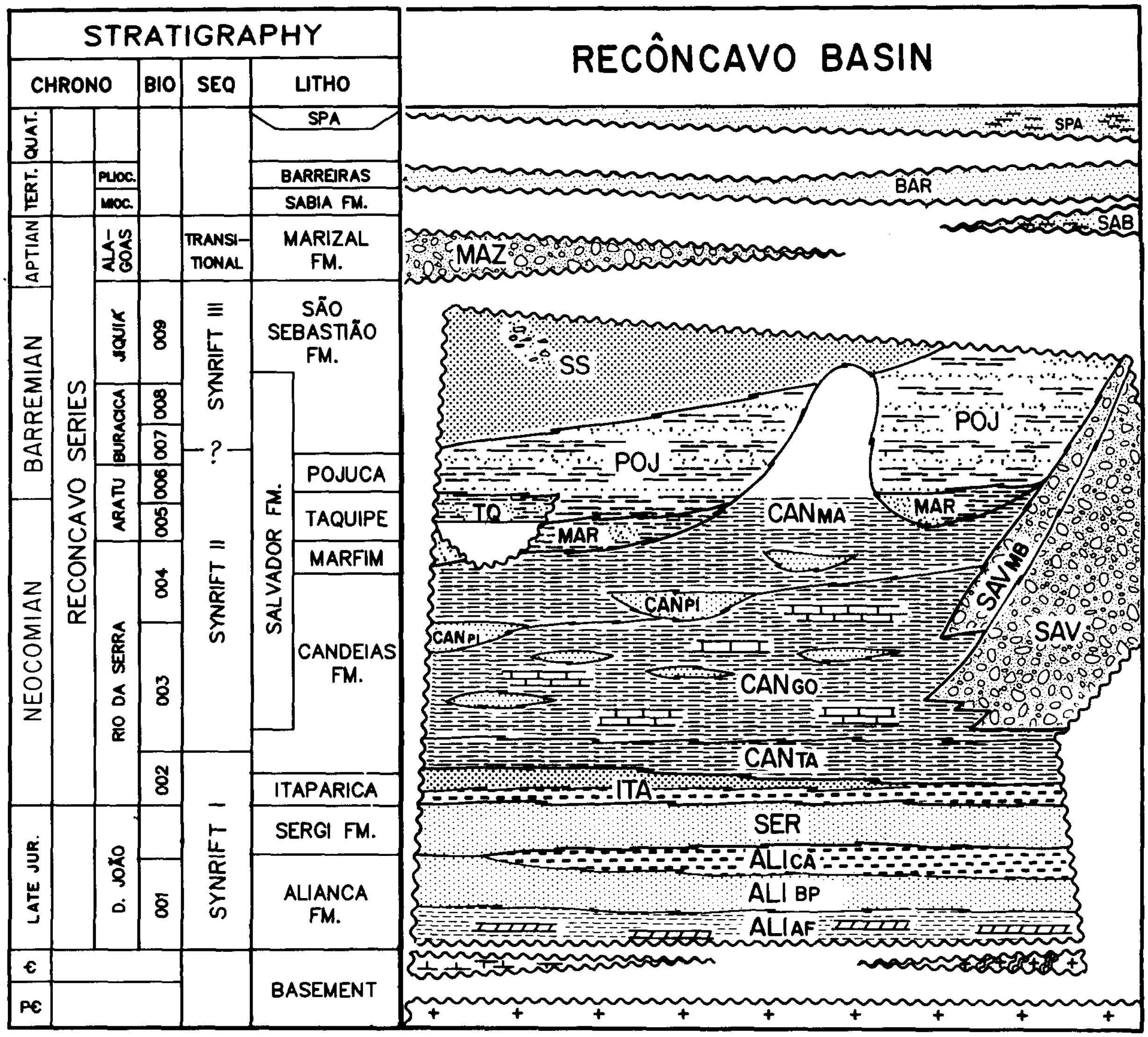

Fig. 7. Stratigraphic chart of the syn-rift sequence in the Recôncavo Basin, giving local stages based on ostracod biozones (modified after Figueiredo et al., in press).

and Bananeiras Formations are the time-equivalent sedimentary package of the Aliança Formation (Fig. 9).

The Sergi Formation is characterized by conglomeratic to fine-grained, light-colored sandstones, and sporadic intercalations of red shales near its base. It was deposited by a very broad system of coalescent alluvial-fluvial fans which wedge out toward the east and overstepped the earlier basin margins (Netto et al., 1982; Bruhn and De Ros, 1987).

As the Sergi sandstones represent the principal reservoir-rock for oil accumulation in the Recôncavo Basin, several detailed lithofacies studies were carried out. Initially, clastic transport was mainly by braided rivers and later by meanderbelt systems. Near the top of the formation, eolian sands occur; they have a regional distribution and form the best reservoirs of this formation. In the Sergipe–Alagoas Basin the equivalent section is represented by the Serraria Formation.

The Itaparica Formation is composed of a cyclical upward fining and thinning sequence of sandstones, siltstones and shales, which is capped by a laterally very persistent, fine to coarse-grained sandstone (Fig. 7), the equivalent of which, in the Sergipe–Alagoas Basin, is the basal Feliz Deserto Formation (Fig. 9). These upper sands were deposited by basin-ward progradational coarse-grained fluvio-deltaic system; its top-parts, showing evidence of intense eolian reworking, form an excellent oil-reservoir.

At the transition from the deltaic Itaparica to the lacustrine Candeias Formation water-depths increased gradually, giving rise to the deposition of the Tauá Member; the latter consists of black conspicuously parallel laminated, dark-gray, micaceous, splintery and hard, fossiliferous shale. These shales were deposited in a marshy lacustrine plain under hot and humid climatic conditions; they mark the top of the Syn-rift sequence-I, and the beginning of the taphrogenic Syn-rift sequence-II.

Syn-rift sequence-II

During the Neocomian, accelerated crustal extension caused rapid subsidence of the Sergipe–Alagoas and Recôncavo–Tucano grabens, in which deep and permanently stratified fresh water lakes developed. These lakes were rapidly filled with fine-grained clastics and sand-grained turbidites, interbedded with coarse mass-flow deposits which were derived from the fault scarp

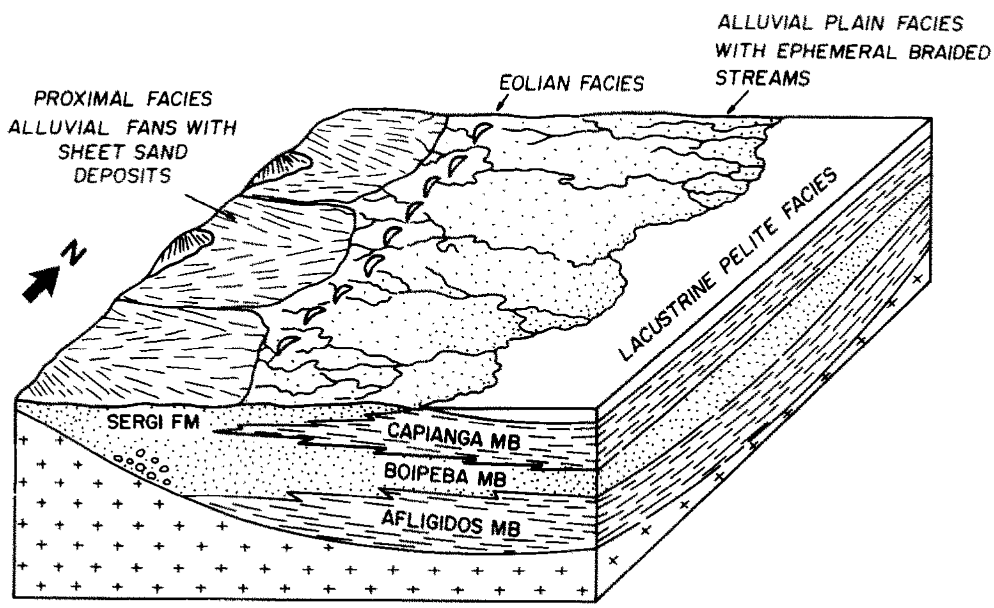

Fig. 8. Schematic depositional model for Syn-rift sequence-I of the Recôncavo Basin (modified after Medeiros and Ponte, 1981).

fanglomerate wedges. The Candeias sequence terminated with an advance of fluvial-deltaic sediments over the entire lake.

In the southern Campos and Santos basins, volcanism was very active during the Neocomian and thick wedges of basaltic lava flows floor the Syn-rift sequence-II in these basins (Fig. 3). Basal coarse clastics, intercalated with volcaniclastics, are overlain by talc-stevensite oolitic siltstones and fine-grained sandstones; these were deposited in shallow euxinic, fresh water and alkaline lakes, rich in magnesium.

In the *Recôncavo Basin* (Fig. 7), the Neocomian is essentially represented by the Candeias Formation which is composed chiefly of shales, containing lenses of sandstones, siltstones, and some lenticular limestones (Fig. 7). This formation attains thickness of up to 3000 m. The basal part of the Candeias Formation consists of dark-gray to greenish-gray, very calciferous and locally

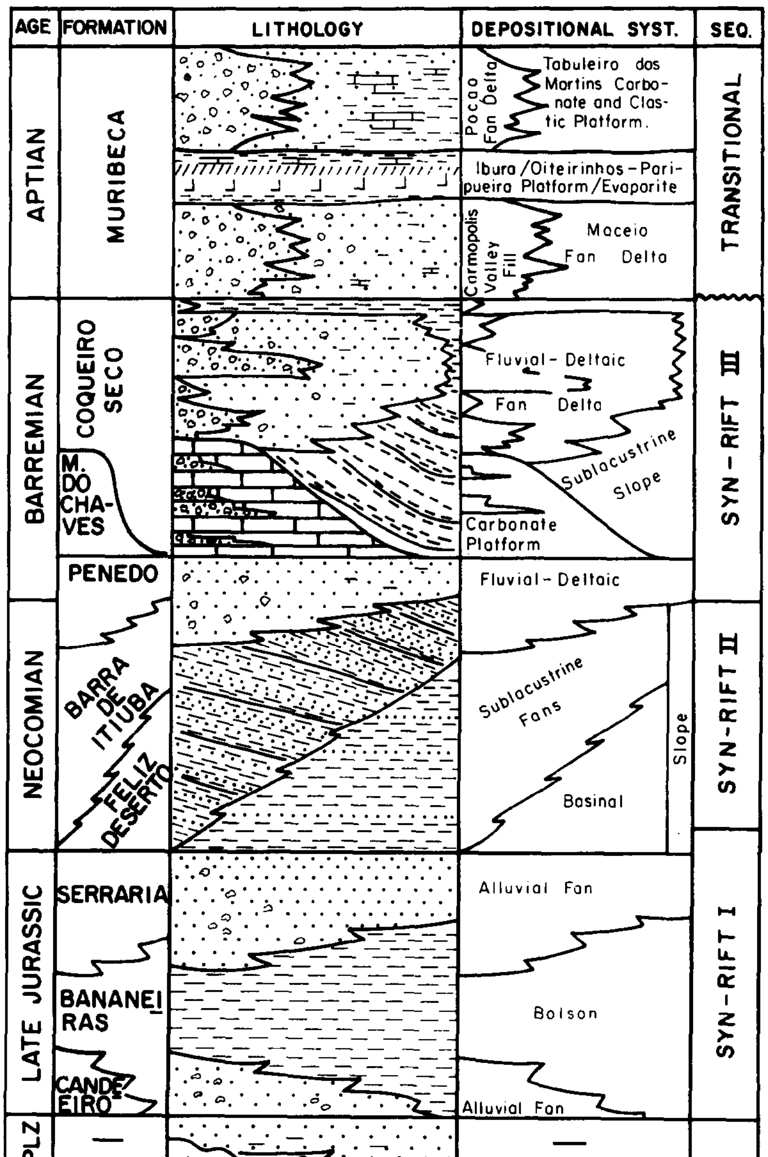

Fig. 9. Stratigraphic chart and depositional models of continental (syn-rift) and transitional megasequences in the Sergipe–Alagoas Basin (modified after Figueiredo, 1981).

dolomitized shales, containing frequent concretions and lenses of micritic limestones. Shales are characterized by a high organic content and represent the main source rock for the oil found in the Recôncavo Basin; their thickness varies from a few hundred meters on platform areas to more than 2000 m in basinal lows (Netto et al., 1984).

The upper part of the Candeias Formation attains thicknesses of up to 1500 m and is characterized by light- to dark-greenish shales and abundant lacustrine turbidites, composed of yellowish and greenish, very fine-grained, highly argillaceous massive sandstones (Netto and Oliveira, 1985). In the deeper parts of the basin, this sequence is deformed by clay diapirs, earthquake triggered slumping, basin tilting and sedimentary loading by the advancing deltaic complexes, which cap the Candeias Formation. Syndepositional deformation and mass flows led to the accumulation of fluid sandstones in basinal

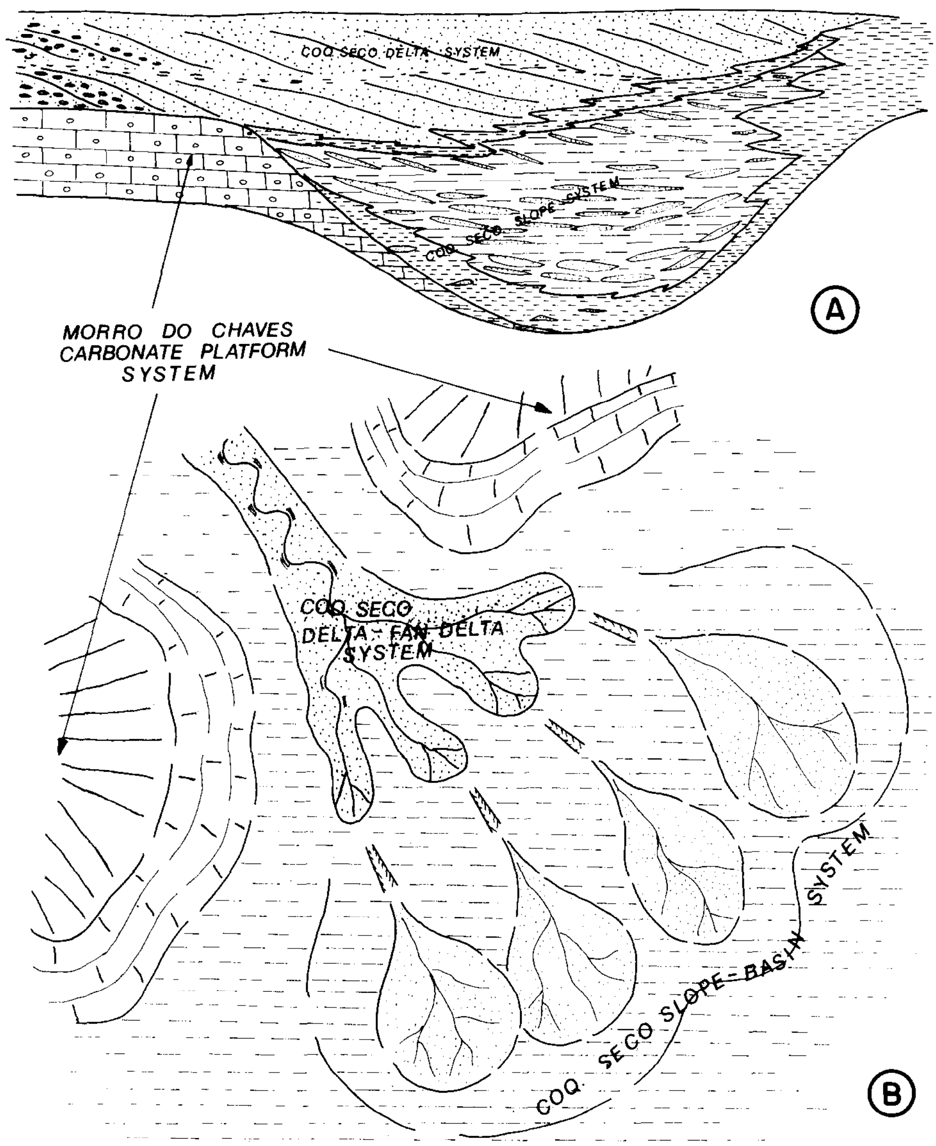

Fig. 10. Schematic facies model in cross section (A) and plan view (B) of the Syn-rift sequence-III in the Sergipe–Alagoas Basin (Figueiredo, 1981). Not to scale.

lows (Rajagabaglia, 1990). Towards the faulted eastern basin margin, the Candeias sediments interfinger with the fanglomeratic wedge of the Salvador Formation.

Progressive infilling of the basin culminated in the advance of the deltaic system of the Marfim and Pojuca Formations; these attain thicknesses of some 1000 m and consist of several deltaic sheet sands, which are interrupted by periodic transgressive shales and ostracodal limestones, correlative throughout the basin. Deposition of this unit reflects a very strong climatic/tectonic control on lacustrine sedimentation.

In the *Sergipe–Alagoas Basin,* the time-equivalent section is represented by the Feliz Deserto, Barra de Itiúba and Penedo Formations (Fig. 9). The entire sequence reaches thicknesses of up to 2700 m and reflects a similar basin evolution as described above for the Recôncavo Basin; the only difference is the apparent absence of shale diapirism.

In the *Campos Basin*, and probably in all south-southeastern basins (Pelotas, Santos, Campos and Espirito Santo), the basal part of the Early Cretaceous consists of basalts intercalated with volcaniclastic and sedimentary rocks (Mizusaki, 1986; Mizusaki et al., 1988). This volcanic sequence is overlain by up to 3000 m thick sediments of the Lagoa Feia Formation, which can be subdivided into the following four depositional units (Dias et al., 1988): basal clastic, talc-stevensitic, coquinas, and clastic-evaporitic sequences. The first two units correspond to the Syn-rift sequence-II (Fig. 11).

The basal clastic unit consists of conglomerates composed mainly of basaltic fragments, quartz and feldspar grains, intensely cemented by calcite; laterally these grade into cross-stratified sandstones and brown mudstones. These were deposited under relatively arid conditions whereby alluvial fan and fan delta systems interfingered with mudflats and sabkhas. Upwards, these clastics grade into marginal and distal lacustrine sediments composed of bioclastic calcarenites and calcilutites.

The talc-stevensite unit is characterized by siltstones and fine-grained sandstones, composed of talc and stevensite oolites and peletoids and intercalated ostracod shell beds, calcilutite and black shales. Sedimentation took place in a restricted, euxinic, relatively shallow alkaline lake rich in magnesium (Rehim et al., 1986). In the Santos and Pelotas basins, this unit was never reached by boreholes.

Syn-rift sequence-III

During the Barremian, sedimentation in lakes of the interior rift system was succeeded by the deposition of the alluvial–fluvial sediments of Syn-rift sequence-III. In the eastern marginal basins, carbonate platforms composed of bivalves and ostracods were accumulated on structural highs, whereas fluvio-deltaic and lacustrine sediments were deposited in the adjacent lows. These contain thick sections of highly organic-rich black shales, which represent periodic highstand of lake level. Salinity in the Syn-rift sequence-III lakes increased gradually to almost the level of normal seawater. This sequence is well developed in the Sergipe–Alagoas and Campos basins.

In the *Sergipe–Alagoas Basin*, the Syn-rift sequence-III is represented by the Morro do Chaves carbonate platform, and by the somewhat younger Coqueiro Seco fluvial-deltaic and slope deposits (Fig. 10; Figueiredo, 1981). This sequence reaches thicknesses of 3000 m and more; it is present in all depositional lows.

Initially, lacustrine carbonates were deposited on shallow positive areas, flanking the principal source points, and consist of massively bedded, high-energy bivalve and minor ostracod limestones. Shore-ward these limestones are interbedded with coarse clastics, grading into coarse-grained sandstones and conglomerates, deposited in cyclical fan delta system. Beyond the fault-controlled shelf edge, the high-energy carbonates give way to marls and very calcareous slope shales, which become condensed toward the center of the lacustrine basin.

Clastic supply to the basin increased at the onset of the Coqueiro Seco Formation when fan-delta and fluvial-deltaic complexes prograded over the earlier carbonate platform and advanced basin-ward, presumably in response to tectonic pulses. In the Alagoas sub-basin, the most prominent low in the area, a cyclical sequence of alternating fluvial-deltaic and slope sediments reflects continued tectonic activity and climatic lake level oscillations.

The fan delta and fluvial-deltaic systems display high sand/shale ratios. Conglomerates predominate in fan delta and coarse-grained meander belt facies. Delta-plain channel-fills, consisting mainly of medium- to coarse-grained sandstones, dominate in fluvial-deltaic system (Fig. 10).

In the *Campos Basin*, the Middle and Late Barremian is represented by a thick section (around 2500 m) of lacustrine carbonates and black shales (Dias et al., 1988), deposited under starved basin conditions.

The relatively saline water lacustrine carbonate is made up of bivalves, minor ostracods and,

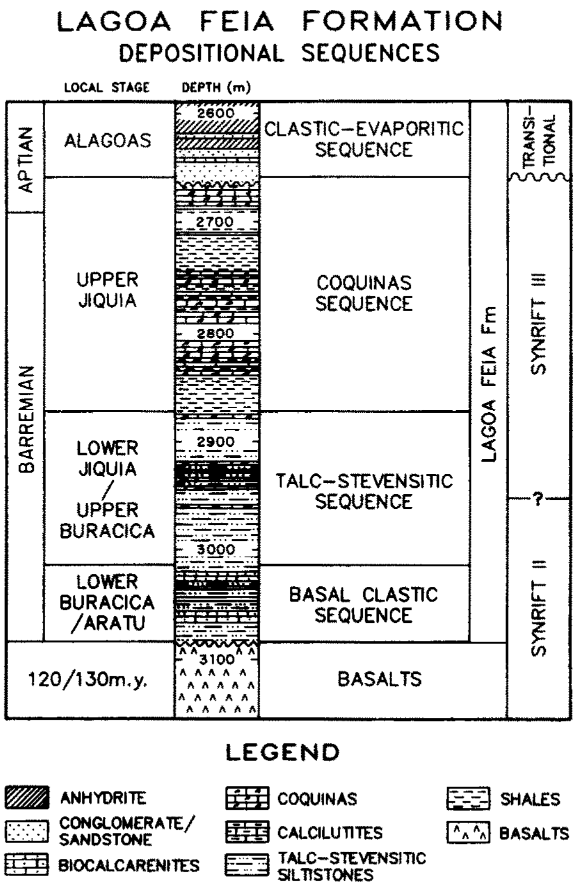

Fig. 11. Stratigraphic units of the Lagoa Feia Formation, Campos Basin, showing the syn-rift sequences, the rift/post-rift unconformity and the post-rift transitional evaporitic megasequence (modified after Dias et al., 1988).

towards the top, by gastropods. Carvalho et al. (1984) recognize several facies in these carbonates, such as peletoidal calcilutites, oolitic and bioclastic calcarenites and calcirudites with or without matrix, that are associated with the development of high-energy carbonate platforms and offshore banks and bars superimposed on syndepositional structural highs (Fig. 12). In the structural lows and in the deeper portions of the basin, marls and organic rich black shales accumulated; these represent the principal source-rock for oils generated in the Campos Basin. In some shallow parts of the rift lake, syn-depositional tectonic activity is held responsible for the development of coastal conglomerate and coarse-grained sandstone fan delta systems.

3.2 Transitional evaporitic megasequence

In most of marginal and interior rift basins, the continental syn-rift megasequence is covered by Aptian (Alagoas stage) sediments of the transitional megasequence (Figs. 3 and 13). A very conspicuous, angular unconformity separates these two megasequences and marks the rift/post-rift boundary. In the northeastern interior rift basins, Aptian sediments are thin, though widespread, and are composed of arid alluvial–fluvial coarse clastic rocks, which are associated with local sabkha deposits.

Along the present East Brazilian continental margin, the syn-rift sediments are overlain by a thick section of clastic and evaporitic sediments forming part of the transitional evaporitic megasequence (Figs. 4 and 5). The extensive development of alluvial and fluvial sediments in the basal part of this sequence testifies to a renewed period of intensified tectonic activity and reactivation of clastic sources. This coarse clastic system was deposited on a peneplane truncating the structural relief of the syn-rift sequences. In basinal areas siliciclastic sediments are replaced by carbonates, composed mainly of stromatolitic and nodular limestones, which were deposited in very shallow waters.

Progressive transgression, originating from the south, led to the development of a narrow seaway along the entire East Brazilian margin north of

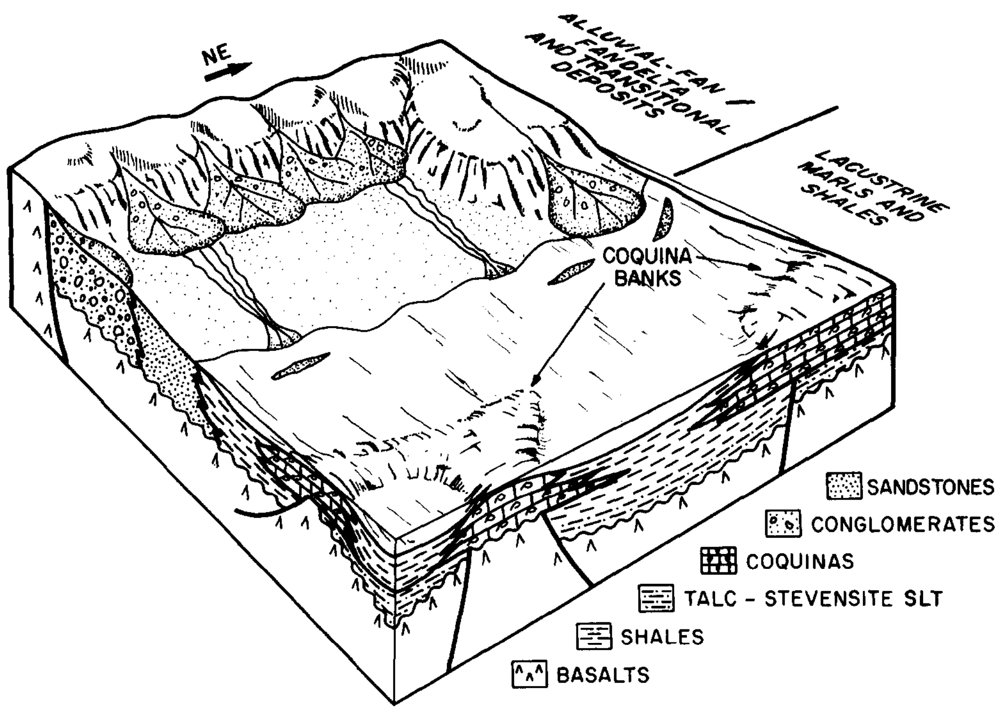

Fig. 12. Schematic depositional model of Syn-rift sequence-III in the Campos Basin (Guardado et al., 1989).

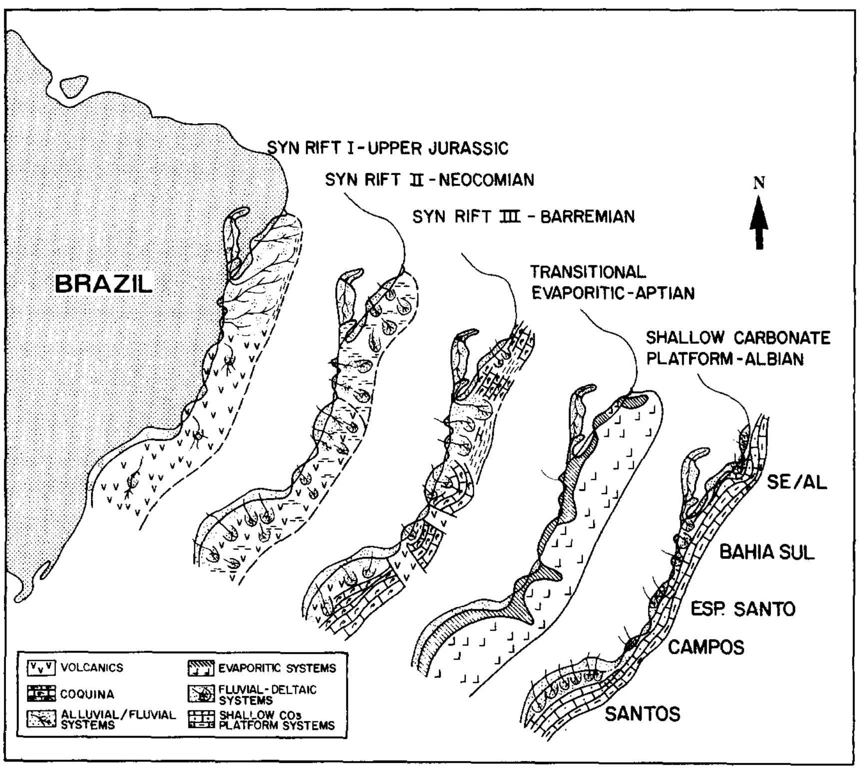

Fig. 13. Paleogeographic reconstruction of the syn-rift sequences and post-rift transitional evaporitic and shallow carbonate platform megasequences in EBRIS (Chang et al., 1988).

the Pelotas Basin. Very restrictive, euxinic and saline conditions predominated in this seaway due to the existence of a natural barrier to water circulation at the Walvis–São Paulo Ridge (Fig. 2). Massive halites, reaching thicknesses of as much as 2000 m accumulated in the most rapidly subsiding parts of the proto-South Atlantic Basin, whereas anhydrite, and in some places even sylvites and tachydrites were deposited on adjacent highs and platform areas.

3.3 Shallow carbonate platform megasequence

The Aptian (Alagoas stage) evaporites are capped by Albian marine carbonate platforms. This reflects a gradual opening of the South Atlantic Gulf, the break-down of the São Paulo–Walvis Ridge barrier and the gradual de-restriction of this evolving seaway (Figs. 3 and 13).

The very continuous and extensive high-energy Albian carbonate platform consists of oolitic, oncolitic, peletoidal and minor bioclastic limestones, reflecting a semi-restricted environment.

The Albian coastlines were associated with coarse clastic fan delta systems which prograded onto the adjacent carbonate platform.

Towards the axial parts of these basins, the carbonate banks give way to deeper water calcisiltites, calcilutites, marls and carbonatic shales; thus the depositional model of a carbonate ramp appears to apply to this sequence (Spadini et al., 1988). Salt tectonics strongly affected the distribution of the high-energy facies, mainly controlling oolite–oncolite banks, developing on the culmination of gentle salt pillow structures.

3.4 Open-marine megasequence

At the end of the Albian, gradual deepening of the basin, accompanied by marine transgressions, resulted in drowning of the high-energy carbonate platform (Chang et al., 1988) and deposition of a low-energy sequence consisting of rhythmically bedded calcilutites, marlstones and shales containing calcispheres and planktonic foraminifers. By the end of the Albian–Cenomanian, bathial conditions were established (Dias-Brito, 1982; Koutsoukos and Dias-Brito, 1987).

In several basins, density flow deposits are recognized in the form of channelized and blanket systems (Barros et al., 1982). Stacked channel sands, characterized by rapid lateral facies and

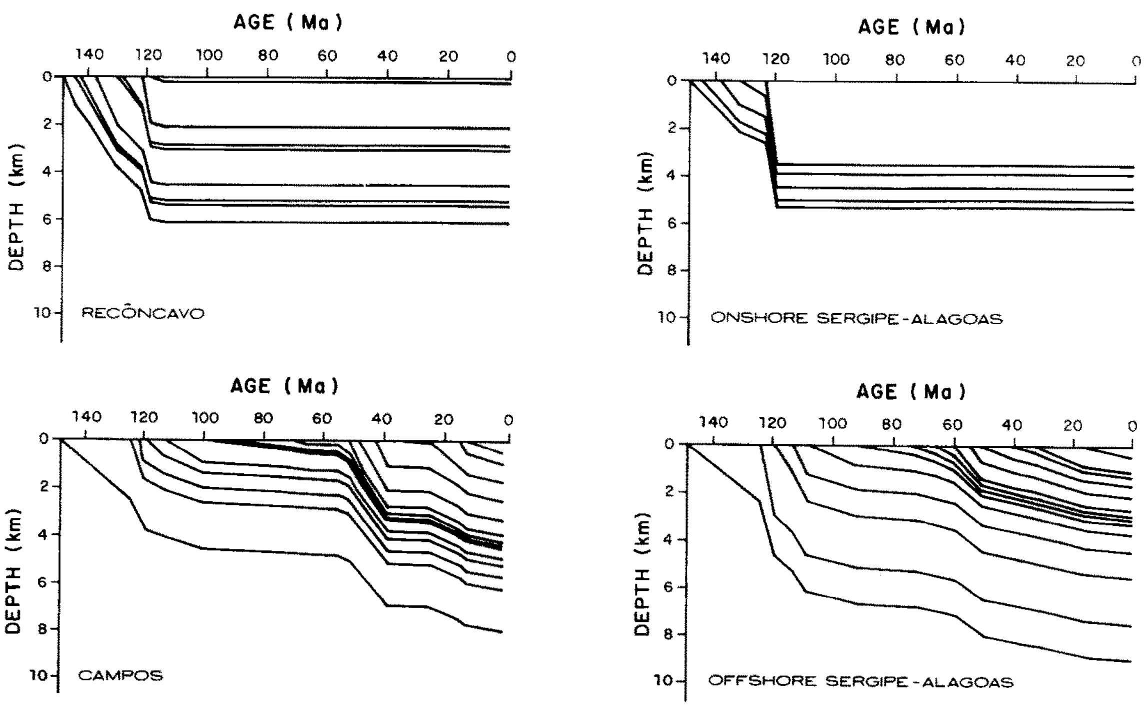

Fig. 14. Burial curves of selected wells in EBRIS.

thickness variations, were deposited during the Albian–Cenomanian and Late Cretaceous in narrow depressions of the basin slopes and floors that developed in response to intense salt flowage. Sheet like sands of turbiditic origin, covering as much as 250 km^2, accumulated during periods of Cenozoic low stands in sea level. These sands were sourced by the coastal Serra do Mar ranges (Carminatti and Scarton, 1991).

The Cenozoic passive margin sedimentary prism has an overall regressive character; it is characterized by a typical offlap configuration whereby top-set beds correspond to fan delta, fluvial-deltaic, siliciclastic shelf and carbonate platform, whereas fore-sets and bottom-sets consist of slope and basinal shales (Figs. 4 and 5). In some areas, the progradational pattern can be replaced by one of vertical aggradation, reflecting the build-up of high-energy carbonate platforms.

The southern Brazilian margin was dominated by siliciclastic systems whereas in the north carbonate platforms are best developed. This difference is also reflected by the amount of clastic turbidites. The Campos Basin contains much more turbidite sandstones than the Sergipe–Alagoas Basin; these sands contain major oil accumulations.

4. Subsidence history

The burial history curves given in Fig. 14 illustrate the subsidence history of the Sergipe–Alagoas, Recôncavo and Campos basins. The Recôncavo and onshore parts of the Sergipe–Alagoas Basin contain only syn-rift sediments whereas the Campos Basin and the offshore Sergipe–Alagoas Basin consist of rift basins which are overlain by a post-rift sedimentary sequence. Dating of the syn-rift sediments is highly uncertain as it relies on the correlation of local endemic ostracod zonation with the international geological time scale (Harland et al., 1982). Such a tentative correlation is given in Fig. 6 (Beurlen and Figueiredo, 1988). Despite these uncertain-

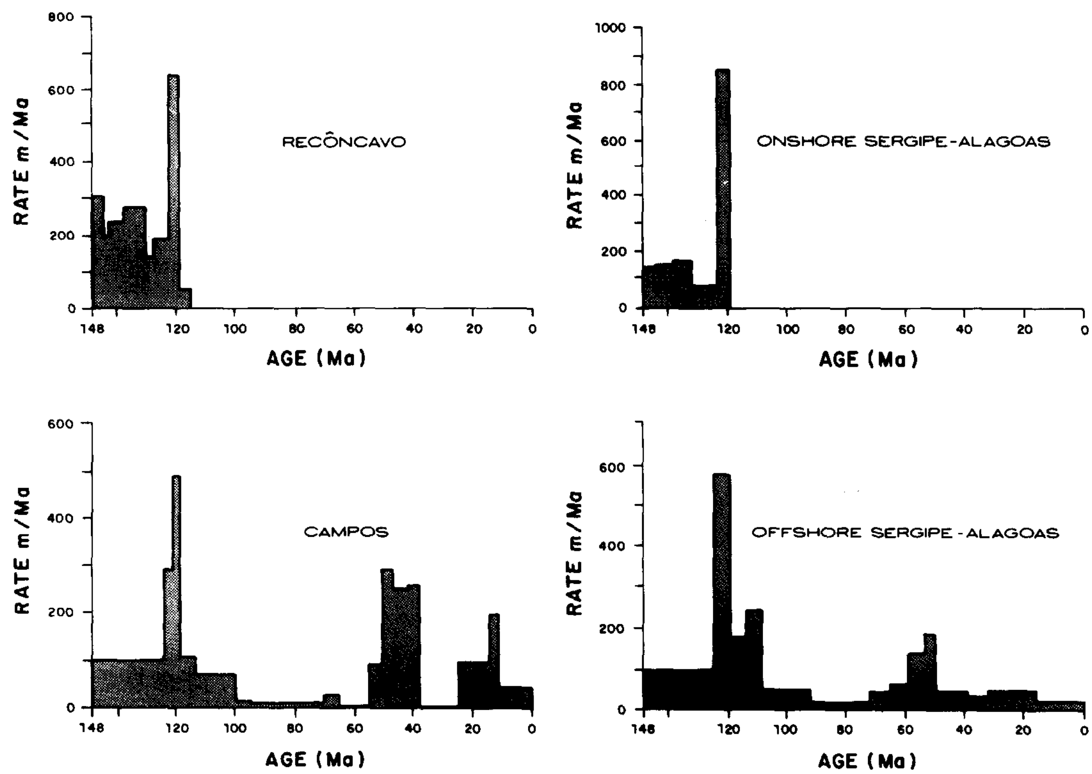

Fig. 15. Decompacted sedimentation rates of selected wells in EBRIS. Common high sedimentation rates occurring between 122 and 119 Ma coincide with last pulse of rifting. Note different vertical scales.

ties, a comparison between individual wells should be valid, since identical dating criteria were used.

Two basic patterns emerge from the burial curves (Fig. 14): subsidence of the Recôncavo and onshore part of Sergipe–Alagoas rift was completed by Early Aptian time, whereas the offshore Sergipe–Alagoas and Campos basins continued to subside until the Present.

Rift subsidence

Rift subsidence of the Recôncavo and onshore Sergipe–Alagoas basins increased sharply at the end of the Neocomian/Barremian, at least 2-fold increase from that of Neocomian (Fig. 15), and terminated abruptly during the Aptian. The synchroneity of these events contradicts the once popular notion that rifting activity shifted from the aborted Recôncavo rift, active during the early Neocomian, to the Sergipe–Alagoas rifts, active mainly during the Barremian, which afterwards evolved to a continental margin. Also in Campos Basin the subsidence accelerated conspicuously during the Barremian, towards the end of the rift stage; however, a comparison with Neocomian subsidence rates cannot be made, since no data is available below the basalts.

Post-rift subsidence

Geohistory diagrams of post-rift stage display a common high subsidence pulse during Aptian–Albian time, immediately following the end of rifting (Fig. 14); this is compatible with the thermal cooling of the attenuated continental lithosphere. Sedimentation rates decreased after the Albian and reached a minimum during the Late Cretaceous first-order highstand in eustatic sea level. The end of this period of basin starvation is not synchronous along the East Brazilian margin and is essentially controlled by sediment supply, driven by local tectonics (Chang et al., 1988). The apparent high Tertiary subsidence rates are mainly the effect of sedimentary loading of the

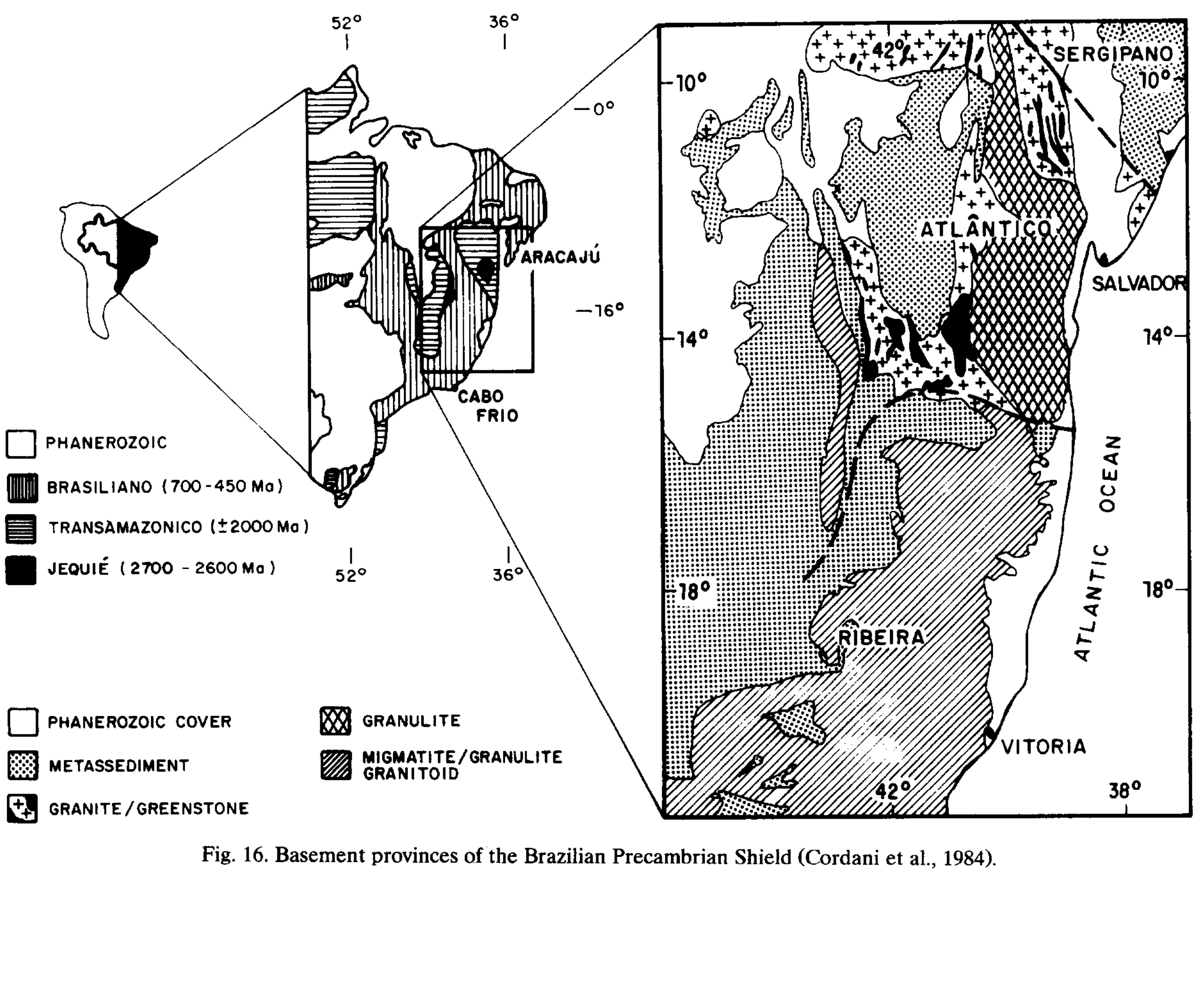

Fig. 16. Basement provinces of the Brazilian Precambrian Shield (Cordani et al., 1984).

lithosphere in response to a high clastic supply, resulting in rapid shallowing of the basin and seaward progradation of the shelf (Fig. 15).

5. Structural styles

EBRIS developed as result of Late Jurassic/Early Cretaceous lithospheric extension. The general trend of grabens and horsts forming this rift system is NE, subparallel to the principal basement lineaments (Fig. 17). Other, locally important trends, either follow local basement fabric anisotropies or crosscut them; the latter occurs particularly when the major basement lineaments parallel the orientation of the stress field (Milani and Davison, 1988).

The structure of EBRIS can be readily observed in the Recôncavo–Tucano–Jatobá (RTJ) aulacogen, which lacks post-rift sediments, but is also evident in the Sergipe–Alagoas and Campos basins on the continental margin. Basement-involved structures are less well defined in the Espirito-Santo, Santos and Pelotas basins, because of thick salts and/or post-rift volcanic layers and limited deep reflection seismic resolution.

5.1 Precambrian basement provinces

The Precambrian shield along the East Brazilian margin, can be subdivided into the Ribeira, Atlântico and Sergipano provinces (Fig. 16; Cordani et al., 1984; Schobbenhaus et al., 1984).

Three major regional tectono-magmatic events are recorded in the Brazilian shield. The oldest, Jequié regional event, goes back to Archean (approximately 2700–2600 Ma); the Early Proterozoic Transamazonian event occurred between 2200 and 1800 Ma. During the Late Proterozoic Brasiliano event (700–500 Ma), major restructuring of the Brazilian shield took place; this major tectonic event is time-correlative with the Pan-African event. Thereafter the Brazilian shield maintained its integrity until the breakup of the Gondwanaland.

The *Ribeira Province* south of 16°S (Fig. 16) is characterized by high grade metamorphic rocks (migmatized amphibolite and granulite facies, gneisses and quartzites) of Archean age, which are juxtaposed to Early Proterozoic schists and Late Proterozoic low grade metasediments (phillites and greenschists). These units were overprinted by the Brasiliano tectono-magmatic events. The high grade metamorphic units dominate north of Cabo Frio, and form the basement of the Espirito Santo Basin (Fig. 16).

The *Atlântico Province* extends from 16°S to 8°S and consists of an Archean granulite core, which is surrounded by granulite and granite/greenstone belts, reworked during the Transamazonian cycle. This is the oldest basement province of the East Atlantic margin; it is limited towards the Ribeira block by a pronounced gravity anomaly, characterized by a gradient as high as 1 mGal/km (Inda et al., 1984).

The Atlantic granulite/migmatite complexes extend northward into the *Sergipano Province* where they are overlain by a thick cover of low grade Late Proterozoic metasediments; these consist of limestones, quartzites, phillites and metagraywackes. Older basement rocks and Late Proterozoic sedimentary cover were intensely deformed during the Brasiliano orogenic cycle and are pierced by very large batholiths particularly in the State of Alagoas.

5.2 Basement fabric

The structural fabric of the East Brazilian shield is characterized by notable distinctions, which do not necessarily coincide with lithologic and basement province boundaries.

The compositional fabric of the high and low grade metamorphic rocks of the southern Ribeira province differs from the fairly homogeneous high-grade terranes of the northern Ribeira and Atlantico provinces; the latter extend from Campos (21°S) northward to Aracaju (11°S; Fig. 16). This change in fabric coincides with a deflection of lineament trends from NE to NNE (Fig. 17). Figure 17 presents a map of basement lineaments for eastern Brazil and the adjacent shelves. This map has been compiled from airborne radar images (Cunha, 1984, 1985, 1987) and reflection seismic data (Falkenhein et al., 1985; Netto and Oliveira, 1985; Dias et al., 1988; Gomes et al., 1988). Most lineaments represent foliations; south

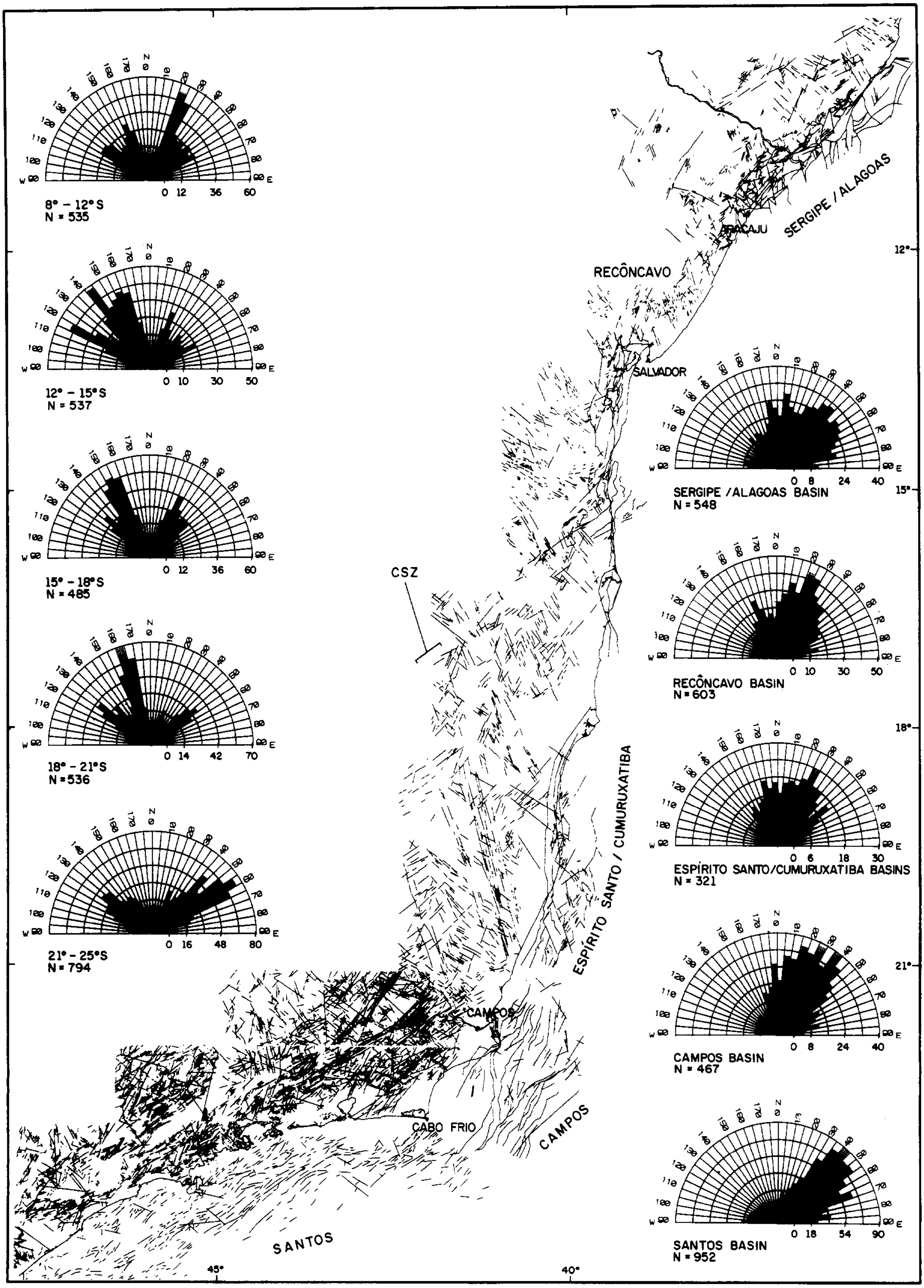

Fig. 17. Lineaments along the East Brazilian borderland. Onshore basement lineaments are based on airborne radar images. Rift basin lineaments are compiled from seismic structural maps of basal rift units except in the Santos Basin, where they originate from airborne magnetic surveys and may represent basement grain rather than rift structures (Sources cited in text). *CSZ* = Colatina shear zone. Rose diagrams represent absolute distribution of the lineaments. *N* = number of lineaments recorded.

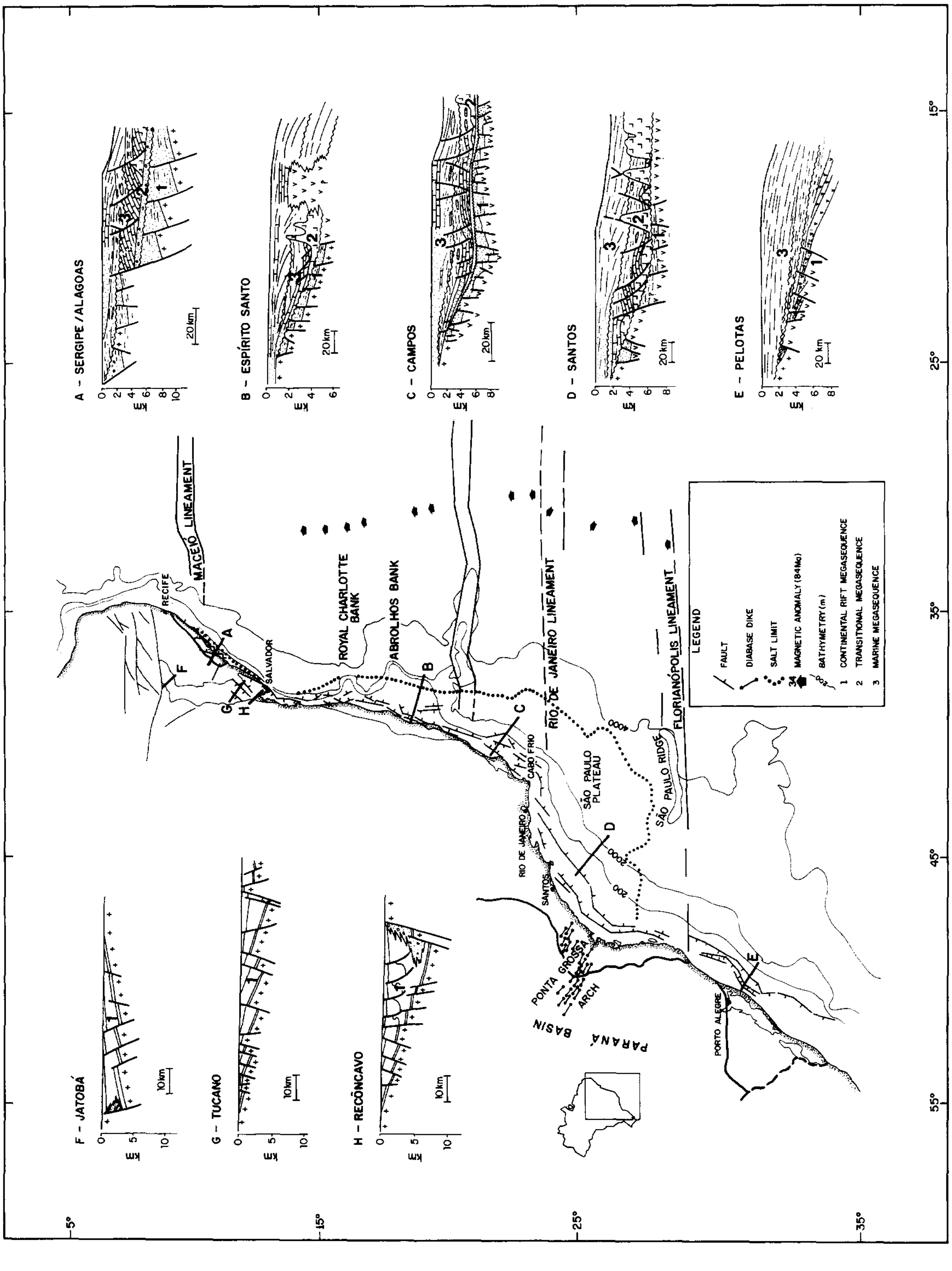

Fig. 18. Tectonic framework of EBRIS. Map and schematic basin cross-sections modified after Asmus and Guazelli (1981).

of 22°S, NE trending features coincide with major subvertical ductile shear zones; NW and NNW trending lineaments include a great number of brittle shear zones, of which the most prominent is the Colatina cataclastic zone projecting into Campos Basin.

North of Aracaju the compositional fabric of the Precambrian shield changes again at least in its upper levels, which are composed of low grade metamorphic rocks, involved in the thin- skinned Brasiliano fold belt. These subhorizontal detachment surfaces contrast markedly with the subvertical discontinuities of the Ribeira province to the south. These contrasting basement structural fabrics, influenced the structural style of the Mesozoic rifted basins.

5.3 Mesozoic rift tectonics

The orientation of the Mesozoic rift lineaments is quite homogeneous along EBRIS and consists mainly of extensional normal faults and transfer faults; the former generally strike NNE to NE whereas the latter trend NNW to NW (Figs. 17 and 18).

The *Recôncavo-Tucano-Jatobá (RTJ)* rift is characterized by a sigmoidal configuration. It consists of the N–S trending Tucano graben which terminates to the north at the NE trending Jatobá Basin and grades in the south into the SW trending Recôncavo Basin. This rift system consists of asymmetric grabens, containing up to 7 km of sediments which are bounded on one side by high-angle planar normal faults and on the other by gentle ramps cut by several minor normal faults (Figs. 18 and 19; Milani, 1985; Magnavita and Cupertino, 1987, 1988; Milani and Davison, 1988; Santos and Braga, 1990; Figueiredo et al., 1990; Magnavita, 1990; Roque, 1990; Ferreira, 1990). Several NW oriented dextral transfer zones segment the RTJ rift (Milani and Davison, 1988). However, a change in polarity occurs only between Central and North Tucano graben, along a reactivated Precambrian transcurrent fault (Vaza–Barris fault; Fig. 19).

Diapirism of lacustrine clays plays an important role in the structuration of the Recôncavo Basin (Fig. 18); individual diapirs are over 1 km

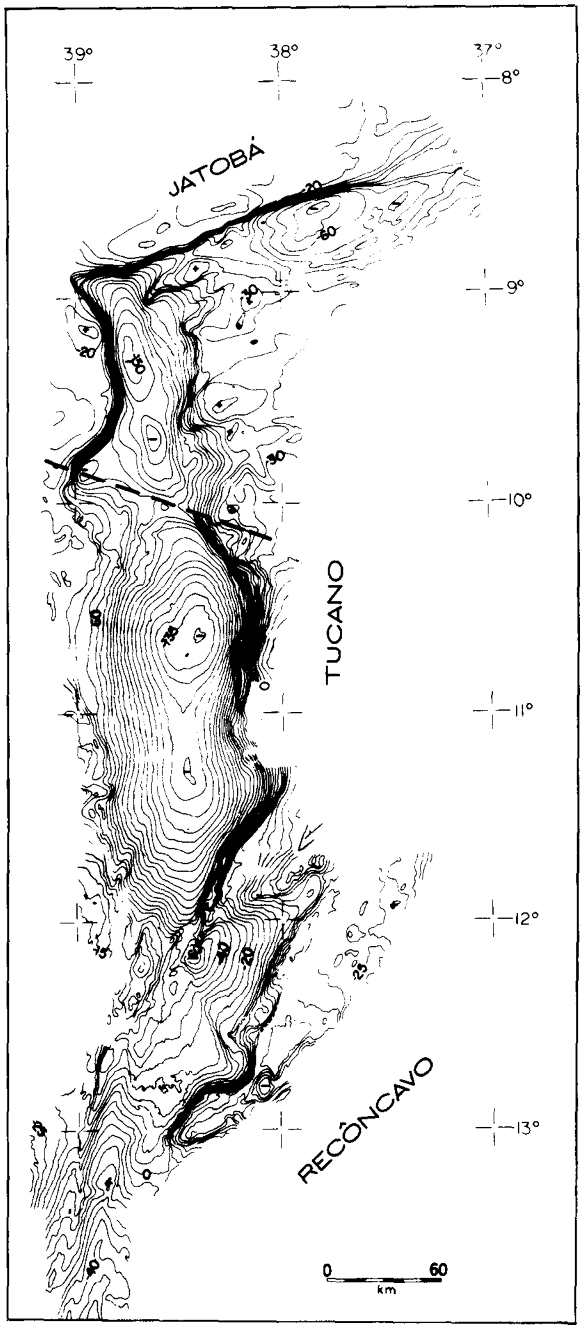

Fig. 19. Bouguer gravity map of the Recôncavo–Tucano–Jatobá rift. Contours in mGal (courtesy PETROBRÁS/SEMEPO). Note the polarity reversal indicated by the positions of the steepest gradient (master faults) at the dashed Vaza Barris Fault.

high and form ridges subparalleling the graben axis.

The structural framework of the *Sergipe–Alagoas Basin* is controlled by fault systems trend-

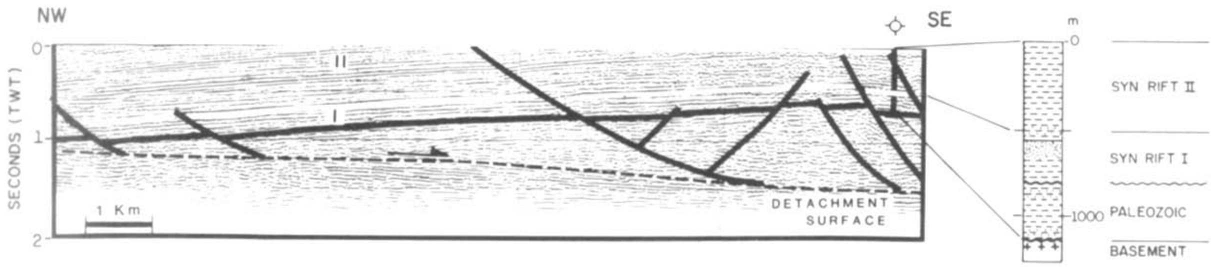

Fig. 20. Seismic section in the Sergipe–Alagoas Basin with well showing the syn-rift stratigraphic column. Note large wavelength rollover produced by basement-involving low angle listric fault (after Guimarães, 1988).

ing N–S, ENE and NE (Figs. 17 and 18). Basement involving synthetic normal planar step-faults predominate. Low angle planar and listric faults, involving sediments or even basement, are more abundant in the Sergipe sub-basin, where they are associated with roll-over structures (Figs. 5 and 20; Falkenhein et al., 1985; Lana, 1985; Guimarães, 1988).

A set of en echelon N–S and ENE striking high-angle normal faults, affecting the syn-rift sequence, separates the shallower onshore grabens from the deeper offshore half-graben. Throws on these faults may reach up to 5 km. Throws on N30°E to N40°E trending faults, which dip 45° to SE and cut through the basal rift section, are in the range of few hundred up to 2000 m (Destro et al., 1990). Statistical analyses indicate that NE striking faults, centered at N30°E, have the highest frequency (Lana, 1985; Guimarães, 1988). Upper crustal extension by faulting was taken up by NW striking faults and along relay ramps (Destro et al., 1990). ENE striking faults are interpreted as transfer faults, accommodating down-throw displacement along the N–S oriented normal faults (Guimarães, 1988).

Post-rift structuration is mainly the effect of basin tilting and salt flowage during thermal subsidence (Fig. 5); associated faults sub-parallel the present coastline.

The rift structures of the *Espirito Santo Basin* is only defined in its shallower portions (Gomes et al., 1988). General fault trend is NE with a wide range of directions from N25°W to N60°E (Fig. 17). Major faults, separating the shallower landward portion of the basin from its deeper

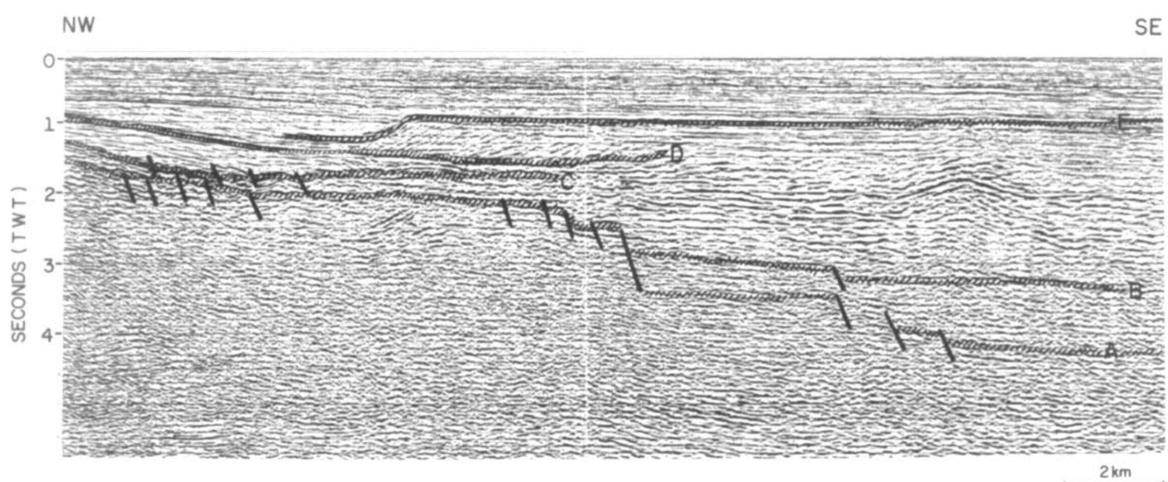

Fig. 21. Seismic section in Espirito Santo Basin across Abrolhos Bank. A = basement; B = base of evaporites; C = top of shallow carbonate platform; D = boundary between marine transgressive and regressive sequences; E = top of Abrolhos volcanics.

offshore parts, strike approximately N–S, whereby synthetic normal faults dominate (Fig. 18). Post-rift structures are very conspicuous and include salt domes and pillows trapped behind the massive Tertiary Abrolhos and Royal Charlotte volcanic banks (Fig. 21).

The rift structure of the *Campos Basin* is characterized by a series of horsts, and grabens and half-grabens filled with volcanics and sediments (Dias et al., 1987). Synthetic and antithetic faults, bounding laterally continuous blocks, have throws up to 2500 m. Orientation of these faults follows the Precambrian basement grain, which trends approximately N30°E in the southern and NNE in the northern portions of the basin (Fig. 17). In the northern extreme of the basin, faults strike NNW and project towards the Colatina cataclastic shear zone of the outcropping shield. Syn-rift faults are truncated by the pre-Aptian unconformity. In a few cases syn-rift faults were later reactivated and affect the post-rift sequence (Fig. 4). Post-rift structures are related to salt movement triggered by the seaward tilting of the basin during its thermal subsidence. The resultant listric faults (Fig. 4) are oriented closely to the rift trends denoting a common fundamental tectonic control (Guardado et al., 1990).

Very little is known about the rift structure of the *Santos Basin* due to the thickness of the evaporites and other post-rift sediments (Pereira et al., 1986). General structural trends are thought to follow a NE direction that was inherited from the basement grain, as suggested by magnetic lineations (Fig. 17). A gentle hinge zone ramp separates the shallower part of the basin, in which Tertiary sediments rest on syn-rift series and the basement, from the deep portion of the basin, which is characterized by small synthetic faults. In the latter, rift structures, consisting of gently tilted blocks, limited by steep planar normal faults, are evident in areas of salt withdrawal. The post-rift structural style of the basin is dominated by salt pillows along the hinge zone and by salt domes further offshore, where the base of the salt is essentially flat. Growth and collapse features, associated with the diapirs are frequent, specially in the deeper parts of the basin.

On the adjacent *São Paulo Plateau*, individual diapirs give way to compressive salt ridges (Cobbold and Szatmari, 1991), which reach almost to

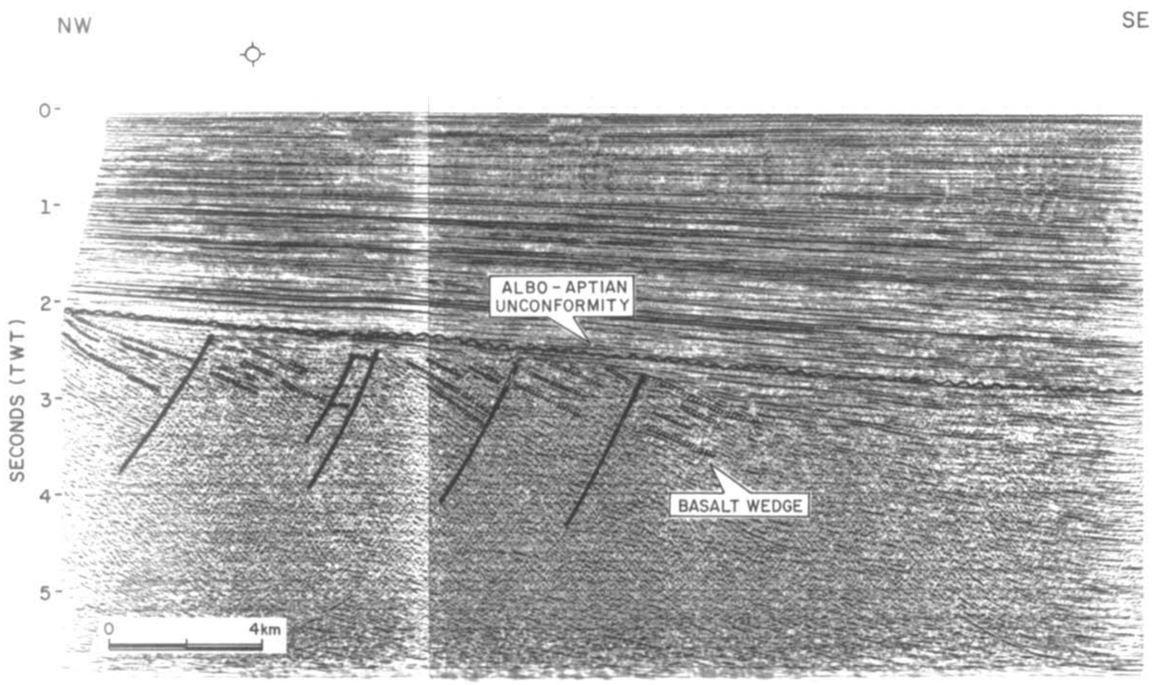

Fig. 22. Seismic section in Pelotas Basin, showing seaward dipping basalt wedges cut by antithetic faults, Aptian–Albian unconformity and post-rift planar reflections above unconformity.

the seafloor. The Plateau ends eastward with a pronounced salt scarp marking the eastern edge of the Aptian evaporite province (Leyden et al., 1976). To the south, the Plateau terminates against the São Paulo volcanic ridge, which represents one of the largest transverse structures in the South Atlantic (Fig. 1), and forms a conjugate to the Walvis Ridge.

The rift structure of the *Pelotas Basin* is characterized by seaward dipping basalt wedges (Fontana, 1987) which are frequently crosscut by antithetic normal faults (Fig. 22). Models proposed for the occurrence of such volcanic wedges include two end-members (Morton and Taylor, 1987), namely that these wedges are rift-related and overlie thinned continental crust (Hinz, 1981),

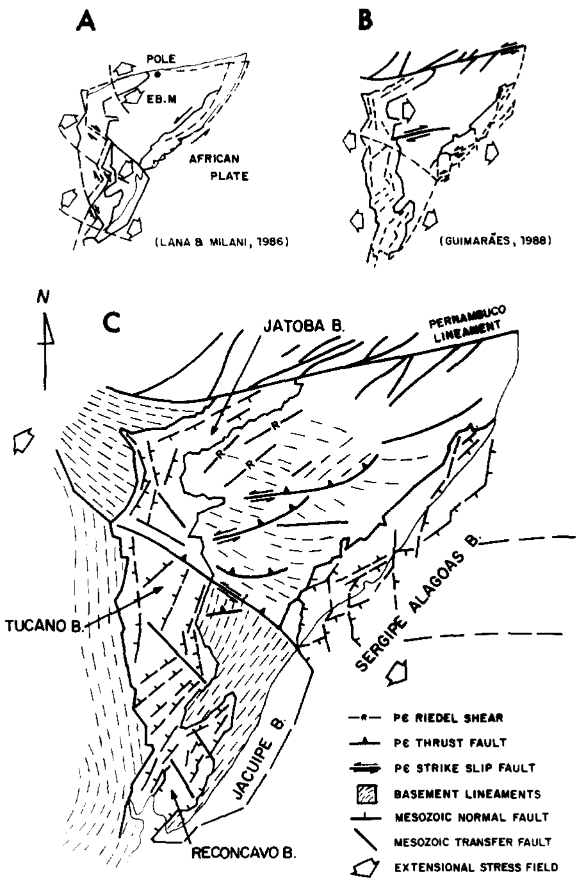

Fig. 23. Kinematic models of northeast EBRIS evolution. (A) Microplate rotation (Lana and Milani, 1986). (B) Homogeneous E–W extension (Guimarães, 1988). (C) Proposed single NW–SE extension. Base map compiled from Milani and Davison (1988) and Magnavita (1990).

or that they formed during an initial phase of subaerial seafloor spreading and thus lie on the ocean/continent boundary (Mutter et al., 1982). The seaward dipping wedges in Pelotas Basin are often cut by antithetic normal faults; they occur close to the hinge zone which separates stretched continental crust, overlain by Tertiary sediments, from stretched crust covered by Early Cretaceous sediments. The oxidized nature of the vesicular basalts, and associated occurrences of conglomerates, consisting of metamorphic and volcanic rocks, indicates that they were subaerially exposed, close to the continent (Fontana, 1987); therefore, a continental rift-related origin of these features appears to be more likely.

Some grabens, of limited width (6 km), occurring at the hinge zone (Fig. 18), have been interpreted as tension cracks formed by the flexing of the basement during the post-rift subsidence (Gonçalves et al., 1979). However, the age of their sedimentary fill indicates that they formed during the rifting stage (Fontana, 1987).

The general lack of structuration of the post-rift sequence can be related to the absence of Aptian evaporites. Monotonous, gently diverging, planar reflections onlap the Aptian–Albian unconformity (Fig. 22). An exception are the gravitational growth faults and downslope compressional features, affecting post-Oligocene sediments in the area of the Rio Grande Cone (Fontana, 1990).

5.4 Kinematic model

A kinematic model for the EBRIS rifting must reconcile the geometry of basins along the entire rift system with the history of the stress field. In particular, compatibility must be sought when dealing with the NE segment, which includes the Recôncavo–Tucano–Jatobá (RTJ) aborted branch and the Sergipe–Alagoas on- and offshore rift (Fig. 23).

Two basic models have been proposed for the structural configuration of the Recôncavo, Tucano, Jatobá, Sergipe–Alagoas and Jacuípe basins (Ussami et al., 1986; Milani et al., 1987; Castro, 1987, 1988). The first model proposes that these rifts delimit a microplate which rotated counterclockwise around a pole located just east of Jatobá Basin (Fig. 23a; Szatmari et al., 1985; Lana and Milani, 1986; Milani and Davison, 1988); the alternate model assumes a homogeneous stress field (Fig. 23b), and regional E–W oriented extension (Guimarães, 1988; Magnavita, 1990).

The *microplate model* implies a concentric extensional stress field, oriented approximately N50°W, acting upon the RTJ rift system, thereby generating a set of NE striking normal faults; at the same time sinistral strike-slip motions are activated in the Sergipe–Alagoas Basin (Fig. 23a). Segmentation of the RTJ rift system was accommodated along the NW oriented dextral transfer faults. According to Milani and Davison (1988), a 2° rotation was sufficient to accommodate the 20% extension estimated for the Recôncavo and Tucano basins. In the Sergipe–Alagoas Basin, N–S oriented faults are interpreted as tensional segments of the NE striking sinistral shear zone, which subparallels the basin margin (Lana, 1985); within this same shear system, NE faults are interpreted as Riedel shears.

According to the *alternative model* of Guimarães (1988), the N–S trending faults in both the Sergipe–Alagoas and the RTJ basins are interpreted as normal faults; in the Sergipe–Alagoas Basin, E–W faults are interpreted as sinistral transfer faults. E–W oriented faults are not significantly present in the RTJ rift system. Nevertheless, Magnavita (1990) showed that an E–W extension across the RTJ system could explain the observed fault pattern. He assumed that the main grabens formed in response to oblique extension and are bounded by two sets of NW and ENE striking transverse accommodation zones, whereby the NW striking fracture system corresponds to the dextral transfer zones of Milani and Davison (1988).

A major problem with the microplate rotation model (Szatmari et al., 1985; Lana and Milani, 1986) is the lack of evidence of Neocomian strike-slip motion in the Sergipe–Alagoas Basin (Guimarães, 1988; Destro et al., 1990). NE striking faults (dominantly N30°E) show essentially dip-slip motion with steeper dips to NW (approximately 60°) and more gentle dips to SE (dominantly 50°) and rotation of major blocks in a SE direction. Moreover, transfer faults in the RTJ rift system do not conform with small circles derived from the pole of rotation of the microplate and do not show a progressive change in

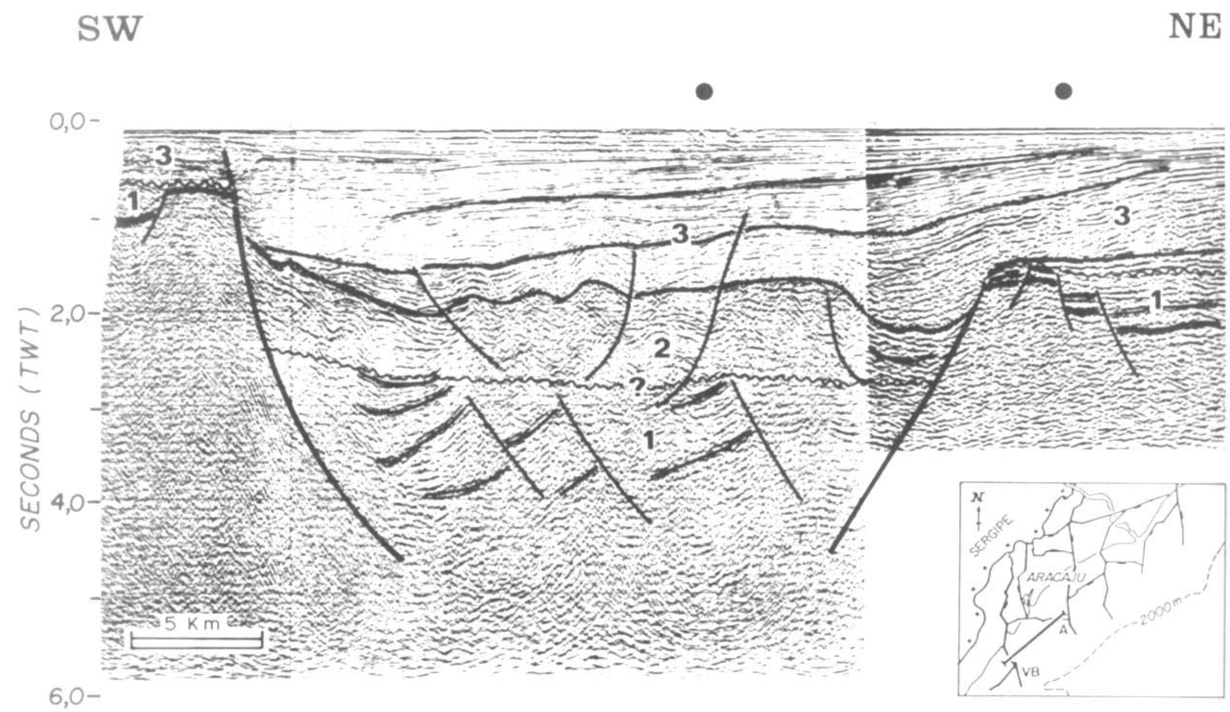

Fig. 24. Seismic section through the Sergipe–Alagoas Basin crossing the Vaza Barris (left) and Atalaia (right) major normal faults. *1* = Continental syn-rift megasequence; *2* = transitional evaporitic megasequence; *3* = marine megasequence.

strike direction towards E–W with increasing distance from the pole (Castro, 1988); in fact, the opposite occurs (Fig. 23c).

The E–W extension model of Guimarães (1988) and Magnavita (1990) has the problems that there is no clear evidence of strike-slip motion and associated pull-apart origin for Jatobá Basin. The NE oriented Mesozoic diabase dike, which Magnavita (1990) interprets as having intruded along tension fractures produced by a sinistral simple-shear movement along the Pernambuco lineament, may be better interpreted as associated with a pure NW–SE extension. Such a homogeneous extensional stress field is more compatible with the N30°E normal fault orientation in the Recôncavo and Tucano grabens.

Our preferred *single homogeneous NW–SE oriented extension model* is compatible with the structural style of the Sergipe–Alagoas Basin (Fig. 23c). It fits the dominant N30°E normal fault pattern of the Basin. The general N–S and E–W trend of the major normal faults, separating the deep parts of the basin from the shallow ones, are interpreted as pre-existing crustal weakness zones, which were reactivated under an oblique extensional stress field. Sand-box experiments, simulating an oblique extension along the bisector of orthogonally bounded blocks, produced two sets of dominantly dip-slip normal faults. These faults trend subparallel to the inferred weakness zones. Similar results were obtained by Withjack and Jamison (1986) in clay models, particularly near the edges of their experiment. Figure 24 gives a NE oriented seismic line, cross-cutting the major N–S (Vaza–Barris) and E–W (Atalaia) faults which exemplifies this situation.

Neocomian (Syn-rift-II) NW–SE extension appears to be applicable also to the Espirito Santo and Campos basins, since their dominant normal fault orientation follows a NE direction (Fig. 17). Furthermore, this single, rather homogeneous extensional stress field is also compatible with the direction of extension during the Barremian/ Earliest Aptian last pulse of rifting (Jiquiá stage, Syn-rift-III).

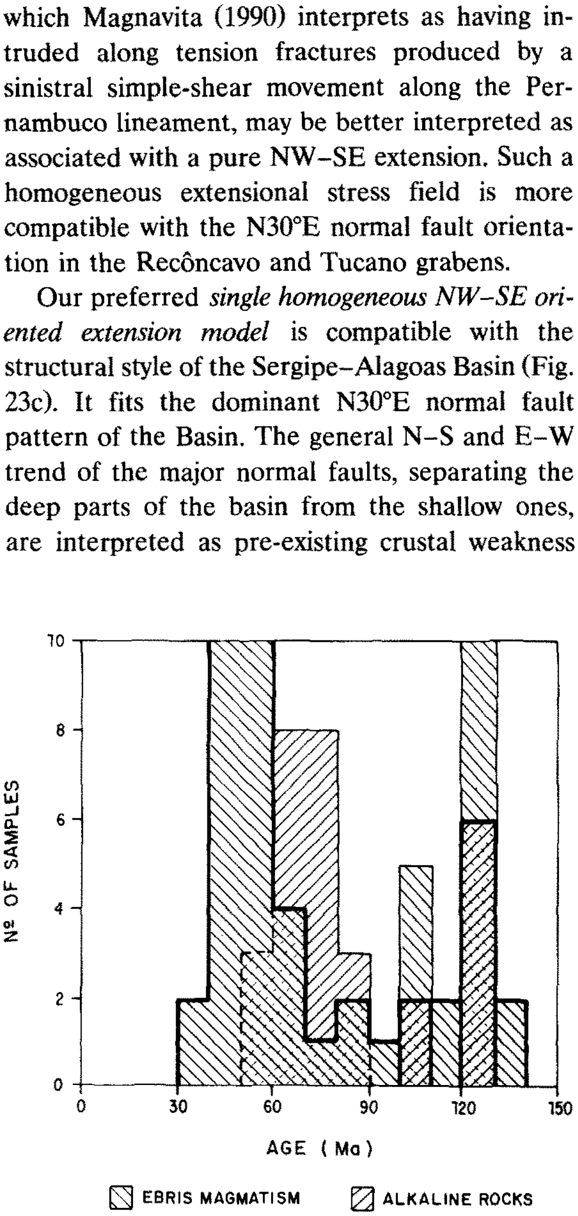

Fig. 25. Histogram of K–Ar ages of EBRIS basaltic rocks and of Mesozoic/Cenozoic alkaline rocks of the East Brazilian shield basement. Sources: EBRIS (Cordani, 1970; Cordani and Blazekovic, 1970; Asmus and Guazelli, 1981; Fodor et al., 1983; Mizusaki and Saracchini, 1991); alkaline rocks (Ulbrich and Gomes, 1981).

6. Magmatism

Magmatic activity in EBRIS spanned Early Cretaceous through Mid-Tertiary times and peaked between 130 and 120 Ma, 110 and 100 Ma, and 60 and 40 Ma (Fig. 25). Magmatism is essentially restricted to areas south of 16°S, and coincides roughly with the distribution of the outcropping Paraná Basin flood-basalt province (Fig. 26).

Syn-rift volcanic rocks occur in the Pelotas, Santos, Campos and Espirito Santo basins and also in the conjugate West African Moçamedes and Cuanza basins (Asmus, 1975). These rift basalts, which are overlain by younger syn-rift sediments, generally represent the acoustic basement.

The available radiometric age determinations indicate that volcanic activity commenced in Paraná Basin around 150 Ma and peaked around 125–135 Ma (Rocha Campos et al., 1988); it thus predates by some 10 Ma the earliest volcanic manifestations (138 Ma) in the EBRIS which peaked around 130–120 Ma (Fig. 25). Although the age of the earliest syn-rift sediments is only poorly constrained (±150 Ma, Tithonian), crustal extension appears to have commenced concomitantly with the first volcanic activity in the Paraná Basin.

Syn-rift basaltic rocks have only been studied in some detail in the Campos Basin (Mizusaki, 1986; Mizusaki et al., 1988). These consist of lava-flows and volcaniclastics. Individual lava-flows have an average thickness of 3 to 4 m whereby their top is marked by vesicular basalts; textures vary from hyaline at the base and top of the flow to hemicrystalline and holocrystalline in the center. Basalts are alkali-rich due to chemical alteration by alkaline lake water. Those samples that experience loss of volatiles on ignition of less than 4%, and are thus considered less altered, display a tholeitic affinity (Mizusaki et al., 1989). The geochemical composition of the Campos Basin rift basalts is similar to the flood-basalts of the southern Paraná Basin province (Bellieni et al., 1984), particularly when elements such as TiO_2, Zr, V, P_2O_5 and Sr are considered (Mizusaki et al., 1989).

Volcaniclastic rocks, associated with the lava-flows consist of autoclastic (friction breccias), pyroclastic (tuffs and hydrovolcanic breccias) and epiclastic (in situ weathering) rocks (Mizusaki, 1986; Mizusaki et al., 1988). Friction breccias and oxidized tuffs with pedogenetic fractures are indicative of subaerial volcanism; associated lava-flows exhibit a reddish color, vesicles, mineral alterations and fractures. Hydrovolcanic breccias, intercalated with sedimentary rocks, are interpreted as being deposited in a subaqueous environment. Well sorted and rounded volcanic arenites lacking a primary matrix formed near the coastline. Thus, volcanic activity in Campos Basin was partly subaerial and partly subaqueous.

Early Cretaceous alkaline rocks (130–120 Ma) occurring onshore of Santos Basin just north of the Ponta Grossa Arch in the area of the Precambrian Ribeira fold belt (Ulbrich and Gomes, 1981) have a similar age as the syn-rift volcanics of the Campos Basin (Figs. 25 and 26). They consist of

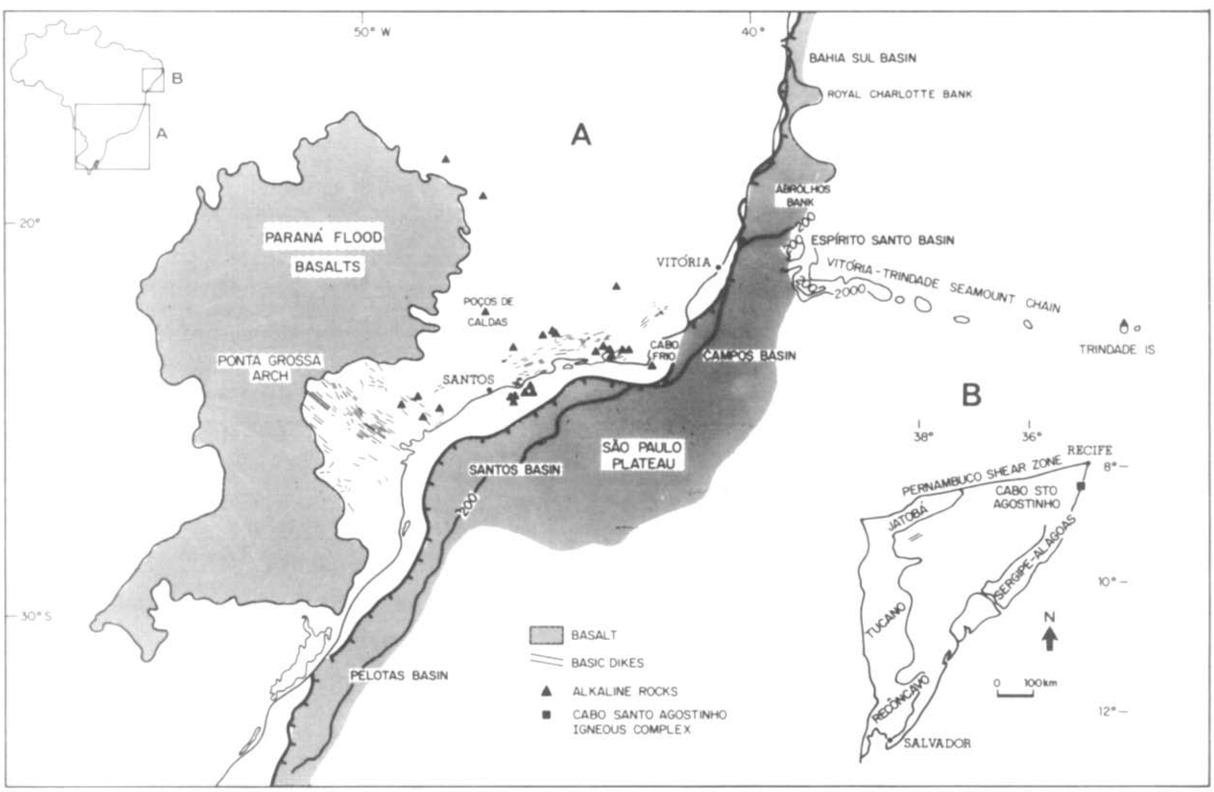

Fig. 26. Distribution of magmatism in EBRIS and adjacent regions. Sources: Borba (1978); Schobbenhaus et al. (1984); Almeida (1986); Magnavita (1990).

mafic to ultramafic, alkali- saturated rocks, such as dunites, peridotites and pyroxenites.

Late Cretaceous and Cenozoic igneous rocks occurring onshore and offshore of East Brazil have contrasting compositions (Asmus and Guazelli, 1981). Offshore, they are characterized by basaltic composition and range in age between 90 and 30 Ma. Onshore, volcanic rocks are alkaline and yield ages in the range 90–50 Ma (Fig. 25).

Late Mesozoic and Cenozoic alkaline rocks occurring around the northern margin of the Paraná Basin, on the Ponta Grossa Arch, on various coastal islands north of Santos city and particularly along a WNW trending line between Poços de Caldas and Cabo Frio (Fig. 26; Ulbrich and Gomes, 1981) are described as syenitic associations of various degrees of saturation (alkali-syenite, nepheline-syenite, phonolite and tinguaite), mafic to ultramafic alkali-saturated to peralkaline associations (dunite, peridotite and pyroxenite) and alkali-granite/alkali-syenite massifs. The alignment of alkaline rocks has often been cited as a hot-spot track, however, so far the pattern of radiometric ages has failed to confirm this hypothesis (Almeida, 1983).

On the shelf, magmatic activity occurred in the vicinity of the Cabo Frio High and, most conspicuously, as the huge basaltic edifice of the Abrolhos and Royal Charlotte banks, in the Espirito Santo and Bahia Sul basins (Figs. 1 and 26). Rocks of the Abrolhos Bank yielded radiometric ages around 50–40 Ma (Fig. 25) and consist of basalts, diabase dikes, tuffs, volcanic breccias and associated sediments. Basalts are characterized by ophitic to subophitic crystalline textures (Cordani and Blazekovic, 1970). The Abrolhos Bank volcanics form part of the Vitória–Trindade Seamount chain, which has been interpreted as a hot-spot track. K–Ar dates from Trindade Island (Fig. 26) range between 4 Ma and Present (Cordani, 1970) and are the only ones supporting this concept.

A Cretaceous volcanic suite in the coastal strip immediately south of Recife, known as the Cabo Santo Agostinho magmatic province (Fig. 26), has been the focus of many studies because of its peculiar composition and its isolated occurrence in EBRIS, north of Espirito Santo. It is composed of olivine basalts, trachites, alkali rhyolites, emplaced as flows, dykes and sills, and granites (Figueiredo et al., 1978). These rocks cut through Precambrian basement and the Cabo conglomerates of Aptian age (Alheiros et al., 1989). K–Ar and Rb–Sr ages range from 90 to 114 Ma (Vandoros and Valarelli, 1976). Geochemical data indicate that the volcanic rocks evolved by a slow fractional crystallization from an alkali olivine basaltic parental magma (Legrand et al., 1978; Borba, 1978). Strontium isotope data ($^{87}Sr/^{86}Sr = 0.702$) supports a mantle origin for the granites, rather than one resulting from crustal contamination (Legrand and Figueiredo, 1979).

A further magmatic occurrence, north of Espirito Santo, are 105 Ma old diabase dikes near the Pernambuco shear zone which are associated with NE–SW oriented mega-Riedel fractures in Precambrian basement rocks (Figs. 23c and 26), indicative of the sinistral movement along the latter (Magnavita, 1990; pers. commun., 1991). However, this pre-existing NE–SW anisotropy lies roughly orthogonal to the Mesozoic extensional stress field and may originate from it. Previous dates obtained from these diabase dikes indicated an Early Jurassic age (212 Ma; Santos and Souza, 1988).

A first attempt to interpret the rift phase magmatism in southern EBRIS was made by Asmus and Porto (1980) who associated crustal updoming to partial melting of the underlying mantle without specifying its driving mechanism. Chang and Kowsmann (1984) proposed that the mechanism for large scale magma generation in southern EBRIS was adiabatic decompression of the asthenosphere and lower lithosphere resulting from lithospheric extension greater than a stretching factor of 2 (Foucher et al., 1982). In contrast, the northern parts of EBRIS are characterized by negligible magmatism, whereby highly stretched crust is limited to a relatively narrow zone. More recently, White and McKenzie (1989) proposed that major rift-related volcanic activity is associated with extension-induced decompression of an anomalously hot asthenospheric mantle, citing the central South Atlantic rift zone as an example.

7. Coastal range uplift

The southeast Brazilian coastal ranges attest that the borders of EBRIS were uplifted during its geological history. This uplift was accompanied by a broad uplift of the hinterland, which now consists of a series of plateaux (Petri and Fúlfaro, 1983).

The Serra do Mar coastal range extends from 28°S to 22°S (Fig. 1); its elevation is generally of the order of 800–1000 m and reaches a maximum of 2400 m. Its tectonic history is reflected in the

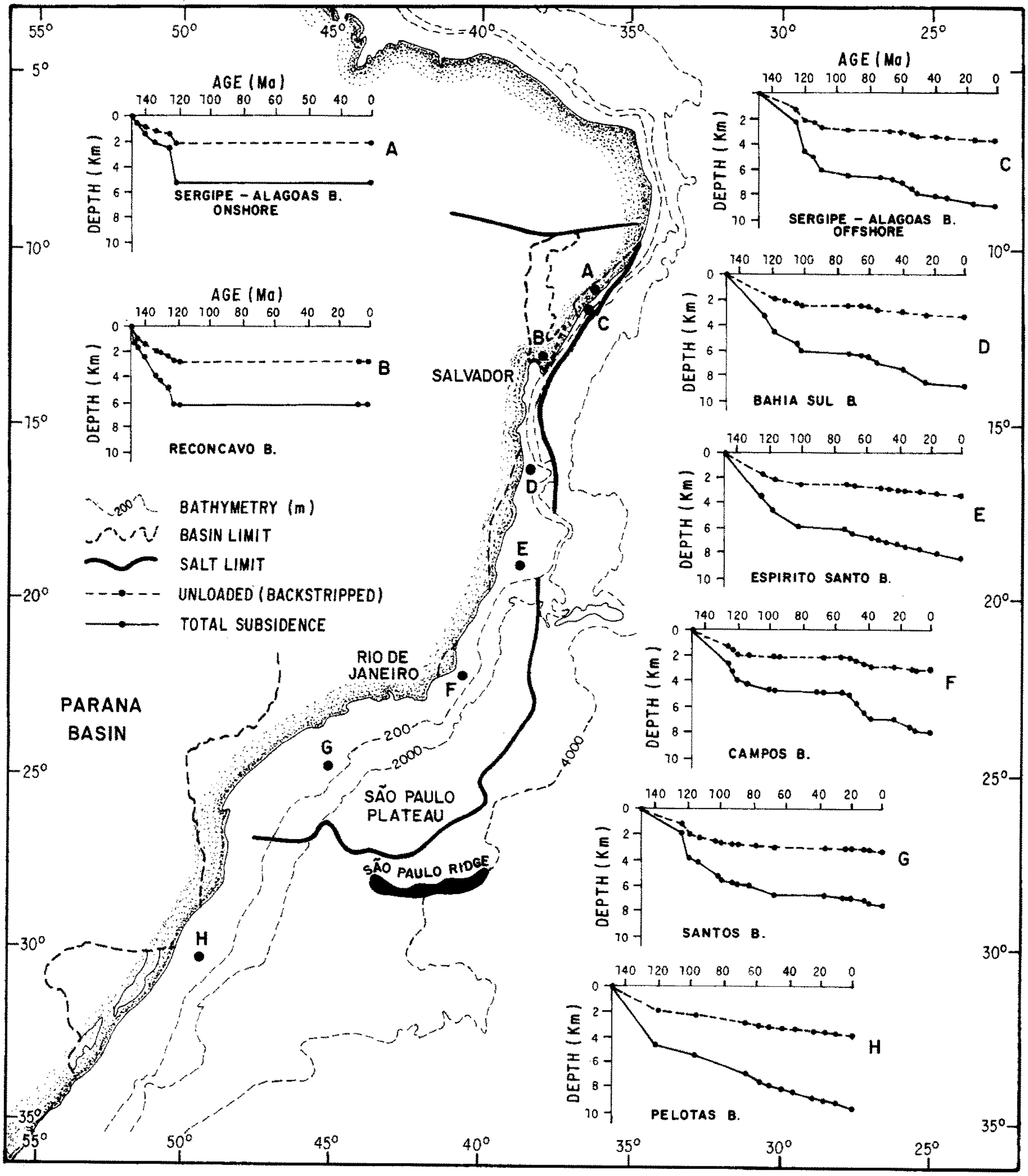

Fig. 27. Backstripped subsidence curves of selected wells from the East Brazil margin basins.

sediments of the adjacent Santos and Campos basins and from the Cenozoic-filled basins which developed between the Serra do Mar and the Serra da Mantiqueira. The latter parallels the Serra do Mar further inland and is characterized by slightly higher elevations.

In the Santos Basin, progradation of deltaic systems commenced during the Late Turonian and culminated in the Campanian when coarse alluvial-fan, fluvial and coastal sediments were deposited, reaching thickness in excess of 1000 m (Pereira et al., 1986). This depositional regime persisted until Early Eocene times. Thereafter, however, marine transgressions dominated the evolution of the Santos shelf. Bacoccoli and Aranha (1984) attribute the increased Late Cretaceous–Early Tertiary clastic supply into the Santos Basin to uplift of the now dissected Carioca Massif and the Serra do Mar, which form part of the eastern-most coast ranges. After the Early Eocene, the Serra da Mantiqueira, which is located more inland, was uplifted and its detritus was trapped in the elongated valleys between it and the Serra do Mar, thus forming the Resende and Taubate basins. These basins contain lacustrine and fluvial sediments of Oligocene and younger age (Melo et al., 1985). This Late Cretaceous–Palaeogene tectonic event was accompanied by alkaline magmatism. Mechanisms of formation of the coastal ranges were proposed by Asmus and Ferrari (1978) and Macedo (1987) who invoked a coupled isostatic model in which downwarping in the Santos Basin is compensated by the adjacent Serra do Mar uplift.

However, fission track studies on metamorphic basement rocks outcropping near Rio de Janeiro reveal that erosion of the rift margin mountains was already in progress as early as 120 Ma (Fonseca and Poupeau, 1984). This uplift was most probably induced by flexural rebound associated with the unloading of the lithosphere during rifting (Weissel and Karner, 1989). The superimposed uplift of the hinterland is possibly the consequence of the underplating of partially melted magma to the base of the crust in the region adjacent to the rift (G. Karner, pers. commun., 1991). The Late Cretaceous–Palaeogene input of coarse-grained siliciclastics in the Santos Basin, which was previously used to date the tectonic uplift, is, in this model, interpreted as a consequence of augmented rates of erosion of the pre-existing high relief composed of less resistant rocks (Azevedo, 1991). These high rates of denudation were possibly due also to increased rainfall associated with the progressive widening of the South Atlantic Ocean (Parrish and Curtis, 1982).

8. Crustal framework

Deep refraction seismic profiles are only available for the Pelotas and Santos basins (Leyden et al., 1971; Kowsmann et al., 1977) whereas deep reflection profiles have been recorded across the Campos Basin (Mohriak et al., 1990). Extensive gravity modelling was carried out in an effort to investigate the deep structure of the Brazilian passive margin basins (Guimarães et al., 1982; Chang and Kowsmann, 1984, 1986; Szatmari et al., 1985; Milani, 1985; Mohriak and Dewey, 1987; Costa, 1988; Gomes and Rizzo, 1988; Mohriak et al., 1990). Backstripping of wells, calibrated by gravity data, was used systematically to map crustal thickness along EBRIS (Chang and Kowsmann, 1984, 1986; Fontana, 1987; Costa, 1988).

Typical backstripped curves of selected wells from EBRIS are given in Fig. 27. Backstripping was performed assuming local isostasy, with an initial crustal thickness of 32 km. For large loads, typical of passive margin sediment wedges and away from the hinge zone, flexural isostatic response within the basin approximates that of local isostasy. Palaeobathymetry and absolute sea-level changes were not included in the computations. The stratigraphic record of offshore wells was extrapolated to basement by means of reflection seismic information. All of the analyzed wells are located seaward of the hinge zone and most of them are situated near the depocenters of the respective basins. For the onshore Recôncavo and Sergipe–Alagoas basins one well each was analyzed.

Between the onshore and offshore wells a distinct difference can be observed. The former correspond to aborted rifting and lack a thermal subsidence phase; the respective subsidence

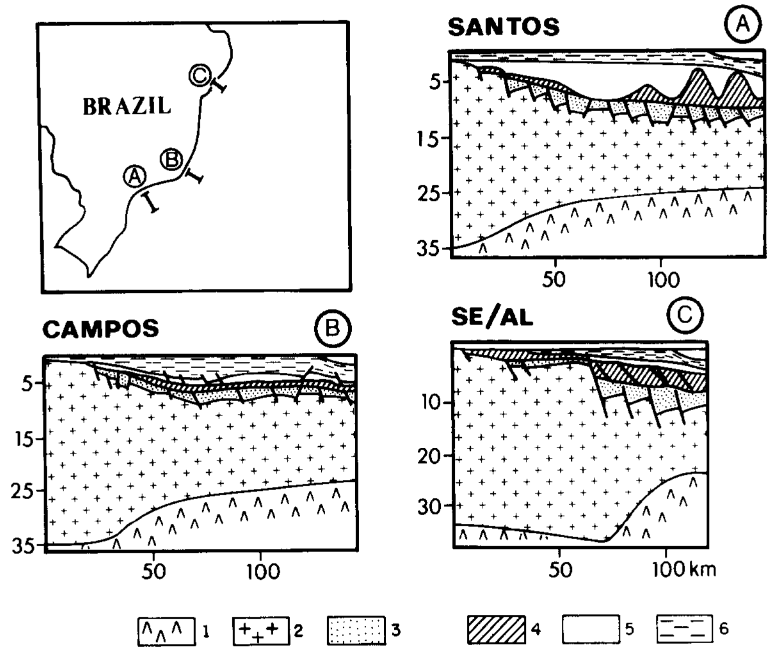

Fig. 28. Crustal structure derived from tectonic subsidence and gravity modelling (from Chang and Kowsmann, 1984, 1986; Mohriak and Dewey, 1987). *1* = Upper mantle; *2* = continental crust; *3* = rift stage continental megasequence; *4* = transitional evaporitic megasequence; *5* = shallow carbonate platform megasequence and marine transgressive sequence; *6* = marine regressive sequence.

curves do not show post-rift uplift and erosion as its magnitude is only of a few hundred meters. The offshore wells show evidence of distinct post-rift thermal subsidence, whereby the Pelotas basin is characterized by the most continuous subsidence. All other basins show a decreasing subsidence immediately after the Albian, which continued in some wells till the beginning of

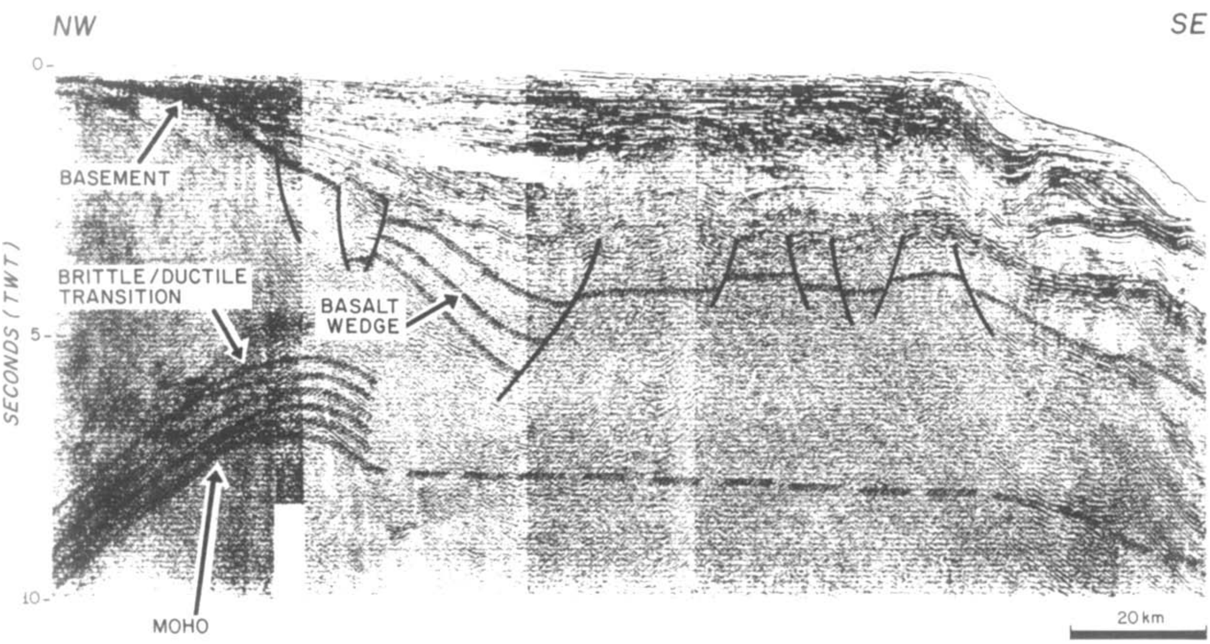

Fig. 29. Regional deep seismic reflection profile through Campos Basin, showing intracrustal and mantle reflections beneath the hinge zone. For further detail see Mohriak and Dewey (1987) and Mohriak et al. (1990).

Tertiary. This pattern is an effect of ignoring paleobathymetry.

The crustal structure of the Santos, Campos and Sergipe–Alagoas basins is illustrated in Fig. 28. These cross-sections, which are based on backstripping and gravity modelling (Campos and Sergipe–Alagoas), indicate that crustal thinning in the Campos and Santos basins is fairly smooth, whereas in the Sergipe–Alagoas Basin it is abrupt across the hinge zone. Common to all sections is the coincidence of a rapid change in Moho elevation coupled with a pronounced deepening of the basins across the hinge zone.

Recently, Ziegler (1989) suggested that the volume of the crust during rifting is not necessarily conserved, since in many basins the amount of upper crustal extension by faulting is considerably smaller than that derived from their crustal configuration. Unfortunately, geometric estimates of surficial extension could not be adequately performed in EBRIS, since basement-involved extensional faulting cannot be resolved on the available reflection seismic data. For the shallower parts of the Campos Basin, an attempt was made to determine an upper crustal extension value; results indicate a stretching factor of 1.43. This is just slightly smaller than the β-value (1.5) derived from backstripping in the same region. Therefore, if volume incompatibility is the norm, then this should occur only in the deeper, more extended areas of the basin. Jackson and White (1989) argue that for large β-values (> 1.7) later generations of superimposed faults may increase the amount of extension. In this case, in order to

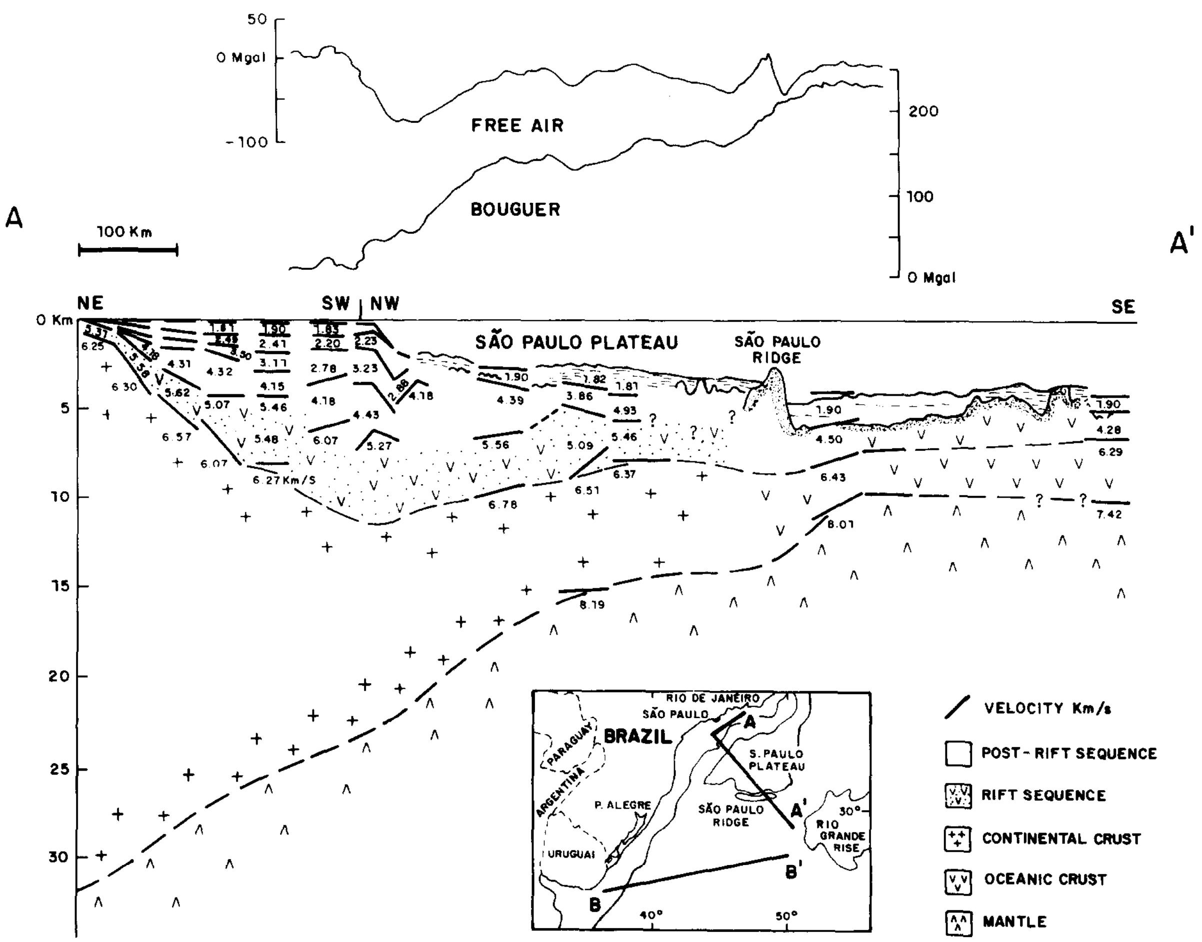

Fig. 30. Crustal structure of Santos Basin and São Paulo Plateau (modified after Leyden et al., 1971). B–B' refers to location of Fig. 31.

reach a β-value of 4.0 (representative of highly stretched crust in EBRIS), three generations of faults producing β-values of 1.6 each, would be required.

Mohriak and Dewey (1987) and Mohriak et al. (1990) presented a seismic image of deep reflectors located beneath the hinge zone of Campos Basin (Fig. 29). The shallowest, gently dipping intracrustal reflections were interpreted as basalt wedges or shear zones associated with antithetic normal faults. The laminated reflection zone above the domal structure near the NW end of the profile may be assigned to heterogeneities marking the transition between the seismically transparent brittle upper crust and the ductile lower crustal layer. The dome-like feature was interpreted as an uplifted Moho which was later depressed by flexural loading in the area of the basin depocenter. The deepest reflections within the domal apex occur at a depth of about 20 km; this is more or less compatible with the estimates of Moho depth, based on gravity modelling and backstripping of wells.

The crustal structure beneath the continental slope and rise is only controlled by refraction lines in the area of the Santos and Pelotas basins (Leyden et al., 1971; Kowsmann et al., 1977).

One refraction line parallels the continental shelf of the Santos Basin and crosses the São Paulo Plateau and Ridge (Fig. 30). Leyden et al. (1971) interpreted the refractions with velocities of ± 4 km/s as salt, ± 5 km/s as sediment interbedded with Serra Geral lavas (Paraná Basin basalts) and those with velocities greater than 6 km/s as continental crust. On the Plateau, the Moho was assumed to be overlain by thickened

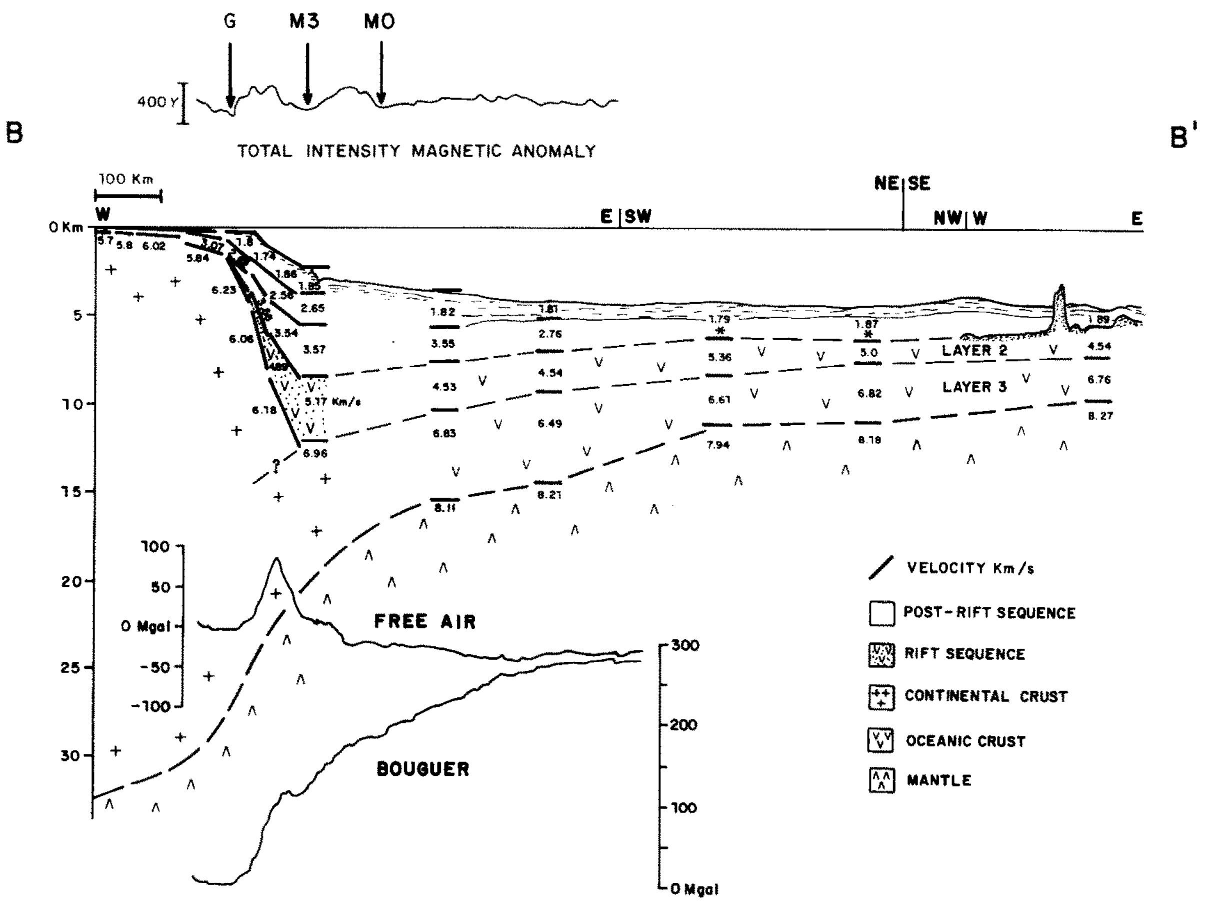

Fig. 31. Crustal structure of Uruguay interbasinal high and Pelotas Basin (modified after Leyden et al., 1971). Mesozoic magnetic anomalies after Rabinowitz and LaBrecque (1979). Location in Fig. 30.

oceanic layers 2 and 3 and evaporites. Oceanic crust thins considerably seaward of the São Paulo Ridge.

A re-interpretation of the profile indicates that rift stage lavas and sediments are represented by the the 5 km/s range layer on the shelf (Chang and Kowsmann, 1984; Pereira et al., 1986). The São Paulo Plateau is probably not upheld by thickened oceanic crust, but is underlain by extended continental crust as indicated by gravity modeling (Guimarães et al., 1982). Parallel reflections, very similar in acoustic character to the rift stage sedimentary sequence imaged in Campos Basin (Lagoa Feia Formation), have been recorded beneath the São Paulo Plateau evaporites (Mascle and Renard, 1976; Lobo and Ferradaes, 1983; Van der Ven, 1983). Backstripping of Plateau sediments reveals a total tectonic subsidence in the order of 4.5 km, which is incompatible with a 100 Ma old normal oceanic crust.

Seismic refraction surveys in the Pelotas Basin, to the south of São Paulo Plateau (Leyden et al., 1971; Kowsmann et al., 1977) suggested that the ocean/continent boundary underlies the lower continental slope to the east of the hinge zone which separates unstretched from highly stretched continental crust. The hinge zone coincides with magnetic anomaly G of Rabinowitz and LaBrecque (1979). At the foot of the slope, the hinge zone with its seaward dipping basement, gives way to a configuration of slowly seaward rising basement refractors, whereas the Moho rises seaward (Leyden et al., 1971).

Isostatic considerations, as well as seaward deepening of the top of the basement and increasing water depths, indicate that attenuated continental crust extends to the foot of the slope. Beyond this point, the basement rises gently and may be transitional between continental and oceanic; this interpretation is supported by refraction data (Kowsmann et al., 1977). Moreover, backstripped total tectonic subsidence of a point located immediately seaward of the basement dip inversion, is compatible with a 100 Ma old oceanic crust (water depth greater than 5.5 km), and thus marks the continent/ocean crustal boundary. This point lies adjacent to magnetic anomaly M3 and seaward of anomaly M4, which was considered by Austin and Uchupi (1982) as representing the oldest seafloor magnetic anomaly in the central South Atlantic (Fig. 31).

In order to investigate the geometry of the extended crust, six regional reflection seismic sections were assembled. Along these, crustal thick-

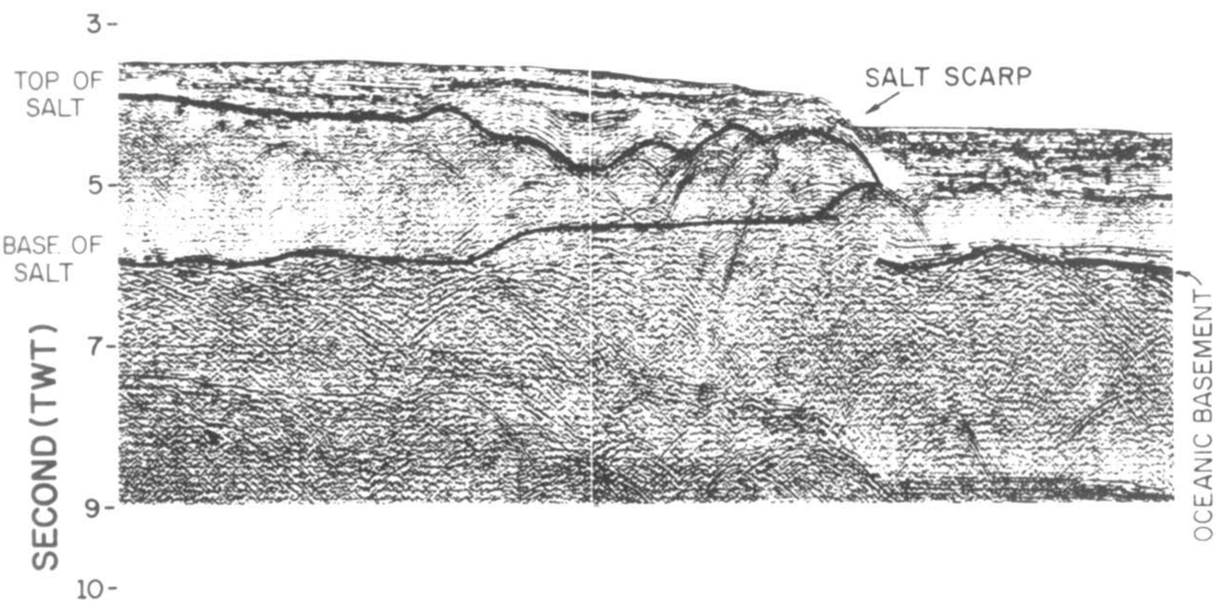

Fig. 32. Seismic section through eastern margin of the Campos Basin, showing limit of the Aptian evaporites. The salt scarp is believed to coincide with the continent/ocean boundary; oceanic crust of Early Albian age lies to the right.

ness variations were determined according to Le Pichon and Sibuet (1981) and Sawyer (1985), by relating total tectonic subsidence to crustal thinning.

Only two of these profiles crossed the ocean/continent boundary. Along the Campos Basin line, the continent/ocean boundary coincides apparently with a bathymetric scarp, limiting the salt dome province to the west from even bedded basinal sediments to the east (Fig. 32). In the Pelotas Basin, as well as in all other regional sections, the ocean/continent boundary was assumed to lie in the area where stretching values, derived from post-rift subsidence, reach values of about 5 (Sawyer, 1985). In the Pelotas Basin, this boundary coincides roughly with the abrupt disappearance of seaward dipping reflectors.

On lines which apparently failed to extend onto oceanic crust, the location of the ocean/continent boundary was estimated on the basis of bathymetry and total sediment thickness as given by Moody et al. (1979) and Kumar et al. (1979).

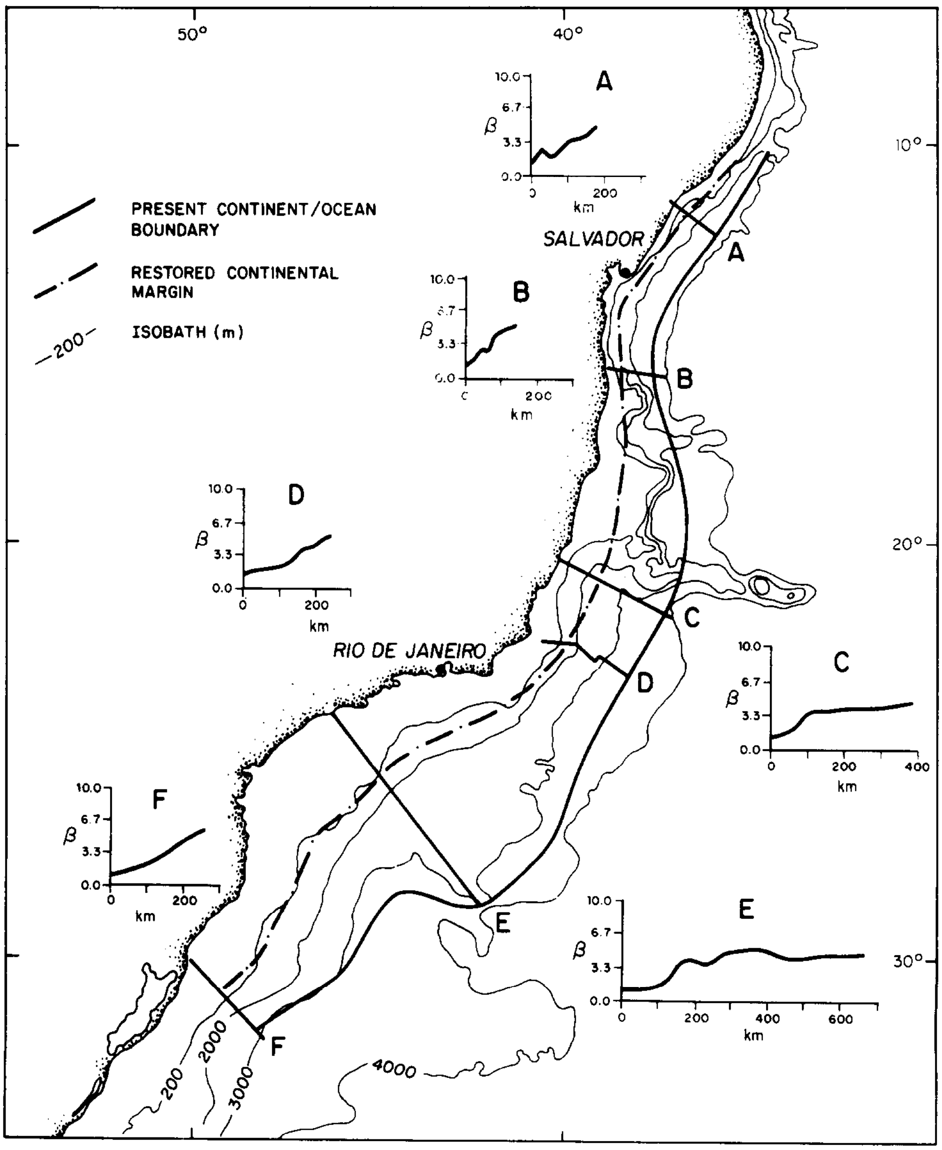

Fig. 33. Selected crustal extension (β) profiles along EBRIS and palinspastic restoration of the extended continental crust to its pre-rift position.

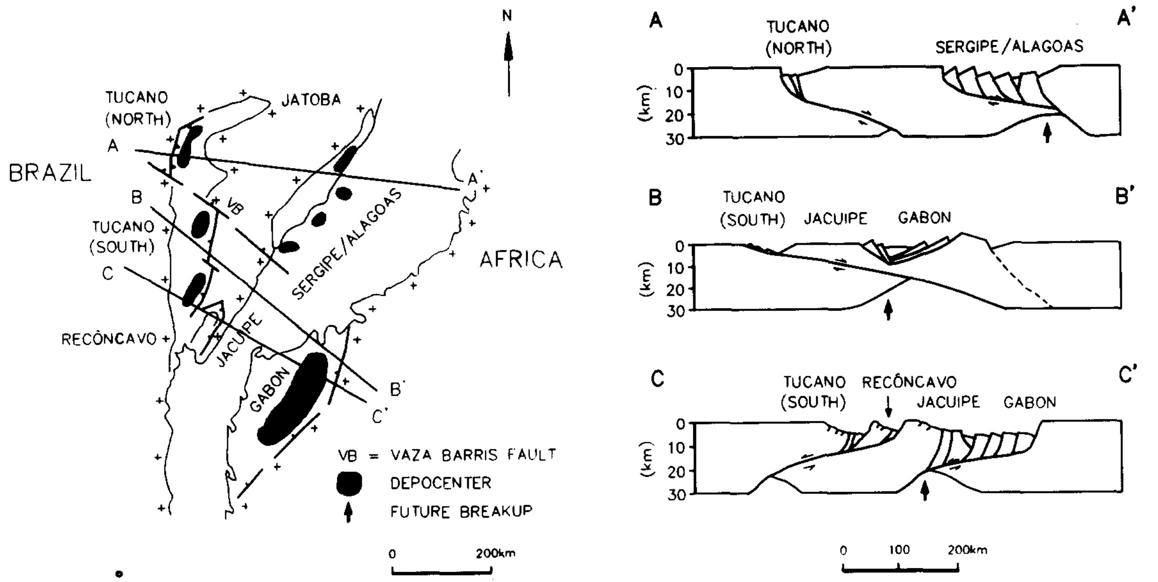

Fig. 34. Simple shear models for the northern segment of EBRIS. Sections A–A' and C–C' after Castro (1987); B–B' after Ussami et al. (1986).

Although total tectonic subsidence cannot by itself determine the ocean/continent boundary, it can give an approximate location of this zone.

Total tectonic subsidence (Sawyer, 1985), was calculated by backstripping seismic sections, assuming local isostasy (Steckler and Watts, 1978). Wells, located mainly in the shallower parts of the basins provided stratigraphic calibration of the seismic lines and also control on porosity and density of the sediments. Despite this, basement depth estimates for the deeper parts of these basins are subject to larger uncertainties, whereby the ratio of syn-rift to post-rift tectonic subsidence of 0.81 (0.45/0.55) provides some constraints. This ratio assumes the applicability of the McKenzie (1978) uniform extension model, an initial crustal thickness of 34 km, a temperature of 1330°C for the base of the lithosphere at the depth of 125 km and a mantle density of 3.3 g/cm^3. In deeper waters (continental slope and rise), the present bathymetry was equally distributed in the rift and post-rift sections, for the purpose of calculating the total tectonic subsidence. The existence of significant waterdepth in the rift is not unreasonable in this area, since it is situated far from clastic sources.

TABLE 1

Present day and restored pre-rift widths of the extended continental crust along six profiles located in Fig. 33. Restoration based on average β-extension factors listed on the right

BASIN	PRESENT WIDTH(km)	RESTORED WIDTH(km)	AVERAGE BETA
SERGIPE/ALAGOAS	173	68	2.53
BAHIA SUL	133	53	2.53
ESPÍRITO SANTO	383	137	2.80
CAMPOS	238	103	2.32
SANTOS	652	226	2.88
PELOTAS	256	119	2.16

In spite of the large uncertainties about the thus determined stretching values, the resultant crustal sections are more or less in agreement with crustal thicknesses observed on the refraction profiles crossing, for instance, the São Paulo Plateau where the crust has been thinned to about 7 km (Fig. 33).

Average β-values (Table 1) obtained from these regional profiles allowed a preliminary reconstruction of the pre-rift position of the eastern margin of EBRIS (Fig. 33). It is interesting to note that the restored position is located between the 200 and 2000 m isobath and thus coincides closely with the reconstruction of Bullard et al. (1965), which assumed the 500 fathom (913 m) isobath as the continent/ocean boundary.

9. EBRIS geodynamic model

The first geodynamic models, for the rifted origin of EBRIS were proposed by Asmus and Porto (1980), who were inspired by the models of Sleep (1971) and Milanovsky (1972). EBRIS was subdivided in two rifting domains, the Campos and Santos basins characterized by major domal uplift of the rift flanks and the inter-domal Bahia Sul and Sergipe–Alagoas basins. Domed areas, characterized by intensive rift volcanism were thought to be related to discrete mantle upwellings. Inter-domal areas, characterized by continuous subsidence since Late Jurassic, are probably not associated with similar mantle disturbances. This concept suggests that domed parts of EBRIS involved active rifting, whereas inter-domal segments were governed by passive rifting.

The conceptual power of McKenzie's (1978) pure shear model of lithospheric extension proved very useful in explaining the development of EBRIS. Applications of the model to the Santos, Sergipe–Alagoas (Chang and Kowsmann, 1984, 1986), Campos (Mohriak and Dewey, 1987; Mohriak et al., 1990), Pelotas (Fontana, 1987) and Espirito Santo basins (Costa, 1988) leads to the conclusion that uniform and non-uniform stretching of the lithosphere, together with two-stage basin subsidence (lithospheric stretching, followed by thermal contraction), amplified by flexural loading of sediments, could account for the evolution of EBRIS.

The simple shear model of Wernicke (1985) and its adaptation to continental margin settings (Lister et al., 1986) was applied to the East Brazilian continental margin by Ussami et al. (1986) and Castro (1987). Ussami et al. (1986) proposed that the onshore Tucano Basin, the offshore Jacuípe High (south of Sergipe–Alagoas Basin) and the Gabon Basin were formed by linked lithospheric extension. Upper crustal extension affected both onshore and offshore basins, but extension at deeper lithospheric levels was concentrated on the area of the offshore basins whereby a low angle, east-dipping detachment surface linked the zones of upper crustal extension (Fig. 34). In this way, the Tucano aborted rift was located on the lower plate, whilst the offshore Jacuípe and Gabon basins rode on the upper plate. The asthenospheric rise and the future zone of crustal separation was located between Jacuípe and Gabon basins. This model proposes that the intra-crustal detachment surface represents a reactivated basal thrust plane of the Brasiliano (Pan-African) Sergipano fold belt.

For the same region, Castro (1987) proposed a model that differs from the interpretation of Ussami et al. (1986) in so far as it invokes a double detachment in an attempt to explain the polarity reversal between Tucano Sul/Recôncavo–Jacuípe–Gabon basins, and the Tucano Norte–Sergipe/Alagoas–Oriental Gabon basins to the north (Fig. 34). The accommodation zone between the two rift segments of opposing polarities, coincides with the Vaza–Barris fault system. In a third profile crossing the South Atlantic rift zone further to the south, a polarity opposite to the one proposed by Ussami et al. (1986) was assumed. According to the interpretation of Castro (1987), the Gabon Basin is located on the lower plate whilst in Ussami's, it lies on the upper (Fig. 34). Furthermore, the Jacuípe High is regarded as a positive feature sandwiched between the west-dipping double detachment surfaces.

Davison (1988), in a comment to Castro's (1987) paper, argued against the involvement of crustal detachments during basin forming processes. His arguments centered on (1) incompatibility of the crustal structure derived from gravity data with such models, (2) the discordance between the Precambrian basement structure and the orientation of the proposed detachments and (3) the balance between measured amount of basement extension and regional extension derived from the crustal configuration. Obviously additional deep seismic reflection lines are needed to test the proposed models and to arrive at a well constrained interpretation.

On a much broader scale, Etheridge et al. (1989) determined the sense of detachment surfaces controlling the development of rifted basins in the South Atlantic domain. An apparent switch in asymmetry between the African and South American margins is suggested across the São Paulo–Walvis Ridge. South of this ridge, plateau uplift and post-rift volcanism, characteristic of an upper plate setting, prevail on the South African side whereas to its north, plateau uplift and post-rift volcanism occurs on the Brazilian side. The presence of antithetic normal faults in the rift section of the Campos Basin is taken as added proof for its upper plate setting.

This picture is over-simplistic and does not comply with the existing data. The Pelotas Basin, which according to Etheridge et al. (1989) should have a lower plate setting, is characterized by the most convincing example of antithetic normal faulting in EBRIS and contains abundant volcanic rocks. On the other hand, in NE EBRIS (e.g., Sergipe–Alagoas Basin), where this model predicts intense volcanic activity is almost devoid of it.

Structural styles along EBRIS (Fig. 18) tend to vary from basin to basin, with the sense of the rift-stage faulting changing even within one basin (e.g., Tucano Basin). In the Campos Basin, the frequency of synthetic and antithetic faults is balanced, giving rise to a typical horst and graben structural style. In Espirito Santo and Sergipe–Alagoas basins, synthetic faults dominate. Thus a complex pattern of alternating upper and lower plate segments would emerge from such a preliminary synthesis of the EBRIS data.

One of the most puzzling observations in EBRIS is, that the syn-rift/post-rift transition does not seem to be synchronous with the emplacement of the first oceanic crust (Chang et al., 1988). For instance, in the Campos Basin, almost all basement-involved faults are truncated by the earliest Aptian unconformity (Dias et al., 1988), whereas in northernmost EBRIS (Alagoas region), rift faults extend well into the Aptian (Lana, 1985). On the other hand, seafloor spreading commenced in Central South Atlantic at the transition from the Aptian to the Albian, as indicated by oceanic basement seaward from the present limit of the Aptian evaporite basins, which were split upon opening of the South Atlantic ocean into the conjugate São Paulo and Angola plateaux (Rabinowitz and LaBrecque, 1979). Variations in timing of the end of the rifting stage in the South Atlantic domain encompasses most of the Aptian and span about 5 Ma. This could perhaps be explained by extensional strain being distributed during the Neocomian and Barremian across the entire EBRIS and concentrating during the Aptian to the zone of crustal separation corresponding to the marginal plateaux. In northernmost EBRIS, where the basins lie much closer to the continent/ocean boundary, extension is seen to continue until the onset of seafloor spreading. This model of progressive focusing of the extensional strain was proposed by Ziegler (1988) for the Norwegian–Greenland sea and by Martinez and Cochran (1988) for the northern Red Sea; evolution of the latter is now at the transitional stage between rifting and seafloor spreading, similar to EBRIS during Aptian/Early Albian time (see Favre and Stampfli, 1992).

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