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A new scheme for the opening of the South Atlantic Ocean and the dissection of an Aptian salt basin

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SUMMARY

We present a revised model for the opening of the South Atlantic Ocean founded on a remapping of the continent–ocean boundaries and Aptian salt basins, the chronology of magmatic activity in and around the ocean basin and on the timing and character of associated intraplate deformation in Africa and South America. The new plate tectonic model is internally consistent and consistent with globally balanced plate motion solutions. The model includes realistic scenarios for intraplate deformation, pre-drift extension and seafloor spreading. Within the model, Aptian salt basins preserved in the South American (Brazilian) and African (Angola, Congo, Gabon) continental shelves are reunited in their original positions as parts of a single syn-rift basin in near subtropical latitudes (10°S–27°S). The basin was dissected at around 112 Ma (Aptian–Albian boundary) when the model suggests that seafloor spreading commenced north of the Walvis Ridge–Rio Grande Rise.

Key words: Gravity anomalies and Earth structure; Plate motions; Marine magnetics and palaeomagnetics; Continental margins: divergent; Sedimentary basin processes; Oceanic hotspots and intraplate volcanism.

1 INTRODUCTION

The rather good geometrical fit between the Atlantic margins of South America and Africa (Fig. 1) was first recognized by Wegener (1915) and dictates a relatively simple rift-to-drift history for the South Atlantic bordering continents incorporating south-to-north opening of the ocean. Rifting began in the south in the late Jurassic, reaching the equatorial rift zone by the late Cretaceous (Rabinowitz & LaBrecque 1979; Emery & Uchupi 1984). Despite the overall simplicity of the opening, the early stages of continental rifting and pre-drift positions of major tectonic blocks remain uncertain. In the literature, the locations and magnitudes of gaps and overlaps between reconstructed continental blocks differ greatly (Vink 1982; Unternehr *et al.* 1988; Nürnberg & Müller 1991; Jackson *et al.* 2000; Eagles 2007). These differences and their implications for intraplate deformation, crustal thinning, basin formation and salt deposition have fuelled lively debate. The fact that early South Atlantic ocean crust was generated during the Cretaceous normal polarity superchron means that early relative motion between South America and Africa must be based on interpolation between positions fixed by magnetochrons C34 and M0 (Fig. 2) and on the geometries of fracture zones (Section 5.3).

Petroleum companies, who have undertaken intensive seismic surveys mostly using standard multichannel seismic techniques, have extensively studied the sedimentary basins on the South Atlantic margins. Although these studies provide substantial knowledge of the upper sedimentary cover, the presence of vast evaporites, igneous flows and intrusive sheets makes the imaging of deeper basin structures difficult or impossible, leaving the early evolution of the rifted margins unclear. In an attempt to shed more light on this we (1) re-examined the positions of the continent–ocean boundaries (COBs) using potential field and seismic data, (2) compiled information on the timing and character of magmatic activity manifested as large igneous provinces (LIPs) and Seaward Dipping Reflectors (SDRs) and (3) constructed regionally and globally balanced plate reconstructions. Our new regional plate tectonic framework better fits the geological evidence on the breakup history of the region.

For this plate tectonic framework, the question of whether the salt accumulations preserved in the Brazilian and Angola–Congo–Gabon margins are pre- or post-breakup is crucial: if salt deposition was pre-breakup (i.e. deposited before initiation of seafloor spreading), then the pre-salt sediment infill was also pre-breakup, and the substratum must therefore have been continental or subcontinental (e.g. thinned continental crust intruded by mantle material immediately prior to seafloor spreading, Moulin *et al.* 2005). Conversely, if the salt was deposited after breakup, then the salt was probably at least partially deposited on a non-continental substratum, and the continuity of the salt basins between South America and Africa

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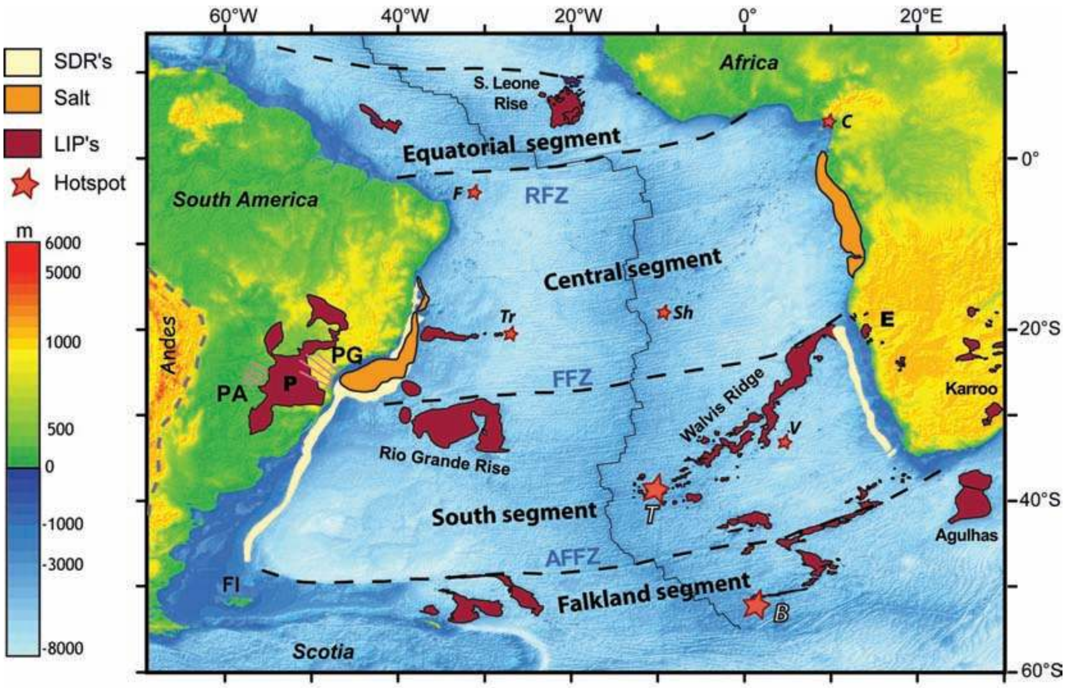

Figure 1. General structural map of the South Atlantic Ocean draped on topographic/bathymetric map from GTOPO 30. Boundaries between the four segments (Equatorial, Central, South and Falkland) are shown by dotted lines (RFZ, Romanche Fracture Zone; FFZ, Florianopolis Fracture Zone; AFFZ, Agulhas–Falkland Fracture Zone). Aptian salt basins (orange), LIPs (P, Parana; E, Etendeka; Karroo, Sierra Leone Rise and Agulhas), Seaward Dipping Reflectors (SDRs, white) and active hotspots (F, Fernando; C, Cameroon; Tr, Trinidade; Sh, St Helena; T, Tristan; V, Vema; B, Bouvet (Meteor) are also shown. Of these hotspots, only Tristan (responsible for the Parana-Etendeka LIP and Rio Grande Rise–Walvis Ridge) is classically considered as a deep plume in the literature (see Torsvik *et al.* 2006). However, Bouvet (Meteor) possibly responsible for Agulhas, and Maud Rise (East Antarctica) and Madagascar Ridge volcanism could possibly also have a deep plume origin (Section 6). PG, Ponta Grossa Dyke System; PA, Paraguay Dyke system; FI, Falkland Islands.

would be in question (Karner *et al.* 1997). Critical constraining factors include the locations of the COBs, the ages of the salt basins, the shapes and extents of the salt basins, amounts of pre- and post-drift extension in the margins and plate reconstruction parameters. We will argue here that the Aptian salt accumulations belonged to a single large pre-breakup (syn-rift) basin confined to continental and/or subaerial basaltic substrates.

2 GENERAL SETTING AND MAGMATISM

The South Atlantic Ocean (Fig. 1) can be divided into four segments (Moulin *et al.* 2005): (1) the *equatorial* segment between ∼10°N and the Romanche Fracture Zone (RFZ), (2) the *central* segment between the RFZ and Florianopolis Fracture Zone (FFZ) hosting the Aptian salt accumulations (Section 3), (3) the ’southern’ segment from the FFZ to the Agulhas–Falkland Fracture Zone (AFFZ) hosting well-documented SDRs and (4) the ’Falkland’ segment south of the AFFZ. The detailed evolution in the equatorial (Pindell *et al*. 1988; Roest *et al.* 1992) and Falkland (Adie 1952; Mitchell *et al.* 1986; Marshall 1994; Dalziel & Lawver 2001; Torsvik *et al*. 2008a) segments is beyond the scope of this paper.

The regional sedimentation histories between the late Jurassic and the present are broadly similar on both margins. Sedimentation proceeded from the deposition of continental pre- and syn-rift facies, through syn-rift deep lacustrine, turbiditic and transgressive shales, to shallow marine carbonate and evaporitic units before the deposition of post-rift marine transgressive (carbonate) and regressive (siliclastic) sequences (Katz & Mello 2000) (Fig. 2).

In contrast, the distribution of Cretaceous magmatic products is asymmetrical with respect to the two rift margins (Fig. 1). The major magmatic event or LIP of interest is the Paraná (Brazil–South America)–Etendeka (Namibia-Africa) province (P–E province), widely linked to the hotspot presently responsible for melt generation beneath the island of Tristan da Cunha (O’Connor & Duncan 1990; Renne *et al.* 1992; Turner *et al.* 1994). The voluminous early Cretaceous continental flood basalts of the P–E province are distinguished by their rapid eruption, mostly between 133 and 130 Ma (Fig. 2; cf. appendix 1 in Torsvik *et al.* 2004). With an estimated original volume of $1.5–2.0 \times 10^6$ km^3, the Paraná rocks of Brazil, Paraguay and Uruguay cover an area of $\sim 1.2–1.6 \times 10^6$ km^2, whereas the Etendeka lavas on the African side cover an area of $\sim 0.8 \times 10^5$ km^2, and with their affiliated lava are scattered along more than 1500 km of the West African coast (Piccirillo *et al.* 1990; Renne *et al.* 1992; Peate 1997).

Dyke swarms associated with the P–E province indicate that the lavas originally covered a greater area than what is preserved today and have two orientations: NW–SE (present orientation) and coast-parallel. NW–SE-oriented dolerite dykes from Brazil, Angola, Eastern Paraguay and Namibia suggest failed rifts or inversions (Jackson *et al.* 2000). Offshore, a large package of basalt flows and sills have been interpreted from seismic data and/or drilled on the South American margin in the Jacuipe, Espírito Santo, Campos, Santos and Pelotas basins (Fig. 3b), and they continue southwards to the Falkland escarpment and offshore South Africa and Namibia (Hinz 1981; Fodor *et al.* 1984; Mizusaki *et al.* 1988; Chang *et al.* 1992; Davison 1997; Bauer *et al.* 2000). Isolated reflectors in offshore Angola and Cameroon have also been interpreted as volcanic rocks or SDRs (Jackson *et al.* 2000; Mohriak *et al.* 2000), but this latter

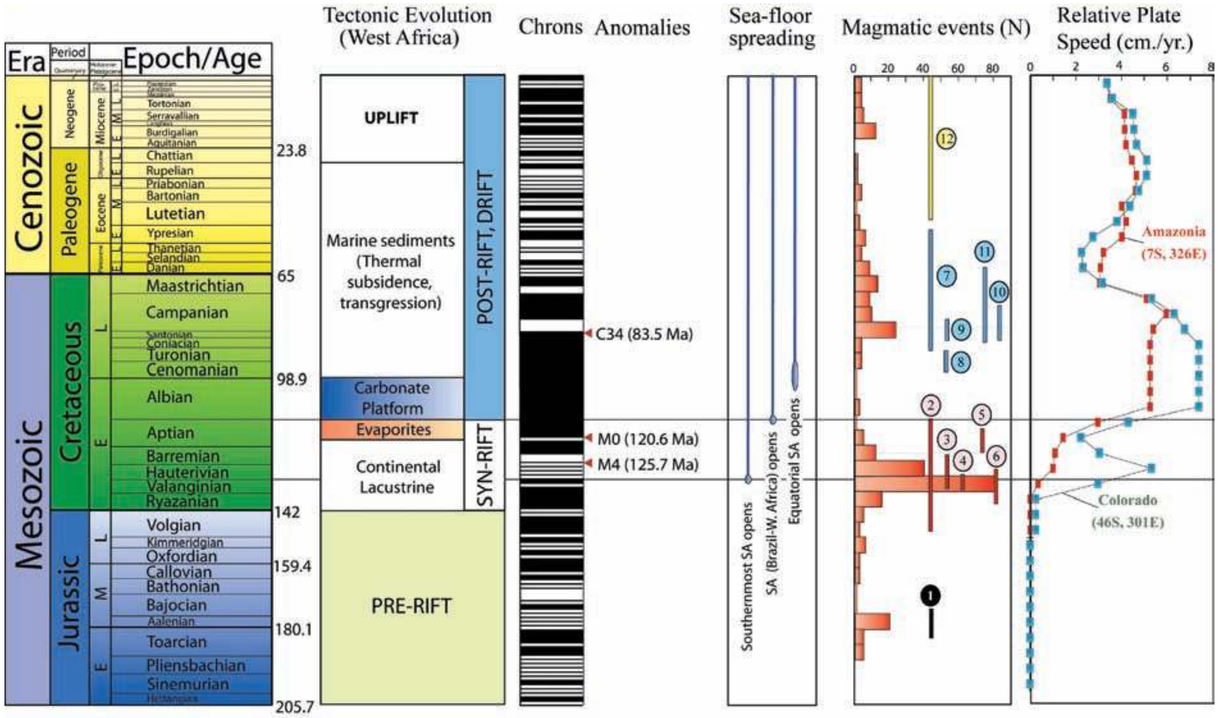

Figure 2. Simplified Jurassic to present chronology of sedimentary/rift, seafloor spreading, magmatic events and full spreading/extension rates between South America and South Africa, calculated for two geographic location in the Central (Amazonia) and South segments (Colorado). Timescale after Gradstein *et al.* (1994), sedimentary/rifting events modified from Marton *et al.* (2000), magnetic anomalies adapted from Cande & Kent (1995) and Chanell *et al.* (1995), M sequence. Numbered magmatic events are (1) Karroo and Patagonia, (2) duration of Parana–Etendeka (P–E) magmatic activity, (3) early P–E dykes, (4) primary phase of P–E CFB bimodal volcanism, (5) late P–E dykes, (6) alkaline/peralkaline magmatism, mobile belts, (7) duration of alkaline/peralkaline magmatism (onshore, near-shore and oceanic), (8) dykes (Benue), (9) dykes (Benue, Brazil), (10) subalkaline, mafic magmatism (Benue, Campos, St Helena, Rio Grande Rise), (11) primary pulse of alkaline/peralkaline magmatism and (12) oceanic and hotspot magmatism (see Torsvik *et al.* 2004 for details).

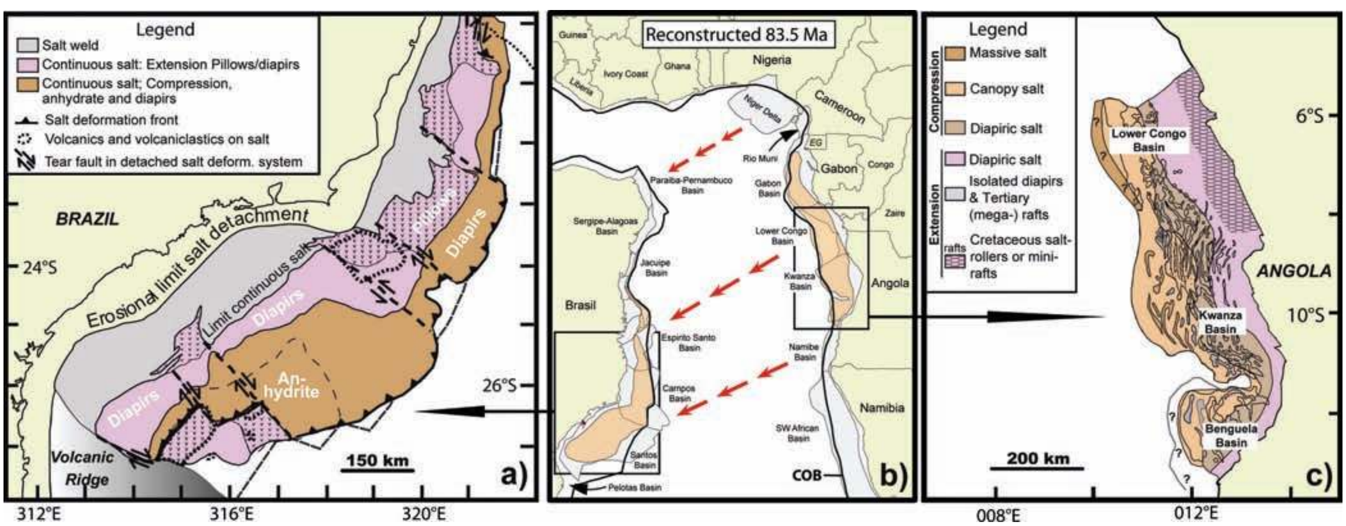

Figure 3. Simplified map of salt-detached structural domains for Campos and Santos basins (a) (after Meisling *et al.* 2001) and offshore Angola (c) (after Marton *et al.* 2000). (b) Distribution of important sedimentary basins along the Brazilian and African margins, COB location (this study, Section 4) and interpreted extent of salt basins on both margins (this study). All features from the Brazilian margin are reconstructed to 83.5 Ma (C34) by keeping Africa fixed (Table 1). The red arrows show that Santos, Campos, Espirito Santos basins (Brazilian margin) were once conjugate basins to Namibe, Kwanza, lower Congo and the Niger Delta before break up. Note that the current shape and extent of some of these basins extend onto oceanic crust. The location of (a, c) are shown as inset boxes in **(b)**.

assessment has not been confirmed by recent seismic interpretations (Contrucci *et al.* 2004; Moulin *et al.* 2005). Onshore exposures of the Kwanza and Namibia basins contain tholeiitic and alkaline volcanic flows that are reported to underlie early Cretaceous sediments and/or to be interlayered with late Cretaceous deposits (Marzoli *et al.* 1999). Excluding the Karroo (Jurassic) and Patagonia rocks and clearly oceanic rocks of late Paleogene and younger ages, the onshore and offshore sedimentary basin magmatism from Central and South segment ranges in age between 147 and 49 Myr, with two dominant concentrations between ∼143–121 Myr and ∼90–60 Myr (Fig. 2; Torsvik *et al.* 2004).

3 THE APTIAN SALT BASINS

3.1 Description

The most significant Aptian salt deposits are located along the Eastern Brazilian margin in the Santos, Campos and Espírito Santo basins (Fig. 3a), and on the African margin between Angola and the Niger Delta (Fig. 3b). The fully marine Albian carbonate systems of both Brazil (Santos Basin) and West Africa (Congo and Kwanza basins) overlie the Aptian salt sequences (Fig. 2). Along the Brazilian margin, the thickest portions of assumed syn-rift Aptian evaporite sequences occur in the Santos Basin. Evaporites rest on a pre-Aptian sequence largely composed of volcanics (basalt flows and tuffs) that are mostly slightly younger than the Paraná Basin continental flood basalts (∼133–130 Ma). The oldest known magmatic rock in the SE-Brazil sedimentary basins is a 138 Ma continental flood basalt, which erupted subaerially onto attenuated crust in the Santos Basin (Fodor *et al.* 1983, 1984). Campos Basin basalts range from 124 to 112 Myr in age and are mostly continental flood basalts (Fodor *et al.* 1984). Early Cretaceous (130–123 Ma) basalts from the onshore Espírito Santo Basin are intercalated with terrigenous sedimentary rocks.

Three main tectono-stratigraphic sequences bounded by major unconformities are identified along the South American (Brazilian) margin, that is, rift, transitional and drift (Mohriak *et al.* 1995). The Neocomian–Barremian rift sequence (Lagoa Feia Formation in the Campos Basin and Guaratiba Formation in the Santos Basin) includes lacustrine sediments (Abrahão & Warme 1990) overlying lower Cretaceous tholeiitic basalts (Mizusaki *et al.* 1988). Landward-dipping normal faults that cut basement, predominate in the proximal less-thinned continental crust, but, in contrast, from shelf to basin (towards the rift depocentre) the faults change polarity and rift blocks are rotated landwards by predominantly seaward-dipping faults with large offsets (Mohriak *et al.* 1995). A major unconformity separates the rotated blocks (filled by Barremian lacustrine sediments) from overlying Aptian conglomerates. The Aptian transitional phase (Lagoa Feia Formation in Campos Basin and Ariri Formation in Santos Basin) is characterized by evaporites deposited in an elongated gulf that extended from the southern Santos to the Northern Sergipe–Alagoas Basin. The salt basin, bounded to the west by the Aptian hinge line (Fig. 3a), is as wide as 300 km and is mostly underlain by rifted continental crust (Chang *et al.* 1992). The western limit of the salt diapir province is closer to the Aptian hinge line in the northern Santos Basin than that in the central Campos Basin. Geophysical data from the Brazilian margin indicate that the area of salt was not significantly increased by post-depositional ocean-ward salt flow during late Cretaceous and Cenozoic times (Szatmari 2000) and thus the seaward limit of the salt basin approximately coincides with the COB (Mohriak *et al.* 1995; Karner 2000; Torsvik *et al.* 2004). The drift phase started in the Albian with a transitional to shallow marine platform that was rapidly flooded in the Cenomanian. The Albian–Cenomanian stratigraphy is characterized by shallow-water limestone in the Campos Basin, grading vertically and distally into calcilutites and marls, whereas the Santos Basin is marked by a major siliciclastic influx (Mohriak *et al.* 1995).

The depositional environment for the South American Aptian evaporites remains the subject of controversy and is mostly inferred from comparison with studies from the African margin (see below; Mohriak *et al.* 1995; Karner 2000; Szatmari 2000). Nevertheless, it is difficult to reconcile a deep-water origin for the evaporites with the fact that, wherever drilled in Brazil, the sediments below the salt are continental lacustrine and those above are typically shallow-water limestones or coarse-grained siliclastics (Mohriak *et al.* 1995).

Along the African margin (Fig. 3c), evaporites are reported from north to south in the Rio Muni, Gabon, Lower Congo and Kwanza Basins (Hudec & Jackson 2006), which were once conjugate basins to the Sergipe Alagoas, Jacuipe, Espírito Santo and Campos basins on the Brazilian margin (Fig. 3b). Even if there are scarce occurrences of basaltic lavas related to the Etendeka province (Seranne & Anka 2005), recent seismic studies (Contrucci *et al.* 2004) show that the crustal structure of the Angolan margin and the <10 km thick oceanic crust display no evidence of thick transitional igneous crust and contradict the presence of SDRs postulated by Jackson *et al.* (2000). Pre-salt sediments on the continental shelf have mostly been identified from industrial wells. Fluvio-lacustrine sediments characterize the lower Carboniferous to Triassic-Jurassic sequence, followed by an episode of increased tectonic activity in Neocomian to mid-Barremian times, which led to the deposition of clastics and clay. This episode is punctuated by the 'Pointe noire' unconformity (Tessereinc & Villemin 1990). The pre-salt wedge is of Barremian to mid-Aptian age. The deposition of these units is characterized by a low tectonic activity without major block faulting (Marton *et al.* 2000; Moulin *et al.* 2005) and is related to the formation of offshore basins such as parts of the Gabon Basin (Tessereinc & Villemin 1990). The lacustrine sedimentation was followed by the deposition of a thin layer of marine sediments known as the Chela Layer (Moulin *et al.* 2005; Nunn & Harris 2007). Most authors agree that the salt was deposited during Aptian time in about 5 Myr (Moulin *et al.* 2005 and references therein). The post-salt history (Séranne *et al.* 1992; Hudec & Jackson 2004, Dupré *et al.* 2007) shows a change in the nature of sedimentation from carbonate deposition to siliciclastic progradation at the base of the Oligocene. This change in sedimentation has been related to uplift of the Southern Africa Platform and resulted in significant erosion and an increase in sedimentary loading (Lucazeau *et al.* 2003; Moulin *et al.* 2005; Dupré *et al.* 2007).

North of Gabon, the oldest drilled portion of the stratigraphic section is a lacustrine sandstone, the Barremian to Aptian aged N'Toum Formation. Overlying this unit are lacustrine facies of lower Aptian age. The transitional sequence is represented both by the highly variable middle Aptian Como Formation, which includes a series of carbonaceous shales, and the evaporitic Ezanga Formation (Katz *et al.* 2000). In the Congo and Gabon regions, salt is most abundant in the compressional belt, due to remobilization of the salt down slope, where it occurs as thick autochthonous salt, salt walls, diapirs, salt tongues and salt canopies (Fig. 3c). In the Rio Muni Basin, the pre-salt sequence also includes Upper Barremian to mid-Aptian terrestrial clastics and lacustrine shales characterized by extensional rollover structures with large-scale listric faults updip, and toe-thrust structures down-dip. That section is overlain by a sequence of well-developed salt with good quality intercalated oil-prone source rocks. An Albian carbonate platform developed over the area, followed by a Cenomanian to Turonian sand-shale sequence (Dupré *et al.* 2007).

To the south, the Kwanza Basin (offshore Angola) has been divided into the inner and outer Kwanza salt basins, separated by a chain of basement horsts on which the salt is thin or absent (Marton *et al.* 2000; Hudec & Jackson 2004). This agrees with the comprehensive stratigraphic description of the Angola passive margin by

Marton *et al.* (2000) (plate 4), and complementary data from recent multichannel seismic reflection and OBS wide-angle seismic data in Moulin *et al.* (2005). The thickest salt body was deposited in the deepest parts of the flooded basin, which are likely to have overlain the most extended crust. The original salt appears to decrease in thickness northwards, implying that the southern basin initially subsided more, enabling more salt accumulation. The salt tectonics of the Kwanza Basin is illustrated in Fig. 3c. Most authors agree that, despite a complicated post-depositional salt tectonic history, seaward allochthonous salt never exceeded the initial limits of the salts basins on both the African and Brazilian margins by more than 50 km (Marton *et al.* 2000; Hudec & Jackson 2004; Hudec & Jackson 2006). Estimates for the initial width of the African salt basins range from about 150 km in the South Gabon Basin in the north (Dupré *et al.* 2007) to more than 200 km in the Angola Basin to the south (Moulin *et al.* 2005). In the lower Congo and Kwanza Basins of Africa, the initial evaporite thickness reached more than 1 km (Jackson *et al.* 2000).

3.2 Depositional environment and origin of the salt basins

It is widely accepted that the salt was deposited in a shallow-water environment, and palaeontological evidence from the Aptian to earliest Albian dolomite and sapropel sequence of the Angolan margin, deposited just above the Aptian salt, suggests no more than 500 m paleo-water depth (Marton *et al.* 2000). The absence of thick marine layers prior to the Aptian (the marine La Chela layer found on both margins is very thin), and of any erosion surfaces contemporaneous with salt deposition also imply no active marine sedimentation and very shallow depositional environments (Moulin *et al.* 2005).

There is controversy as to whether the Brazilian and Gabon–Angola salt was deposited in the post-rift drift succession (Jackson *et al.* 2000; Tari *et al.* 2003, Dupré *et al.* 2007), or was part of the syn-rift sequence (Karner *et al.* 2003; Hudec & Jackson 2004; Moulin *et al.* 2005) (Fig. 2). The continuity of the salt from the South American to African margins is also debated. The controversy partly stems from an *apparently* large overlap of salt on the two margins when the South Atlantic is closed and the margins juxtaposed (Jackson *et al.* 2000). The early Aptian overlap has been estimated to be as much as 200 km, clearly too much to have been caused by post-breakup (post-Aptian) continental stretching, or allochthonous mobilization of salt. This apparent overlap (see Section 5.3) has led to models suggesting that the Aptian salt basins in the South Atlantic did not form in a single basin (Jackson *et al.* 2000), but accumulated independently on both margins during and after breakup of the South Atlantic. However, according to Moulin *et al.* (2005) and based on recent seismic interpretations, the salt cover is continuous from the western boundary of the basin to the unthinned continental platform, thus refuting the hypothesis of two different salt formations (Karner *et al.* 1997). Geochemical differences between salt on the platform and salt in the deep basin (Jackson *et al*. 2000) can then be explained by petrological differences in the underlying substrate.

To reconcile shallow-water depositional environment and continuity of the salt with a post-rift scenario, a spectacular hypothesis was proposed by Burke & Sengör (1988); Szatmari (2000) and Burke *et al.* (2003) in which it was postulated that the great evaporite sequences resulted from spills of oceanic waters over the young Rio Grande Rise–Walvis Ridge, which (combined with a period of sea-level drop) created a barrier preventing the invasion of the northern deepening post-breakup basins by the sea. They propose that pre-Aptian subaerial seafloor spreading had already occurred north of the proto-Walvis Ridge and the Ponta Grossa dyke swarm (∼130 Ma), and that most of the salt was deposited in pre-existing rift depressions, reaching more than 2 km below sea level. It was also suggested that the area that eventually became the Aptian salt basin was continuously dammed, subsided rapidly and experienced subaerial magmatic spreading, perhaps similar to that ongoing today in the Afar depression. Burke *et al.* (2003) proposed that later catastrophic flooding by oceanic water occurred over the topographic barrier represented by the proto-Walvis Ridge simultaneous with a global sea-level drop (Haq *et al.* 1987). As well as sounding implausible, the hypothesis is not corroborated; first, by the lack of active subsidence inferred from the apparent continuity of salt from the un-thinned continental domain until the deep basin (Moulin *et al.* 2005) and, secondly, by recent sea-level reconstructions that predict a maximum in sea level around at 118 Ma followed by a gradual decrease in sea level (Müller *et al.* 2008).

On the other hand, Dupré *et al.* (2007) found that subsidence trends derived from backstripping calculations from the South Gabon Basin could correlate with a syn-rift origin for the evaporites. Nunn & Harris (2007) also underlined that the thick underlying Barremian-lower Aptian section from both margins north of the Rio Grande Rise–Walvis Ridge contained an extensive record of deposition from saline lakes, with a complete absence of any marine signature from the Neocomian to the middle Aptian. These authors proposed a model of subsurface seepage of sea water across a barrier to explain this apparent geological dilemma, which also would constitute a possible additional mechanism for the syn-rift deposition of the great evaporite sequence.

4 REEVALUATION OF COB LOCATIONS IN THE SOUTH ATLANTIC MARGINS

The position of the COB is commonly deduced from gradients in bathymetry, gravity and magnetic fields and seismic investigations. SDRs in seismic data are subaerially extruded basalts normally located landwards of the oldest ocean floor magnetic anomaly (Fig. 1). On volcanic margins, the COB is often placed within the lateral bounds of the SDR sequence, and therefore we do the same on the margins of the South Atlantic. Evaporite sequences have a seismic 'masking effect' that usually restricts analysis of the underlying structures and hinders localization of the COB. Factors, such as exhumed continental mantle at non-volcanic margins, or depth-dependent stretching at both volcanic and non-volcanic rifted margins, also make locating the COB difficult (Davis & Kusznir 2004). In reality the COB is probably better described as a continental ocean transition (COT) zone between typical continental and oceanic crust, and is usually some tens of kilometres wide. Nevertheless, despite these limitations, properly establishing the location of the COB as a line on a large-scale map brings valuable insights into plate-reconstruction-derived estimates of pre-drift extension/lithospheric stretching (Sections 5 and 6).

4.1 Data processing

We have produced new COBs for the entire South Atlantic based on gravity, magnetic, bathymetric and seismic and other geological information. The gravity data set proved particularly useful, consisting of a proprietary grid (now available publicly) of Bouguer anomaly values with a cell size of 2 min of arc (Torsvik *et al.* 2004). The Bouguer gravity field over a COB is dominated by the effects of

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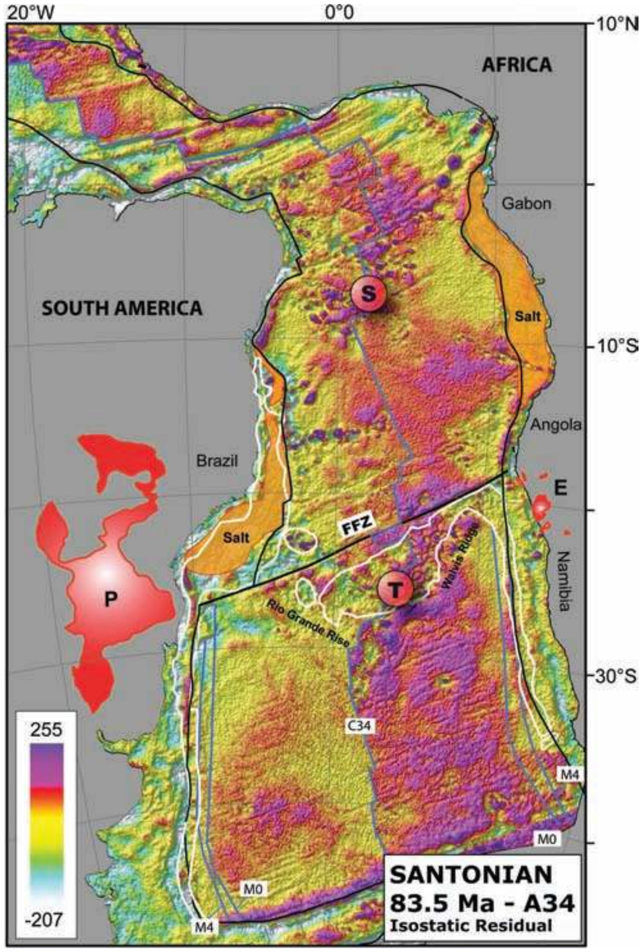

Figure 4. Shaded isostatic residual maps. New interpretation of COBs (black lines) is shown together with LIPs (P, Parana; E, Etendeka), SDRs/hotspot-related magmatism (white lines) and the outline of the salt basins from both margins. Isostatic residuals have been reconstructed to 83.5 Ma by keeping Africa fixed (Table 1). At that time the Tristian (T) and St Helena (S) hotspots were located close to the spreading axis (blue thick line, C34). We also show the location of M0 and M4 (Muller *et al.* 1997). FFZ, Florianópolis Fracture Zone. P, Parana; E, Etendeka.

deep crustal sources that isostatically support the topography. The long-wavelength gravity field associated with deep-seated compensating masses was determined and removed using GMT (Wessel & Smith 1991) to leave the isostatic residual anomaly field associated with shallower crustal density contrasts. For the calculation, the effective elastic thickness was set to 20 km, normal crust density to 2800 kg m^{-3}, mantle density to 3300 kg m^{-3}, water to 1030 kg m^{-3} and infill to 2300 kg m^{-3}.

The isostatic residual map is shown in Fig. 4, together with our COB interpretation, the interpreted extents of the salt basins, LIP and SDR outlines and magnetic anomalies M0/M4 reconstructed to C34 time (83.5 Ma).

In this relative reconstruction Africa is kept in its present position. A wide waterway connected the South Atlantic and Central Atlantic Ocean basins and the Tristan and St Helena hotspots were erupting volcanics near the spreading axis. The African plate moved over the Tristan hotspot shortly after this time resulting in the formation of the Walvis Ridge. Fig. 4 shows the clear asymmetry in spreading the north of the FFZ: more oceanic crust having been generated on the African plate than on the South American plate. To the south of the FFZ, spreading appears to have been more symmetric. The reconstructed isostatic residuals show some interesting differences between the African and South American plates. Offshore northern Brazil, the residual gravity field over the African plate is significantly higher than over the adjoining South American plate. We interpret this to indicate differing sediment thickness from one area to another and differences in dynamic support of the lithosphere from below.

The gravity data were further processed to accentuate the different signatures of continental and oceanic crust by first upward continuing the isostatic residual anomalies 20 km and then calculating the second vertical derivative (Fig. 5) and gradient amplitude.

4.2 COB on the West African margin

We have positioned the COB along the landward side of clearly defined gradients in the residual gravity field (Fig. 4, see also Fig. 5a). In the Namibia region (Fig. 4), the COB is placed within the narrow band of mapped SDRs (Gladczenko *et al.* 1997), in accordance with data from an integrated geophysical study (Bauer *et al.* 2000). The residual gradient probably results from lateral density contrasts in both the mantle (large Moho relief) and the crust (basalts versus continental crust). Along the Namibian region, 'breakup related' rifts vary in width from 75 to 150 km (Gladczenko *et al.* 1997; Bauer *et al.* 2000). Offshore Angola, the interpreted COB position coincides with the position implied by recent seismic experiments (Contrucci *et al.* 2004; Moulin *et al.* 2005; Dupré *et al.* 2007), and also with our knowledge of the extent of the salt basin. Breakup related rifts range in width from 160 to 180 km offshore Angola (Moulin *et al.* 2005), and are approximately 120 km wide on the Gabon margin (Watts & Stewart 1998).

4.3 COB on the South American margin

We have positioned the COB along the landward side of the defined residual gradients (Figs 4, 5a) south of the Rio Grande Rise. Regional multichannel seismic profiles show igneous activity during rifting and breakup on more than 2400 km of the Argentina, Uruguay and Brazil margins (Gladczenko *et al.* 1997). In particular, wedges of SDRs are present along the entire margin south of the Santos Basin, and voluminous extrusive units have been mapped in and along the salt basins up to ∼20°S (Mohriak *et al.* 1990; Chang *et al.* 1992).

Isostatic residual maps in concert with bathymetric maps, seismics and oceanic fracture zones proved very helpful in a more reliable determination of COBs. However, tracing the COB is not always straightforward, and the Brazilian margin and the area immediately north of the Rio Grande Rise are examples of this. When reaching the Santos Basin, the near-shore belt of positive anomalies (on the regional map of Bouguer-corrected gravity) has been interpreted as a broad Moho uplift in the footwall of Neocomian extensional faults, on the basis of a review of existing geophysical and geological data by Meisling *et al.* (2001). The belt of positive anomalies further offshore correlates with a pre-salt ridge of eroded volcanic or basement anticlines covered by thin evaporites interpreted as a possible failed spreading centre.

Along the Campos Basin, the COB remains reasonably well defined by features in the gravity data, whilst in the SE and S margins of the Santos Basin the position of the COB is unclear. To help clarify these uncertainties we rotated the interpreted African COB into its likely pre-drift position against the South American margin

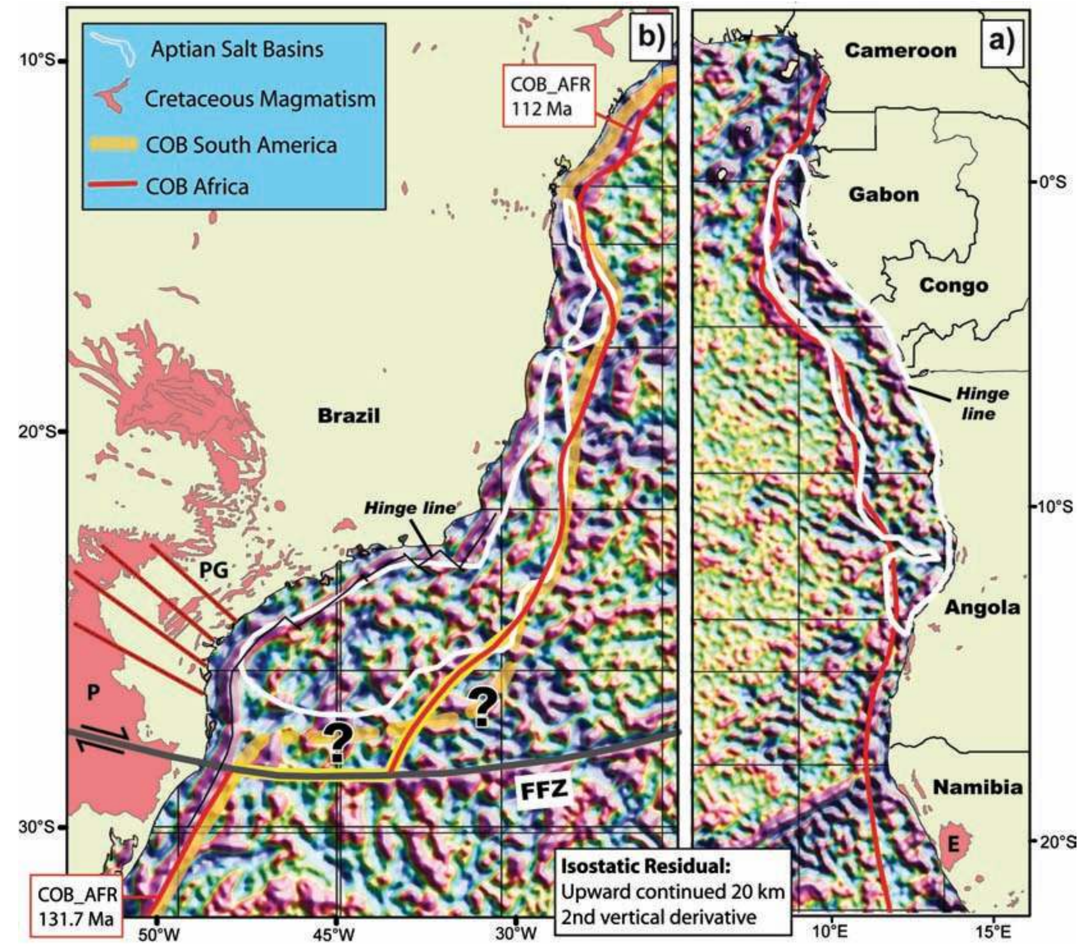

Figure 5. Second vertical derivative of upward continued isostatic residual anomalies along with new COB interpretations offshore Namibia to Cameroon (a) and Brazil (b). We also show salt outlines (white lines) and details in the faulted hinge-line (thin black line) on parts of the Brazilian margin. Rotated COBs from the African margin (red lines) are also shown (112 Ma north of the FFZ; 131.7 south of the FFZ). COB interpretations along the Brazilian margin are thick orange transparent lines. The definition of the COB in the SW Santos basin is complex: originally it was interpreted to follow the stippled orange line marked with question marks but subsequently simplified to match the conjugate African margin with a breakup age of 112 Myr. In Northern Santos and northwards, and south of the FFZ, there is good match between the two conjugate COBs at 112 and 131.7 Ma, respectively (see the text). FFZ, Florianópolis Fracture Zone; P, Parana; E, Etendeka; PG, Ponta Grossa Dyke System.

(Fig. 5b). The Africa and South America COBs south of the FFZ match well at ∼130 Ma, around the estimated breakup time immediately south of the FFZ. The South American Campos region COB aligns well with the African COB at 112 Ma, which we consider the rift-to-drift time north of the FFZ.

Our interpretation of the COB on the Brazilian margin is more outboard than most published models (e.g. Müller *et al.* 1997; McDonald *et al.* 2003), lying up to 400 km more ocean wards. In the Campos Basin, many previous COB interpretations require that salt must have been transported several hundred kilometres onto oceanic crust (Torsvik *et al.* 2004). The most recent COB solution of Müller *et al.* (2008) reduces the implied width of allochthonous salt to 175 km (Fig. 6), but this is still unrealistically high.

5 NEW PLATE RECONSTRUCTIONS AND INTRAPLATE DEFORMATION

Over recent years it has been recognized that to obtain realistic pre-drift fits between South America and Africa one or both continents must have undergone internal deformation, probably focused along a small number of lineaments (Unternehr *et al.* 1988). These internal relative motions are required to eliminate unrealistic continental overlaps in plate reconstructions, implying huge pre-drift extension ($\beta = 4$ in Fig. 7b), and equally unrealistic gaps between the margins, implying pre-drift compression. Large overlaps and gaps of this kind are implied by the rigid plate fits of Bullard *et al.* (1965) and Vink *et al.* (1982) as illustrated in Figs 7a and b, respectively. We therefore proceed to present our new model where we attempt to identify the most significant intraplate boundaries and assign motion histories to each block enclosed by them. The rotation models (Table 1) attempt to account for the shapes and fits of the conjugate margins, geometries of fracture zones, magnetochrons and geological evidence. In our models we use the Aptian M-sequence timescale of Channel *et al.* (1995). The magnetic anomaly M0 (base Aptian) is set to 120.6 Ma (Fig. 2) and not to 125 Ma as in the most recent timescale of Gradstein *et al.* (2004). We do not use this timescale since it inevitably leads to an age-shift of the M sequence back to the end of the Jurassic (see also Gee & Kent 2007; He *et al.* 2008).

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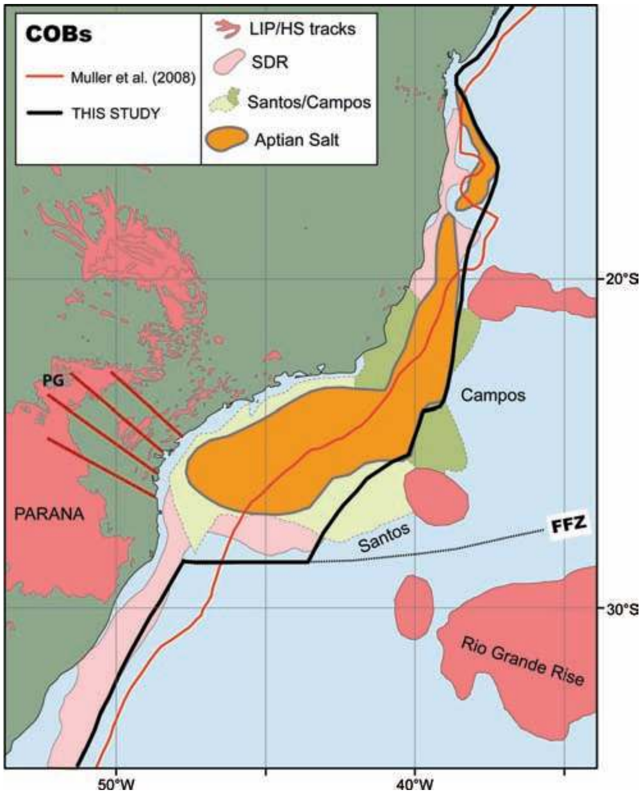

Figure 6. Comparison of our COB (black line) offshore Brazil with that of Müller *et al.* (2008; red line) along with LIPs, SDRs and salt outlines. Note that our COB interpretation in SE Santos is dictated by the conjugate COB in the Angola region (Namibe Basin) using a breakup time at 112 Ma (see Fig. 5b). FFZ, Florianópolis Fracture Zone; PG, Ponta Grossa Dyke System.

5.1 Intraplate boundaries

Nürnberg & Müller (1991) formulated a model for the opening of the South Atlantic in which South America was divided into four domains where the boundaries (Pan-African belts) could have acted as shear zones during the early opening of the South Atlantic. These boundaries could be fitted to a small circle (pole: 47°N, 327°E), which is quite similar to the pole of the younger, nearby oceanic fracture zones. We largely follow the Nürnberg & Müller (1991) model, but since we have revised the COBs and the intraplate boundaries (Figs 8 and 9), we have also altered their original Euler poles, but essentially in rotation angles and timing rather than pole position. We divide South America into four blocks, which we denote by Amazonia, Paraná, Colorado and Patagonia (Appendix A; Fig. 8). The two most import boundaries are between Patagonia and Colorado, where we have invoked 500 km dextral movement between 180 and 160 Ma (along the AFFZ), and between Amazonia and Paraná (along the inland continuation of what we denote by the Paraná-Etendeka Fault Zone; PEFZ) that in our model defines a transtensional boundary with a total lateral offset of 175 km (Fig. 10). In our South America model, Patagonia moved with respect to Colorado between 190 and 160 Ma whilst the Falkland Islands (FI) were being detached and rotated away from South Africa. Colorado/Patagonia moved with respect to Paraná until ∼132 Ma, and dextral movements between Paraná and Amazonia ceased at around 126 Ma: after which time South America became a single rigid plate.

We have divided Africa into five tectonic domains: South Africa, NE Africa, NW Africa, Somalia and Lake Victoria Block (Appendix A; Fig. 9). However, the latter two blocks were firmly attached to the South Africa domain for the period discussed here and are not detailed further. We use the NW Africa versus South Africa fits (dominantly extension between ∼120 and 84 Ma) of Nürnberg & Müller (1991), whilst the NW Africa–South Africa fits (Table 1) were generated by juxtaposing the Guinea and Demerara plateaus, aligning Archean and Pan-African mega shear zones and orogens [Eburnean-Transamazonian and West Congo-Araçuaí (Brazil) orogens; Bozoum-délé-Ngaoundéré-Pernambuco and Benoue Through-Patos lineaments; and Hoggar-Dahomeyide belts with the Maranhao Basin; Popoff, 1988; Ledru *et al*. 1994], and ensuring only modest pre-rift overlaps (50–130 km; Fig. 10) with NE Africa and South Africa so that all the intra-African boundaries are characterized by Cretaceous extension. We invoke an early rift phase in the NW Africa/Benoué trough area between 131.7 and 120 Ma to accommodate extension north of the PEFZ during that time (Milner *et al.* 1995; Renne *et al.* 1996; Raposo *et al.* 1998; Ernesto *et al.* 1999). COB overlaps across Campos/Santos and the conjugate Kwanza/Namibia basins are typically in the order of 350 km, amounting to a β-factor of ∼2.1 across the Santos–Namibia basins (see also Section 6).

Our closure fits for the South Atlantic show only a moderate COB overlap, implying moderate pre-drift extension, south of the PEFZ (maximum ∼100 km; Figs 10 and 12a). We emphasize that this is a schematic model constructed in such a way that the COBs in the southernmost South Atlantic (i.e. south Colorado block immediately north of the AFFZ) match at ≥150 Ma—if we include a substantially larger overlap here to reflect more late Jurassic/early Cretaceous pre-drift extension, this will increase the amount of strike-slip along the PEFZ but reduce the amount of strike-slip required along the GFS/AFFZ.

5.2 Reconstruction of the salt basins

The age of South Atlantic salt basins must be younger than M0 and the lower Aptian (Fig. 1). A realistic reconstruction of the South Atlantic salt basins should therefore use mid-Aptian (116 Ma) or younger Euler poles. Fig. 11 shows different reconstruction schemes for the salt basins, the first two are based on rotations derived from traditional linear interpolation between M0 (120.6 Ma) and C34 (83.5 Ma) and imply no salt basin overlap in the period (a) middle Aptian (116 Ma) to (b) early Albian (110 Ma) times. Arguments for a post-breakup origin for the salt basins due to substantial Aptian overlap between the salt basins (Jackson *et al.* 2000, see their Fig. 14) can therefore be rejected. A syn-rift depositional environment is favoured from all available evidence. To reflect initial breakup and seafloor spreading north of the Rio Grande Rise–Walvis Ridge/FFZ around the late Aptian–early Albian boundary (112 Ma), we juxtapose the salt basins at that time in Fig. 11(d) (Euler pole latitude = 52.4°N, longitude = 325°E and angle 51.3°—denoted as the AA pole). Reconstructions for the South Atlantic during the Cretaceous Normal Superchron (CNS) are therefore interpolated between M0 (120.6 Ma) and the position implied by AA (112 Ma) and then between AA and C34 (83.5 Ma). This leads to lower pre-break extension rates but somewhat higher seafloor spreading rates during the CNS (Fig. 11c; ∼5.5 cm yr^{-1} versus 4.75 cm yr^{-1} with constant interpolation between M0 and C34). In our new model, a subordinate salt basin overlap of ∼75 km is indicated at 116 Ma but before the basin was fully developed, and little or no overlap in late Aptian (Fig. 11e) and early Albian (Fig. 11f) times, when carbonate platform sedimentation began.

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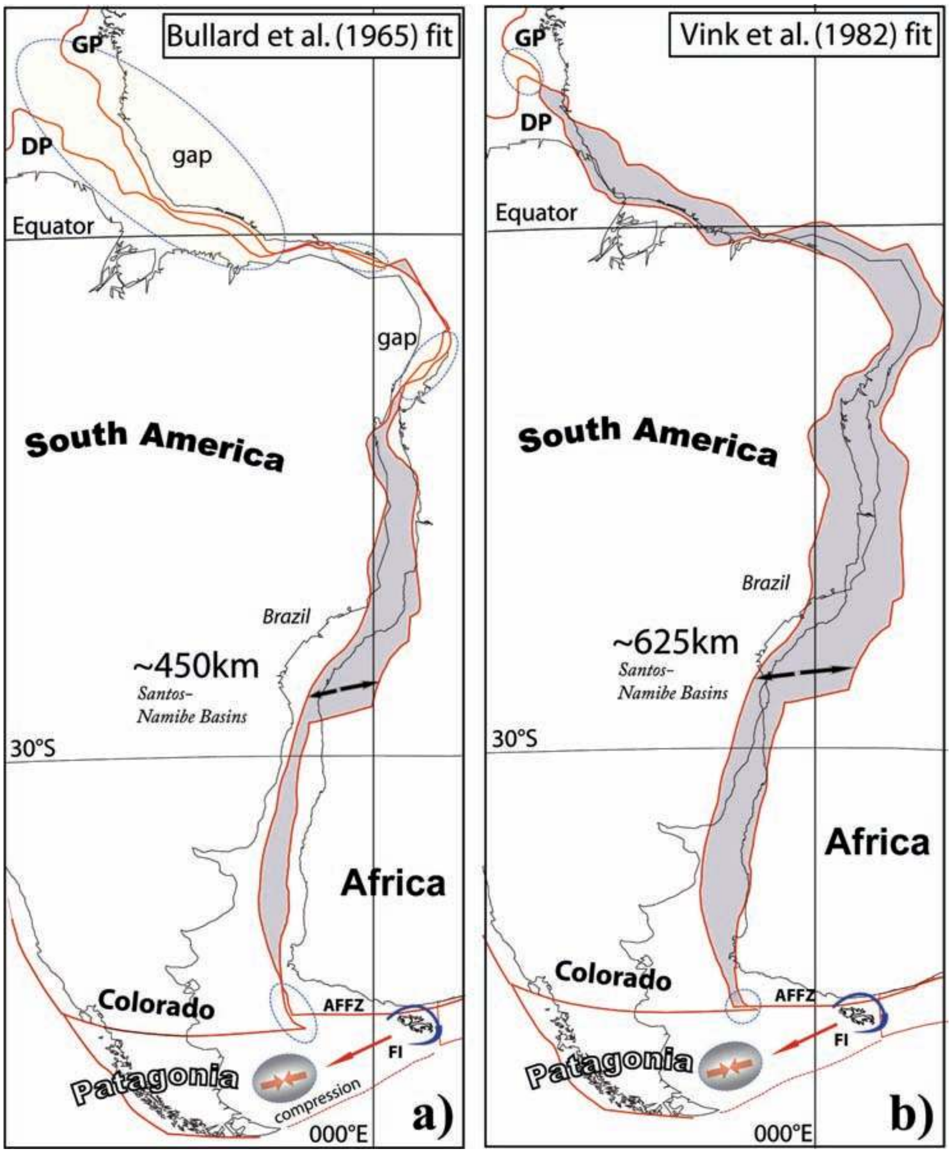

Figure 7. Examples of treating South America and Africa as two rigid blocks based on closure fits of (a) Bullard *et al.* 1965: latitude = 44°N, longitude = 329.3°E and angle = 57°; (b) Vink (1982): 48°N, 327°E and 58°. We show the continents outlined by our own COB interpretations. This figure illustrates some of the earlier noted problems of treating South America and Africa as rigid plates; for example, extreme continental overlap [625 km in (b); dark grey area is overlap] and β-factors [~4.0 in (b)] along the Brazil/Angola margin (Santos/Campos and Namibe Basins), and considerable compression (100–250 km) between Guinea and Demerara Plateaux (equatorial segment and beyond) during the early opening of the South Atlantic. Many reconstructions of the South Atlantic also ignore or exclude the history of the Falkland Islands (FI). The FI must have originated off the SE coast of Africa and subsequently rotated ~180° from their current orientation in the Jurassic (Adie 1952, Mitchell *et al.* 1986; Marshall 1994; Dalziel & Lawver 2001) as required by Jurassic palaeomagnetic data, correlations between the basement and the overlying Paleozoic strata of South Africa with the stratigraphy of the FI and by the structural correlation of the eastern Cape Fold Belt with fold and thrust trends on the FI. The disconnection and rotation away from South Africa probably took place between 182 and 160 Ma (Torsvik *et al.* 2008a). Assuming that South America was a single plate therefore creates a severe problem in the interpretation of the southern segment. Unless Patagonia occupied a much more easterly position, detaching FI from South America would create a convergence and compression of 500–600 km along the Argentinean continental margin, for which there is no geological evidence.

The main salt basins accumulated on mainly extended continental crust at subtropical latitudes between 27° and 10°S when reconstructed with the global hybrid reference frame of Torsvik *et al.* (2008b). Indeed for the last 150 Myr, South America has been extremely stable in latitude and the Campos Basin has remained in the subtropics since the early Cretaceous (Fig. 11c). Our Aptian model is independent of intraplate deformation in South America, since that had ceased at around 126 Ma.

6 CONCLUSIONS AND IMPLICATIONS OF THE NEW RECONSTRUCTIONS

Robust COBs and reconstruction parameters are essential constituents for understanding the detailed geological evolution of divergent margins. Our interpreted COB positions for the Atlantic margins of South America and Africa are based on gravity and magnetic fields, bathymetry, seismic interpretations, the position

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Table 1. Relative reconstruction parameters versus a fixed South Africa (/Amazonia).

Plate A	Plate B	Time	Latitude	Longitude	Angle	Reference
Amazonia	S Africa	2.7	62.20	−39.40	0.83	Cn2a—NUVEL-1
Amazonia	S Africa	9.7	62.05	−40.59	3.18	Cn5—Müller *et al.* (1997)
Amazonia	S Africa	19.0	58.77	−37.32	7.05	Cn6—Müller *et al.* (1997)
Amazonia	S Africa	25.8	57.59	−36.27	9.96	Cn8—Müller *et al.* (1997)
Amazonia	S. Africa	33.1	56.17	−33.64	13.41	Cn13—Müller *et al.* (1997)
Amazonia	S Africa	38.4	57.10	−33.00	15.91	Cn18—Müller *et al.* (1997)
Amazonia	S Africa	46.3	56.95	−31.15	19.11	Cn21—Müller *et al.* (1997)
Amazonia	S. Africa	52.4	58.89	−31.18	21.38	Cn24—Müller *et al.* (1997)
Amazonia	S Africa	55.9	61.35	−32.21	22.27	Cn25—Müller *et al.* (1997)
Amazonia	S Africa	65.6	63.88	−33.61	24.76	Cn30—Müller *et al.* (1997)
Amazonia	S. Africa	71.6	63.41	−33.38	26.57	Cn32—Müller *et al.* (1997)
Amazonia	S Africa	79.1	62.92	−34.36	30.99	Cn33—Müller *et al.* (1997)
Amazonia	S Africa	83.5	61.88	−34.26	33.51	C34—Müller *et al.* (1997)
Amazonia	S Africa	112.0	52.40	325.00	51.30	Forced break North of PEFZ/FFZ—this study
Amazonia	S. Africa	120.6	51.60	−35.00	52.92	M0 Nürnberg & Müller (1991)
Amazonia	S Africa	125.7	50.40	−33.50	54.42	M4 Nürnberg & Müller (1991)
Amazonia	S Africa	131.7	50.00	−32.50	55.08	M11 Nürnberg & Müller (1991)
Paraná	Amazonia	125.7	0.00	0.00	0.00	This study
Paraná	S Africa	125.7	50.40	326.50	54.42	This study
Paraná	S Africa	131.7	47.50	326.70	56.00	This study
Paraná	S Africa	150.0	47.50	326.70	56.20	This study
Colorado	Amazonia	125.7	0.00	0.00	0.00	This study
Colorado	S Africa	125.7	50.40	326.50	54.42	This study
Colorado	S Africa	131.7	47.50	326.70	57.00	This study
Colorado	S Africa	150.0	47.50	326.70	57.30	This study
Patagonia	Amazonia	125.7	0.00	0.00	0.00	This study
Patagonia	S Africa	125.7	50.40	326.50	54.42	This study
Patagonia	S Africa	131.7	47.50	326.70	57.00	This study
Patagonia	S Africa	150.0	47.50	326.70	58.00	Torsvik *et al.* (2008a)
Patagonia	S Africa	160.0	47.50	326.70	58.00	Torsvik *et al.* (2008a)
Patagonia	S Africa	190.0	47.50	326.70	63.00	Torsvik *et al.* (2008a)
NW Africa	S Africa	120.0	0.00	0.00	0.00	This study
NW Africa	S Africa	131.7	33.65	26.02	2.34	This study
NE Africa	S Africa	83.5	0.00	0.00	0.00	This study
NE Africa	S Africa	120.6	40.50	298.60	−0.70	This study

of SDRs and salt basins, and are tested against different plate-fitting scenarios. From plate models and the amount of COB overlap (350 km in average along the Campos/Santos margin; Fig. 10) we can estimate plate-scale β-factors. Used in this sense, β depends on the distance between conjugate hinge lines where the crust/lithosphere thickness is taken to be normal in their pre-rift positions (COBs overlap) and their breakup positions (COBs match). The hinge-line is commonly positioned near the shelf-break and is often associated with a pronounced free air gravity anomaly (Fig. 5). As an example, reconstructing the conjugate Santos/Namibia Basins to break up (Fig. 12f) implicates a zone of ∼685 km of thinned continental crust for a central Santos basin location; given a pre-rift separation (closure fit) of ∼325 km (Figs 10 and 12; 180 and 150 Ma reconstruction), this results in a β factor of ∼2.1. It should be noted that plate-model β-factors cited here assume that the entire lithosphere has undergone stretching; such estimates are typically larger than those derived from, for example, subsidence analysis.

Fig. 12 shows our new reconstructions from ∼180 to 83.5 Ma with a focus on South America/Africa and their margins, but we also include the location of parts of East and West Antarctica on our maps (Torsvik *et al.* 2008a). Rotation parameters relative to a fixed South Africa are listed in Table 1 and the 'absolute' reconstructions and plate motion vectors (grey arrows in Fig. 12) are based on the hybrid reference frame of Torsvik *et al.* (2008b).

6.1 Jurassic reconstructions (180 and 150 Ma)

At around 180 Ma (Fig. 12a) the South American and African blocks spanned 60°S–30°N, and the FI was located offshore SE Africa, almost 180° rotated at latitudes of around 48°S, and situated next to the Ellesworth-Whitmore Mountains (today part of West Antarctica; see also Grunow *et al.* 1987; Dalziel & Lawver 2001; Torsvik *et al.* 2008a). At this time, we consider that FI had just started its convoluted translation and rotation away from South Africa and became an integrated and stable part of Patagonia before 150 Ma (Fig. 12b). Unless Patagonia occupied a more easterly position at 180 Ma (relative Colorado), the departure of the FI from South Africa would result in severe compression along the Argentinean continental margin; we therefore keep FI loosely connected with Patagonia by postulating ∼400 km dextral movement along the AFFZ/GFS between 180 and 150 Ma (compare Figs 12a and b), and ∼100 km of younger dextral movement that had essentially ended by Hauterivian times (compare Figs 12b and c).

At 180 Ma, the velocity field for south Pangea was dominated by west-directed motions (Fig. 12a), and tentatively linked to a combination of plume activity impinging the south Pangea lithosphere (Karroo LIP) and subduction rollback (Torsvik *et al.* 2008a). The velocity vectors for South America are slightly oblique to the 'Andean' subduction margin but nearly parallel to the GFS (/AFFZ),

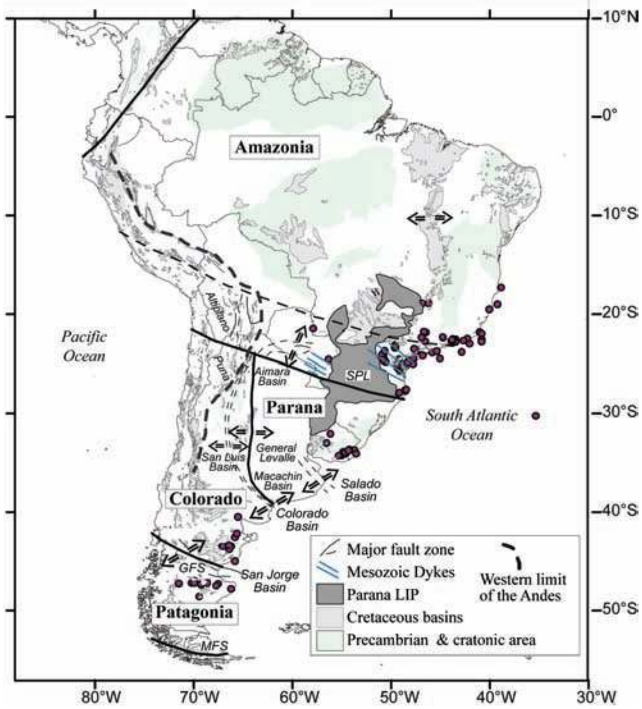

Figure 8. Simplified structural map of South America showing the Cretaceous rifts and the chosen block limits (after Ramos 2005; Franke *et al.* 2006; Monaldi *et al.* 2008). Red dots show our age-data compilation of magmatic products in South America and offshore areas. See Appendix A.

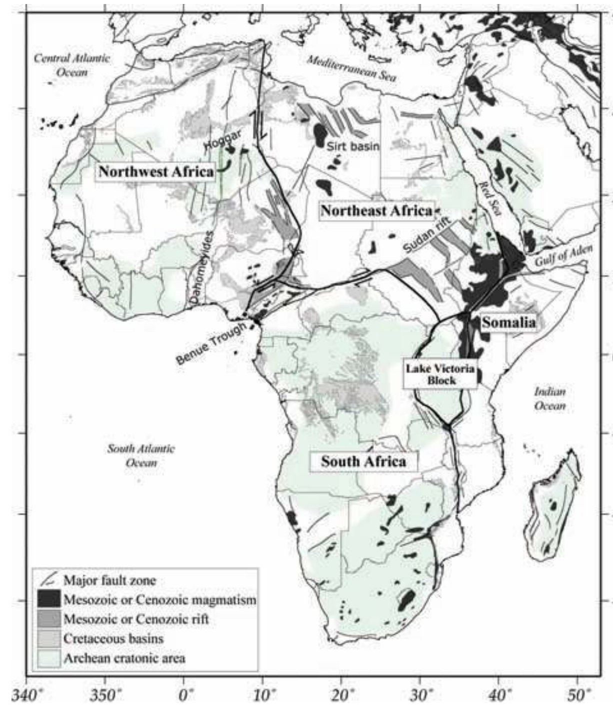

Figure 9. Simplified structural map showing block limits and Mesozoic-Cenozoic rift-related basins and magmatism associated with the breakup of Gondwana. Tectonic and geologic features follow Guiraud & Maurin (1992), the UNESCO Geological map (2000) and the USGS Africa map (2002). See Appendix A.

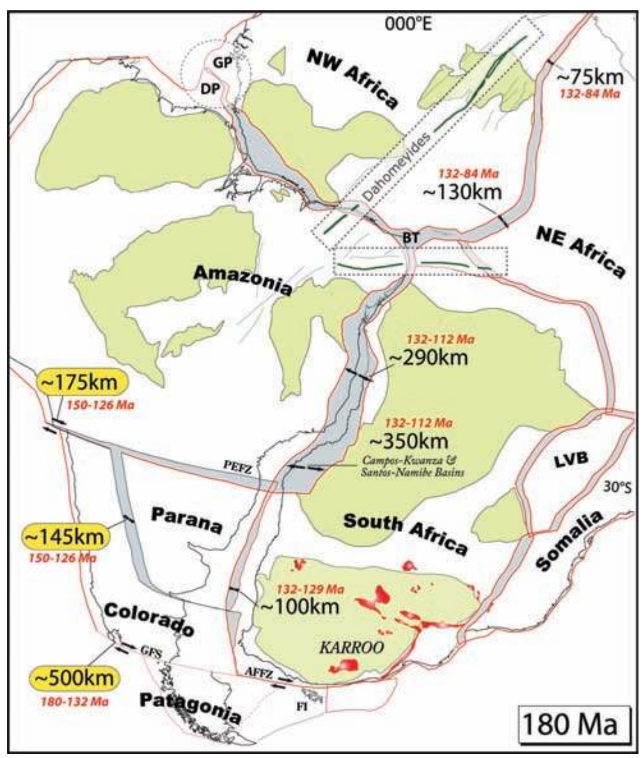

Figure 10. A pre-drift reconstruction of the South Atlantic realm at ∼180 Ma. GP, Guinea Plateau; DP, Demara Plateau; BT, Benue Trough; PEFZ, Parana–Etendeka Fracture Zone; AFFZ, Agulhas–Falkland Fracture Zone; GFS, Gastre Faults System; FI, Falkland Islands; MEB, Maurice Ewing Basin; LVB, Lake Victoria Block. Ages in red denote the total length of pre-drift extension/strike-slip.

which separates the Patagonia and Colorado blocks. If the subduction angle varied, rollback may have been differential and we argue that Patagonia experienced a much stronger rollback effect (driven by the extinct Phoenix plate) that partly explains large movements along the GFS/AFFZ, and was mostly accommodated between 190 and 160 Ma.

From the middle Jurassic to the late Cretaceous, South America, Africa and the Falkland segment to the south experienced at least three major magmatic events that can be classified as LIP events. The oldest, the Karroo flood basalts (∼182 Ma; Fig. 12a) and related dyke swarms in South Africa, the Falklands (FI) and East Antarctica partly coincided with the more prolonged Chon Aike rhyolite volcanism that affected South America (Colorado, Patagonia) and parts of West Antarctica (see Torsvik *et al.* 2008a, their Fig. 7), and probably related to subduction of the now extinct Phoenix and Farallon plates (Fig. 12a)

6.2 Hauterivian and Barremian reconstructions (130 and 126 Ma)

By ∼132 Ma, our COB and rotation parameters demand that seafloor spreading was initiated in the southernmost South Atlantic and propagated along the total length of the Colorado block by 130 Ma (Fig. 12c). We note that magnetic anomaly M11 (∼132 Ma) has been claimed to have been identified on the Southern margin of Africa but not on the South American margin, so its existence remains to be proven (Eagles 2007). Thus the strongest evidence for the time of opening of the southern segment relies on the M4 anomaly (∼126 Ma, Fig. 12d), but we consider that opening may very well have started a few million years before M4, coinciding with the Paraná-Etendeka magmatic event that peaked at around 132 Ma (linked to the Tristan plume). The Paraná igneous province

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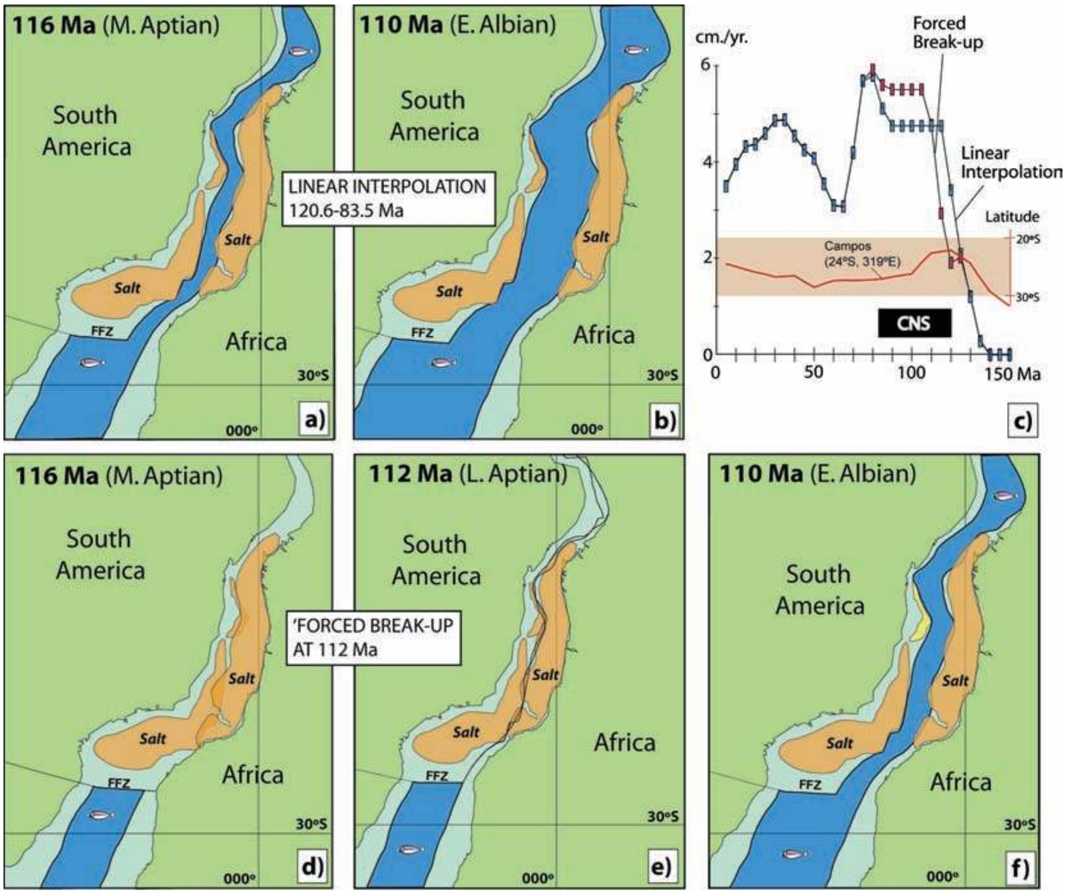

Figure 11. Aptian to early Albian reconstructions using salt outlines and new COBs (black lines). The current shape of the salt basin reflects a combination of extension and inversion as well as a moderate flow of allochthonous salt. Opening the South Atlantic by linear interpolation between 120.6 and 83.5 Ma (M0, C34) does not lead to overlap of the salt basins in mid-Aptian to early Albian times (a, b), which are realistic times to compare the current size of the basins. Note that intraplate movements within South America are assumed to have ceased before the Aptian. (d)–(f) show our preferred model for the South Atlantic where the salt basins are juxtaposed in the late Aptian (e) and seafloor spreading north of the FFZ commenced in the early Albian (f). (c) shows full plate spreading/extension rates for Amazonia versus South Africa (mean plate speeds) using linear interpolation and 'forced' breakup at 112 Ma. We also show the latitude history for a location in the Campos Basin for the last 150 Myr, which demonstrate that Campos was located in the subtropics throughout this period.

is several orders of magnitude larger than Etendeka in extent and presumed volume, and it is therefore reasonable to argue that volcanic outpouring associated with the Tristan Plume occurred closer to Paraná (Fig. 12c), and potentially within the PEFZ. From the onset of seafloor spreading in the extreme south at around 132 Ma most of the South Atlantic margins underwent extension/crustal thinning, and our model incorporates dextral (∼175 km) transtension along the PEFZ. In other words, whilst seafloor spreading propagated northwards (propagating from both south and north?) and reached the PEFZ within 3–4 Myr, the area north of PEFZ (e.g. Santos/Campos at the Brazilian margin) underwent extension that was partly accommodated along the PEFZ until Barremian times (Fig. 12d). At the same time intraplate rifts were active within both Africa and South America.

6.3 Aptian reconstructions (120.6 and 112 Ma)

By the early Aptian (∼121 Ma), seafloor spreading was well established south of the PEFZ/FFZ (Fig. 12e). The ages of the SDRs in the south segment are not well constrained, but we can only assume that they were intruded during the early phases of seafloor spreading (shown on our maps from Aptian and later). The Santos and Campos areas were covered with subaerial basalts at the same time. Rifting in the San Jorge and Colorado Basin (Fig. 8) and northwards continuation up to 22°S had already accommodated most of this opening, with a displacement of 115 km along an E–W trend and an extension up to 80 km. Across the Santos/Namibe Basins the COB overlap was reduced to ∼100 km by the early Aptian; hence our model suggests that the bulk of pre-drift extension (∼250 km) is of pre-Aptian origin. It is interesting to note that the intraplate deformation in South America as well as in Africa, although known for a long time, mostly reactivated the older suture zones from the Pan-African assembly of Gondwana. In South America these structures also had a subdued expression in the more recent Andean Orogeny (e.g. the limit between the Altiplano and Puna plateau); we suggest that they exerted a structural control on the segmentation of the Andean mountain range as it developed.

The Aptian salt basins are clearly pre-breakup (syn-rift) and developed as a single basin within less than 5 Myr towards the end of the Aptian. The basins partly overlie subaerial basalt and accumulated at latitudes between 27°S and 10°S. Near the Aptian/Albian boundary, ∼112 Ma (Fig. 12f), seafloor spreading propagated north of the FFZ, and the Aptian shallow-water salt basin was split into two.

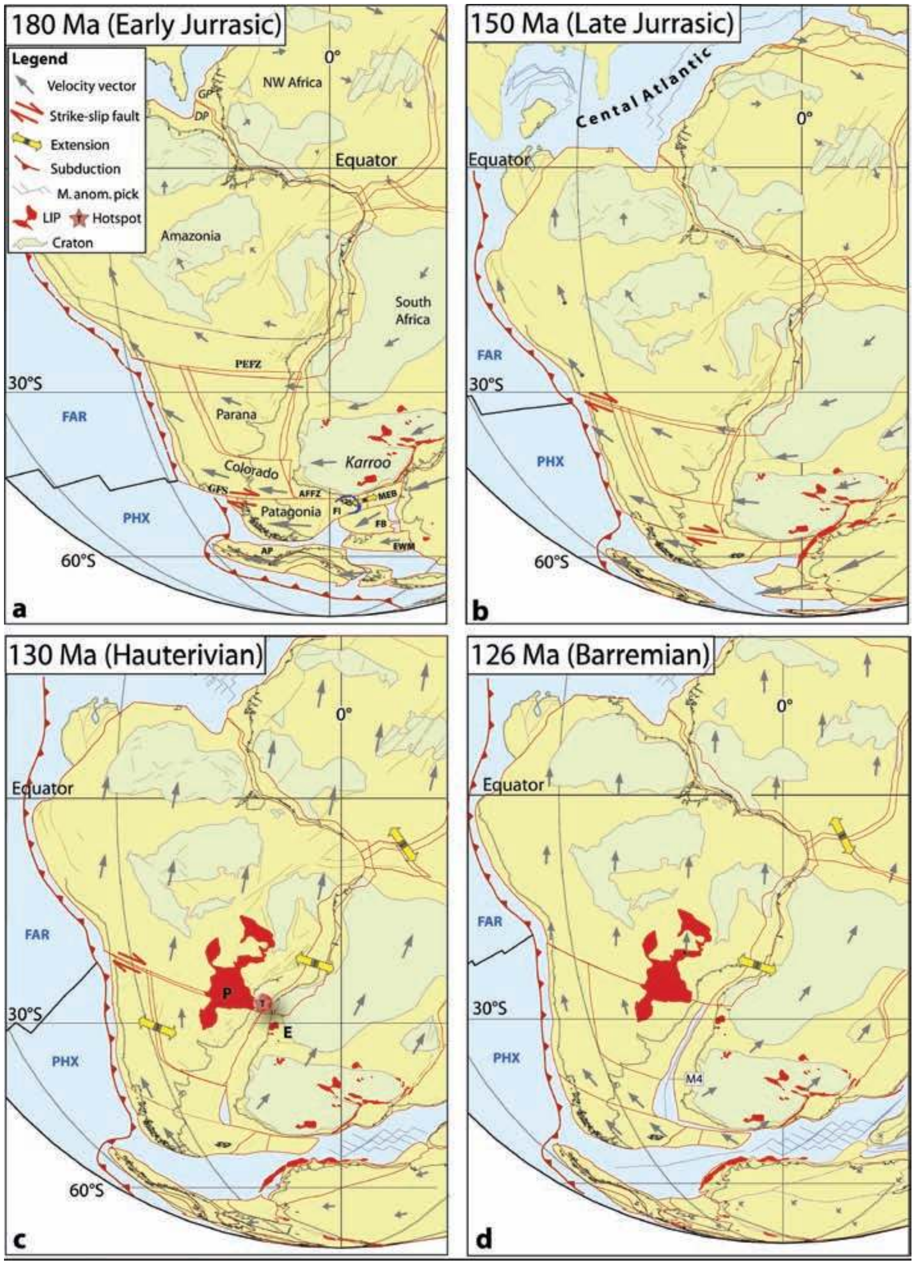

Figure 12. Palaeogeographic reconstructions from (a) the early Jurassic (180 Ma) to (h) the early Campanian (83.5 Ma) based on revised relative reconstruction parameters (Table 1), COBs and Aptian salt basin outlines. South America, Africa and parts of East and West Antarctica were restored to their 'absolute' palaeopositions on the globe with the hybrid plate motion reference frame of Torsvik *et al.* (2008b). Velocity vectors (grey arrows) are also calculated from this reference and represent a 10 Ma average velocity vector (reconstruction time ±5 Ma). Extinct plates: FAR, Farallon; PHX, Phoenix. Major faults/fracture zone: PEFZ, Parana–Etendeka Fracture Zone; GFS, Gastre Fault Systems; AFFZ, Agulhas–Falkland Fracture Zone. Smaller blocks: FI, Falkland Islands (now part of Patagonia); MEB, Maurice Ewing Basin (now part of Patagonia); FP, Falkland Plateau (extended continental crust/possibly oceanic crust now part of Patagonia); EWM, Ellesworth–Whitmore Mountains (now part of West Antarctica); FB, Filchner block (extended continental crust now part of West Antarctica); AP, Antarctic Peninsula (West Antarctica). LIPs: P, Parana; E, Etendeka, MR, Maud Rise. Possible other LIP-related magmatism include AP, Agulhas Plateau; GR, Georgia Rise (see Parsiegla *et al.* 2008); MZR, Mozambique Ridge. Hotspots: T, Tristan; B, Bouvet (Meteor). The Karroo LIP (a) is possibly linked to present day Bouvet hotspot, and the Parana–Etendeka (c) is linked to Tristan hotspot whilst Maud Rise (e) is linked to the Bouvet hotspot. The latter may also been responsible for the Mozambique Ridge [MZR in (e)], which we consider volcanic (at least partly) and not a continental fragment. The combined Georgia Rise and the Agulhas Plateau (Torsvik *et al.* 2008a; Parsiegla *et al.* 2008) probably also represent a LIP event (∼100 Ma, g), and once again related to the Bouvet hotspot. See the text and legend (a) for more details.

6.4 Albian and Campanian reconstructions (100 and 83.5 Ma)

By mid-Albian times (∼100 Ma, Fig. 12g), seafloor spreading propagated north of the Niger segment and connection was established with the Central Atlantic. During breakup north of the PEFZ, the African intraplate boundaries became less active, but tectonic activity did not end until the early Campanian (Fig. 12h). In the early Campanian, ∼83.5 Ma, connection with the Central Atlantic was firmly established, the Rio Grande Rise and the Walvis Ridge were

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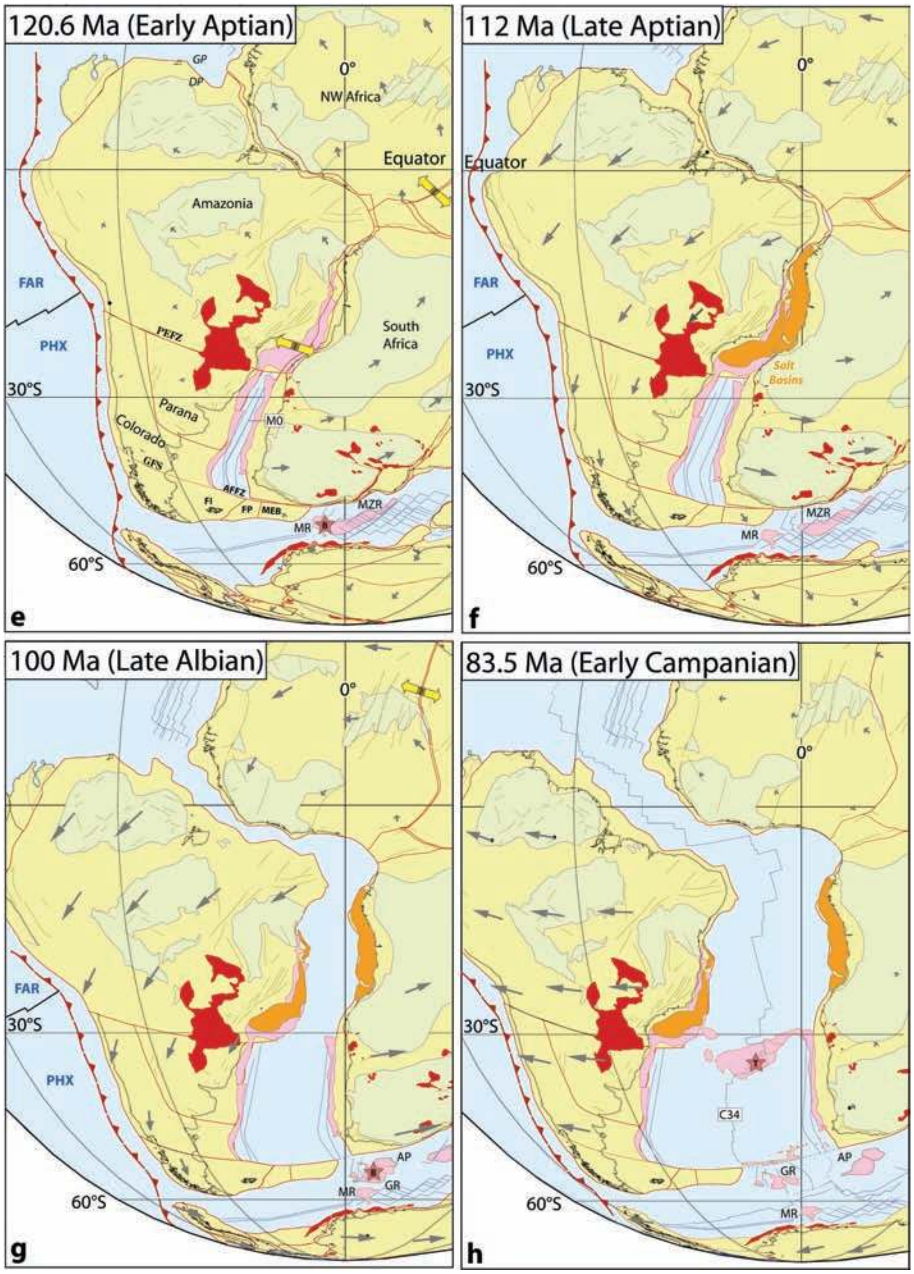

Figure 12. (*Continued.*)

well developed and the Tristan hotspot was located near the spreading ridge. By the early Tertiary, Tristan-related magmatism had crossed the spreading ridge and became located on the African plate.

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APPENDIX A: INTRAPLATE DEFORMATION IN AFRICA AND SOUTH AMERICA

A1 Africa

In the Cretaceous, rifting episodes over large domains occurred in Central and Northern Africa and linked to the breakup of Pangea (Western Gondwana). This intracontinental rifting reused older Pan-African zones of lithospheric weakness (Daly *et al.* 1989), and exhibits both strike-slip and extensional basin geometries accompanied by magmatism in some places (Wilson & Guiraud 1992).

From the late Berriasian to the early Aptian, rifting was very active in Africa–Arabia (Guiraud & Maurin 1992), and rifting was already ongoing in the South Atlantic, tearing three major blocks apart: the Western, the Arabian–Nubian and the Austral blocks (Guiraud & Maurin 1992), here referred to as NW Africa, NE Africa and S Africa, respectively (Fig. 9). Basins trending approximately E–W and NW–SE developed over Central Africa and along the northern African–Arabian Tethyan margin. The N–S trending fault zones of the Algeria–Libya–Niger were rejuvenated and acted as sinistral strike-slip faults (Guiraud & Maurin 1992; Coward & Ries 2003), which resulted in the northward displacement of NE Africa relative to S Africa. This first phase of rifting ended with a regional unconformity (the Austrian unconformity), mostly identified in the Central African Rift system and along the Atlantic Ocean African margin, but also in the Brazilian marginal basins (Chang *et al.* 1992).

The middle Aptian to the late Albian period started with a rapid change in the intraplate stress field. The extension direction, formerly N160°E to N–S, changed to NE–SW, resulting in rapid subsidence along the NW–SE trending Sudan and Ténéré troughs. Active faulting also affected the Sirte Basin in Libya (Guiraud *et al*. 2005). Dextral transtension was initiated along the Central African rift zone (Browne & Fairhead 1983), activating minor NW–SE trending rifts or pull-apart basins in southern Chad, and rifting also continued along some subbasins in the Benue Trough. Strike-slip movement stopped or decreased along the N–S trending trans-Saharan fault zones. During this period, NE Africa tended to move more northeastwards. Although large continental basins covered Africa, two transgressions occurred in the mid-Aptian and late Albian. This second episode of rifting terminated in a regional unconformity, identified both along the Central African Rift System (Genik 1992), where it often marks the cessation of rifting, and in numerous domains in the northern African margin (Guiraud *et al.* 2005). In the late Albian, a major tectonic event, recorded on the margins, coincided with the opening of the Equatorial Atlantic (Mascle *et al.* 1988; Binks & Fairhead 1992; Pletsch *et al.* 2001; Basile *et al.* 2005), whereas in the intracontinental basins of Africa it corresponds with a halt in rifting activity. The latest Albian to the middle Santonian was characterized by decreasing tectonic activity and a marine transgression over the Northern and Western Central Africa, but rifting continued in the Sudan troughs. Locally, and mainly during Cenomanian times, rifting also persisted or was initiated in the Doba Basin (Southern Chad), the Upper Benue Basin, the Ténéré Basin and near the eastern Mediterranean margin (Sirte Basin of Libya). In the Central Africa domain, the subsidence was due to NE–SW directed extension (Guiraud *et al.* 2005) sometimes associated with thermal relaxation.

A simplified five rigid-block model was adopted in our new reconstructions, that is NW Africa, Africa, South Africa, Lake Victoria and Somalia (Fig. 9). The two latter blocks (Somalia and Lake Victoria) are shown on our reconstructions, but not considered in detail in this analysis since they were essentially non-active in the time-interval we have explored (180–83 Ma; Fig. 12); rifting along these boundaries only appeared in the Cenozoic (Guiraud *et al*. 2005).

A2 South America

Like the African continent, important early Cretaceous extension and strike-slip motions also occurred in South America. Important rift deposits accumulated in Northwestern depocentres along northwesterly trending regional zone of crustal weakness in central and southern Argentina (Tankard *et al.* 1995). The syn-rift deposits are mostly late Triassic–early Jurassic and consist of red beds and alkaline basalts. A younger reactivation of these rift system occurred in the early Cretaceous. However, there is no real consensus on the localization of the intraplate deformation in the South American continent (MacDonald *et al.* 2003; Chernicoff & Zappettini 2004; Webster *et al.* 2004; Ramos 2005; Eagles 2007). This is partly due to the less abundant quantitative estimates of onshore deformation available for this continent and to the more tortuous Cenozoic evolution.

The four roughly E–W trending rift basins that were most active in Cretaceous times are (from south to north) the San Jorge Basin (middle Jurassic to early Cretaceous), the Colorado and Salado Basins, two aulacogenic basins that formed during the opening of the South Atlantic and were filled with sequences coeval with the development of the continental platform in the Atlantic margin and the Chaco–Paraná basin. The Salado Basin opened towards the east, and the Colorado Basin, developed along the suture between Patagonia and the South-Western Gondwana continent, to the north; the Chaco–Paraná Basin records a complex history from the Palaeozoic and contains Cretaceous red beds and basalts (Franke *et al.* 2006).

Following the opening of the Atlantic Ocean, the onshore southern limit between the South American Plate and the Scotia Plate lies along the Magellanes Fault System (MFS) (Fig. 8). Simultaneous left-lateral strike-slip motion and transform-normal extension have been documented along this fault system (Lodolo *et al.* 2003). The MFS of Tierra del Fuego is prolonged offshore in the North Scotia Ridge (Bry *et al.* 2004), and with it, forms the boundary between the Scotia Sea, which has only formed since the late Paleogene (Barker 2001), and the Patagonia block. To the West, the MFS links with Pacific-South American subduction zone at around 52°S. On the North Scotia Ridge, the plate boundary itself is an active oblique thrust fault, which has controlled the growth of the frontal front.

Moving north, the E–W trending San Jorge Basin developed at the time of the opening of the Weddell Sea and the South Atlantic Ocean. Following dominantly compressional settings in the Palaeozoic, rifting began around the early Jurassic along the southern margin of Africa. Rifting progressed southwestwards in Patagonia along the Algulhas–Falkland Transform Fault, whose trend can be projected into the San Jorge Basin. This basin formed in a backarc environment and became restricted in the mid-Jurassic (Dalziel *et al.* 1987). During the late Cretaceous, thick sequences of molasse deposits were emplaced unconformably on older sediments. Subsequent sequences of clastic deposits and tuff layers record the volcanic activity in the Andes during the Cretaceous and Paleogene. Continental deposits filled much of the basin, with the first Atlantic marine transgression occurring in Maastrichian to Paleocene times. Coincident timing between the development of the San Jorge Basin and the earliest stage of breakup of western Gondwana suggests that part of the deformation needed to obtain the necessary tight fit between the South American and African plates can be accommodated inside the South American continent, along the Gastre Fault System (GFS), which is an inland continuation of the Algulhas–Falkland Fault (Fig. 1). The GFS, a 30 km wide fault system associated with the setting and emplacement of major granite batholith and crustal block rotation (Geuna *et al.* 2000), formed the northern intracontinental limit of the Patagonian block. It has also been suggested that this limit coincides with a suture zone linked to Cambrian to Carboniferous subduction collision of the Patagonia block with SW Gondwana (Pankhurst *et al.* 2006). However, that did not necessarily accommodate the entire deformation; for example, extensive lower-middle Jurassic tectonism associated with Falkland Plateau Basin rifting during Gondwana breakup has also been postulated as a cause. That tectonic activity ceased in the Upper Jurassic, since post-rift sedimentation occurred on the Maurice Ewing Bank (MacDonald *et al.* 2003).

Further north, Franke *et al.* (2006) provided a comprehensive study of the offshore Colorado Basin based on multichannel seismic reflection data, onshore–offshore refraction profile and gravity data, and argued that the basin developed in conjunction with an early phase of South Atlantic opening (150–130 Ma) and represents a typical rift basin instead of an intracontinental sag basin. The Colorado Basin could have been connected with the Orange Basin of South Africa at the initial stage of evolution, and the early Jurassic extensional phase recorded in the San Jorge Basin probably affected the region. The observed E–W trend of the Colorado Basin, perpendicular to the shelf, may be the product of strike-slip generated pull-apart: the opening direction was initially NNW, towards the Macachin Graben in onshore Argentina. The latter is a linear belt of transtensional subsidence that runs from the Colorado Basin in a NW direction and may be explained by rifting (Tankard *et al.* 1995). The post-Upper Jurassic Colorado and Fortin formations of mostly non-marine origin that fill the Colorado Basin, and the presence of basaltic intrusions of upper Jurassic/lower Cretaceous age in the Salado Basin, as well as comparable intrusion in the Colorado Basin, also support the model for rift movements in both basins (Franke *et al.* 2006). According to these interpretations, the Colorado Basin was chosen as a main intracontinental limit accommodating the deformation linked to the opening.

From the Colorado Basin, the northward continuation can be chosen either along a thrust fault, which represents the western boundary of a basement high that comprises most of the outcrops of low grade metamorphic rocks from the La Pampa Province of Argentina (Chernicoff & Zappettini, 2004), or along the potential Cretaceous rift of the Macachin Basin (Tankard *et al.* 1995; Ramos 2005). The thrust structure is clearly visible in aeromagnetic surveys (Chernicoff & Zappettini 2004), and partly coincides with the proposed location of the suture between the Pampia Terrane and the Rio de la Plata Craton (Ramos 2005). In the southern segment, the thrust is truncated by apparently undeformed Permian to early Triassic intrusive bodies (Chernicoff & Zappettini 2004), so the Macachin option is preferred. To the north, the structures link with the General Levalle Basin in southern Cordoba Province (Webster *et al.* 2004). In the General Levalle Basin, the lowest rift-fill section is a middle Neocomian siliciclastic and evaporite package deposited in an arid restricted rift basin, whereas the uppermost rift-fill sequence is a series of Aptian basalt flows and sills with some thin red-bed intervals (Webster *et al.* 2004). From the General Levalle Basin the limit is extended into the southeastern Pampean Range, where the inverted Sierras Chica Cretaceous Basin crops out (Schmidt *et al.* 1995). Although more poorly defined, this trend extends northwards to the northwest Aimara Basin of Salta and JuJuy Provinces at around 22°S (Webster *et al.* 2004; Monaldi *et al.* 2008). West and subparallel with these trends several other structures and inverted Cretaceous basins are aligned (Chernicoff & Zappettini 2004; Webster *et al.* 2004; Ramos 2005), but the trend described above is one that can be understood on a large geographical scale. This curved limit, though the closest to the known palaeogeography of the Cretaceous intracontinental rifts, is not often used as it is, because it complicates the reconstructions. Thus the authors have usually preferred to introduce deformation in South America along more or less latitudinal limits (Nürnberg & Müller 1991; Eagles 2007), but evidence for these limits cutting the Pacific coast between 22°S and 42°S is rare, although these hypotheses cannot be ruled out.

Of the intraplate deformation being described here, the Southern Paraná limit (SPL or PEFZ, Figs 8 and 10) is less documented from surface studies; however, its presence can be suspected from several lines of evidence. For example, calculations of the effective elastic thickness (Te) of South America show that intracontinental deformation is focused within relatively thin, hot (and hence weak) lithosphere and that the cratonic interiors are strong enough to inhibit tectonism (Pérez-Gussinyé *et al.* 2007, 2008). Within the Precambrian basement, low Te that usually coincides with sutures, rifts or hotspot magmatism is clearly identified in the Paraná Basin at ∼27°S, west of the region covered by flood basalts, where NW–SE-oriented dyke swarms that fed the ∼130 Ma Paraná flood basalts are exposed (Milner *et al.* 1995). The large area of low Te confirms the large scale of the feature evident from aeromagnetic surveys (Milner *et al.* 1995), and correlates with low shear wave velocities (Feng *et al.* 2007), relatively high heat flow (Hamza *et al.* 2005) and the presence of intracontinental seismicity (Assumpçao *et al.* 2004). These features are enhanced towards the West where the SPL apparently extends into the Chaco Basin up to the foothills of the Andes. Near the Andes, the chosen limit connects with the E–W trending structures of the Cretaceous Salta Rift (Monaldi *et al.* 2008), and in the Andes, it also roughly coincides with the separation between the Altiplano and Puna plateaux that have developed contrasting tectonic histories (Allmendinger *et al.* 1997), and appear to be characterized by a more mafic as opposed to a more felsic crust. The SPL also coincides with the southern limit of the Arequipa block, based on Pb and Nd–Sr isotopes (Mamani *et al.* 2008). On the other side, close to the Atlantic Ocean, the Rio Grande and Ponta Grossa arches that bound the chosen limit to the north and south are known to have been uplifted in early Cretaceous time (Ernesto *et al.* 1999).

On the basis of these data, south of the SPL South America was divided into three rigid blocks: Parana, Colorado and Patagonia that accommodated our scenario for the opening of the South Atlantic. North of the SPL, several other sutures probably exist that accommodated the last steps of the separation of South America and Africa. Their location is subject to debate (e.g. the relative importance of the Amazon Basin versus the Transbrasiliano Lineament; Eagles 2007; Pérez-Guyssiné *et al.* 2007) and is beyond the scope of this paper. Similarly, the earliest steps of the Africa–South America separation also pose their own complex geodynamic problems (Ghidella *et al.* 2007), and will be discussed in a different paper.