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Paraná Magmatic Province–Tristan da Cunha plume system: fixed versus mobile plume, petrogenetic considerations and alternative heat sources

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Abstract

Paleomagnetic reconstructions demonstrate that the Tristan da Cunha (TC) plume, which is usually related to the genesis of the high- and low-Ti flood tholeiites of the Paraná Magmatic Province (PMP), was located ∼1000 km south of the Paraná Province at the time of the magma eruptions. Assuming plume mobility, and considering the low-velocity zone identified in the northern portion of the PMP as the TC 'fossil' plume (∼20° from the present TC position), the plume migrated southward from 133–132 (main volcanic phase) to 80 Ma at a rate of about 40 mm/yr. From 80 Ma to Present the plume remained virtually fixed, leaving a track (Walvis Ridge) compatible with the African plate movement. However, geochemical and Sr–Nd–Pb isotopic data do not support that the tholeiites from Walvis Ridge, Rio Grande Rise and Paraná can result from mixing dominated by the TC plume and mid-ocean ridge basalt components. The similarity among the high-Ti basalts from Rio Grande Rise, part of Walvis Ridge (525A) and the Paraná Province suggests that delaminated subcontinental lithospheric mantle must be considered in their genesis. Regional thermal anomalies in deep mantle mapped by geoid and seismic tomography data offer an alternative non-plume-related heat source for the generation of intracontinental magmatic provinces.

Keywords: Paraná Magmatic Province; Tristan da Cunha; mantle plumes; paleomagnetic reconstructions; geoid anomalies

1. Introduction

The origin of the Early Cretaceous flood tholeiites of the Paraná Magmatic Province (PMP), including the equivalents in Etendeka (Namibia) and Angola (Fig. 1), is commonly related to the Tristan da Cunha (TC) plume (e.g. Richards et al., 1989; White and McKenzie, 1989; Milner

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and Le Roex, 1996; Gibson et al., 1995; Courtillot et al., 1999), although geochemical evidence led Peate (1997), Comin-Chiaramonti et al. (1997) and Marques et al. (1999) to discard the role of the TC mantle source in the generation of the Paraná melts except as a source of heat. The concept of mantle plumes since the initial proposition by Wilson (1963) became so popular that virtually all large igneous provinces are dogmatically assumed to have been associated with a mantle plume, although the weaknesses of this model have already been evidenced by Anderson (1996, 1998), Smith and Lewis (1999) and Sheth (1999), among others. In the case of the Paraná Province, geological, geochemical and petrological data (e.g. Piccirillo and Melfi, 1988; Peate and Hawkesworth, 1996; Comin-Chiaramonti et al., 1997; Peate, 1997) demonstrated that the source mantle signatures of TC volcanics are not appreciable in the Paraná Large Igneous Province (LIP). This indicates that the 'plume head–hotspot' system, formerly proposed by Wilson (1963), cannot easily be assumed a geodynamic model in order to explain the genesis of the Paraná magmatism, particularly if the genesis of the Paraná flood tholeiites is considered compatible with models involving 'edge-driven convection' processes of upper mantle (e.g. Anderson, 1994a,b; King and Anderson, 1998; Sheth, 1999; King and Ritsema, 2000; Tackley, 2000), as recently documented for the Central Atlantic Magmatic Province (e.g. McHone, 2000; De Min et al., 2002).

The Paraná flood tholeiites have often been used in the literature as one of the best examples of a continental LIP genetically related to the plume-head, mainly because the oldest part (113 Ma) of the Walvis Ridge volcanic chain is located close to the Etendeka plateau (e.g. O'Connor and Duncan, 1990), which represents the easternmost extent of the PMP in a pre-Atlantic configuration. Therefore, there is a strong tendency to interpret the Walvis Ridge as a trace left by the TC hotspot, in spite of the fact that it is very difficult to interpret the expected western symmetrical magmatic chain (i.e. Rio Grande Rise) in terms of a simple mantle plume–hotspot model (see below).

However, the location of the TC plume relative to the South American plate by the time of Paraná volcanism is still unconstrained as the plume head emplacement under the lithosphere predates the South Atlantic opening. The oldest oceanic magnetic anomaly (M4), at Paraná latitudes dates to 126.5 Ma (Nürnberg and Müller, 1991), and no hotspot trace was recognized onshore. Geochronological data, based on the $^{40}Ar/^{39}Ar$ method (Renne et al., 1992; Turner et al., 1994, Renne et al., 1996a; Ernesto et al., 1999), show that the main volcanic activity in the PMP took place at 133–132 Ma, before the South America and Africa separation (Piccirillo et al., 1988). Therefore, between 113 Ma (Walvis Ridge) and 133–132 Ma (main Paraná volcanism), there is a time interval of about 20 Ma of unknown history regarding the location of the plume relative to the continents.

Various attempts to locate the TC plume have been made, and there is a consensus that it must be placed beneath the PMP since this province would represent the major surface expression of the plume. In doing so, most of the authors admit TC as a fixed point in relation to the mantle (e.g. O'Connor and Duncan, 1990; Turner et al., 1996; Courtillot et al., 1999), and the Western Gondwana supercontinent (South American and African plates) is reconstructed by forcing the PMP over the TC plume. Paleolongitudes and paleolatitudes of the supercontinent are therefore set arbitrarily in terms of the TC–PMP relationship.

The mantle plume geodynamic model was originally conceived as a fixed thermal mantle anomaly generated at lower mantle–outer core depths over which the lithospheric plate moved, leaving a time-related volcanic track. However, this model has shown inconsistency in the case of some LIPs believed to be related to mantle plumes and, therefore, a mobile plume with velocities exceeding 100 mm/yr (e.g. Molnar and Stock, 1987; Vandamme and Courtillot, 1990; Duncan, 1990) must be considered.

Local scale and high resolution tomography from VanDecar et al. (1995) has mapped a low-velocity zone in the mantle, extending from 100 to 600 km, which has been interpreted as the 'fossil' TC plume in the northeastern Paraná Province. However, no geoid anomaly or surface expression

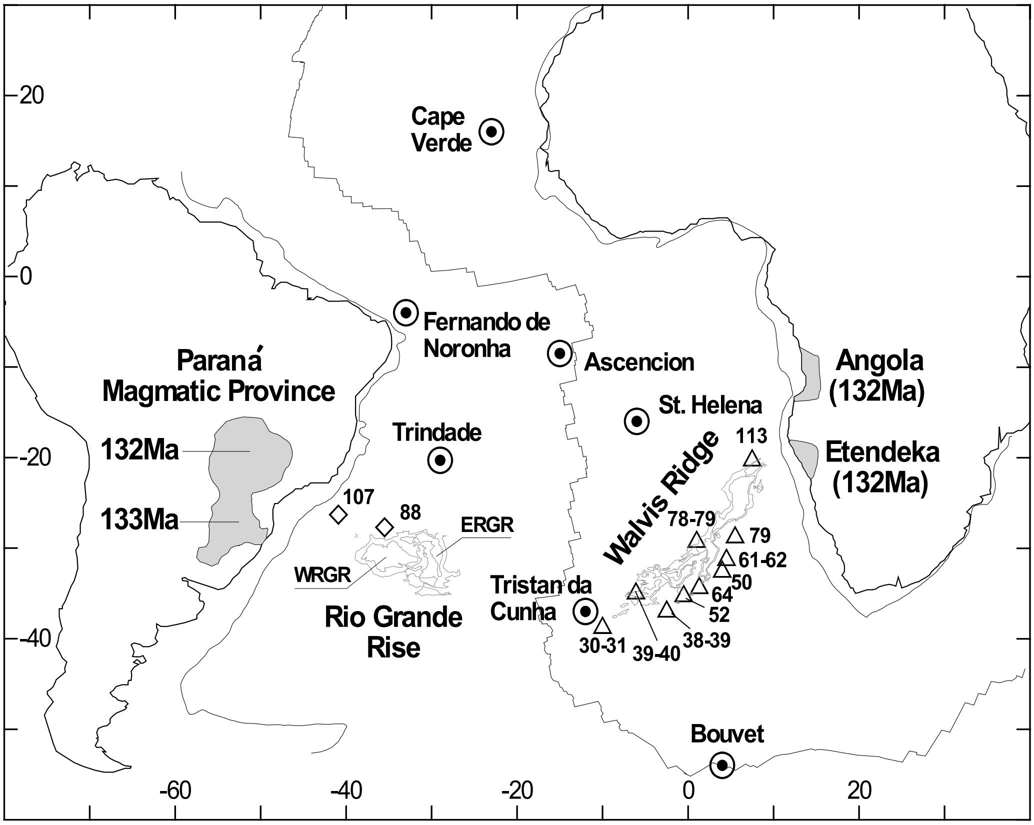

Fig. 1. Location of the PMP in South America, and the corresponding Angola and Etendeka provinces in Africa. The TC hotspot is shown along with other Atlantic hotspots. Bathymetric data and ages for Walvis Ridge and Rio Grande Rise are according to Müller et al. (1993).

of the TC thermal anomaly is recognized in this region (Molina and Ussami, 1999), except for the Iporá and Alto Paranaíba Late Cretaceous alkaline provinces (Comin-Chiaramonti et al., 1997, and references therein) further to the north.

In addition, according to the plume model, it would be expected that geochemical and isotopic signatures would be compatible with a dominant contribution of asthenospheric mantle materials in the basalt genesis (Campbell and Griffiths, 1990; Arndt and Christensen, 1992). For the PMP, in spite of the large amount of petrological, geochemical and isotopic data available, there is no consensus about the mantle sources involved in the basalt genesis. An origin in the lithospheric mantle, without significant geochemical and isotopic plume signatures, is proposed to explain the geochemical and isotopic characteristics of the low- and high-TiO_2 PMP tholeiites, their spatial distribution in the province and the association with coeval alkaline and alkaline-carbonatitic magmatism (e.g. Piccirillo and Melfi, 1988; Peate and Hawkesworth, 1996; Comin-Chiaramonti et al., 1997; Marques et al., 1999). On the other hand, according to other interpretations (Gibson et al., 1995, 1999; Milner and Le Roex, 1996), some basalt compositions could reflect the geochemical imprint of the TC mantle plume.

The purpose of this paper is to (1) summarize the geochemical and isotopic data of the Paraná basalts and their comparison with those of Rio Grande Rise, Walvis Ridge, South Atlantic Mid-Ocean Ridge and TC volcanic rocks, particularly for their mantle source characteristics; (2) test the

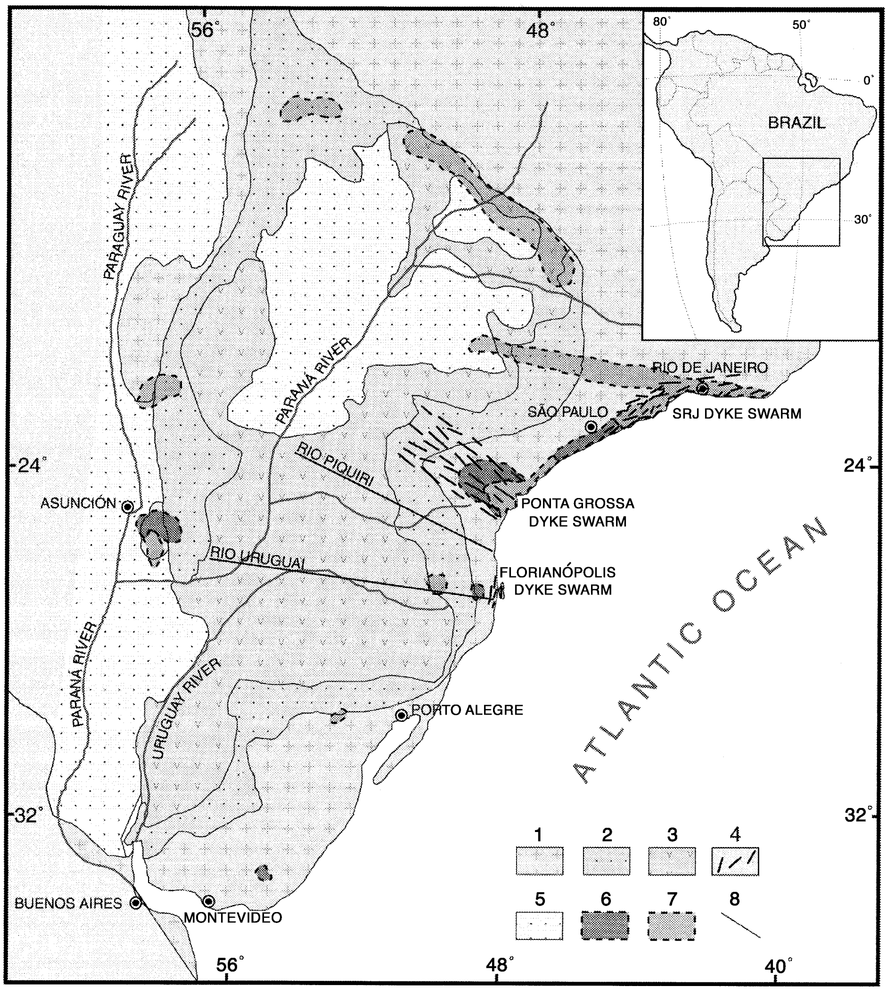

Fig. 2. Generalized geological map of the Paraná Basin. 1 = Pre-Devonian crystalline basement; 2 = pre-volcanic sediments (mainly Paleozoic); 3 = flood volcanics of the PMP; 4 = dyke swarms associated with the PMP; 5 = post-volcanic sediments (mainly Late Cretaceous); 6 = main areas of Early Cretaceous alkaline and alkaline–carbonatitic rocks; 7 = main areas of Late Cretaceous alkaline and alkaline–carbonatitic rocks; 8 = tectonic and/or magnetic lineaments. Data sources: Piccirillo and Melfi (1988); Comin-Chiaramonti et al. (1997); Alberti et al. (1999).

hypotheses of both the fixed and the mobile TC plume, and set constraints on the location of the TC plume using an updated paleomagnetic database for the Early Cretaceous from South America and Africa; (3) discuss an alternative hypothesis for the Paraná magmatism heat source, integrating new evidence of the existence of an anomalous mantle region, given by short to intermediate geoid anomalies, and recently published global seismic tomography results (Zhou, 1996; King and Ritsema, 2000).

2. The PMP

The volcanic rocks of the PMP are of Early Cretaceous age (Renne et al., 1992; Turner et al., 1994, Ernesto et al., 1999) and were emplaced on a large intracratonic Paleozoic sedimentary basin (Paraná Basin) that started subsiding in the Early Paleozoic (Fig. 2). The Paraná Basin forms part of the South American platform that was affected by the metamorphic and magmatic episodes related to the Brasiliano Cycle (ca. 700–450

Ma). The crystalline basement of the Basin is probably formed by different cratonic nuclei surrounded by mobile belts (Cordani et al., 1984).

The PMP comprises continental flood basalts, sills and dyke swarms concentrated towards the eastern continental margin. The total area flooded by the magmatism considerably exceeded that of the present occurrence of the volcanics (1 200 000 km^2). The Ponta Grossa arch was the site of the most important dyke swarm in the province and is characterized by hundreds of basaltic dykes (mainly NW–SE trending). The Florianópolis dyke swarm in Santa Catarina Island, as well as those exposed from Santos to Rio de Janeiro, parallels the coast (ca. NE–SW). The dykes intruded mainly the Paleozoic sediments and the Proterozoic–Archean crystalline basement.

The PMP volcanics and intrusives are represented by dominant flood tholeiites (Fig. 2) that are differentiated (Mg# < 0.56), and are divided into: (1) LTiB basalts, low in TiO_2 (< 2 wt%) and incompatible elements (e.g. P, Sr, Ba, Zr, Ta, Y and light rare earth elements) and (2) HTiB basalts, high in TiO_2 (> 2 wt%) and incompatible elements. HTiB tholeiites dominate the northern Paraná Basin (north of $\sim 26°S$), whereas LTiB rocks prevail in the southern PMP (south of $\sim 26°S$). Minor HTiB and LTiB are found in the southern and northern PMP, respectively.

The tholeiites from the sills and from Ponta Grossa, Florianópolis and Santos–Rio de Janeiro dyke swarms show geochemical and isotopic characteristics close to those of the flood plateau Paraná basalts, in spite of most being of HTiB-type (Comin-Chiaramonti et al., 1983; Bellieni et al., 1984; Piccirillo et al., 1990; Hawkesworth et al., 1992; Ernesto et al., 1999; Marques, 2001; unpublished data).

High-precision $^{40}Ar/^{39}Ar$ dating, along with paleomagnetic studies, allowed tight constraints on the age of PMP rocks, indicating that the main magmatic activity occurred within a few million years. These data show that the flood volcanism (mainly 132–133 Ma; Renne et al., 1992, 1996a; Turner et al., 1994; Ernesto et al., 1999) was followed by the emplacement of the dyke swarms. The Ponta Grossa dykes were intruded during a narrow interval (131–129 Ma; Renne et al., 1996b), although dykes as young as 120 Ma can be found towards the continental margin. The radiometric ages of Florianópolis dykes vary between 129–119 Ma (Raposo et al., 1998; Deckart et al., 1998), whereas for the Santos–Rio de Janeiro swarm they span from 132 to 119 Ma (Renne et al., 1993; Turner et al., 1994; Deckart et al., 1998).

It is worth noting that Early Cretaceous K-alkaline and alkaline-carbonatitic magmatism (Comin-Chiaramonti et al., 1997, 1999) is older (e.g. northeastern Paraguay, 145 Ma), coeval (e.g. southeastern Brazil: Jacupiranga, Juquiá, Anitápolis, and Uruguay, 130–132 Ma) or younger (e.g. central eastern Paraguay, 128–126 Ma) than the Early Cretaceous tholeiitic magmatism in the PMP, and occurred along tectonic lineaments (e.g. Ponta Grossa Arch, Rio Uruguay and Rio Pilcomayo) which were also sites of Late Cretaceous alkaline magmatism. The latter is also found along other important tectonic lineaments such as Alto Paranaíba–Iporá and Taiuvá–Cabo Frio lineaments.

3. The Rio Grande Rise and Walvis Ridge

The Rio Grande Rise is composed of two distinct portions described by Gamboa and Rabinowitz (1984) as an elevated western (WRGR) portion of elliptical shape and an eastern (ERGR) portion, about 600 km long, trending north–south and parallel to the trend of the present Mid-Atlantic Ridge. These authors claim that ERGR and its conjugate portion of Walvis Ridge were formed at the same time by the same processes, and could represent an abandoned spreading center. However, ERGR and WRGR are morphologically different and may have had distinct origins. LePichon and Hayes (1971) suggested that WRGR represents a transverse ridge formed along a fracture zone trend and that the north–south ERGR is due to a modification in the stress field during the opening of the South Atlantic.

Coring at site 516F indicates that the basement of WRGR consists of tholeiitic basalts of about 85 Ma, but rocks dredged from the escarpments of the guyots, towering over the platform of the

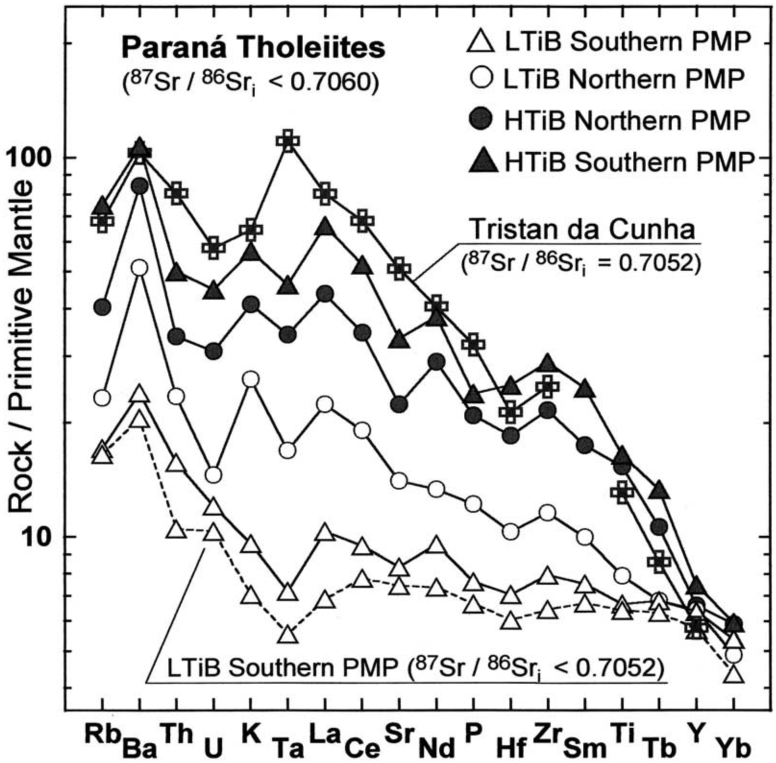

Fig. 3. Mantle-normalized (Sun and McDonough, 1989) incompatible trace element distribution patterns relative to the low- and high-Ti Paraná tholeiites. In contrast to the PMP basalts, the trace element pattern of the TC alkaline volcanics (MgO > 5.5 wt%) shows a distinctive Ta positive anomaly. Data sources: PMP (Mantovani et al., 1985; Petrini et al., 1987; Piccirillo and Melfi, 1988; Piccirillo et al., 1989; Marques et al., 1989, 1999); TC (Le Roex et al., 1990).

rise, indicated the presence of Eocene (K–Ar date of 47 Ma; Bryan and Duncan, 1983) alkaline basalts (Fodor et al., 1977).

4. Mantle sources of the PMP

In order to investigate the mantle source characteristics and the contribution of different mantle components involved in the PMP basalt genesis, only tholeiites with SiO_2 < 53 wt% and initial $^{87}Sr/^{86}Sr$ < 0.7060 were considered. Therefore, effects caused by different extents of partial melting and/or fractional crystallization processes, and low-pressure contamination were minimized.

The incompatible trace element distribution patterns normalized to primordial mantle (Sun and McDonough, 1989) for the LTiB and HTiB (TiO_2 > 3 wt%) from the northern and southern PMP are shown (Fig. 3). The HTiB and LTiB patterns from the northern PMP are very similar, despite the different abundances of incompatible elements among the two groups. On the other hand, the LTiB from the northern PMP are characterized by a strong U negative anomaly which is not observed in the LTiB from the southern PMP. The latter are also distinct due to their relatively low La/Ce ratio, especially those with initial Sr isotopic ratios lower than 0.7052. All the PMP tholeiites have a Ta negative anomaly, which may be considered a mantle source signature. This Ta (Nb) negative anomaly is also systematically present in Early Cretaceous K-alkaline magmatism from Paraguay (Comin-Chiaramonti et al., 1999).

The geochemical differences are accompanied by systematic changes in Sr, Nd and Pb isotopes

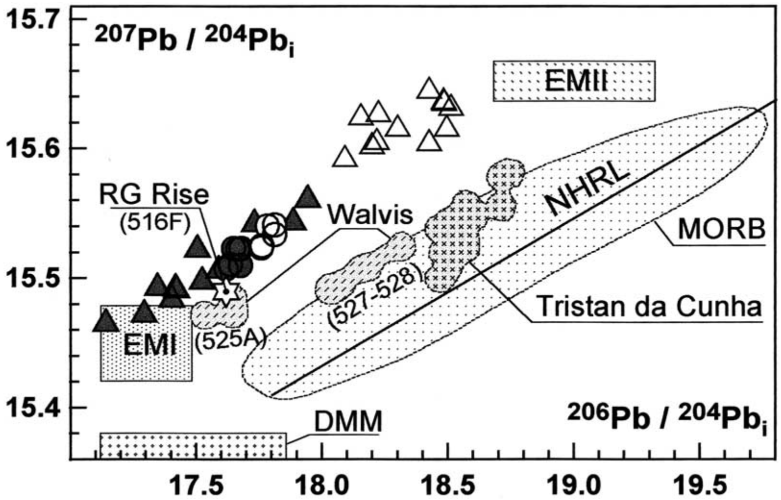

Fig. 4. Lead isotopic compositions for the PMP tholeiites in comparison to the fields for MORB, TC volcanics, Walvis Ridge and Rio Grande Rise. According to mixing systematics, TC plume and/or N-MORB components did not play a significant role in the PMP basalt genesis. Symbols as in Fig. 3. Data sources: PMP (Peate and Hawkesworth, 1996; Peate et al., 1999; Marques et al., 1999); MORB (Hamelin et al., 1984; Ito et al., 1987, and references therein); Walvis Ridge (Richardson et al., 1982); TC (Sun, 1980; Le Roex et al., 1990); Rio Grande Rise (Hart, 1984); EMI, EMII and DMM components (Zindler and Hart, 1986).

(Peate and Hawkesworth, 1996; Marques et al., 1999; Peate et al., 1999). The significant variations between the northern and southern PMP tholeiites indicate the contribution from different mantle source materials. Considering the hypothesis that the PMP tholeiites were generated by the impact of the TC plume head beneath the Western Gondwana, the mantle components potentially involved in the basalt genesis could be the depleted upper mantle (N-MORB type), plume material (ocean island basalt type) and the lithospheric mantle.

The remarkable differences in incompatible trace element ratios (e.g. La/Th, Nb/La, Zr/Ta, Ce/Pb) among Paraná tholeiites, N-MORB and TC least evolved alkaline volcanics are indicative that MORB and TC asthenospheric mantle components did not play a substantial role in the Paraná basalt generation (Marques et al., 1999). This is strengthened by radiogenic isotopic ratios that do not show evidence of significant involvement of either mantle components (Figs. 4 and 5). In addition, the geochemical and Sr–Nd–Pb isotopic data are different from those of Walvis Ridge basalts, which are commonly believed to be the traces left by the continuous magmatic activity of the TC plume. However, samples of site DSDP 525A (used to define the EMI mantle component; Zindler and Hart, 1986) of Walvis Ridge and basalts of Rio Grande Rise (site 516F) are very similar to the HTiB, especially those from the southern PMP, suggesting the involvement of a common depleted end member source. It is worth noting that the Walvis Ridge basalts of sites 527, 528, 530A and 524, lower portion, cannot be explained simply by binary mixing between N-MORB and TC mantle sources, as evidenced in Fig. 6.

It is also important to stress that the mantle heterogeneity involved in the Paraná magmatism is not confined to the tholeiites, but also applies to PMP Early (and Late) Cretaceous alkaline magmatism. Even the carbonatites have on the whole a Sr–Nd–Pb isotopic imprinting close to that of the related alkaline rocks and the spatially associated tholeiites (Comin-Chiaramonti et al., 1997, 1999; Alberti et al., 1999; Peate et al., 1999; Marques et al., 1999), indicating similar mantle components in their genesis.

Recently, it has been suggested that the present-day plume compositions underwent substantial variations over the last 130 Ma due to the inter-

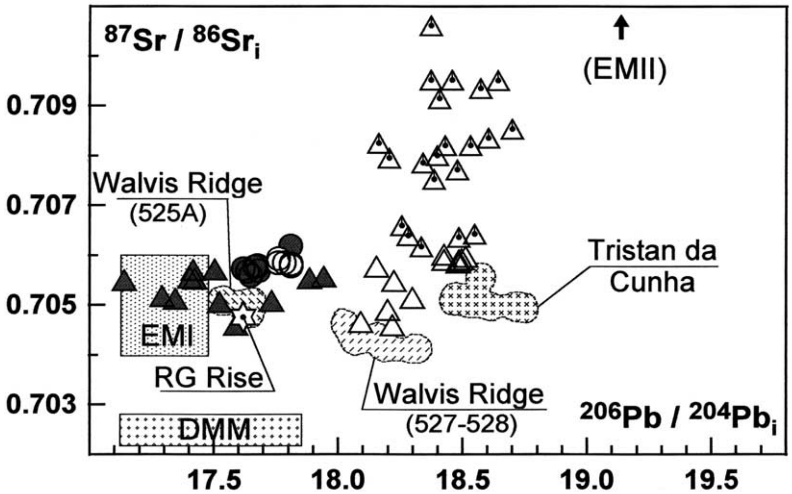

Fig. 5. Initial Sr and Pb isotopic compositions for PMP basalts in comparison to the fields of Walvis Ridge, Rio Grande Rise and TC. The isotopic signatures of HTiB, Rio Grande Rise (516F) and Walvis Ridge (525A) are very similar and indicate a significant involvement of the EMI mantle component in their genesis. Symbols as in Fig. 3; open triangles with a dot correspond to the LTiB with $^{87}Sr/^{86}Sr_i$ >0.7060. Data sources as in Fig. 4.

action of plume material with surrounding lithospheric mantle (Gibson et al., 1995; Ewart et al., 1998). In this case, the LTiB and HTiB from the Paraná would require plume-derived melts to be almost completely dominated by lithospheric mantle components in order to explain the generation of low- and high-TiO_2 basalts in southern and northern Paraná regions, respectively.

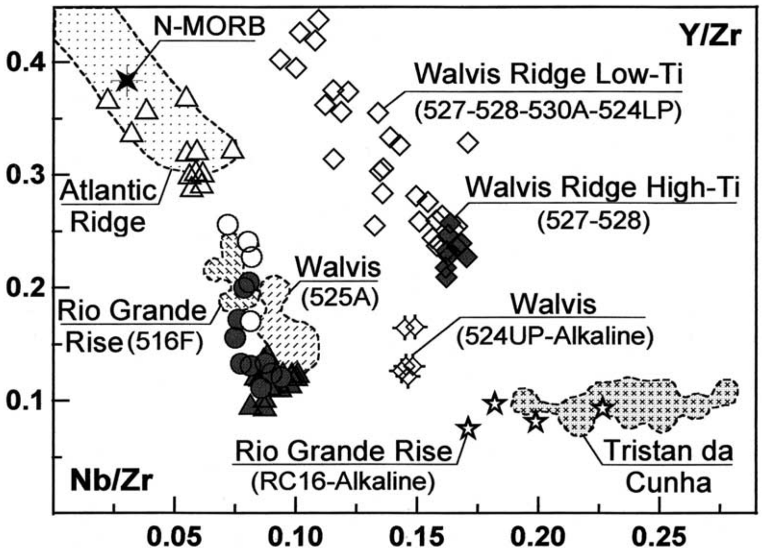

Fig. 6. Zr, Y and Nb relationships for the PMP tholeiites, N-MORB, Walvis Ridge, Rio Grande Rise and TC. Note that Walvis Ridge (site 525A) and Rio Grande Rise (site 516F) basalts, all of high-Ti type, are very similar to the PMP (all HTiB and northern LTiB) tholeiites. The alkaline rocks of Walvis Ridge (65 Ma; site 524, upper portion) and Rio Grande Rise (46 Ma; RC16) are also shown for comparison. Symbols as in Fig. 3. Data sources: PMP (Piccirillo and Melfi, 1988; Peate and Hawkesworth, 1996; Marques et al., 1999; Peate et al., 1999); N-MORB (Sun and McDonough, 1989); Atlantic Ridge (Dietrich et al., 1984; Humphris et al., 1985); TC (Le Roex et al., 1990); Walvis Ridge (Humphris and Thompson, 1983; Dietrich et al., 1984; Richardson et al., 1984; Thompson and Humphris, 1984); Rio Grande Rise (Fodor et al., 1977; Thompson et al., 1983).

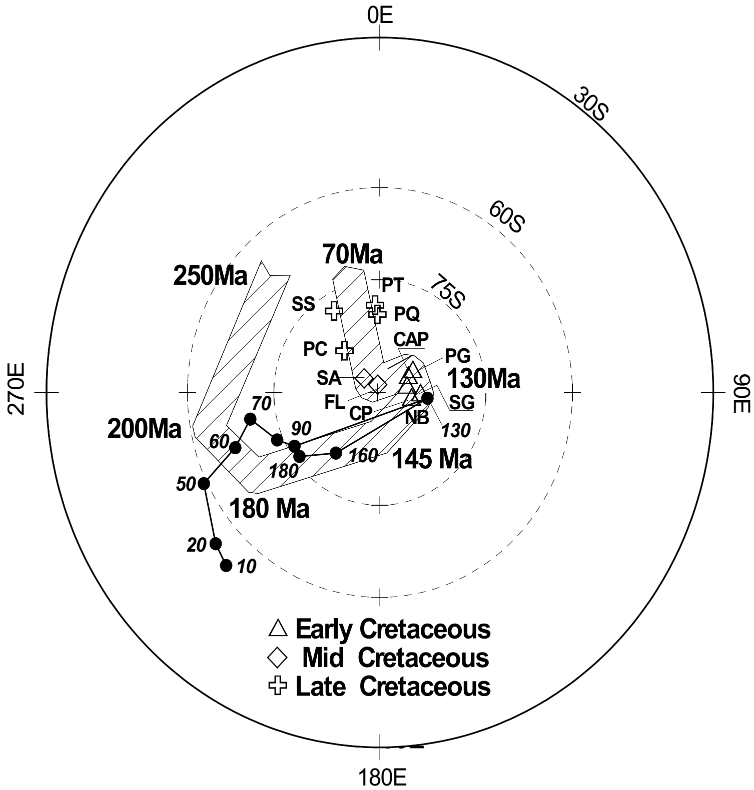

Fig. 7. Wulff stereogram displaying selected Cretaceous paleomagnetic poles for South America and the APWP for the Mesozoic as proposed by Ernesto et al. (2002). Superimposed is the APWP (Besse and Courtillot, 1991) since 180 Ma for Africa (rotated to South America). Codes as in Table 1.

In conclusion, all the geochemical data indicate that the genesis of the PMP tholeiites mainly reflects melting of heterogeneous lithospheric mantle reservoirs (cf. Comin-Chiaramonti et al., 1997). In addition, the geochemical and isotopic signatures of Walvis Ridge and Rio Grande basalts may be explained by detached continental lithospheric mantle left behind during the continental break-up processes (e.g. Hawkesworth et al., 1986; Peate et al., 1999).

5. Paleomagnetic constraints for the PMP–TC plume relative position

Although paleomagnetism is insensitive to longitude variations since it is based on a geocentric axial dipole model of the Earth's magnetic field, paleomagnetic data are very useful for fixing paleolatitudes and rotations experienced by tectonic blocks. If additional information about longitudes exists, then an absolute paleogeographic reconstruction can be approximated. In the case of the separation of the South American and African plates during the Mesozoic, the spreading center of the ocean floor (the Mid-Atlantic Ridge) constitutes a very good indicator for the longitudes occupied by Western Gondwana just prior to the ocean opening.

The Mesozoic apparent polar wander path (APWP) for South America (Fig. 7) is now well established, especially for the Cretaceous (Ernesto et al., 2001). High quality paleomagnetic poles based on a large number of independent sites and satisfying rigorous criteria of selection are available. The selected paleomagnetic data

Table 1
Selected Cretaceous paleomagnetic poles for South America

Formation	Age (Ma)	Paleomagnetic pole N	Long. (°E)	Lat. (°S)	α_{95} (°)	Code	Reference
Patagonia volcanics, Argentina	79–64	18	358.4	78.7	6.5	PT	Butler et al. (1991)
Passa Quatro–Itatiaia Complex, Brazil	70	18	331.9	79.4	4.9	PQ	Montes-Lauar et al. (1995)
Poços de Caldas Complex, Brazil	84	47	320.1	83.2	2.7	PC	Montes-Lauar et al. (1995)
São Sebastião Island, Brazil	81	18	331.9	79.4	4.9	SS	Montes-Lauar et al. (1995)
Mean Late Cretaceous	**∼80**	**4**	**343.9**	**80.3**	**4.6**		
Santo Agostinho Cape, Brazil	99–85	9	315	87.6	4.5	SA	Schult and Guerreiro (1980)
Florianópolis dykes, Brazil	∼121	65	3.3	89.1	2.6	FL	Raposo et al. (1998)
Mean Mid Cretaceous	**∼110**	**2**	**327.6**	**88.5**	**4.2**		
Central Alkaline Province, Paraguay	127–130	75	62.3	85.4	3.1	CAP	Ernesto et al. (1996)
Ponta Grossa dykes, Brazil	129–131	115	58.5	84.5	2.0	PG	Ernesto et al. (1999)
Serra Geral formation, Paraná Basin	133 ± 1	339	90.1	84.3	1.2	SG	Ernesto et al. (1999)
Cordoba Province, Argentina	133–115	55	75.9	86.0	3.3	CP	Geuna and Vizan (1998)
Northeastern Brazil magmatism	125–145	44	97.6	85.2	1.8	NB	Ernesto et al. (2002)
Mean Early Cretaceous	**∼130**	**5**	**76.6**	**85.3**	**1.5**		

N = number of sites; α_{95} = Fisher's (1953) confidence circle.

(Table 1) are restricted to magmatic units to assure highest fidelity to paleofield directions and better age constraints. The Early Cretaceous (∼130 Ma) is marked by a set of poles (Table 1) dating ∼145–127 Ma, all in good agreement with the Serra Geral (SG; 133–132 Ma) pole from the Paraná Basin based on more than 330 independent sites. The Late Cretaceous poles cover a time interval of approximately 90–70 Ma, and the path between Early and Late Cretaceous is defined by the Florianópolis (FL) pole, based on ∼121 Ma basic dykes ($^{40}Ar/^{39}Ar$ dating; Raposo et al., 1998), and the magmatic rocks from the Cabo de Santo Agostinho (SA; Schult and Guerreiro, 1980), ranging in age from 90 to 110 Ma. These two poles are close and almost coincide with the Earth's geographic pole, indicating that the South American plate remained virtually at its present latitude between 120 and 100 Ma. According to the APWP, from Jurassic (200–145 Ma) to Early Cretaceous the main axis of the South American plate underwent a clockwise rotation, causing slight changes in latitudes. From Early to Late Cretaceous the same sense of rotation persisted but the plate moved southwards. The APWP for Africa (Fig. 7) since 180 Ma (Besse and Courtillot, 1991) indicates that the Jurassic and Early Cretaceous segments of this curve are in good agreement with the South American one. The 130-Ma mean pole includes the data from the Etendeka volcanics (Gidskehaug et al., 1975) in Namibia. From 90 Ma to Present, the independent behavior of the African plate is clear in relation to the South American plate.

Paleogeographic reconstructions of the South American and African plates at the time of the onset of the PMP can be achieved by gathering the two plates in a pre-drift position and rotating both plates as a unique block by the rotation pole (Table 2) prescribed by the corresponding paleomagnetic pole. The resulting reconstruction (Fig. 8a) shows longitudes that were set by considering (a) that rifting between the two plates was already taking place at southern latitudes (reaching 38°S at about 130 Ma; Nürnberg and Müller, 1991), and therefore the area should be near the present Mid-Atlantic Ridge, and (b) the TC hotspot (taken here as an anchored point in the mantle) must be placed as near as possible to the Paraná Province.

Using the fixed hotspot reference frame (Fig. 8a), the PMP area occupied latitudes of about

Table 2
Rotation poles for South America and Africa

Reference	Age (Ma)	Long. (°)	Lat. (°)	Angle (°)	Description
Mean Early Cretaceous	~130	180.1	0.0	5.7	paleomagnetic rotation
Mean Mid Cretaceous	~110	57.6	0.0	1.5	paleomagnetic rotation
Mean Late Cretaceous	~80	73.9	0.0	9.7	paleomagnetic rotation
Pocos de Caldas Complex	~84	50.1	0.0	6.8	paleomagnetic rotation
Nürnberg and Müller (1991)	131.5	−32.5	50.0	−55.08	AF to SA, pre-drift
Nürnberg and Müller (1991)	126.5	−33.5	50.4	−54.42	AF to SA, anomaly M4
Nürnberg and Müller (1991)	118.7	−35.0	51.6	−52.92	AF to SA, anomaly M0
Nürnberg and Müller (1991)	80.17	−34.0	63.0	−31.00	AF to SA, anomaly AN33 R

5° north of the present position and was not located above TC plume when the Paraná volcanics were being emplaced. The TC plume was well south of the PMP. Even considering a plume head spreading to about 2000 km in diameter (White and McKenzie, 1989), only the southern edge of the PMP would be under the influence of the plume. A distance of about 20° separates the fixed plume from the expected plume centered in the PMP (e.g. Turner et al., 1996). Therefore, plume mobility is required in order to maintain the PMP–TC relationship. Trying to avoid an unrestrained location of the drifting plume, VanDecar et al.'s (1995) plume position in the northeastern PMP can be assumed. These authors interpreted the low-velocity zone identified by seismic tomography as the 'fossil' TC plume that moved with the lithospheric plate. This fossil plume head in the same reconstruction at 130 Ma (Fig. 8b) with a 2000-km-diameter zone of influence reaches most of the PMP (including Etendeka and Angola). However, the southern PMP, where ages are slightly older, is not encompassed by this plume head.

If the plume impacted the plate beneath the northeastern PMP then TC might have migrated southward to reach its present latitude. However, no north–south hotspot trace was left on the eastern border of the South American plate. The oldest trace left by TC is on the oceanic crust, near the African continental shelf, and dates to 113 Ma (O'Connor and Duncan, 1990), where it marks the beginning of the Walvis Ridge, assumed in literature to be an undisputable hotspot trace.

The reconstruction of the two continents for an age, as close as possible to the time of the beginning of the WR construction (Fig. 8c), can be achieved by using the Euler pole given by Nürnberg and Müller (1991), based on the oceanic magnetic anomaly M0 of 118.7 Ma. A subsequent clockwise rotation of 1.5°, predicted by the Mid-Cretaceous paleomagnetic pole (mean of Florianópolis and Cabo de Santo Agostinho poles), was applied to the continental ensemble. Longitudes were fitted close to the Mid-Atlantic Ridge as the sea floor spreading had already began and a narrow oceanic crust separated the southern parts of the two continents. The oldest portion of the Walvis Ridge would soon be formed (113 Ma), and therefore the plume should be at compatible latitudes. Considering that the WR segment of 113 Ma should coincide with TC at the corresponding age, these two features are displaced by about 5°. Therefore, the TC plume did not reach its present location before 80 Ma, as seen in Fig. 8d.

In a Late Cretaceous (~80 Ma) reconstruction (Fig. 8d), Africa is rotated to South America in order to fit magnetic anomaly AN33 R (80 Ma); a rotation pole based on the Late Cretaceous mean paleomagnetic pole was calculated. However, the rotation pole derived from Poços de Caldas Complex (84 Ma) produces a better result in adjusting the ~80-Ma segments of Walvis Ridge and Rio Grande Rise to the TC hotspot. As a result, the ~80-Ma trace of the TC hotspot coincides well with the present position of TC. This implies that, in the last 80 Myr, the TC plume has been fixed relative to the mantle. Coincidentally, in this reconstruction the Trindade plume (plotted in its present coordinates), which is supposed to

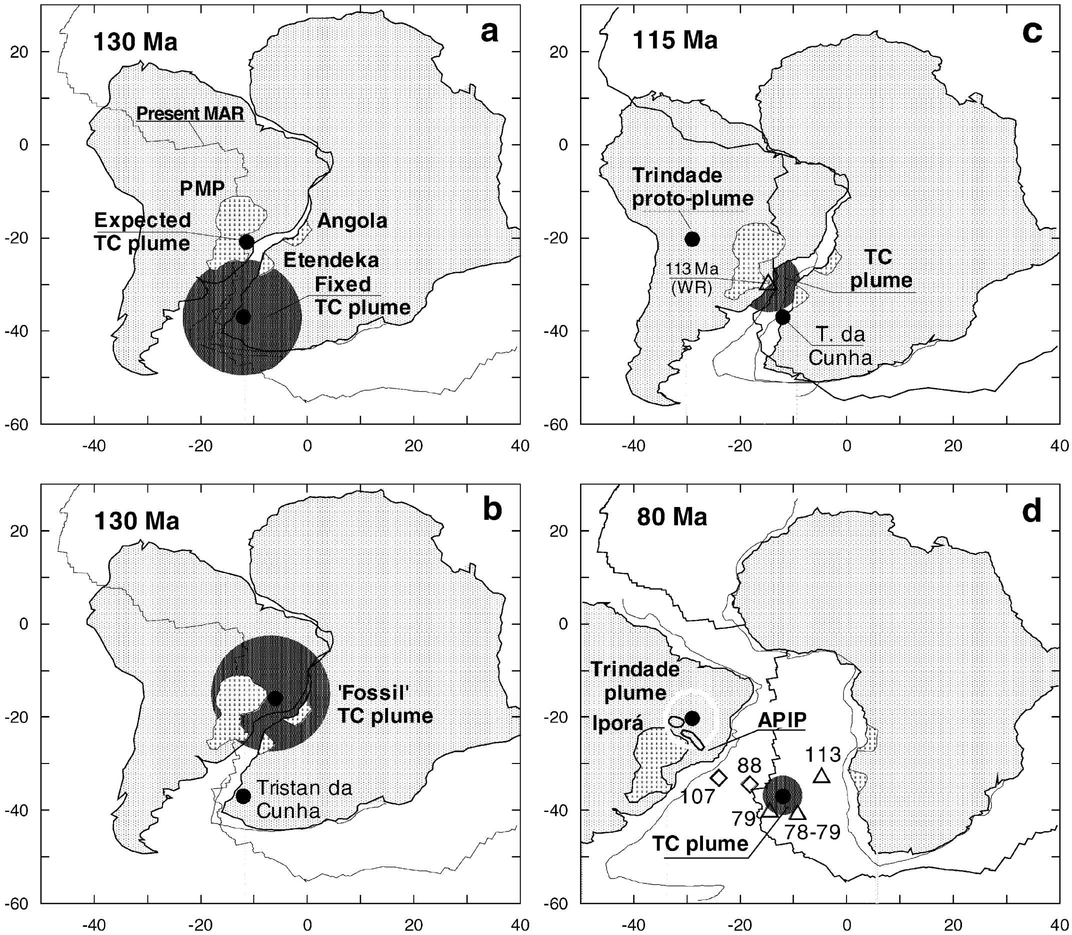

Fig. 8. South American and African plates (Western Gondwana) reconstructed in a pre-drift (130 Ma) configuration during initial stages of the South Atlantic opening (115 Ma) and as two independent lithospheric blocks (80 Ma). Rotation poles as in Table 2. Longitudes are constrained by the Mid-Atlantic Ridge (MAR) and oceanic magnetic anomalies. (a) Relative position of the TC plume (in present coordinates) and the PMP during Early Cretaceous (∼130 Ma), considering the fixed plume model; expected TC plume according to Turner et al. (1996); shaded circle represents the 2,000 diameter plume-head. (b) Same reconstruction as in (a) with the TC plume head coinciding with the location of the 'fossil' plume of VanDecar (1995). (c) Initial stages of oceanic crust; the TC plume migrated southwards and is near the present TC hotspot with a smaller area of influence (shaded circle); the 113-Ma portion of Walvis Ridge (triangle) marks approximately the location of the plume at 115 Ma. Trindade proto-plume corresponds to the present coordinates of Trindade island. (d) At 80 Ma the corresponding trace of Walvis Ridge (WR) and WRGR coincides with the TC hotspot (present coordinates); the Iporá and APIP alkaline provinces were being formed and the Trindade plume is located in the same area.

have originated in the Iporá (80–90 Ma) and Alto Paranaíba (average 85 Ma) alkaline provinces (Thompson et al., 1998), is located beneath those provinces, implying that this plume has also been anchored in the mantle and therefore constitutes a fixed hotspot.

In conclusion, TC behaved as a mobile hotspot from Early (∼130 Ma) to Late (∼80 Ma) Cretaceous, moving more than 20° southward at a rate of about 40 mm/yr. Such velocity is greater than that given by Molnar and Stock (1987) for TC moving in relation to the Hawaiian hotspot. An even higher drifting velocity is required (∼120 mm/yr) if the magmatic activity commenced in the northwest PMP, as proposed by Turner et al. (1996).

6. True polar wander and paleomagnetic reconstructions

Citing Sheth (1999): "If measured paleolatitudes along a hotspot track, which is created by a plume anchored in the deep mantle, are not constant, it is thought by some that the mantle reference frame itself must be moving ('rolling') relative to the Earth's spin axes (true polar wander, TPW)." Thus, depending on the considered time interval, paleomagnetic data would fail in correctly locating large igneous provinces in relation to its correspondent plume head. This could be the case for the Paraná Province, where the TC plume head appears displaced from the various locations required by some authors (O'Connor and Duncan, 1990; Duncan and Richards, 1991; Müller et al., 1993; Turner et al., 1996) based on several different reasonings.

The subject has been debated by many authors (e.g. Gordon and Livermore, 1987; Andrews, 1985). Calculations of the amount of TPW rely on the basic concept of hotspots, that is, that the associated volcanic chains exactly reflect the movement of the lithospheric plates. Therefore, they do not constitute an independent source of information, not from paleomagnetic data itself nor from hotspot traces. Most of the analyses are restricted to the last 100 Ma, when hotspots were more easily traceable. Andrews (1985) presented a TPW path for the past 180 Ma, and found a relative displacement of about 14° (with ~9° of confidence limits) between the hotspot and the geomagnetic frameworks for the 170–131 Ma interval and displacements within 10° for the past 85 Ma. Prévot et al. (2000), to the contrary, concluded that no TPW occurred after 80 Ma, nor from 200 to 150 Ma. A single period of shifting existed between 150 and 80 Ma, culminating at 110 Ma with an abrupt tilting of 20°. It is interesting to note that this large tilting corresponds to the misfit between the PMP and the TC plume at ages close to 130 Ma. Investigations of Mesozoic–Cenozoic TPW are strongly based on the assumption of well-defined hotspot tracks and, therefore, it is not illegitimate to wonder whether such TPW rates are actually a consequence of an ill-postulated problem.

7. The PMP and the plume model paradox

In previous sections it has been demonstrated that the conventional plume model fails to explain the generation of the Paraná continental magmatism, unless several alternative propositions are admitted: (a) the plume contributed with heat, but not with material, in which case it is not a plume model in the classical sense; (b) plumes are not tightly anchored in deep mantle but move in relation to the others at velocities that can exceed 100 mm/yr; (c) if plumes appear displaced from the large igneous provinces, they are supposed to have generated ~500 mm/yr fast (Larsen et al., 1999) laterally spreading plumes that may give an explanation; (d) plumes would be channeled for a few thousand kilometers to the appropriate position (Gibson et al., 1999); (e) paleomagnetic data may be affected by TPW, mislocating the magmatic provinces in relation to the corresponding plumes at the time the main magmatic phase was taking place.

In the case of the Paraná–TC system, these assumptions must be used if the genesis of the PMP is to be kept linked to the plume model, mainly because of the lack of geochemical identity between the PMP and TC. These indicate that the classical mantle plume model can hardly be invoked in this case. Moreover, if this plume really existed before the South Atlantic opening, its tail would have been about 1000 km from the southernmost part of the PMP at ~133 Ma, and only a very small portion of the province would have been affected (Fig. 8). In this case there is no clear justification as to why the plume should not have generated magmatic activity over its entire area of influence. Conversely, the PMP tholeiites do not show significant contribution from both TC (present-day) and MORB sources. The geochemical signature indicates generation from melting of lithospheric mantle reservoirs, considered here to be physically isolated from convecting stirred mantle with different chemical characteristics.

8. Thermal anomalies in the mantle

An alternative source for continental flood basalts and recurrent intraplate magmatism may be indicated by geophysical data of the deep mantle, i.e. from indirect evidence of a long-living thermal anomaly in the mantle. Constraints on plume activity under hotspots or thermal anomalies within the mantle are sought using seismic data. Especially for deep mantle, velocity distribution models based on seismic tomography techniques using both P- and S-waves have improved in resolution in the last decade (e.g. Zhang and Tanimoto, 1992; Zhou, 1996; Li and Romanowicz, 1996; vanderHilst et al., 1997).

Another source of information on the thermal state of the Earth's interior is the geoid. Geoid anomalies are defined as the difference in geoidal height between the measured geoid and some reference model (Anderson, 1989). Bowin (1991) proposed the use of a truncated geopotential model, typically at degree 10 or degree 12, as the reference model, and showed that the geoid anomalies so obtained have a good correlation with major oceanic structures. For the EGM96 (Rapp and Lemoine, 1998) geopotential model, lateral resolution may be as good as 50–60 km if the model representation in spherical harmonics is expanded up to degree 360. Therefore, the next step was to try to find evidence in both geoid and seismic data for a thermal anomaly which could have provided heat to initiate the magmatic activity in South Atlantic.

The geoid residual anomalies for the South Atlantic, including the continental lithosphere of South America and Africa (Fig. 9), has been obtained from the separation of the EGM96 geoid, expanded up to degree 10, which accounts for mass anomalies in the lowermost part of the mantle, from EGM96 fully expanded to degree 360. This map is overlaid by a shaded-relief map of the $5' \times 5'$ digital topography/bathymetry. The most important residual geoid anomalies clearly correlate with major lithospheric scale features, such as a belt of subtle positive anomalies over the mid-ocean ridges, negative anomalies over deep oceanic basins, thick cratons and, more importantly, over the Paraná Province.

Two conspicuous intermediate-wavelength positive anomalies are observed over the South Atlantic. The first starts in the equatorial transform fault system which separates the Central and South Atlantic, obliquely cross-cuts the ridge axis, and passes over the Ascencion and St. Helena hotspots and the Walvis Ridge. The second one extends from southern Brazil, continues over the Rio Grande Rise and links with the Antarctic Ridge System. The extent of these anomalies indicates that their source must lie within the mantle. In order to set further constraint on the depth of mass anomalies, this geoid map has been compared with the P1200 seismic tomography model of Zhou (1996), which provided the velocity distribution for the mantle between 165 and 1200 km within cells with a nominal resolution of 1°. Positive residual geoid anomalies have a good correlation with low-velocity regions in the mantle. For the first 150-km upper mantle interval, Tanimoto and Zhang (1992) proposed an S-wave seismic tomographic model in which it is shown that the mid-ocean ridges are characterized by a focussed and shallow (~100 km) depth interval of low-velocity distribution. This may explain why geoid anomalies are also not very strong over these features, except in those places where an extra deep-seated mantle thermal anomaly is present, as it is the case for those two plate scale positive geoid anomalies. In the P1200 model, these two large positive anomalies correlate with a low-velocity zone mapped within the mantle, extending from 165–210 km (see Zhou, 1996, plate 1a) to 660–710 km (plate 1f). Below this depth, the low-velocity zones almost disappear.

Anderson et al. (1992) also observed that the upper mantle is characterized by vast domains of high temperatures rather than small regions surrounding hotspots, and that low-velocity anomalies record previous positions of migrating ridges, or in Tanimoto and Zhang's (1992) view, mark the place where the Western Gondwana break-up occurred. Whatever the interpretation, this is a clear indication that this anomaly persisted for over 100 Ma and that the existence of long-living deep-mantle thermal anomalies is very likely.

9. Discussion

When the South American plate is plotted in

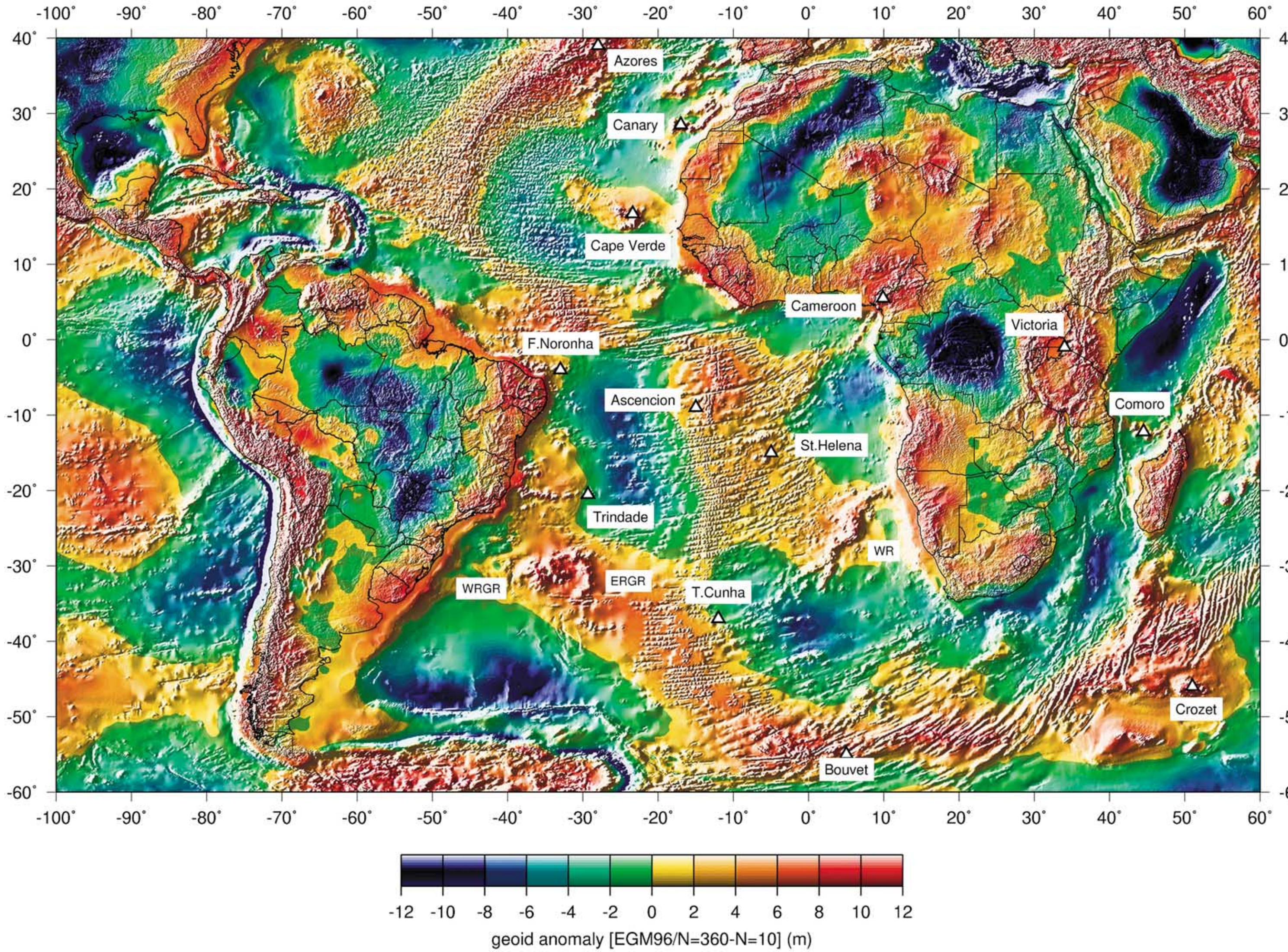

Fig. 9. Geoid anomalies for central South Atlantic and South American and African continental lithosphere based on the EGM96 (Rapp and Lemoine, 1998) geopotential model expanded up to degree 360 minus EGM96 expanded up to degree 10 shown in color (interval 2 m), overlaid by the shaded-relief map of topography (illuminated from NW) and bathymetry (30″ resolution). Anomalies in meters.

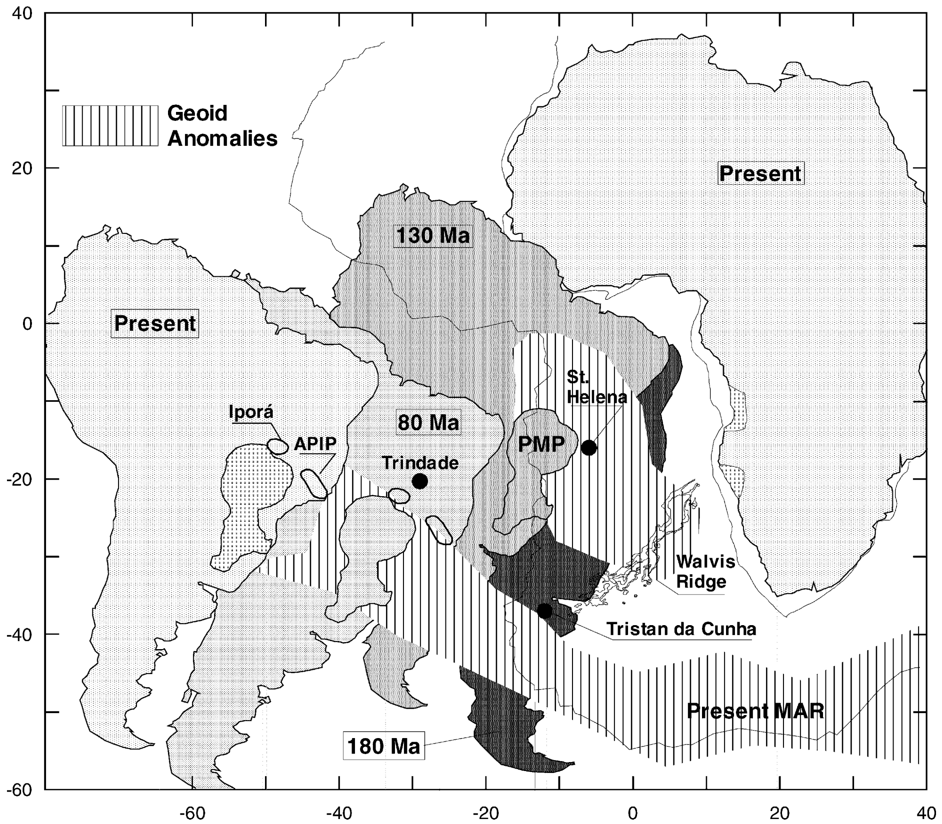

Fig. 10. Superposition of the various reconstructions of South America since Early Jurassic. Africa and Walvis Ridge in present coordinates. Striped areas correspond to the geoid anomalies of Fig. 9.

successive paleopositions calculated in 8. Thermal anomalies in the mantle (Fig. 8), we see that for ~130 Ma, the Paraná Province was above the NW–SE geoid/mantle thermal anomaly located on the western African plate (Fig. 10). This is in accordance with the observation and suggestion of Tanimoto and Zhang (1992), as discussed previously. From the Early Jurassic to Early Cretaceous (Fig. 10), the South American plate occupied similar paleolatitudes (Besse and Courtillot, 1991; Courtillot et al., 1999; Ernesto et al., 2001). Therefore, the area corresponding to the PMP remained quasi-stationary over a mantle thermal anomaly for about 50 Ma (from approximately 180 to 130 Ma). This elapsed time was long enough for the lithosphere to incorporate the necessary heat to produce voluminous tholeiitic melts when suitable tectonic conditions developed. Considering that the Paraná Basin occupies mainly cratonic areas, involving heterogeneous lithospheric domains surrounded by mobile belts, models such as those proposed by Anderson (1994a), King and Anderson (1998) and Tackley (2000) can explain the genesis of the Paraná magmatism. This tectonic situation also favors the mechanism envisaged by King and Ritsema (2000) of edge-driven convection of the upper mantle due to lateral contrasting physical conditions (temperature and viscosity) beneath the craton edges.

A low-velocity region mapped on the African side of the South Atlantic by Zhou (1996) has, on average, P-wave velocity perturbation of 0.5%. If this velocity decrease is due entirely to a thermal anomaly, the mantle temperature in this region would be ~100°C higher than the surrounding mantle. Temperature perturbations of 200°C are enough to trigger volcanism and large volume of melts in lithosphere under extension (White and McKenzie, 1989). Since the break-up and opening of South Atlantic, 120 Ma have passed, so part of the heat has been dissipated and the temperature perturbation may have decreased since then. It may also be possible that this thermal anomaly is part of the edge-driven convective system at the border of the western African continental lithosphere (King and Ritsema, 2000), whereby the low velocity region indicates the region of mantle upwelling and high velocity region under the cratonic area and the downwelling flow of mantle material.

Observing the movement of the South American plate from 130 to 80 Ma (Fig. 10), the main drift component was almost parallel to the Walvis volcanic lineament. This strengthens the hypothesis previously raised by some authors (e.g. Popoff, 1988; Unternehr et al., 1988) that the Walvis Ridge and the western portion of the Rio Grande Rise represent the accommodation of stresses during the initial opening of the South Atlantic. According to the authors cited, this zone has its prolongation within the South American plate between the Campos and Pelotas basins as a dextral deformation. The eastern sector of the Rio Grande Rise is also in accordance with the change of South American movement at about 80 Ma when a north–south drifting component is apparent, as already pointed out by LePichon and Hayes (1971). Similar interpretation was given by Ferrari and Riccomini (1999) for the Vitória–Trindade chain, believed to be the Trindade–Martin Vaz hotspot trace.

The strong geoid anomaly that is located over the Rio Grande Rise and continues along the mid-ocean ridge is probably responsible for the recurrent volcanism in the western sector of this rise. The same anomaly continues toward the South American eastern continental margin, describing a path that nearly coincides with the plate drift path itself. Therefore, the alkaline magmatism that developed in the time interval 90–70 Ma bordering the Paraná Province does not necessarily need an associated plume to be explained because, as in the case of the PMP, all the affected region was (and part of it still is) over a mantle thermal anomaly. The Trindade–Martin Vaz hotspot (Thompson et al., 1998), presently located eastward in the oceanic crust, was close (Figs. 8 and 10), if it already existed, to the alkaline provinces of Iporá and Alto Paranaíba at the time these provinces were formed. However, the St. Helena hotspot (Figs. 9 and 10) was once close to the PMP (at ~130 Ma), and there is no attempt in the literature to associate this plume with the Paraná Province, except that the area where this plume was supposed to be is practically the same as where VanDecar et al. (1995) proposed a seismic low-velocity zone. Therefore, some of the volcanic islands, if not all, in the South Atlantic that are currently seen as hotspots may have a very recent history, rather than represent the final stage (tail) of a mantle plume. It is important to note that both intraplate and near-plate-boundary hotspots in South Atlantic, as defined by King and Ritsema (2000), can be explained by the edge-driven convection involving the mantle as deep as the transition zone (600 km).

The hypothesis, raised here, of large, long-living thermal anomalies in the mantle, capable of transferring sufficient heat to a segment of lithosphere that stays stationary or quasi-stationary over it can also be used to explain the Tertiary magmatism in Ethiopia (Courtillot et al., 1999; George et al., 1998) that is associated with a rift process under development. $^{40}Ar/^{39}Ar$ ages indicated the existence of two distinct magmatic phases at 45–35 Ma and 19–12 Ma in southern Ethiopia, whereas in northern Ethiopia eruptions took place in the 31–29-Ma interval (George et al., 1998, and references therein). In order to explain these age distributions, a two plume model (the Kenyan and Afar mantle plumes) was proposed (George et al., 1998). The area where the flood basalts occur, the Ethiopian plateau (Mohr and Zanettin, 1988; Piccirillo et al., 1979), is not marked by important geoid anomalies nor by low-velocity zones deep in the mantle (Zhou, 1996). However, a prominent positive geoid anomaly is noted to the south, near Lake Victoria, and centered approximately at the equator, where no expressive magmatism has been reported. The reconstruction of the African plate at 80 Ma (Fig. 8d) indicates that the paleolatitudes of this rift region was about 10° southward, coinciding with the geoid anomaly of about 50 Myr ago, based on the apparent polar wander path (Fig. 7) calculated by Besse and Courtillot (1991). Therefore, the area where the Ethiopian basalts were emplaced had been under the influence of this thermal anomaly for at least 20 Myr, based on drifting velocities inferred from the same polar path.

10. Concluding remarks

(1) The geochemical and Sr–Nd–Pb isotopic data do not support the tholeiites from Walvis Ridge, Rio Grande Rise and the PMP having resulted from mixing dominated by the TC plume and MORB components.

(2) The similarity among the high-Ti basalts from Rio Grande Rise (site 516F), part of Walvis Ridge (site 525A) and the Paraná analogues suggests that delaminated subcontinental lithospheric mantle has to be considered in their genesis.

(3) Paleogeographic reconstructions of the Paraná–TC system, assuming this hotspot is a fixed point in the mantle, indicates that the TC plume was located $\sim$800 km south of the PMP. Therefore, plume mobility would be required in order to maintain the PMP–TC relationship.

(4) Assuming that TC was located in the northern portion of the PMP ($\sim$20° from the present TC position), the plume migrated southward from 133–132 Ma (main volcanic phase in the area) to 80 Ma at a rate of about 40 mm/yr. From 80 Ma to Present the plume remained virtually fixed, leaving a track compatible with the African plate movement. Notably, the southward migration of the plume is in opposition to the northward migration of the main Paraná magmatic phases (133 Ma in the south, and 132 Ma in the north).

(5) Regional thermal anomalies in the deep mantle, mapped by geoid and seismic tomography data, offer an alternative non-plume-related heat source for the generation of intracontinental magmatic provinces.

(6) The 'hotspot tracks' of Walvis Ridge and Rio Grande Rise as well as the Vitória–Trindade chain might reflect the accommodation of stresses in the lithosphere during rifting rather than continuous magmatic activity induced by mantle plumes beneath the moving lithospheric plates.

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