Tectonophysics, 212 (1992) 59–76
Elsevier Science Publishers B.V., Amsterdam

Analysis of MAGSAT magnetic contrasts across Africa and South America

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(Received June 20, 1990; revised version accepted July 31, 1991)

ABSTRACT

Ravat, D.N., Hinze, W.J. and von Frese, R.R.B., 1992. Analysis of MAGSAT magnetic contrasts across Africa and South America. In: R.R.B. von Frese and P.T. Taylor (Editors), Lithospheric Analysis of Magnetic and Related Geophysical Anomalies. *Tectonophysics*, 212: 59–76.

Comparisons of MAGSAT magnetic contrasts and geology across the Mesozoic assembly of Africa and South America provide new insight into the interpretation of the long-wavelength magnetic anomalies near the present continental margins. Across continental Africa and South America, the MAGSAT magnetic contrasts can be correlated with geologic provinces formed before the Mesozoic separation of the continents. On the continents, areas affected by significant Mesozoic hotspot tectonism display negative magnetic contrasts suggesting a causative relationship between hotspot tectonism and the origin of the observed magnetic contrasts. The magnetic characteristics of a portion of the lower crust in these areas appear to have been significantly altered during the Mesozoic hotspot epeirogeny. By analogy with the processes of continental rifting, it is suggested that the magnetic mineralogy of the intruded lower crust may be dominated by weakly- to non-magnetic titanomagnetites.

Oceanic magnetic contrast comparisons show that the positions of magnetic anomaly highs over the Rio Grande–Walvis Ridge System of the South Atlantic are consistent with the interpretation of their evolution over the Walvis Hotspot. However, only the parts of these ridges that were formed during the Cretaceous normal polarity geomagnetic epoch bear strong magnetic contrasts. Remanent magnetization thus appears to be an important contributor toward the MAGSAT magnetic anomalies of these features.

Introduction

Limited anomaly and source resolution as well as deficiencies in our knowledge of magnetic properties of deep crustal and upper mantle rocks significantly affect geologic interpretation of long-wavelength magnetic anomalies observed by satellites. Although it has been shown that long-wavelength magnetic anomalies originate from geologic and tectonic units in the earth's crust (Regan et al., 1975; Wasilewski et al., 1979; Frey, 1982; von Frese et al., 1982, 1986; Arkani-Hamed and Strangway, 1985a; LaBrecque and Raymond, 1985; Mayhew et al., 1985; Johnson, 1985; Hayling and Harrison, 1986; Taylor and Frawley, 1987; Thomas, 1987; Bradley and Frey, 1988; Fullerton et al., 1989) and possibly upper mantle (Arkani-Hamed and Strangway, 1985b, 1986; Haggerty and Toft, 1985; Toft and Haggerty, 1986), the above limitations pose difficulties in extracting information on the origin, nature, and evolution of regional magnetic sources, the principal objectives of long-wavelength magnetic mapping.

Interpretation of the magnetic anomalies near continental margins may be facilitated by considering the magnetic patterns and geology on the opposing margins of rifted continents, especially if pre-rift magnetic patterns can be isolated from

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0040-1951/92/$05.00

magnetic effects of rift-stage and post-rift tectonic manifestations. In the past, Frey et al. (1983), Galdeano (1983), and von Frese et al. (1986) have used regional magnetic anomaly data from POGO and MAGSAT missions to show the consistency of the long-wavelength magnetic anomalies across the Pangean assembly of the continents. Recent data processing improvements (e.g., Arkani-Hamed and Strangway, 1985a; Ridgway and Hinze, 1986) permit more detailed comparisons between the MAGSAT anomalies and their respective regional geologic sources. In addition, the equatorial regions require special processing strategies; these include corrections to equatorial ionospheric effects (Yanagisawa and Kono, 1985; Ravat, 1989) and equivalent source matrix inversions that allow computation of stable radially-polarized (reduced-to-pole) magnetic anomalies as well as effective susceptibility variation of the lithosphere (Langel et al., 1984; von Frese et al., 1988; Ravat et al., 1991). Thus, with these data processing improvements, it may be possible to isolate magnetic anomalies of pre-rift geologic provinces (e.g., continental shields) from rift-stage and post-rift geologic sources.

In this study, we investigate additional information on the nature and origin of the MAGSAT magnetic anomalies by comparing geology of the adjoining portions of the African and South American continents. The principal assumption made here is that magnetic anomalies (or sources) will be similar over geologic provinces that have been disrupted from their previous continuity on Pangea during the rifting process, unless the magnetic anomalies (or sources) have been disturbed by later geologic events on one or both of the lithospheric plates. The validity of this assumption across adjoining portions of the continents can be qualitatively tested by comparing magnetic anomalies (or sources) with regional tectonic provinces and by reviewing their geologic history. It is not expected that this assumption will be met in every instance (e.g., in comparing two different shield areas) because sources of long-wavelength magnetic anomalies are sensitive to various parameters controlling the rock magnetic properties (lithology, oxidation state, temperature, and changes in these parameters through the geologic history) as well as variations in structural differences and past and present lithospheric temperatures. In addition, the observed magnetization variation depends on magnetic properties of surrounding tectonic provinces, and therefore, neither the sign nor the amplitude of the magnetization variation can be easily related directly to tectonic provinces. Despite these complexities, there is overwhelming evidence that the African and South American tectonic provinces are associated with characteristic magnetic anomaly patterns (Frey, 1982; Hastings, 1982; Ridgway and Hinze, 1986; Ravat, 1989). For example, after reduction-to-pole, the long-wavelength magnetic maps of the African and South American lithospheric plates show that continental shields are generally magnetically positive, rifts and basins are magnetically negative, and oceanic plateaus are magnetically positive with respect to surrounding ocean basins (e.g., see Figs. 1 and 2). Thus, the correlation, or lack thereof, of magnetic anomalies across the lithospheric plates pair

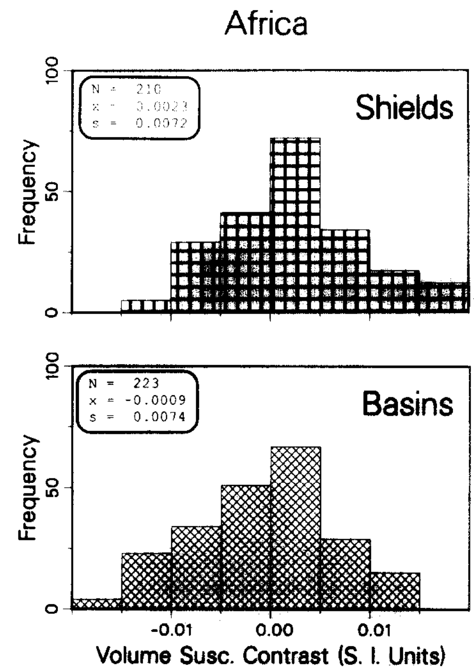

Fig. 1. Comparison of the volume magnetic susceptibility contrasts of the shields and basins in Africa. N = number of points; x = mean; s = standard deviation. The arithmetic means of the two distributions are statistically significantly different from each other.

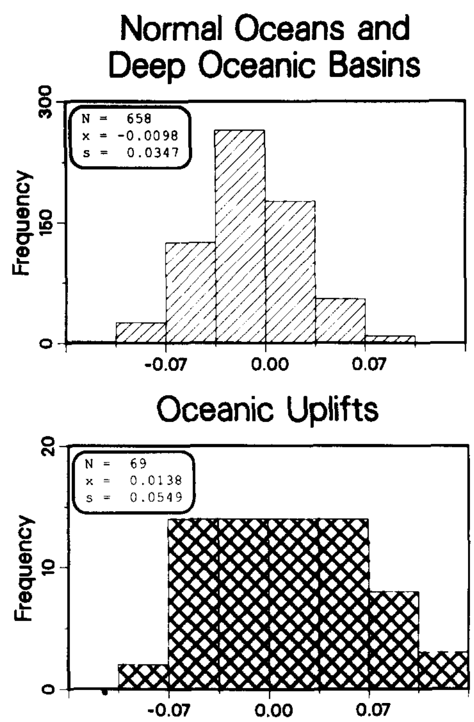

Fig. 2. Comparison of the volume magnetic susceptibility contrasts of the oceanic uplifts and the normal oceans and basins in the Atlantic and Southwest Indian Oceans. N = number of points; x = mean; s = standard deviation. The two distributions are statistically significantly different from each other.

geomagnetic field over a geographically large region is removed by this process. However, the inherent assumption of the inversion process, that most magnetic sources observable at satellite altitudes are induced and/or viscoremanently magnetized (Wasilewski et al., 1979) may not strictly hold for all geologic sources. Moreover, potential problems in total intensity magnetic anomalies, such as any residual ionospheric field effects, will be overly emphasized in radially-polarized magnetic anomalies. In spite of these drawbacks, radially-polarized magnetic anomalies and effective susceptibility contrasts are employed here because of the ease of their use in directly comparing the geologic source regions and the magnetic anomaly sources.

We make detailed comparisons between the magnetic contrasts and geologic provinces across the African and South American continents to better understand the origin of magnetic anomalies near the continental margins, consider the modifying effect of hotspot magmatism on the pre-rift continental anomalies, and investigate the origin of magnetic contrasts associated with the oceanic Rio Grande–Walvis Ridge System that evolved under the influence of one of the prominent hotspots in the South Atlantic Ocean. To

and its association with geology is capable of providing information on the likely origin of the magnetic sources.

Hotspot tectonism has played an important role during the Mesozoic breakup of western Gondwana (Burke and Dewey, 1973; Duncan, 1981; Morgan, 1983; White and McKenzie, 1989; Duncan and Richards, 1991). The associated voluminous magmatism is regionally extensive and is likely to compositionally alter parts of the crust. Thus, it is necessary to consider the effect of this magmatism in understanding the origin of magnetic contrasts across the continental assembly.

In an anomaly–source correlation study such as this, conversion of total intensity magnetic anomalies into the radially-polarized magnetic anomalies or susceptibility variation is important because the variability in the total intensity magnetic anomalies due to changing attributes of the

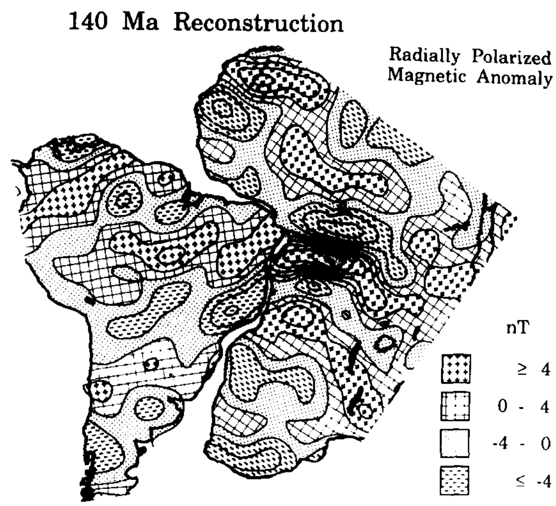

Fig. 3. The radially-polarized MAGSAT anomalies of Africa and South America rotated to the 140-Ma reconstruction based on Smith et al., 1981. Rigid plate rotations are shown because match of only the conjugate coast-lines of Africa and South America is pertinent to this study.

facilitate the comparisons, radially-polarized magnetic anomalies and effective susceptibility contrasts (Figs. 3 and 4; for methodology, see von Frese et al., 1982, 1988; Ravat et al., 1991) are shown on the reconstructed positions of Africa and South America at 140 Ma (reconstruction from Smith et al., 1981). Both representations are useful because the magnetic signatures of individual geologic and tectonic provinces are better isolated on the susceptibility variation mapping, whereas the correlation of the magnetic sources across the rifted margin is aided by the smooth variations in the high-altitude anomalies. In Figure 5, major tectonic features in Africa, South America, and South Atlantic Ocean are shown that are pertinent to this study and Figure 6 shows the susceptibility contrasts over the corresponding region.

Magnetic contrasts of continental shields

There are numerous geologic links between Brazil and Africa (Hurley et al., 1967; Herz, 1977; Torquato and Cordani, 1981, and references therein) in accordance with Wegener's (1912) continental drift hypothesis. However, with the exception of large geologic age provinces, most of the geologic evidence is in the form of structural belts that have nearly north–south orientation (e.g., Paraguay–Araguaia belt, Brasiliano age (550–1200 Ma) belt in the eastern São Francisco Craton, Ribeira belt in South America, and Rockelide belt, Pan-African belts of Nigeria, West Congolian belt in Africa; see fig. 10 of Torquato and Cordani, 1981). Because of the low present-day magnetic inclination in the region, this predominantly N–S-trending structural evidence will not be magnetically observable. The problem of the relatively small, north–south features is exacerbated because of the high orbital inclination of the MAGSAT tracks (nearly north–south).

Within the southern West African Craton, the distinction between the Archean (> 2500 Ma) and the Eburnian age (~ 1850 ± 250 Ma) provinces is readily apparent (Fig. 3, and Mn in Fig. 5). The Eburnian age province in the West African Craton displays a radially-polarized magnetic anomaly low which can be traced into northeastern Brazil in the Transamazonian age (~ 2000 Ma) São Luiz Craton (SLC in Fig. 5). The São Luiz Craton is a small remnant of the West African Craton that was separated from its African counterpart during the Mesozoic separation of the continents (Hurley et al., 1967;

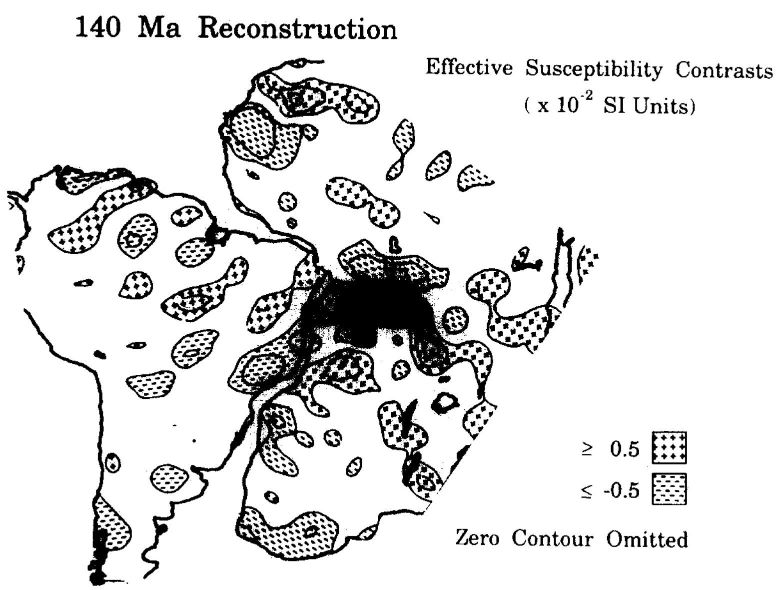

Fig. 4. The effective susceptibility contrasts of Africa and South America rotated to the 140-Ma reconstruction based on Smith et al., 1981.

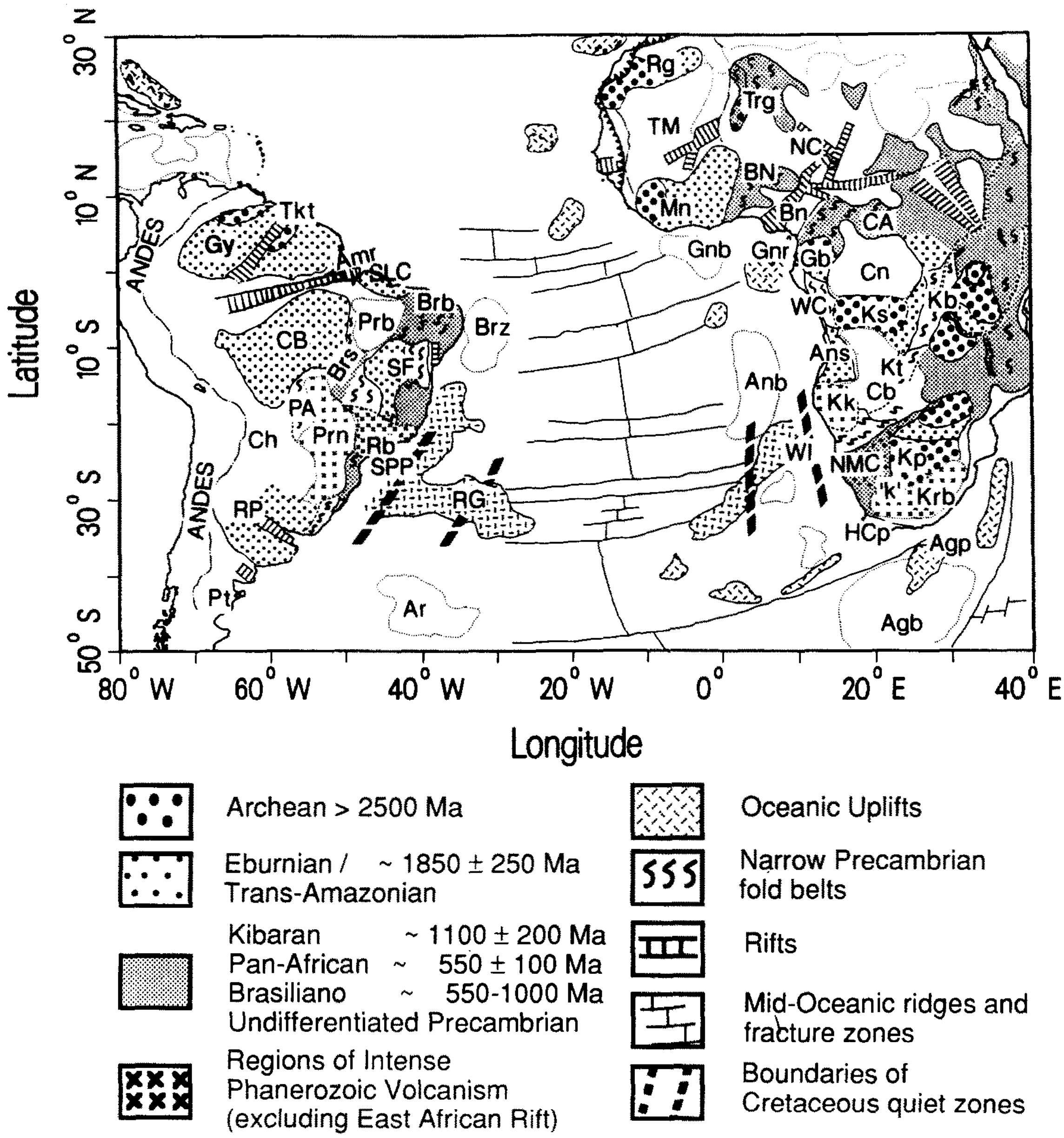

Fig. 5. Major tectonic features of Africa, South America, and the South Atlantic Ocean with tectonic provinces identified. *Africa: HCp* = Hercynian Cape Foldbelt; *Rg* = Reguibat Rise; *Mn* = Man Shield; *Trg* = Tuareg Shield; *BN* = Benin Nigeria Shield; *CA* = Central African Shield; *Gb* = Gabon Nucleus; *Ks* = Kasai Nucleus; *WC* = West Congolian Belt; *Ans* = Angola Shield; *Kb* = Kibarides; *Kt* = Katangan Orogeny; *NMC* = Namaqua Metamorphic Complex; *Kp* = Kaapvaal Craton; *TM* = Taoudeni–Mali Basin; *NC* = Niger–Chad Basin; *Bn* = Benue Rift; *Cn* = Congo Basin; *Cb* = Cubango Basin; *Krb* = Karroo Basin; *Kk* = Kaokoveld Province; *'k'* = Kalahari line. *South America: Gy* = Guyana Shield; *CB* = Central Brazilian Shield; *SLC* = São Luiz Craton; *SF* = São Francisco Craton; *RP* = Rio de la Plata Craton; *Pt* = Patagonian Platform; *Brb* = Borborema Province; *Brs* = Brasiliano Foldbelt; *PA* = Paraguay–Araguaia Foldbelt; *Rb* = Ribeira Foldbelt; *Tkt* = Takatu Rift; *Amr* = Amazon Rift and Basin; *Prb* = Paranaiba Basin; *Prn* = Parana Basin; *Ch* = Chaco Basin. *South Atlantic: Ar* = Argentine Basin; *Brz* = Brazil Basin; *Gnb* = Guinea Basin; *Anb* = Angola Basin; *Agb* = Agulhas Basin; *RG* = Rio Grande Ridge; *SPP* = São Paulo Plateau; *Gnr* = Guinea Rise; *WI* = Walvis Ridge; *Agp* = Agulhas Plateau. On oceans, the bathymetry approximately > 5000 m is considered basins while the bathymetry approximately < 2500 m is considered plateaus (except in the South Atlantic where the depth anomaly contour of 750 m (McClain, 1984) is used to outline the plateaus. The approximate boundaries of the Cretaceous quiet zones near Rio Grande Rise and Walvis Ridge are shown with heavy dashed lines..

Torquato and Cordani, 1981). The negative magnetic anomalies continue westward along the Brazilian coastal provinces of the Transamazonian age, where the magnetic low may be further enhanced by the possible thinning of the magnetic crust in the Amazon Rift (Amr) in comparison to the Guyana (Gy) and Central Brazilian Shields (CB) (Hurley et al., 1967; Longacre et al., 1982).

The positive radially-polarized magnetic anomalies over the western São Francisco Craton (SF) and the Borborema province (Brb) at the northeastern tip of Brazil appear to merge at the high observation altitude (Figs. 3 and 5). However, susceptibility variation mapping (Fig. 4) suggests that sources of these anomalies are isolated from each other. The Brasiliano age (~ 550–1000 Ma) Borborema province in Brazil and the Pan-African age (~ 550 Ma) province in Nigeria and Cameroon are structurally similar; however, the magnetic contrasts over these provinces are opposite. While it is not possible to entirely rule out a fundamentally different nature for these two terrains, it is more likely that the African provinces were significantly affected by later large-scale tectonic activity that has a prominent magnetic signature. Indeed, geologic and geophysical data over the African portion suggest that this region has been involved in a prominent Mesozoic rifting episode (Burke and Dewey, 1973; Fairhead, 1988). Moreover, thermal effects of the Cenozoic hotspot activity in this region may have contributed further toward thermal demagnetization of the magnetic crust (Arkani-Hamed and Strangway, 1985b). Thus, it appears that the low magnetic contrasts on the African side may be associ-

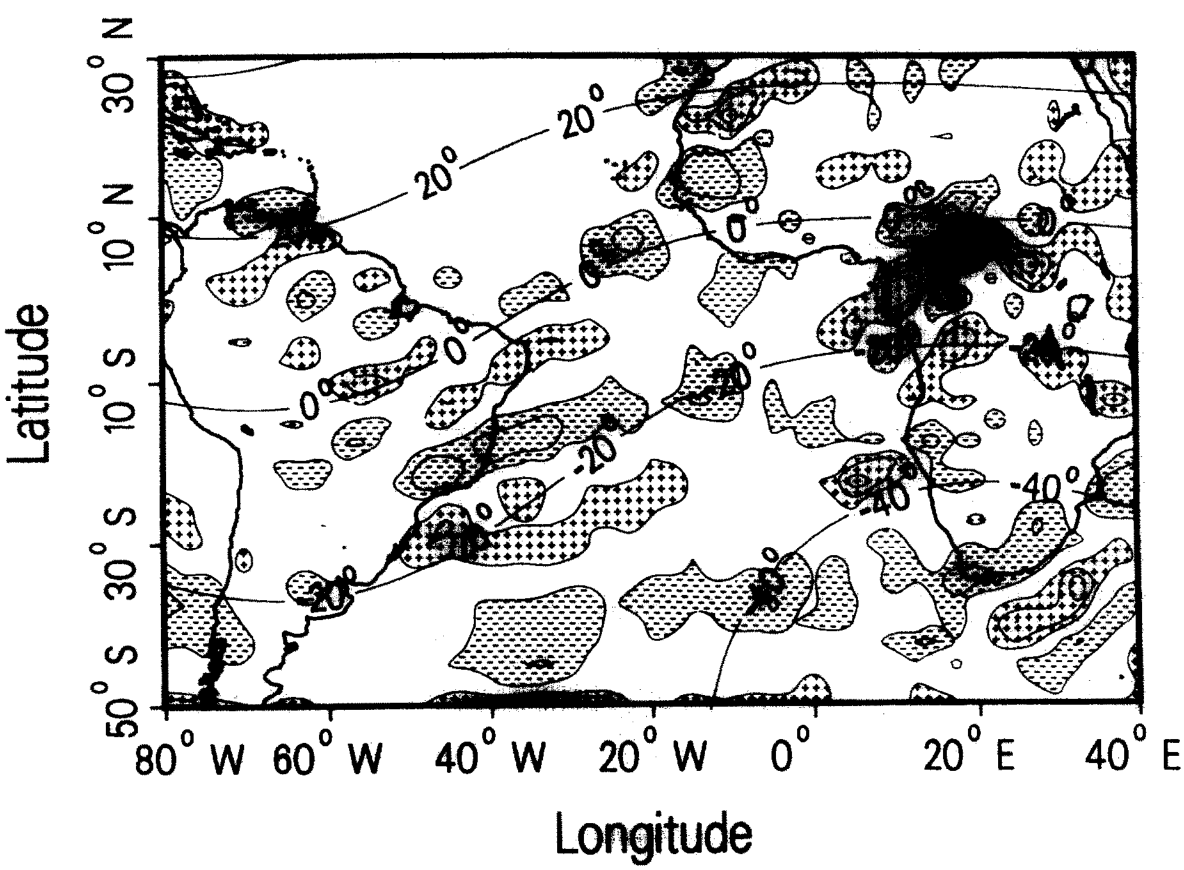

Fig. 6. Effective susceptibility contrasts of the 'crust'. The continental susceptibility contrasts are normalized to a 40-km thick 'crustal layer', whereas the oceanic susceptibility contrasts are normalized to a 6-km thick 'crustal layer'. Cylindrical Equidistant Projection. Continental C.I. 5×10^{-3} (in SI units). Oceanic C.I. 3×10^{-2} (in SI units). Zero contour omitted. Isoclines are also shown at C.I. 20°. Positive susceptibility contrasts are patterned by +, whereas negative susceptibility contrasts are patterned by −.

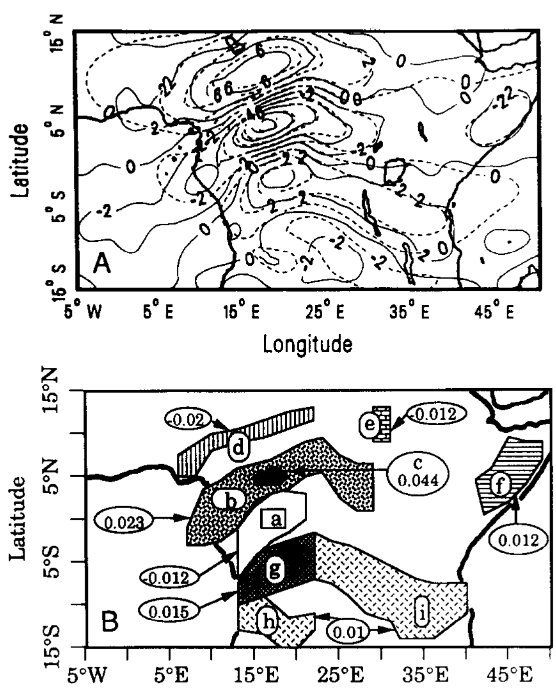

Fig. 7. (A) The observed total intensity (continuous) and computed (dashed) MAGSAT anomalies of central Africa computed from the model in (B). Cylindrical Equidistant Projection. C.I. 2 nT. (B) A three-dimensional magnetic model of central Africa. Numbers inside the ellipses are effective susceptibility contrasts (in SI units) for 40-km thick polygons. The susceptibility contrasts of the polygons should be proportionately inflated if the interpreted thickness of the polygon is lower than 40 km. Cylindrical Equidistant Projection. a = Congo Basin; b = Central African Shields; c = Bangui; d = Benue Rift; e = Central African Rift; f = Somali Basin (Shield); g = Kasai Nucleus; h = Angola Shield; i = cannot be constrained or explained by geologic or geophysical data, but the source is required to model the magnetic anomalies.

ated with the magnetic and structural modifications of the crust that have taken place during Mesozoic rifting and later tectonic activity. Using constraints from gravity and seismic evidence, the low magnetic contrasts in Nigeria and Cameroon can be consistently modeled as ~ 12 km of thinned lower crust (Stuart et al., 1985; Fairhead and Okereke, 1988; Fairhead, 1988) or a non-magnetic basal crustal pillow (von Frese et al., 1981) that is associated with the western portion of the West African Rift System (Figs. 3, 5, and source d in Fig. 7; see Ravat, 1989, for the geologic and geophysical constraints on the model). In the eastern portion of the West African Rift System, where there is no evidence of a surface graben, we interpret an intruded, more ductile, and non-magnetic basal lower crust. This interpretation is not only consistent with the regional magnetic anomalies and models (at ~ 10°N in Fig. 7), but the rift-related basal crustal intrusions also could have facilitated the large-scale Mesozoic wrenching and deformation of the Central African Shear Zone discussed by Fairhead (1988).

South of the West African Rift System, the high-amplitude Bangui anomaly in the Central African Shield (CA) and its side-lobes to the north and south (Figs. 3 and 5) dominate most of the magnetic anomaly signal in this region. In contrast to previous interpretations of the Bangui anomaly (Regan and Marsh, 1982; Hastings, 1982), we interpret about half of the central portion of the Bangui anomaly to originate from a highly concentrated, possibly near-surface magnetic source such as buried Precambrian iron formations (source c in Fig. 7; see Ravat, 1989, for the geologic and magnetic justifications of the model construction); the remaining portion of the modeled high magnetic contrast zone is constrained by the boundaries of the Central African Shield and the Gabon Nucleus (Gb) (source b in Fig. 7). In the Gabon Nucleus in central Africa and the São Francisco Craton (SF) in eastern Brazil (Fig. 5), the basement rocks are remobilized Archean terrane with imprints of Eburnian–Transamazonian age activity (Clifford, 1970; de Almeida et al., 1981). In view of the similarity of their evolutions, the positive magnetic anomalies over the western São Francisco Craton and the Gabon Nucleus (western portion of the Central African Shield) appear to be consistent. The eastern São Francisco Craton consists of nearly north–south structural belts of predominantly Brasiliano age (550–1100 Ma; fig. 7 in de Almeida et al., 1981). Although we would like to believe that the resolution of our maps can discern the disruption (due to the Brasiliano age belts) of the anticipated magnetic anomaly continuity between the western stable São Francisco Craton and the Gabon Nucleus in Africa, we cannot verify that the magnetic anomalies are truly disrupted by the Brasiliano age deformations.

Moving southward, the magnetic anomaly low that extends from northeast of the Parana Basin (Prn) (mid-Paleozoic) towards the Atlantic coast of South America in an east-northeast direction (at 20°S, 45°W in Fig. 4) has been difficult to interpret because the low crosses several tectonic provinces within South America (Parrot, 1985). Although a similar magnetic low occurs over the Congo Basin (Cn) (also mid-Paleozoic) in Africa that can be visually correlated across the continental assembly (Figs. 3 and 5), the structural style and the older ages of intervening provinces (N–S-trending Brasiliano–Pan-African age Ribeira (Rb) and West Congolian (WC) foldbelts) suggest that the origin of the sources of these anomalies may not be related. An alternate interpretation of the South American ENE-trending continental magnetic low is discussed in the next section.

The positive magnetic anomaly over the Angola–Kasai Craton (Ans–Ks) in Africa (Figs. 3 and 5, and sources h and g in Fig. 7) has no Brazilian counterpart; however, no corresponding "stable" region of Eburnian or older age has been detected near Rio de Janeiro in Brazil (Torquato and Cordani, 1981).

Farther south, Reeves (1985) has interpreted a 1800-Ma suture zone between Archean rocks of the Kaapvaal Craton (Kp) to the east of 22°E and the Namaqualand granites and gneisses of middle and upper Proterozoic age rocks to the west (NMC). Within the suture zone, the so-called "Kalahari line", the metamorphosed sedimentary and volcanic rocks of the Kheis group have been dated at 1800 Ma in the southern portion, whereas about the same age has been obtained on granitic gneisses in the northern part of the line (Cornell, 1977; Key and Rundle, 1981; Reeves, 1985). The aeromagnetic and gravity anomaly levels are higher to the west of the suture and their longer-wavelength character suggests deeper, more mafic sources (Reeves, 1985). Although the Kalahari line itself is not well-resolved on the MAGSAT data (Figs. 3 and 5), the high magnetic anomaly level to the west of the line is reflected in the MAGSAT maps and can be modeled accordingly

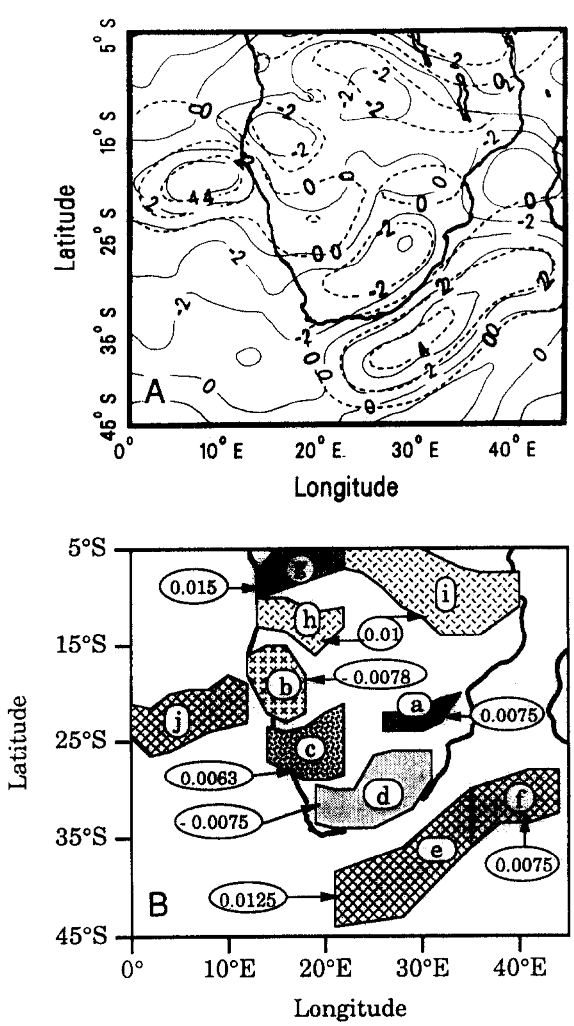

Fig. 8. (A) The observed total intensity (continuous) and computed (dashed) MAGSAT anomalies of southern Africa generated from the model presented in (B). Cylindrical Equidistant Projection. C.I. 2 nT. (B) A three-dimensional magnetic model of southern Africa. Numbers inside the ellipses are effective susceptibility contrasts (in SI units) for 40-km thick polygons. The susceptibility contrasts of the polygons should be proportionately inflated if the interpreted thickness of the polygon is lower than 40 km. Cylindrical Equidistant Projection. a = Limpopo dikes; b = altered lower crust due to Walvis Hotspot; c = interpreted Eburnian 'basement' west of the Kalahari line; d = deep crustal magnetization minima of the Karroo Basin; e, f, and j = preliminary source configuration related to Agulhas Plateau, Mozambique Ridge, and northeastern Walvis Ridge (not constrained by geologic and geophysical parameters); g, h, and i = same as Figure 7.

(Fig. 3 and source c in Fig. 8). The interpreted age of the suture suggests that the province to the west of the Kalahari line is either of Eburnian age (~ 1850 Ma) or older and has been reworked by middle Proterozoic activity. Thus, the age of the region is consistent with the Transamazonian age (~ 2000 Ma) of the Rio de la Plata Craton (RP) in South America (de Almeida et al., 1973). The continuation of the MAGSAT anomaly high across the margins of the continents is, therefore, consistent with the available geologic and geophysical information (Figs. 3, 5, and 8).

The above correlations between the MAGSAT anomalies and the major shield regions between Africa and South America suggest that the geologic/magnetic provinces, formed prior to the Mesozoic separation of the continents, generally have correlatable magnetic signatures across the continental margin. The lack of correlation between the magnetic sources on the two continents is mainly in regions where the magnetic crust on one continent apparently was extensively altered during or after the separation (e.g., the West African Rift System). Because hotspots have played a major role in the Mesozoic breakup of western Gondwana (Burke and Dewey, 1973; Morgan, 1983; Duncan and Richards, 1991), we now examine spatial correlation between the major hotspots and the magnetic anomalies.

Hotspots and MAGSAT magnetic anomalies

Continental anomalies

Many magnetically negative regions across the continental assembly described in the previous section were found either incongruous with respect to pre-rift geology (subject to assumptions described in the introduction) or crossed multiple unrelated geologic provinces. Because occurrences of hotspots and associated magmatism and rifting entail large-scale crustal modifications, we examine the origin of these anomalies in the context of hotspot epeirogeny. Based on geologic observations of Mesozoic age volcanics and/or alkaline igneous activity in southeastern Brazil, southwestern Africa, the Karroo Basin and the oceanic extension of the magmatism in the form of oceanic uplifts, it has been inferred that a large portion of regions across the continental boundaries were underlain by hotspots during the Mesozoic (Burke and Dewey, 1973; Herz, 1977; Duncan, 1981; Morgan, 1983; Cahen and Snelling, 1984; White and McKenzie, 1989; Duncan and Richards, 1991). White and McKenzie (1989) have recently suggested that rifting over mantle plumes can generate huge quantities of melt due to decompression. The melt can be either extruded as basalt flows or trapped within or beneath the crust. Moreover, significant surface occurrences of alkaline igneous activity inferred to be related to hotspots and the nearly flat-top, mushroom-head shapes of the experimentally and numerically modeled hotspots (White and McKenzie, 1989; Duncan and Richards, 1991) probably indicate that areally vast, deep crustal regions may also have been affected.

Magnetic signatures of near-surface volcanic rocks originating in hotspot magmatism (or otherwise) depend on the product of their volume and their effective magnetization (to a rough approximation, the sum total of the normally magnetic components minus the reversely magnetic components). In regions where the products due to normal and reversed magnetization are roughly equal (e.g., Deccan Traps, based on a synthesized log in Vandamme et al., 1991; and possibly Parana basalts based on the oceanic reversal chronology in Cox and Hart, 1986, and a slightly tenuous assumption that the outpouring of basalts in the Parana Basin occurred intermittently from 135 to 120 Ma), only a weak magnetic anomaly signature is likely to persist to the high MAGSAT elevations. However, if large volumes of melt are solidified within the lower crustal regions affected by hotspots and if the magnetic properties of these rocks are different than adjacent lower crust, there is a strong likelihood that such lower crustal regions could be detected at satellite altitudes. The magnetic properties of the modified (by mantle intrusions) lower crust, however, are subject to speculation. For the Mississippi Embayment rift environment in the U.S.A., Thomas (1983) has suggested, based on data from Haggerty (1978, 1979), that under relatively low oxygen fugacity conditions of the lower crust, the newly formed titanomagnetites will be dominated by ulvospinels. Magnetic investigations of granulite xenoliths from Kilbourne Hole in the southern Rio Grande Rift, U.S.A., by Wasilewski and Mayhew (1982) also suggest that the lower crustal rocks beneath the rift regime contain abundant ilmenite and titanomagnetites with composition closer to ulvospinels (Curie temp. < 300°C). Unless these rocks are oxidized and later were exsolved to more magnetic forms (presumably by later thermal pulses), they will be paramagnetic (or essentially non-magnetic) at the lower crustal temperatures even under the normal continental geotherm (Haggerty, 1978; Thomas, 1983). If we assume that the magnetic state of the intruded lower crust has not been altered, it is possible that deep continental crustal regions intruded by hotspot magmatism will be weakly- to non-magnetic. We apply this hypothesis to the regions affected by Mesozoic hotspots on the African and South American continents.

In addition to the above mechanism, the results from Arkani-Hamed and Strangway (1985b) suggest that regions that have undergone early- to mid-Cenozoic hotspot magmatism and rifting are likely candidates for the thermal demagnetization of parts of the lower crust by elevation of the Curie isotherm. Although Arkani-Hamed and Strangway (1985b) indicate that thermal demagnetization may last for 200–300 Ma within the upper mantle, they assume a magnetic upper mantle with the same intensity of magnetization as the lower crust. If the mantle is essentially non-magnetic (Wasilewski et al., 1979), the thermal demagnetization mechanism alone may not be adequate for demagnetization of parts of the lower crust affected by Mesozoic hotspot activity and rifting dating ~ 120–140 Ma ago. The thermal regime of the crust, after this time, would have returned to temperatures below the Curie isotherm.

Possible regions affected by prominent Mesozoic hotspots across the African and South American margins are shown in Figure 9. The locations are based on dated volcanic and alkaline igneous intrusives (Siedner and Miller, 1968; Herz, 1977; Ulbrich and Gomes, 1981) and the constraints from consistently fitting the continental motions to the above igneous occurrences as well as oceanic aseismic ridges (Duncan, 1981; Morgan, 1983; Duncan and Richards, 1991). Comparison of these regions (Fig. 9) and the MAGSAT magnetic contrasts (Fig. 4) suggests that many of the regions coincide with observed negative magnetic contrasts.

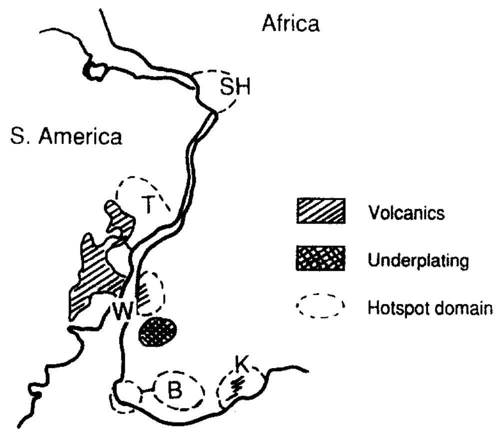

Fig. 9. Continental locations of major Mesozoic hotspots associated with the breakup of the continents (from Duncan, 1981; Morgan, 1983; White and McKenzie, 1989). $SH = 100$–120 Ma; $T = 50$–80 Ma; $W = 120$–130 Ma; $K = \sim 178$ and ~ 193 Ma; $B = 60$–75 Ma and 80–135 Ma.

The negative magnetic contrast associated with the Benue Trough (Bn) (Figs. 5 and 9; source d in Fig. 7) was discussed earlier in conjunction with Precambrian shields of northeast Brazil and Nigeria and Cameroon in Africa. Farther south, on the South American continent, Herz (1977) and Ulbrich and Gomes (1981) postulate two widespread (diffuse) hotspot-related episodes around ~ 120–145 Ma and 50–80 Ma (Fig. 9) based on the dated alkaline rocks northeast of the Parana Basin, the southernmost São Francisco Craton, and the Brasilia and Ribeira belts. The earlier episode (~ 120–145 Ma) of the igneous activity was also accompanied by voluminous outpouring of flood basalts in the Parana Basin (Prn) in South America (~ 120–135 Ma) and the Kaokoveld region (Kk) of Africa (~ 114–136 Ma) (Siedner and Miller, 1968; Herz, 1977; Cahen and Snelling, 1984; White and McKenzie, 1989, and references therein). Although the basalts and dolerites and alkaline complexes in the Kaokoveld region are significantly less extensive than the Parana basalts and alkaline intrusions in South America, the known occurrences in the Kaokoveld region are diffuse and are distributed over a large areal extent (Cahen and Snelling, 1984). Thus, it is possible that a significant amount of intruded material associated with the Walvis Hotspot (White and McKenzie, 1989) may be trapped within the crust under the Kaokoveld region. There are no seismic data on the deep crust of the Kaokoveld region; however, seismic evidence exists for a high-velocity (7.0–7.5 km/s) and 5–10 km thick (variable) layer above the Moho in the Damara belt of southwest Africa about 250–300 km away from the continental margin (Baier et al., 1983). South of the Damara belt, under the northern Kalahari Craton (NMC in Fig. 5), the high-velocity basal crustal layer is absent (Baier et al., 1983). Based on available data, it cannot be resolved whether the high-velocity basal crustal layer is encountered only in the Damara belt or can be associated with the mantle intrusions anticipated from the Walvis Hotspot. Our hypothesis, in conjunction with the magnetic contrasts and the models (Fig. 4 and source b in Fig. 8), indicates that roughly 5 km of non-magnetic intruded material within the lower crust is sufficient to explain the observed magnetic anomalies. This is a first-order estimate; the actual thickness will depend on the exact magnetic mineralogy as well as the significance of contributions from the upper crustal intrusives and extrusives. Despite these difficulties, it is worth noting that the magnetically derived estimate of the thickness of the interpreted hotspot intruded material within the lower crust is comparable to the seismically obtained thickness of the high-velocity basal crustal layer (Baier et al., 1983).

In the region covered by the Parana basalts (Fig. 9), no significant negative magnetic contrasts are observed (Figs. 3 and 4). Although it is possible that the lack of a magnetic low in our maps may be an artifact of our anomaly contouring cutoffs, we feel that in the South American regions under the domain of the Walvis Hotspot (120–135 Ma), most of the melt was extruded as Parana basalts (over 1.2 million km^3 according to White and McKenzie, 1989) and very little melt was trapped within the crust. Considering the geomagnetic field reversal chronology for this period (Cox and Hart, 1986), it is conceivable that the Parana basalts would yield only a weak magnetic anomaly signal at MAGSAT elevations that is not resolvable in our maps.

Based on the dated alkaline intrusives in the Minas Gerais–Goias belt (~50–80 Ma) within the Brasiliano (Brs) and Ribeira (Rb) belts (Fig. 5) in South America, Herz (1977) and Ulbrich and Gomes (1981) suggest a second diffuse hotspot that existed much after the Mesozoic continental split. The extensive alkaline igneous activity between 50 and 80 Ma suggests that the region was possibly underlain by a mantle plume (the present Trindade Hotspot; Crough et al., 1980). As mentioned in the previous section, the low MAGSAT contrasts centered over the Brasiliano and Ribeira belts (Figs. 4 and 5) cross over multiple geologic provinces and have been found difficult to interpret (Parrot, 1985). However, in view of our hypothesis, the observed low magnetic contrasts can be interpreted consistently as due to weakly- to non-magnetic mantle intrusions within the lower crust associated with the hotspot.

Similarly, the Karroo Basin (Krb) was under the influence of several hotspots during the Mesozoic as determined by volcanic activity of 193 ± 5 and 178 ± 5 Ma age in the northeastern Karroo Basin (Fitch and Miller, 1984, in White and McKenzie, 1989) and kimberlites in the central Karroo Basin of 80–145 Ma age (Hargraves, 1989). These igneous activities have been associated with the present Crozet and Bouvet Hotspots, respectively (Duncan, 1981; Morgan, 1983; White and McKenzie, 1989; Duncan and Richards, 1991). The continental locations of these hotspots and the prominent negative MAGSAT contrasts over the Karroo Basin (Figs. 4 and 9; source d in Fig. 8, which may be interpreted as approximately 5 km of intruded, non-magnetic material within the lower crust), also indicate a possible genetic relationship between the MAGSAT anomalies and lower crustal rocks generated during the hotspot activity.

The above comparison of the MAGSAT magnetic contrasts and the continental regions affected by Mesozoic magmatic activity (Figs. 4 and 9) suggests that there is a consistent relationship between these features. However, at the present time, the genetic correlation is only a hypothesis because there is little regional gravity, deep crustal seismic, and rock magnetic property information available to verify the contention. Nonetheless, many observed regional negative magnetic contrasts that are otherwise difficult to interpret can be consistently interpreted using the hypothesis. Thus, locations of hotspots in the recent past and the geologic record may be used to infer the origin of the regional magnetic anomalies.

The inferred correlation of the hotspots and the associated magnetic contrasts may also be geophysically useful in judging the magnitude and extent of deep crustal modifications. At the present time, there are only a few ways of estimating the amount of melt that was trapped in the lower portion of the crust. Because MAGSAT data generally emphasize crustal sources of large dimensions and high magnetizations (such as lower crustal sources), source regions interpreted to have formed due to decompression melting may be useful in estimating the amount of melt that was trapped in the lower crust. Using the regional magnetic models over these regions, the average magnetic properties of the normal lower crust (~ 0.063 SI units), and assuming that the intruded material added to the lower crust is weakly- or non-magnetic (Haggerty, 1978; Thomas, 1983), it is possible to compute estimates of the volume of the intruded material within the lower crust. Based on the susceptibility contrast maps and especially the magnetic models derived from available geologic constraints (Figs. 4, 7, and 8), we have estimated the volume of material added to the lower crust during the Mesozoic hotspot epeirogeny. For the regions discussed in the paper these estimates range from 1.0 to 2.8 million km^3 (~ 2.8 million km^3 or roughly a 10-km thick layer within the lower crust northeast of the Parana Basin, ~ 1.0 million km^3 or roughly a 5-km thick layer in the Kaokoveld region, ~ 2.2 million km^3 or roughly a 12-km thick layer in a relatively narrow zone in the Benue Rift region and under the Central African Shear Zone, and ~ 2.1 million km^3 or roughly a 5-km thick layer in the Karroo Basin region). Because of the simplifying assumptions made here (i.e., magnetizations of the normal and the intruded lower crust, neglecting the effect of the upper crustal intrusives and extrusives), the estimates will undoubtedly contain errors. Nonetheless, the same simplifying assumptions yield estimates of the basal, high-velocity, non-magnetic lower crust that are satisfactory with seismic refraction results in the Benue Rift and the Damara belt. Moreover, these estimates are consistent with the order of magnitude calculations of the amount of added material to the crust due to decompression melting (White and McKenzie, 1989; Hutchinson et al., 1990). Few geophysical data are useful in directly estimating the effective contribution of the large-scale intrusions within the continental lower crust in this manner, and although the MAGSAT magnetic anomalies have their limitations, they can be useful in computing first-order estimates of the volume of these intrusions.

Magnetic anomalies of the Rio Grande–Walvis Ridge System

A large portion of the Rio Grande Rise (RG) and the northeastern Walvis Ridge (Wl) are well-correlated with effective susceptibility contrast highs (Figs. 2, 5, and 6). These aseismic oceanic ridges initially were formed by the Tristan da Cunha Hotspot (or the Walvis Hotspot) under or near the mid-Atlantic Ridge. The mid-Atlantic Ridge has a long history of westward migration away from the hotspot and its repeated capture by the hotspot resulting in eastward jumps (at 106 Ma, 91 Ma, and 71 Ma according to Barker, 1983; between 70 and 50 Ma according to Duncan and Richards, 1991). After the last ridge jump, no ridge capture occurred and the westward migration of the mid-Atlantic Ridge was uninhibited. The hotspot trace after this time can only be observed on the African plate, where the active Tristan da Cunha Hotspot presently lies under the ~ 25-Ma old ocean-floor.

Fullerton et al. (1989) have modeled the northeastern portion of the Walvis Ridge with constraints from bathymetry and an oceanic magnetic model (Thomas, 1987). They argue that remanent magnetization acquired during the Cretaceous quiet epoch is necessary to explain the large intensity of the magnetic anomalies of the northeastern portion of the Walvis Ridge because susceptibility contrasts of the individual oceanic layers alone is inadequate. For similar high magnetic anomaly intensities in the western Rio Grande Rise, we have previously suggested that the mantle directly underneath the rise could bear positive magnetic contrasts, which, when supplemented by the viscous remanence acquired since the last geomagnetic reversal and the coherent remanent magnetization within the Cretaceous quiet zone, could explain the anomaly (Ravat et al., 1991). It is possible that both crustal and mantle magnetic sources could be responsible for some of the high magnetic anomalies observed near oceanic rises created by hotspots.

The correspondence of the small positive Free-air gravity anomalies (not shown) and the strong depth anomalies over the eastern Rio Grande Rise and the eastern Walvis Ridge suggests that these topographically uplifted regions may be nearly compensated at depth by low-density contrasts (possibly with respect to the surrounding mantle rocks), while the small negative gravity anomalies and the high depth anomaly over the western Rio Grande Rise suggest that this region is overcompensated at depth by similar low density contrasts. For the eastern Walvis Ridge, gravity modeling (Detrick and Watts, 1979) indicates Airy-type compensation with a total oceanic crustal thickness of 15–25 km, whereas geoid-height anomalies suggest Pratt-type compensation at ~ 30 km (Angevine and Turcotte, 1980; Turcotte and Harris, 1984). Chave's (1979) analysis of Rayleigh wave paths indicates both crustal thickening as well as low upper-mantle shear velocities related to a compositional change to the depth of 45 km under the eastern Walvis Ridge. In the western Walvis Ridge, the gravity anomalies are best explained by support of the topographic load by flexure of the lithosphere (Detrick and Watts, 1979; Kogan, 1979).

The correspondence of the thickened crust under the uplifts and the high effective susceptibility contrast suggests that at least a portion of the magnetic anomaly over the eastern Walvis Ridge (and possibly the Rio Grande Rise) may be explained by a highly magnetic root under the uplift. Hayling and Harrison (1986) and Fullerton et al. (1989) suggest a remanently magnetized source mainly in the thickened portion of the crust of the eastern Walvis Ridge. This interpretation is reasonable for the northeasternmost Walvis Ridge because most of the magnetic source occurs in the Cretaceous quiet zone, and therefore, may have coherent remanent magnetization direction. There are a number of reasons for the lack of a positive magnetic anomaly associated with the southwestern portion of the Walvis Ridge. First, the portion beyond the Cretaceous quiet zone may not have a coherent remanent direction because of the rapidly changing geomagnetic field, and consequently, may have both normal and reverse components. Second, the thickened crustal root within the southwestern part of the ridge may be thermally demagnetized at the present time. And third, as suggested by the flexural model of Detrick and Watts (1979) and Kogan (1979), there may not be a compensating low density (and highly magnetic) root in the southwestern portion of the Walvis Ridge.

Our simple model constrained by the boundaries of the bathymetric uplift of the northeastern Walvis Ridge (Ravat, 1989) poorly approximated the location of the anomaly maximum as well as the anomaly trends. The computed anomaly trends are northeast even in the extensive model of Fullerton et al. (1989), whereas the observed anomaly trend is prominently east–west. Hence, we suggest minor modifications to these models that could overcome the apparent disagreement in the trends between the observed and modeled anomalies. In the present model, depth anomalies (McClain, 1984) were used to constrain the lateral boundaries of the magnetization model because the magnetic contrasts in the conceptual model are dominated by the thickened root under the uplifted portion of the northeastern Walvis Ridge and the limits of the deeper inhomogeneities are better defined by depth anomalies rather than isobaths. In the eastern portion of the model where no depth-anomalies are available, the model is based simply on geologic and geophysical considerations.

Sibuet et al. (1984a, b) have studied the near-surface and deeper aspects of the structure of the eastern Walvis Ridge from a large quantity of seismic reflection, gravity, magnetic, and drill-hole data compiled and collected during the Leg 75 of the DSDP. Based on these data, they suggest that the continent–ocean boundary occurs at about 10°E longitude and the provinces to the east of the boundary have pre- and syn-rift continental origins. Their models suggest that the stretched continental crust east of 10°E is approximately 10–15 km thicker than normal oceanic crust. White and McKenzie (1989) suggest that the thickened transitional crust may be explained by a considerable amount of igneous intrusions in this magmatically active (due to the Walvis Hotspot) stretched margin (White and McKenzie, 1989).

As in Fullerton et al. (1989), our present model considers the remanent magnetization acquired during the Cretaceous quiet epoch. The remanent magnetization direction in this model (inclination of −67° and declination of −23°) was inferred from the paleomagnetic data from the mean of 81–100 Ma South African kimberlites (Hargraves, 1989). The direction of the paleo-remanent vector is not resolvably different (by modeling) from the present-day geomagnetic field inclination and declination at the eastern Walvis Ridge, and hence, susceptibility and viscous remanence cannot be resolved from thermal remanence from the magnetic anomalies and the natural remanent magnetizations estimates of Harrison (1987).

Even after the above modifications which help broaden the east–west extent of the previously modeled magnetic sources, if we were to honor the constraints from the observed MAGSAT anomalies, the southern extent of the modeled sources must be terminated near roughly 24°S latitude (Fig. 10). The only explanation for such a termination (which is 3°–4° north of the extent of the bathymetric uplift and the high depth anomalies) is that the thickened root under the southern part may have cooled below its Curie isotherm after the end of the Cretaceous quiet epoch, and consequently, the crustal column may have roughly equal normal and reversed volume magnetizations. The model presented here (Fig. 10) does not necessarily negate the possibility of up-

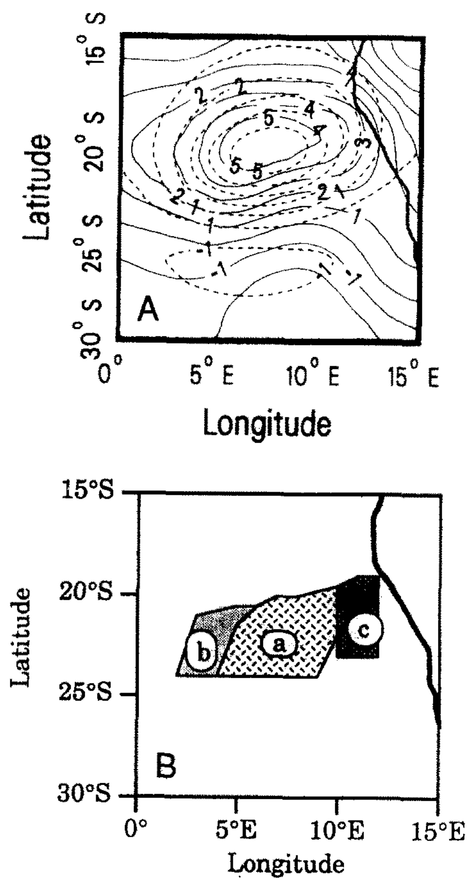

Fig. 10. (A) The observed total intensity (continuous) and computed (dashed) MAGSAT anomalies of the northeastern Walvis Ridge generated from the model in (B). Cylindrical Equidistant Projection. C.I. 1 nT. (B) The three-dimensional magnetic model of the northeastern Walvis Ridge in which the lateral boundaries of the model are based on the depth anomalies (McClain, 1984). See text for details. Model parameters—*magnetic susceptibility:* $a = 1$ km basaltic uplift and 10-km thick serpentinized gabbroic root (both 0.012 SI units), b = 7-km thick serpentinized gabbroic root (0.012 SI units), c = 5-km thick layer of thickened transition zone basalts and gabbros (0.01 SI units); *remanent magnetization:* a = 1-km basaltic uplift (~ 2.4 A/m) and 10-km thick serpentinized gabbroic root (~ 1.51 A/m), b = 7-km thick serpentinized gabbroic root (~ 1.51 A/m), c = 5-km thick layer of thickened transition zone, basalts and gabbros (~ 1.2 A/m). Remanent inclination −67°; remanent declination −23°. Magnetization estimates are derived from table 4 of Harrison (1987) as follows: basaltic uplift—oceanic Layer 2 values; serpentinized gabbroic root—assuming a volume ratio of 4 to 1 for oceanic Layer 3 gabbros and serpentinite, respectively; transition zone basalts and gabbros—assuming a volume ratio of 2 to 1 for gabbros and basalts. Cylindrical Equidistant Projection.

per mantle magnetization contrasts under the oceanic ridges created by hotspots (and the possibility exists that the uppermost mantle regions of the anomalously low shear velocities observed by Chave, 1979, under the northeastern Walvis Ridge could be more magnetic than the surrounding mantle) (Ravat et al., 1991). However, it is not possible to discriminate between the two models (magnetic contrasts mainly in the thickened crust or the combination of the thickened crust and the upper mantle) based on the available geologic, geophysical, and the magnetic property data; in approximating the details of the observed MAGSAT anomaly (considering the ± 1 nT error bars for the maps), both models face similar problems and have similar apparent solutions. Certainly, the remanent magnetization model suggested by Hayling and Harrison (1986) and Fullerton et al. (1989) and the minor modifications suggested here appear to satisfy the available geologic, geophysical, magnetic property, and magnetic anomaly modeling constraints for the northeastern Walvis Ridge.

The situation in the Rio Grande Rise is more complex. A portion of the eastern Rio Grande Rise susceptibility contrast zone occurs beyond the Cretaceous quiet zone and is unlikely to have a coherent remanent magnetization in a rapidly reversing magnetic field. However, beyond the Cretaceous eastern Rio Grande Rise, the susceptibility contrasts continue to the northeast where there is no surface expression of an oceanic geologic source. Thus, highly magnetic crustal rocks are required to explain the high observed magnetization northeast of the Rio Grande Rise.

Another source for the local high magnetization (both normal and reversed) is suggested by the magnetic properties of the gabbro drilled in the Ocean Drilling Program and reported in EOS (1988). Drilling on a "plateaulike area" in the Atlantis II Fracture Zone on the southwest Indian Ridge has recovered gabbros that have "... up to 1000 times more magnetization than a normal midocean ridge rock". At present, it is not known whether these "very magnetic" gabbros are characteristic of anomalously elevated ocean crust or not. Nonetheless, some of the magnetic contrast highs and lows observed by the MAGSAT on the ocean-floor may have their origin in such highly "anomalous" rocks.

Concluding remarks

Comparison of the high-resolution radially-polarized magnetic anomalies and the geologic sources across the rifted continental margins of Africa and South America suggests that there is a general consistency between the magnetic anomaly sources across the conjugate margins when the timing of their possible geologic sources is considered and allowances are made for singular regional magnetic sources (e.g., Bangui anomaly). The consistency, in itself, constitutes the geologic verification of the MAGSAT magnetic anomalies.

Based on the correlation of continental regions affected by Mesozoic hotspots and negative MAGSAT magnetic contrasts, we suggest that there is a genetic relationship between them. We relate the negative magnetic contrasts to the solidified melt within the crust that evolved over the mantle plumes. At present, there are only a few geophysical techniques that can directly estimate the extent and amount of deep crustal modification due to hotspot magmatism. Because the regional magnetic contrasts mainly represent the large-scale, highly magnetic contrast sources, they can be useful in estimating the lateral extent and amount of melt that was trapped within the crust.

The magnetization of the well-correlated Rio Grande–Walvis Ridge System that evolved over the Tristan (or Walvis) Hotspot can be partially explained by an induced and remanent magnetic source in the intruded thickened crust under the uplifted portion of the ridge. However, local highly magnetic sources may be required to explain some of the oceanic magnetization contrasts.

Acknowledgements

We greatly appreciate the comments of Kevin Burke, Paul Toft, and two anonymous reviewers. This research was funded by NASA grant NAGW-1819 and NSF grant EAR-8617315.

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