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MOHO AND LITHOSPHERE-ASTENOSPHERE BOUNDARIES IN EAST AFRICA FROM REGIONAL GRAVITY DATA

Abstract. The analysis, processing, and interpretation of gravity anomaly fields between East Africa and Southern Arabia are here oriented to modelling of Moho and LAB (Lithosphere-Asthenosphere Boundary) morphologies as a contribution to the reconstruction of the Afro-Arabian plates geodynamical evolution.

The great interest in East Africa and surrounding regions is due to the presence of more or less advanced phases of extension tectonics, as evidenced by lithospheric extent, LAB upward migration, crustal rifting and sea floor spreading. In this study, importance was given to estimation of the density variation in asthenospheric upwellings. The adoption of theorethical models that are able to estimate the vertical distribution of temperature in the upwelling mantle and to calculate the density of melt silicate depending on its pressure, temperature, and chemical composition, allowed us to predict the vertical density variation in the upwelling mantle related to crustal rifting in East Africa. These results were adopted to perform regional gravity 2.5-D modelling along three main geotransects in East Africa, in order to describe the morphology of mantle uprises in this region.

INTRODUCTION

The great interest in the geodynamics of East Africa and surrounding regions is due to the presence of more or less advanced stages of extension tectonics, such as, for instance, crustal stretching, lithospheric thinning, LAB (Lithosphere-Asthenosphere Boundary) upward migration and sea floor spreading (Fig. 1). By contrast, the Western Somali Basin represents an old ocean basin where sea floor spreading failed 83-100 My BP. The oceanic basement is hidden by a thick sedimentary cover and displays only partially recognisable small reliefs (DHOW, ARS and VLCC Fracture Zones) (Bunce and Molnar, 1977). The Gulf of Aden is one of the most significant examples of young ocean basin, spreading from the Sheba Ridge from 10 My ago to the present (Laughton et al., 1970; Cochran, 1981). Moreover, along the Ethiopian Plateau, the great East African Rift Valley shows and interesting example of a continental rift: it is connected to the Gulf of Aden and to the Red Sea through the Afar Depression, which represents a classic example of a triple junction among Somali, African and Arabian plates, recording a complex evolution from continental to oceanic rifting (Roberts and Whitmarsh, 1969; Courtillot, 1980; 1982; Makris and Ginzburg, 1987).

A study based on the analysis, processing and interpretation of gravity anomaly fields in these areas is here presented. It is oriented to modelling the asthenosphere/lithosphere (LAB) and the mantle/crust (Moho) boundaries in the East African rifting zones; it was done following suggestions by the Global Geotransect Project (G.G.P.). The studied area (Fig. 1) is located between Longitude 36°E-59°E and Latitude 2°S-17°N. It includes the Somali region, the Ethiopian region (Ethiopian and Harar plateaux), the Afar Depression, South-Yemen, the Gulf of Aden and the North-Western Somali Basin.

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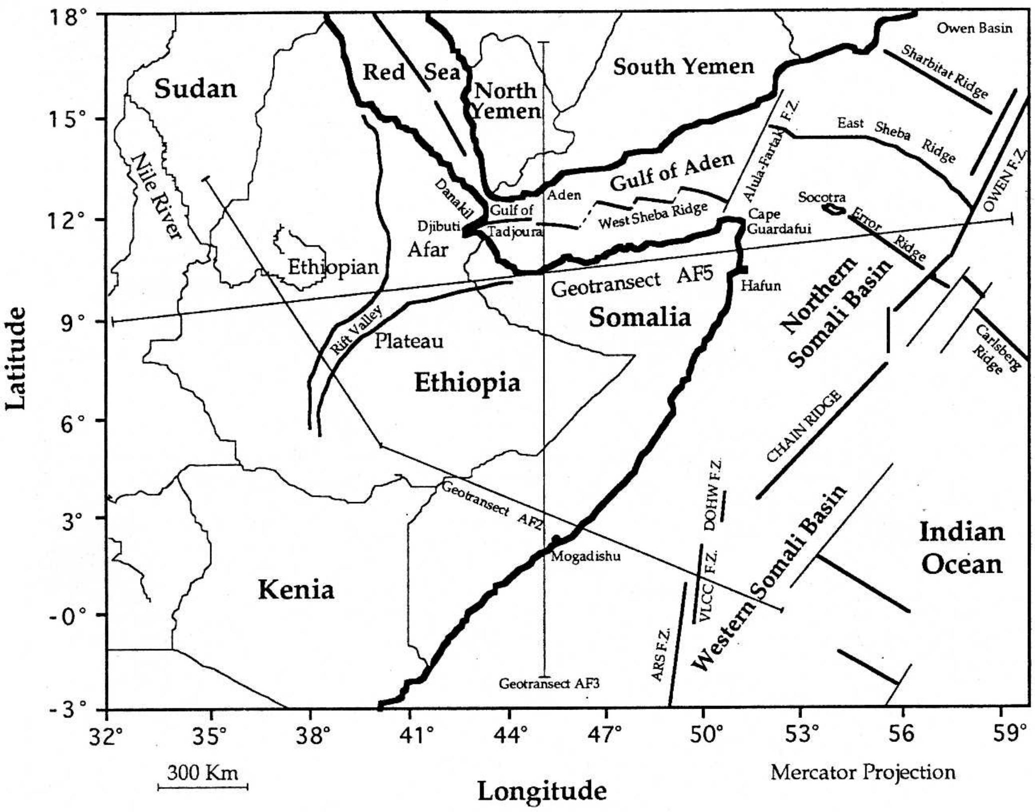

Fig. 1 — Sketch map of the studied area showing main geographic and structural features. Locations of geotransects AF2, AF3, and AF5 are here indicated.

GRAVITY DATA

Gravity data processing

The study is based on three different data sets:

1) The AGP (African Gravity Project) Somali Gravity Anomaly Map, published in 1988 by the University of Leeds Industrial Services and based on data from oil companies and research centers, such as the Dipartimento di Geofisica e Vulcanologia dell'Università "Federico II" di Napoli (Rapolla et al., 1985; 1986) (Fig. 2);

2) the East Africa and Red Sea Gravity Anomaly Map, (Makris et al., 1991);

3) the Gravity anomaly map of the east Sheba Ridge and Owen Fracture Zone, (Stein and Cochran, 1985).

Gravity data were digitized using a 5'x5' (average 9.4 km) step grid. A total of 26208 (168x156) values were obtained. This sampling was chosen due to the regional aim to this study and for quick processing but good detail.

Because the purpose of this research is reconstruction of the geometrical pattern of sub-crustal sources (Moho and LAB), removal of short-period signals due to shallow crustal sources was necessary. The evaluation of the gravity effects related to structures at different depths may be performed by recognizing the well defined frequency ranges in the anomaly fields. This was done my means of bidimensional spectral analysis, to obtain the power spectrum of the whole gravity field (Fig. 3). These frequency ranges were chosen by noting local sharp

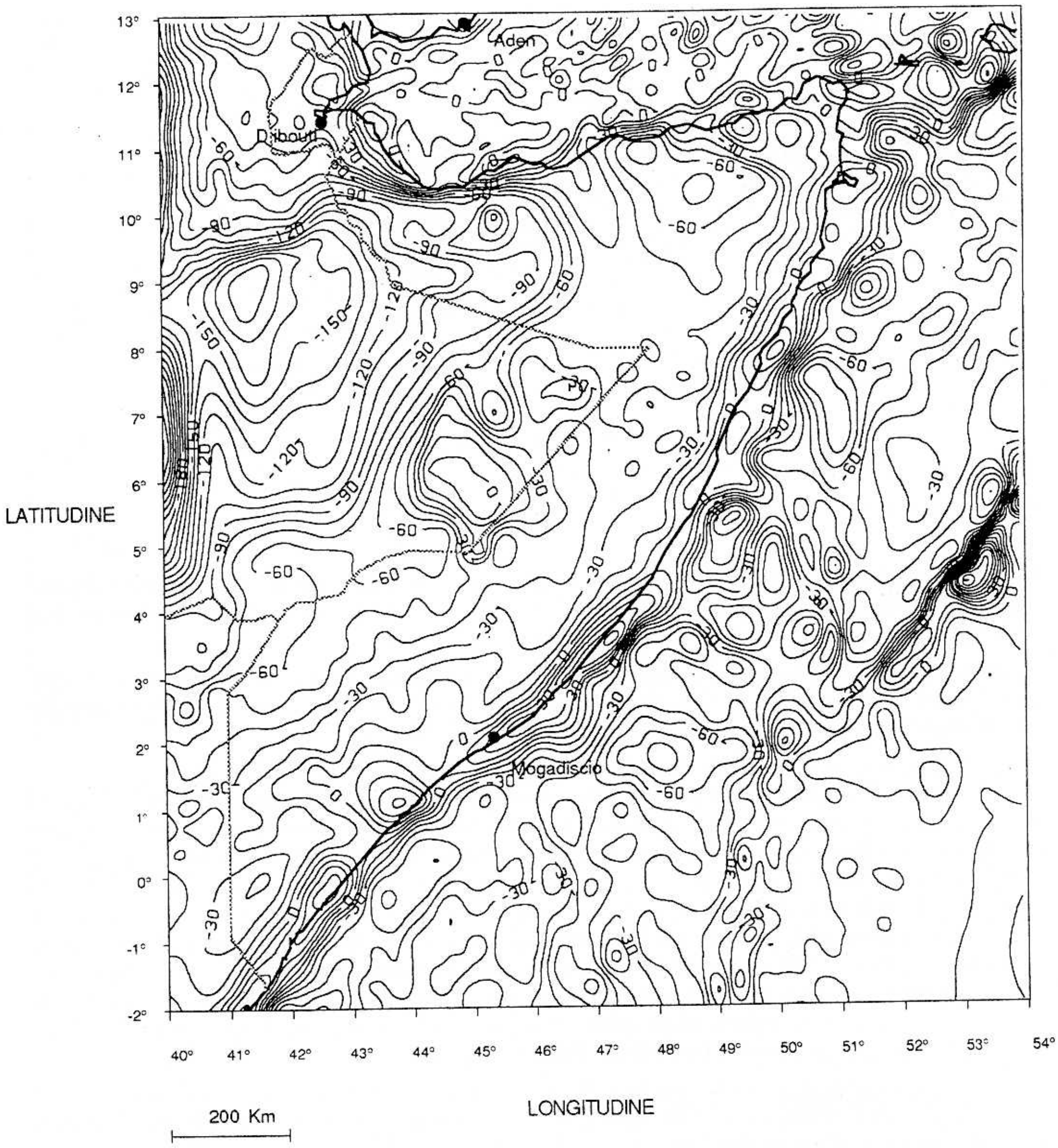

Fig. 2 — Gravity anomaly map of Somali and surrounding regions (modified from: University of Leeds Industrial Services, 1988); contours at 10 mGal; Bouguer anomalies on-shore; Free Air anomalies off-shore; density for Bouguer correction = 2.67 g/cmc; Equatorial Mercator Projection. Gravity calculated using the International Gravity Formula, 1967 (IGF67) and referred to the International Gravity Standard Network, 1971 (IGSN 71).

decreases in the signal, as these should indicate the boundaries between different groups of anomalies with well-defined wavelength ranges.

The following phase was numerical filtering of the signal, using a frequency-domain filter with a Hamming-Tukey window. The "boundary effect" problem (generation of parasitic signals along boundaries of the map during numerical filtering, with the appearance of false anomalies and distortion of the real ones) was avoided by extending the original gravity anomalies field by 200 km over the limits of the grid using a predictive filter. In order to obtain the best results, a careful direct evaluation of the filtered fields using a series of different cut-off wavelengths can, however, be useful to identify the best ones, and to differentiate several anomaly patterns.

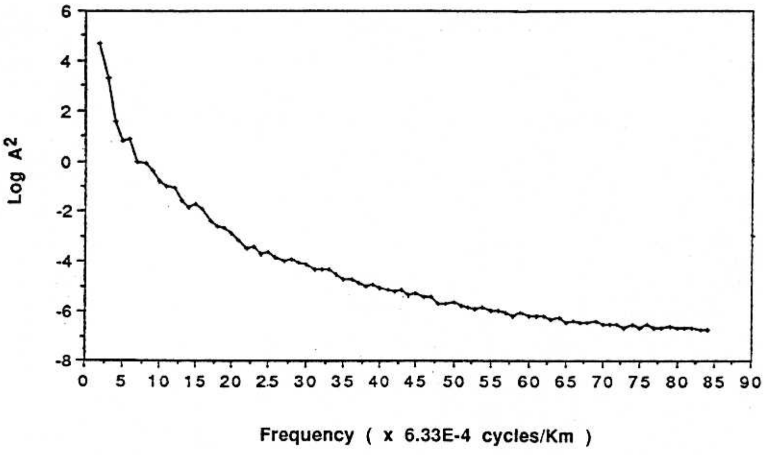

Fig. 3 — 2-D power spectrum of the gravity anomaly field shown in Fig. 2.

Thus, a set of low-pass filtered maps was computed using many cut-off wavelengths ranging from 140 km to 1200 km, in order to control the accuracy of the choice based on the spectral analysis.

Previous results (Pinna and Rapolla, 1979; Gasparini and Mantovani, 1984; Dorre, 1989) indicated the wavelength characterizing the variation in Moho depth and showed that this generally causes some hundreds of km wavelength Bouguer anomalies.

For instance, by examining several anomaly profiles collected in the Somali area, Dorre (1989) suggested a range of possible cut-off wavelengths from 150 to 250 km. It was also observed by Dorre (1989) that while filtered maps obtained using cut-off wavelengths lower than 180 km still show effects caused by local shallow sources, filtered maps computed with cut-off wavelengths higher than 225 km instead show excessive smoothing. Therefore, the author suggested the use of a 200 km cut-off wavelength. Analysis of the power spectrum - for the whole area or parts of it - and of the filtered maps indicates here also a range of likely cut-off wavelengths between 160 and 270 km. Filtered maps computed using cut-off wavelengths closest to the 300 km value show an excessive smoothing effect; therefore, a 200 km cut-off wavelength was also chosen here to differentiate anomalies produced by sources deeper than the Moho (Fig. 4) from those generated by crustal sources (Fig. 5).

Numerical filtering gave informations about the presence of an anomaly source deeper than Moho. As a matter of fact, most of the anomalies disappear or their amplitude strongly decreases in maps filtered with more than 400 km wavelengths. The only anomaly that almost preserves its shape and amplitude is the well known negative one located on the Ethiopian Plateau. In fact its shape does not change starting from 600 km cut-off wavelength (Fig. 6), and its amplitude (130 mGal at 600 km cut-off wavelength) still shows high values (75 mGal) at a 1200 km cut-off wavelength, appearing by now on an almost flat field (Fig. 7).

Gravity data interpretation

All Authors agree in ascribing the large negative anomaly of the Ehtiopian Plateau and related Rift Valley to lithospheric thinning and to an asthenospheric upwelling process (Brown and Girdler, 1980; Bermingham et al., 1983). The latter has produced several recognizable effects on the earth surface, such as uprising, a deep axial depression, and a strong bimodal volcanism.

Lambeck (1972, 1973) proposed that the very long-wavelength gravity anomalies recognized along several oceanic spreading axes are due to deep sources, probably related to change in LAB depth. Kaula (1972) suggests that asthenospheric vertical flows can generate gravity anomalies with wavelengths ranging from 1200 km to 3500 km. The great negative anomaly

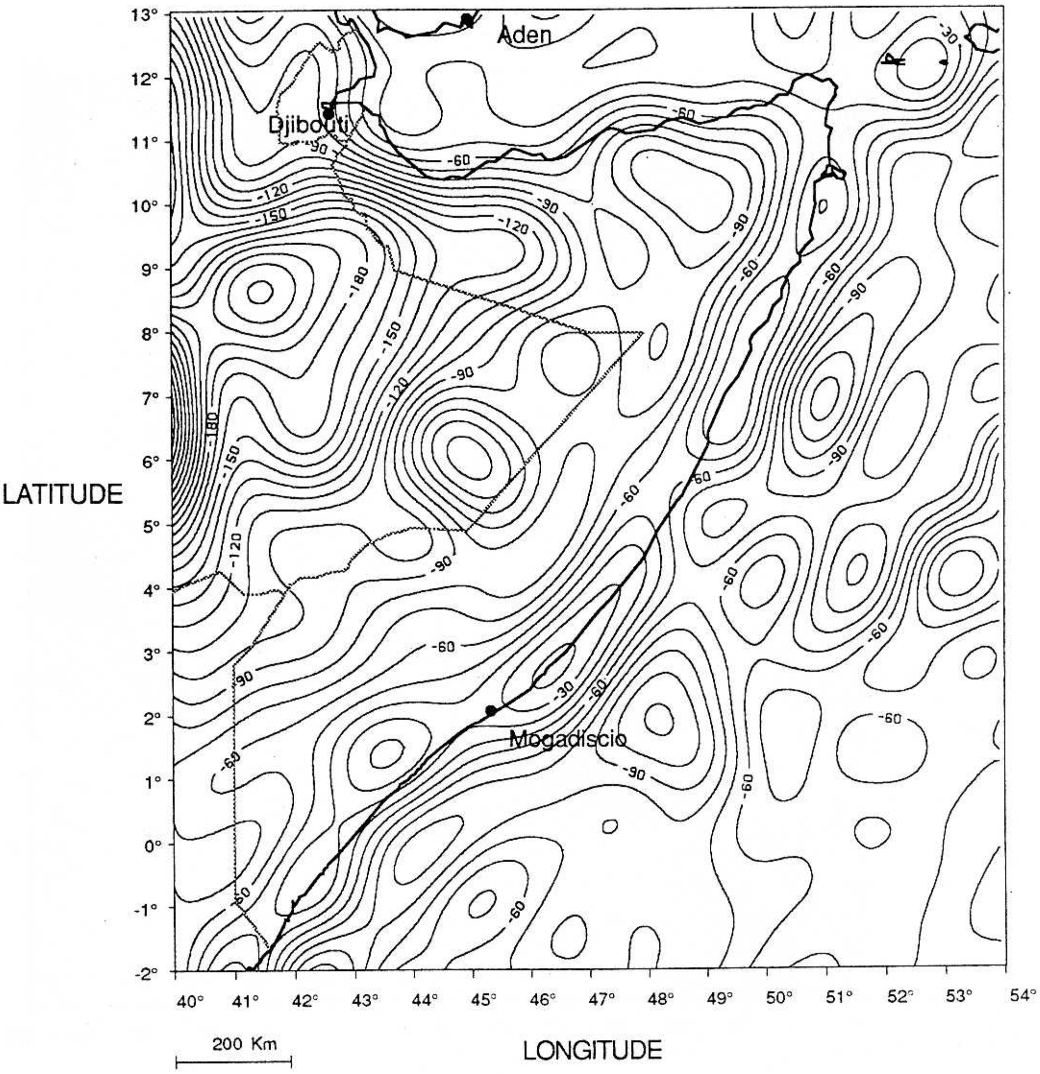

Fig. 4 — Low-pass filtered gravity anomaly map of Somali and surrounding regions; cut-off wavelength = 200 km; contours at 10 mGal; Equatorial Mercator Projection.

in East Africa, as already observed, might be one of them. One might also expect a correlation between this feature of the gravity field in East Africa and a deep region (450-650 km of depth) of low seismic velocities detected under the Red Sea and the Gulf of Aden using tridimensional inversion methods (Woodhouse and Dziewonski, 1984).

The mass deficiency at the origin of the anomaly is explained by the upwelling of hot asthenosphere and consequent change in its physical properties, such as density. The gradual decrease in density is caused by dilatation due to adiabatic decompression and by increasing melting processes. The asthenospheric upwellings are surrounded by areas of different physical properties, thus explaining the appearance of lateral density contrasts and, consequently, long period anomaly fields.

Most of the Authors who have previously studied this problem with a gravity approach have described the morphology of the asthenosphere uprising, and consequent density decrease, only in terms of a good fit between observed and computed model. However, it is well known that it is possibile to compute many density distributions which are all compatible with the anomaly

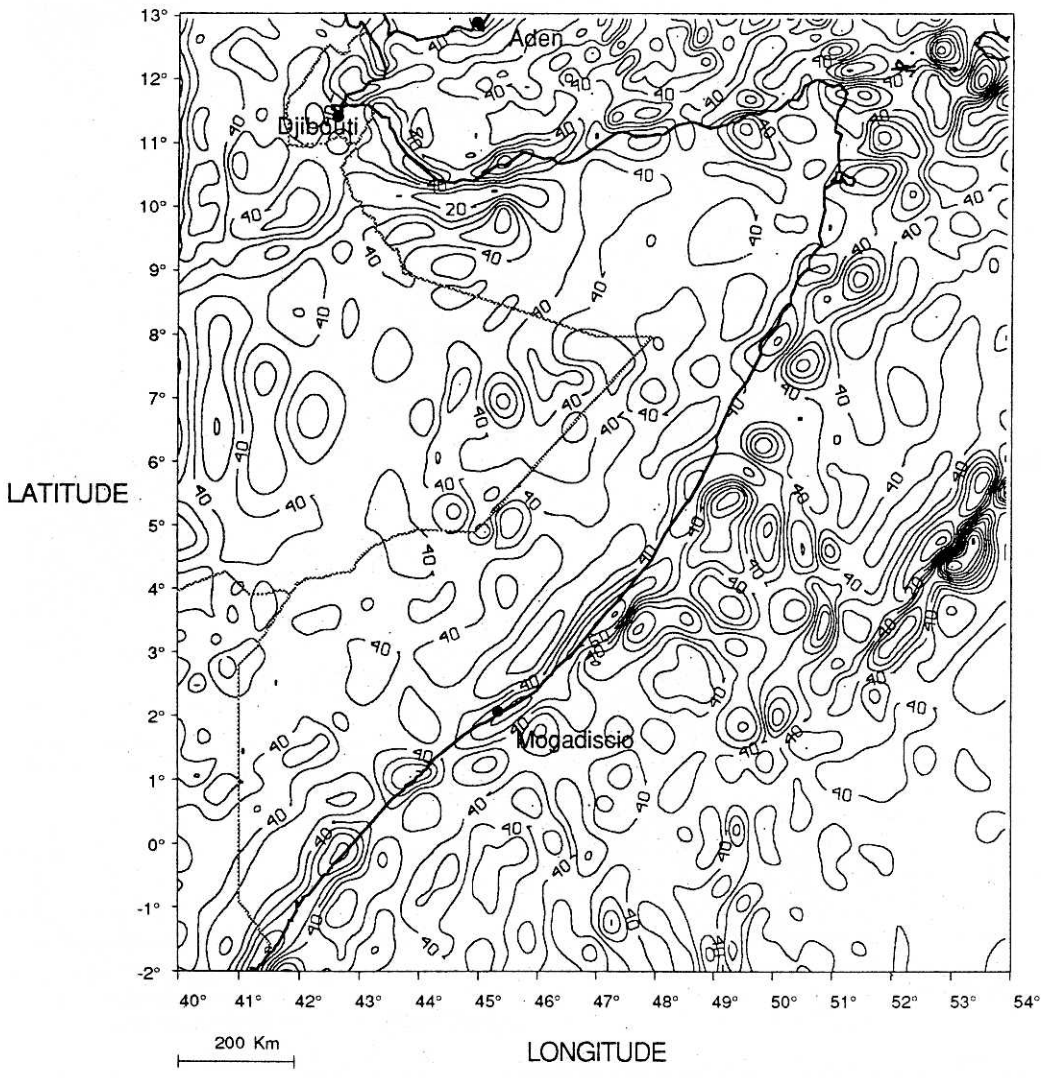

Fig. 5 — Residual gravity anomaly map of Somali and surrounding regions obtained from low-pass filtered map (cut-off wavelength = 200 km); contours at 10 mGal; Equatorial Mercator Projection.

field. This inconvenience can be reduced if the gravity modelling can be constrained by other physical or petrological data.

We attempted to satisfy this requirement by independently estimating the vertical density decrease in an upwelling asthenosphere, which depends on the variable petro-physical environment (temperature, pressure and chemical composition).

Consequently, two main problems arise:

a) The need to define the thickness of the lithospheric and crustal structures in surrounding areas not involved in extentional tectonics, and to estimate their correct density values;

b) The need to recognize areas where lithospheric thinning and asthenospheric upwelling are located, and to estimate correct density values in the uprising mantle.

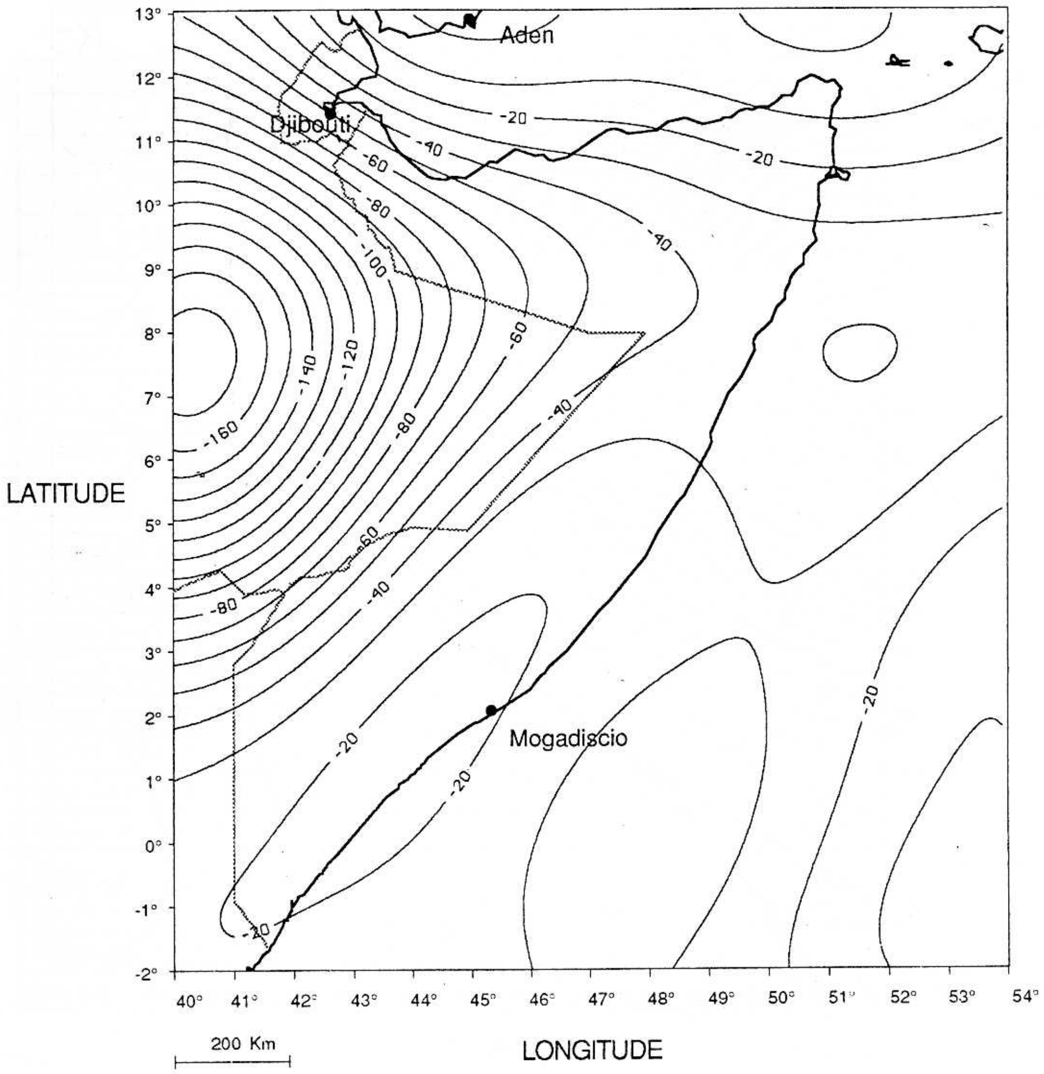

Fig. 6 — Low-pass filtered gravity anomaly map of Somali and surrounding regions; cut-off wavelength = 700 km; contours at 10 mGal; Equatorial Mercator Projection.

LITHOSHPERE AND ASTHENOSPHERE CHARACTERISTICS

Lithosphere density and thicknesses in undisturbed areas

Upper mantle phase transition from garnet peridotite (3.36 g/cm ^{3}) to spinel peridotite (3.32 g/cm ^{3}) is hypothesized at a depth of 50-60 km (Bottinga and Allegre, 1976; Green and Libermann, 1976; Ringwood, 1976), and a good agreeement between petrological and seismological data is generally found. The "AFRIC" model (Gumper and Pomeroy, 1970) suggested a 3.35 g/cm ^{3} LID density, and crustal density values decreasing towards the surface from 2.9 g/cm ^{3} to 2.67 g/cm ^{3}. These results have been confirmed by Press (1973). The PEM (Parametric Earth Model) model (Dziewonsky et al., 1975) assigns a 3.37 g/cm ^{3} density to the asthenosphere near the LAB (at a 70-80 km depth in oceanic lithosphere, at 90-120 km depth in young continental lithosphere, and at 200 km depth in cratonic lithosphere). This

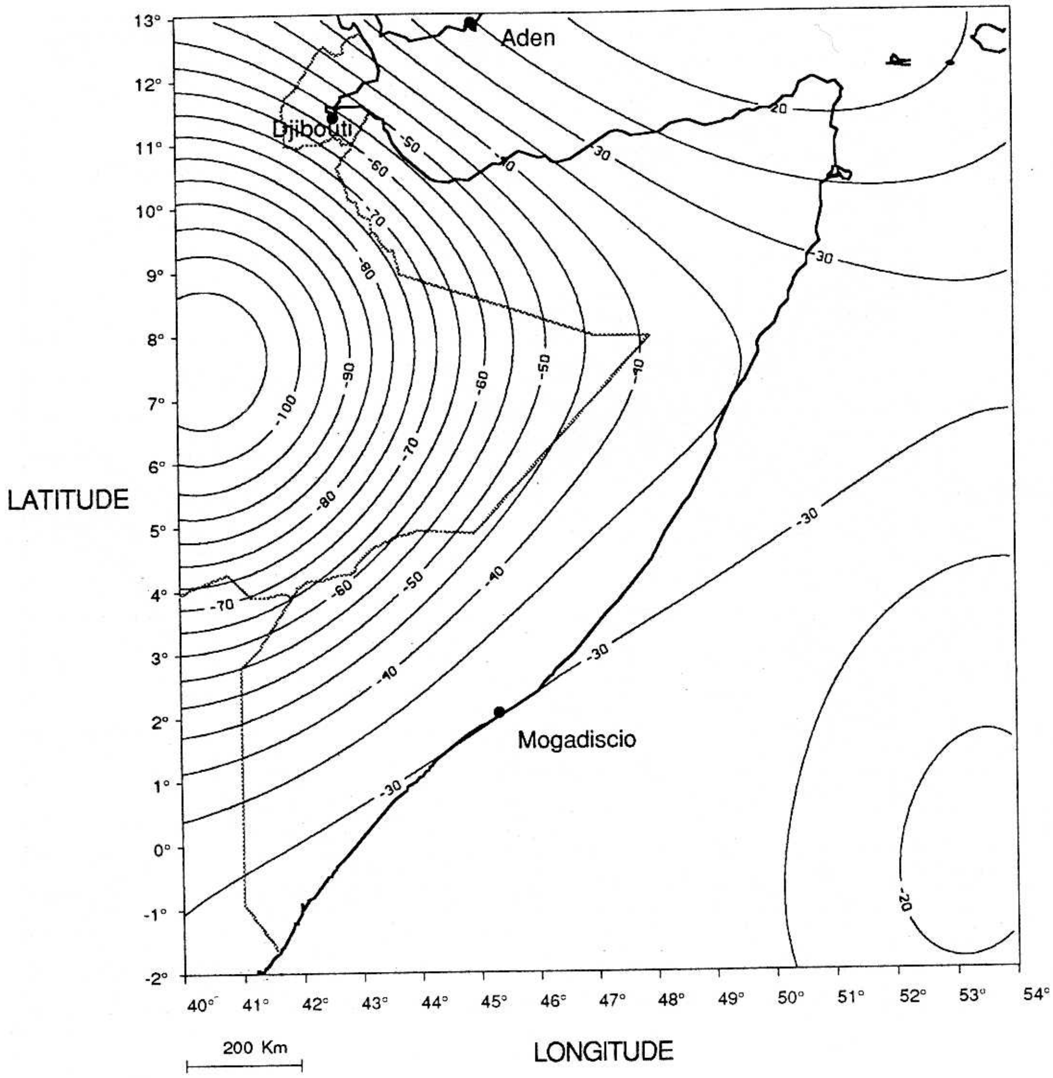

Fig. 7 — Low-pass filtered gravity anomaly map of Somali and surrounding regions; cut-off wavelength = 1200 km; contours at 5 mGal; Equatorial Mercator Projection.

model does not imply a strong density variation near the LAB; instead it suggests a gradual decrease from about 3.37 g/cm ^{3} to 3.3 g/cm ^{3} near the Moho (30-40 km depth).

Taking into account the density values suggested by several Authors (Table 1), a mean value of 3.34 g/cm ^{3} was chosen in this study as the density of the undisturbed LID material. This value represents the average lithosphere density predicted by Dziewonski et al. (1975) for the depth range between 30 km and 100-110 km. For the crust, intracrustal density discontinuities have not been taken into account in this study, except the lateral transition between oceanic and continental crust, since only this causes a detectable long-period gravity anomaly field. Moreover, again on the basis of the data reported in Table 1, a density value of 2.9 g/cm ^{3} was assigned to the oceanic crust, and 2.85 g/cm ^{3} to the continental crust.

As far as thickness is concerned, the results of several previous studies on crustal and lithospheric thicknesses, inside and outside the studied region, were considered for this work, and are reported in Table 2.

Table 1 — Crustal and LID densities in undisturbed areas (g/cm^3).

AUTHORS	Oceanic Crust	Continental Crust	Lithospheric Mantle
Talwani et al. (1965)	2.84		3.4
Matthews and Davies (1966)	2.83	2.9	3.4
Sowerbutts (1969)	2.8 - 2.86 - 2.94		3.4
Gumper and Pomeroy (1970)		2.9-2.67	3.35
Baker and Wohlemberg (1971)		2.9	3.3
Fairhead and Girdler (1972)	3	2.7 (aver.)	3.34
Darracott et al. (1972)		2.837	3.34
Searle e Gouin (1972)		2.82	
Corrado et al. (1974)	2.9		3.3
Green and Liebermann (1976)			3.32-3.36
Fairhead (1976)		2.83 (aver.)	3.34
Withmarsh (1979)	2.9-2.6		3.35
Pinna and Rapolla (1979)		2.65-2.9	3.3
Seidler and Jacoby (1981)			3.3-3.05
Bott and Mithen (1983)		2.81	3.3
Browne and Fairhead (1983)		2.7	3.27
Tamsett (1984)	2.96-2.57		3.32
Stein and Cochran (1985)	2.8		3.33
Gettings et al. (1986)	2.7	2.82-2.95	3.33-3.38
Makris and Ginzburg (1987)	2.8	2.88	3.2-3.35
Izzeldin (1987)	2.9	2.8	3.32
Makris et al. (1991)	2.9	2.845	3.35
Greene et al. (1991)		2.9-2.83	3.275

Most of the quoted Authors assigned a 30-40 km thickness to continental crust not involved in extentional tectonics, and a 6-10 km thickness to the oceanic crust. A total thickness of undisturbed lithosphere of 110 km is suggested by these Authors.

Modelling: uprising asthenosphere densities

Some studies have shown (i.e.: McKenzie and Bickle, 1988) that when hot anomalous mantle upwelling occurs (as in hot plumes) the rate of advective heat transfer exceeds the rate of cooling by heat conduction. Thus, the phenomenon occurs in quasi-adiabatic conditions and generates an anomalous thermal positive gradient toward the earth's surface. Consequently, mantle rocks involved in the asthenospheric upwelling reach the solidus, and partial melting occurs with upward increasing melting degree and gradual change in melt composition, the latter depending on depth, temperature, and melting degree. Because the density of melt and of residual solid fraction are smaller than that of the original mantle rocks, the density of the upwelling mantle shows a greater decrease than that due only to thermal expansion. Moreover, the density decrease is complicated both by the much greater compressibility of the melt fraction than the solid one, and by the presence of fluids. Thus, a very complex evolution of the petro-physical features

Table 2 — Crustal and lithospheric thicknesses.

	Crustal thicknesses (km)		Lithosphere thickness (km)
AUTHORS	Continental Crust	Oceanic crust	
Sowerbutts (1969)	25-45		
Gumper and Pomeroy (1970)	36		
Kanamori and Press (1970)			70
Baker and Wohlemberg (1971)	33		80
Darracott et al. (1972)	35		90
Dziewonsky et al. (1975)			125
Fairhead (1976)	35		100
Withmarsk (1979)		12-19	
Brown and Girdler (1980)	35		100
Bott (1981)	35		
Cochran (1982)	30		105
Bermingham et al. (1983)	35		100
Tamsett (1984)		7-9	
Stein and Cochran (1985)		5	125
Gettings et al. (1986)	40-45	10	
Izzeldin (1987)	40		
Makris et al. (1991)	30-35		
Greene et al. (1991)	22-23 (rift)		

of mantle rocks involved in the hot plume can be expected. The influence of pressure, temperature, and partial melting through time must therefore be taken into account to estimate the density change of the upwelling mantle in gravity modelling on extensional structures.

Unfortunately, the agreement between density contrasts, obtained by gravity data fitting, and the modalities of change in the physical and petrological features of the upwelling asthenosphere that cause the density contrasts, has often not been sufficiently analysed.

Table 3 reports the values of density contrasts between asthenospheric plumes and surrounding lithospheric mantle given by several Authors. A large scatter can be observed in the data, thus confirming the necessity of an independent estimate of the density values.

Referring to the model proposed by McKenzie and Bickle (1988) and White and McKenzie (1988), the change in melt percentage and, from this, the change in density contrast can be evaluated. This model involves parameters that can be evalueted on the earth's surface. These are the β "stretching factor" (McKenzie, 1978) and the resultant total vertical displacement, subsidence or uplift. The β factor indicates (McKenzie and Bickle, 1988) the areal extension which some part of the lithosphere has undergone, and is expressed as the ratio of final to initial surface area involved in the rifting process. The Authors of the model assign values ranging from 1 to 4 to regions involved in continental rifting, whereas values higher than 5 indicate oceanic spreading.

Knowledge of the β value, of the total vertical displacement and of the original lithospheric thickness allow on estimate of the interior potential temperature (Tp), which is defined as "the temperature the fluid mass would have if it was compressed or expanded to some constant reference pressure" and, in our case, as "the temperature the mantle material would have at the Earth's surface if melting does not occur" (McKenzie and Bickle, 1988).

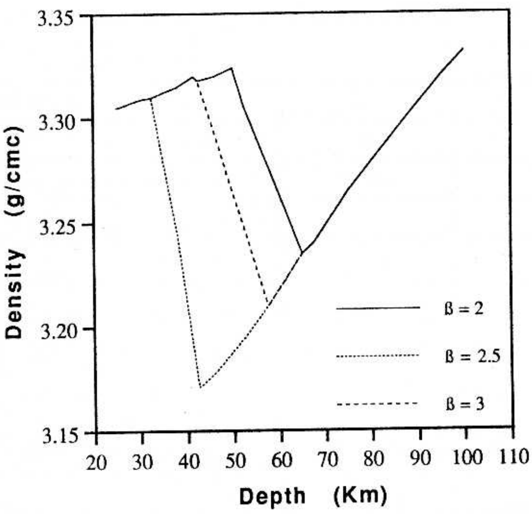

Fig. 8 — Density of an adiabatic upwelling mantle as a function of the depth, calculated for different β values Tp=1480°C; lithospheric thickness=110 km.

High Tp values (1450-1500 °C) characterize active rifting dynamics (hot rising jet), while low Tp values (1280-1350 °C) would be typical of passive rifting, caused by an extensional stress regime induced by driving forces external to the rifting system and related to the plate drift. The model proposed by McKenzie and Bickle (1988) allows calculating in both cases the geotherms and the changes in melting degree. Such melting processes can only start at the depth where the geotherm crosses the solidus of the mantle rock. Moreover, this model suggests that part of the melt previously extracted from the uprising mantle rocks consolidates into or beneath the crust, causing the generation of igneous rocks whose densities and thicknesses depend on Tp values.

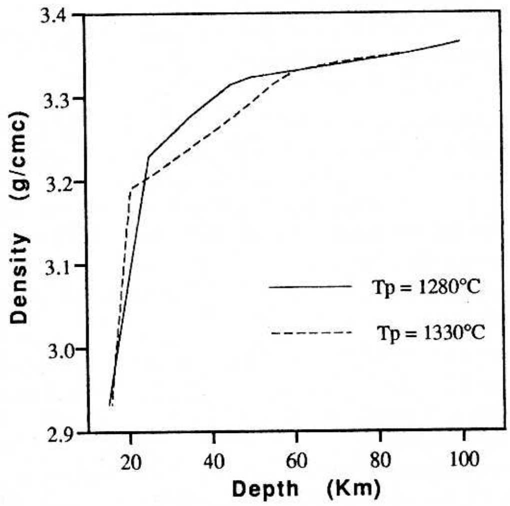

Fig. 9 — Density of an adiabatic upwelling mantle as a function of the depth, calculated for $\beta >> 5$; Tp=1280°C and Tp=1330°C; lithospheric thickness=110 km.

Table 3 — Mantle lithosphere - uprising asthenospere density contrast (g/cm^3).

AUTHORS	(g/cm^3)	AUTHORS	(g/cm^3)
Talwani et al. (1965)	−0.25	Baker and Wohlemberg (1971)	−0.05
Darracott et al. (1972)	−0.1	Yungul (1976)	−0.15
Fairhead (1976)	−0.05	Brown and Girdler (1980)	−0.05
Neugebauer (1983)	−0.1	Bermingham et al. (1983)	−0.05
Tamsett (1984)	−0.6-0.12	Izzeldin (1987)	−0.05
Girdler (1991)	−0.07		

Gravity modelling

The model of McKenzie and Bickle (1988) was used to study both the Ethiopian rift and the Gulf of Aden.

In the first case, a 1000 m uplift, a β value of 2.5 and lithospheric thickness of 110 km were assumed. The first value was chosen on the basis of various geological evidences. Most authors (Le Bas, 1971; Burke and Whiteman, 1973; Spohn and Schubert, 1982; Mohr, 1986) suggest an average uplift of about 1000 m. A higher value (2000 m) was proposed by Almond (1986), but seems to be excessive, since when applying the model suggested by White and McKenzie (1988), an uplift not greater than 1300-1400 m is expected, even taking into account the highest expected β values (4-5) for a continental rift such as the Ethiopian. A 1000 m uplift is coherent with a β value between 2 and 3. Finally, the uplift values adopted for the Ethiopian Plateau lead to a Tp value of 1480 °C.

In the second case (Gulf of Aden), a β value higher than 5, and a subsidence value of 2300 m, lower than that characterizing the average depth of sea-bottom in oceanic basins, were chosen. In fact the narrow ocean floor in the Gulf of Aden is a little shallower than in the larger oceans. By adopting these values, a Tp of 1330 °C was deduced. This is slightly higher than those (1280 °C) generally assigned to other ocean floors cases. The value of 1330 °C was chosen also taking into account the local thermal regime, which is higher than usual. This is due to the presence of the spreading Sheba Ridge and the closeness of the Afar hot spot, as suggested by Tamsett (1984). It is confirmed also by the local heat flow (ranging from 125.3 mW/m^2 to 250.8 mW/m^2) which is on average higher than that of older ocean floors (Sclater, 1966; Langseth and Taylor, 1967; Evans and Tammemagi, 1974). A typical value (Tp=1280 °C) for oceanic ridges was assigned to the upwelling mantle at the Aden/Indian Ocean boundary, near the link between the Owen F.Z. and the Carlsberg Ridge.

With these Tp and β estimates, a reconstruction of the likely geotherms for the two studied cases and therefore the trend in melting degree with depth was possible. This information was then used to evaluate the density change in the asthenospheric uprise. As the density of silicate melt is lower than that of the solid fraction (in this case a garnet peridotite), a strong decrease in the density of the asthenospheric plume can be expected with increase in melting degree. However, the higher compressibility of the liquid fraction and, consequently, the different rate of density change between liquid and solid fraction under different pressure and temperature conditions must also be taken into account.

The change of pressure with depth was evaluated as suggested by Green and Libermann (1976) and by McKenzie and Bickle (1988). Therefore, the normal increase in density of the solid mantle lithosphere, which depends only on the pressure increase, was taken into account following the suggestions by Dziewonsky et al. (1975) on the basis of seismological data. The variation in the melt density with temperature, pressure and composition was estimated using the Herzberg (1987) method. In our modelling, a liquid fraction with MORB composition was chosen as melt, but its density is very similar to that of a liquid fraction generated by melting of other suggested mantle rocks in the pressure and temperature range of interest (alcali basalts and olivine tholeiites) (Takahashi and Kushiro, 1983). Finally, the vertical density variation,

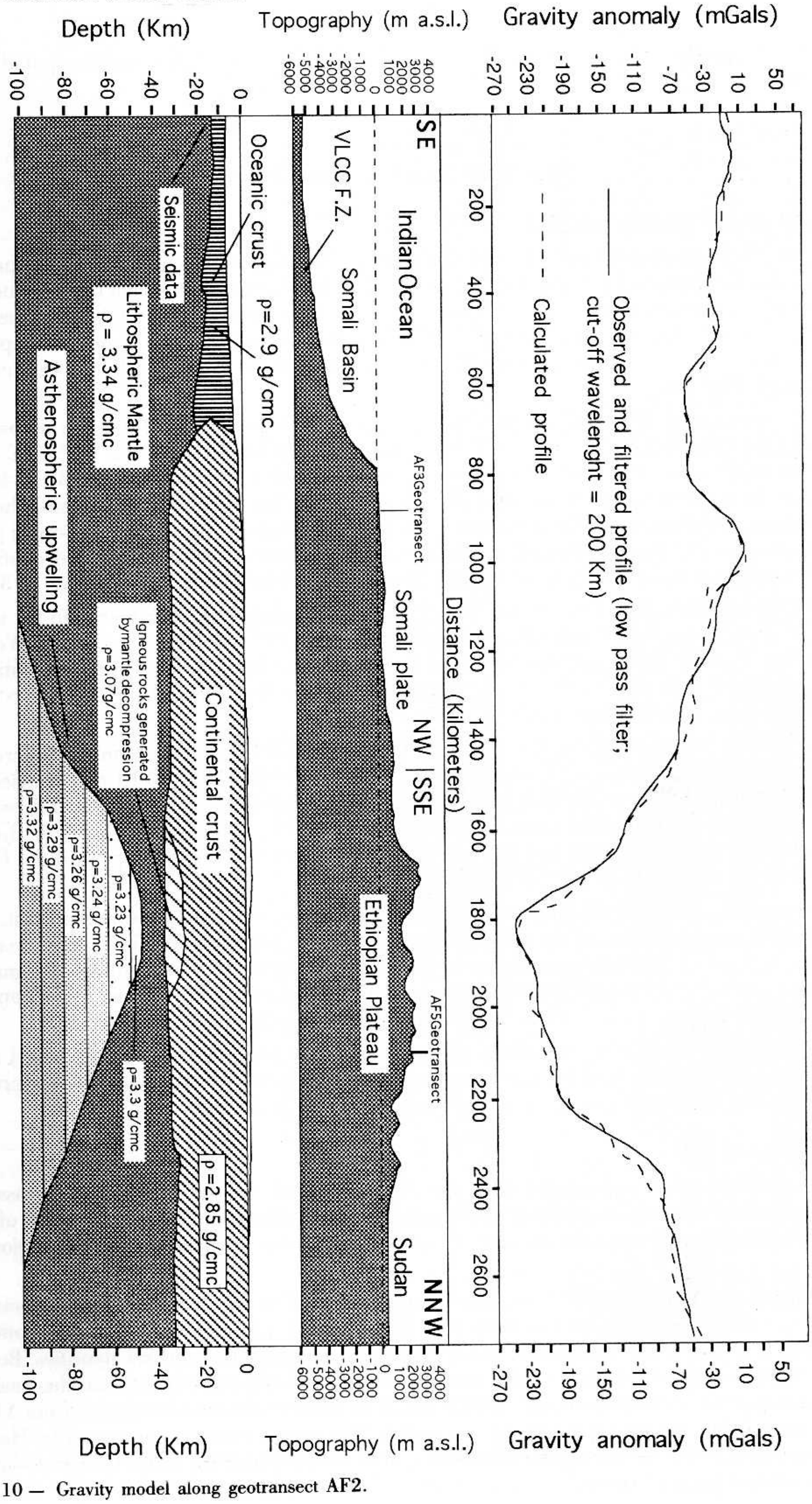

Fig. 10 — Gravity model along geotransect AF2.

depending on both pressure and temperature and on chemical composition, was computed (Figs. 8 and 9).

RESULTS AND DISCUSSIONS

Location of the gravity profiles

The modelling was done along three main geotransects suggested and afterwards approved by the Global Geotransect Project (G.G.P.) (Fig. 1). Their locations are significant from a structural point of view, and constrain the interpretation anomaly profiles with geophysical data from seismic refraction and/or reflection. The interpretation of the anomaly profiles was done by comparing the calculated gravity anomaly profile with the observed profile low-pass filtered with cut-off wavenlength 200 km.

AF2 Geotransect (Fig. 10) extends in a WNW-ESE direction for 2300 km across the Somali/Africa plate margins. It starts from the Ethiopia-Sudan boundary (Long. 35°05'E - Lat. 12°15'N) and crosses the whole Ethiopian Plateau and Rift Valley as far as 4°35'N-40°E; from here the transect continues in a S-E direction crossing the Mandera Lugh Basin, che Bur Acaba region, the Mogadishu Basin and the eastern continental escarpment of the Somali plate. Then, it crosses the Western Somali Basin at its deepest part (over 5250 m) and continues towards the Seychelles plateau in the Indian Ocean to the end (Long. 52°E - Lat. 0°37'S).

AF3 Geotransect (Fig. 11) extends in a N-S direction along the 45th meridian for about 2000 km. It starts from the Yemen-Saudi Arabia boundary (17° Lat. N) and crosses the Yemen Plateau, the Gulf of Aden and the Northern Somali passive margin. The transect continues southwards through the Ogaden Basin, the Oddur Arc and the Mogadishu Basin. Then it extends into the Western Somali Basin to 2° Lat. S.

The AF5 Geotransect (Fig. 12) is 2780 km long, has a N80°E direction and again crosses the Africa/Somali plate margin. It starts from the Sudan (8°50'N-33°30'E) and crosses the Ethiopian Plateau (65 km South of Addis Ababa) and the southern part of the Afar Depression. Then it continues parallel to the northern passive margin of the Somali plate, enters the Indian Ocean crossing the Error ridge and the intersection between the Sheba Ridge and the Owen Fracture Zone, and ends at 12°53'N-59°15'E.

The densities and the thicknesses of the LID, of the oceanic and continental crust, and the vertical density variation of the upwelling asthenosphere beneath the rifting and spreading axes were estimated. The latter was discretized for computer purposes by modelling the mantle plume as a set of several superimposed layers, each 10 km thick; a density value, corresponding to the average depth at which each layer is located, was assigned to each one.

The results of the gravity modelling, using a 2.5-D inversion method are discussed zone by zone in the following section, focussing mainly on the results relative to the sub-crustal structures. The results on crustal structures are discussed in Rapolla et al. (1995).

The Ethiopian Rift

The anomaly profiles along the AF2 and AF5 geotransects (Figs. 10 and 12) cross the large long-period negative anomaly centred on the Ethiopian Plateau. The interpretation of this anomaly is linked to the assumed presence of an asthenospheric upwelling under the Ethiopian Rift.

Fairhead and Reeves (1977), relating Bouguer anomalies and the time delay of seismic waves, suggest the presence, in the Ethiopian Rift area, of a thinned lithosphere (about 50 km of thickness) and of an "anomalous mantle" with high temperatures and low densities. Bonjer et al. (1970) suggested a 40-km - thick crust north of Addis Ababa whereas, on the basis of seismic prospecting, Berckhemer et al. (1975) predict a crustal thickness decreasing from 41-42 km, south of Addis Ababa, to 30-31 km westward. The same values are proposed by Hebert and Langston (1985). Analyzing the same data, Makris and Ginzburg (1987) suggest sligthly larger thickness crustal values.

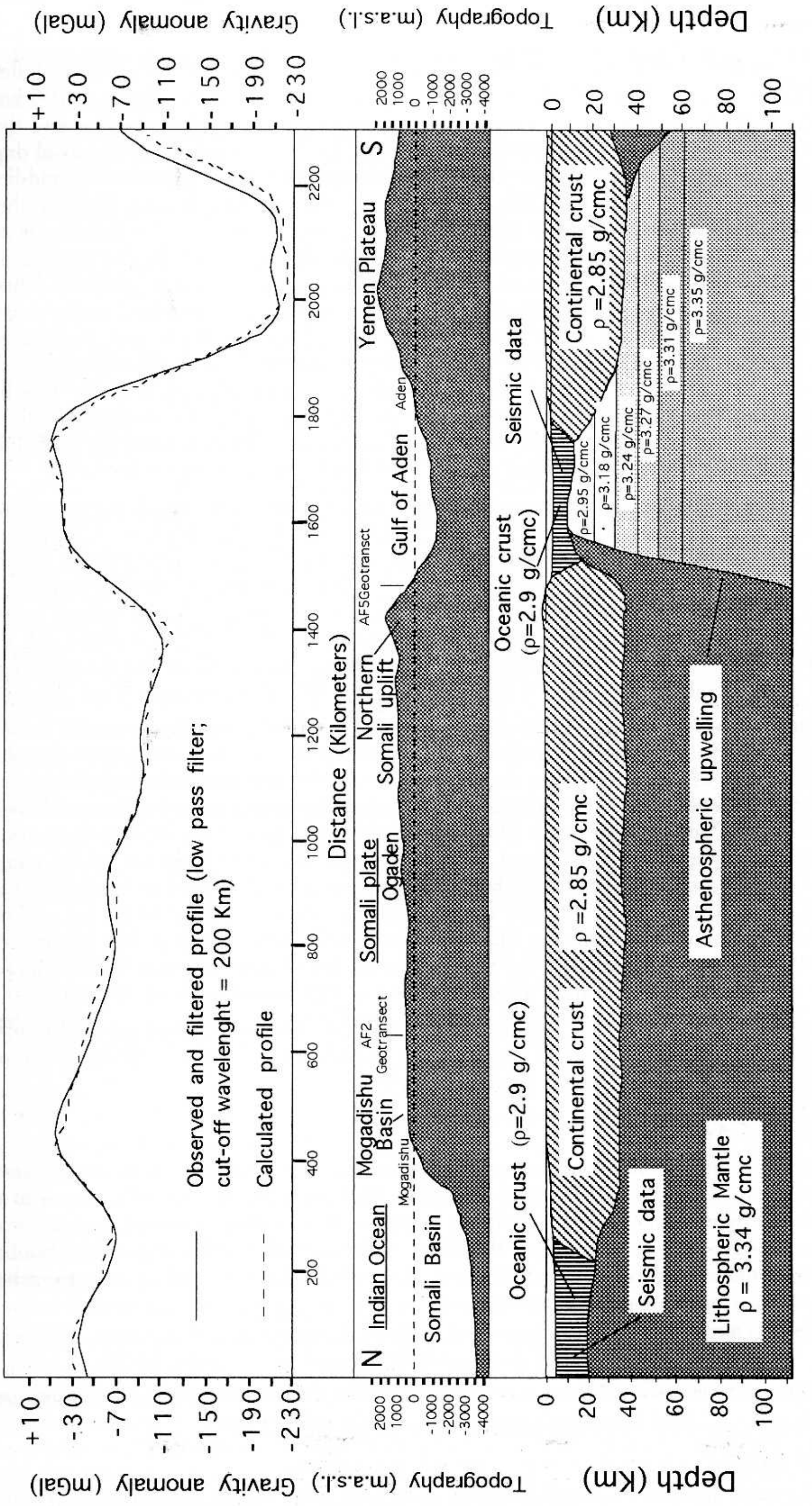

Fig. 11 — Gravity model along geotransect AF3.

In agreement with the above mentioned data on the structure of the area, modelling along the AF2 Geotransect (Fig. 10) shows a gradually uprising LAB from a depth of 100 km, which was assumed for the undisturbed areas. At this depth, partial melting begins and the plume is about 1200 km wide. The minimum depth of the LAB is 35 km under the axial depression of the Rift Valley. Therefore, more than 50% lithospheric thinning under the middle part of the Ethiopian Rift is suggested. The LAB does not reach crustal depths, because the crustal thickness locally maintains values of around 30 km. Hence, most of the thinning is recorded by lithospheric mantle since the expected crustal thinning is hidden, in the modelling, by the presence at the base of the crust of a body with an about 3.07 g/cm ^{3} density. This should correspond to the underplating igneous (mafic/ultramafic) rocks generated at the crust/mantle boundary, as expected in the McKenzie and Bickle (1988) and White and McKenzie (1988) model, by consolidation of part of the upward migrating melt. This implies the addition of large volumes of new crustal rocks to the previously thinned and torn original crust. This body has a thickness of about 8-9 km and extends for over 500 km along the profile direction, which is orthogonal to the rift axis. These rocks are considered to be generated by consolidation of magma extracted from the uprising mantle and which subsequently migrated.

The progressive upward decrease in density of the asthenospheric plume comes out in the modelling as 0.14 g/cm ^{3}, ranging from 3.35 g/cm ^{3} at 100 km depth to 3.23 g/cm ^{3} at 50 km depth. The trend inverts upwards, showing a strong decrease in melting up to a depth of about 40 km, where density reaches 3.3 g/cm ^{3}. From here on, the melting reduces and the density decrease of the mantle material only depends on the dilatation caused by decrease in pressure. Besides being very small, this is uniform in the horizontal plane and therefore has no influence on the gravity anomaly. The resulting density values for the uprising mantle in this area are shown in Table 4 for the AF2 and AF5 Geotransects (Figs. 10 and 12):

Along the AF5 Geotransect (Fig. 12) also, modelling the subcrustal structure of the Ethiopian Plateau results in a gradual LAB uprise, from a depth of over 100 km, where the plume has a lateral extent of 1200 km, to a depth of 55 km under the southern boundary of the Afar Depression. Therefore, a 50% lithospheric thinning also characterizes this area. Moving from the Ethiopian Plateau to the Afar Depression, the crust thins from 40 km to about 30 km. Here the gravity model assigns about 30% of the total crustal thickness to a 3.07 g/cm ^{3} body located at the base of the crust, which is also here interpreted as igneous basic rock extending laterally for 200 km. Thus, the Afar crustal thickness first decreased to 20 km and then was thickened by igneous underplating. This structure has also been recognised by seismic refraction prospecting recently carried out in the Afar Region by Makris and Ginzburg (1987) who proposed a new interpretation of the data already studied by Berckhemer et al. (1975).

Finally, it can be stressed that the good gravity fit obtained on the basis of the previously chosen high Tp value (1480 C°) confirms the hypothesis of "active" rifting as the origin of the extensional dynamics in the Ethiopian region.

The Gulf of Aden

The structure of litho-asthenosphere system in the Gulf of Aden, as deduced from interpretation of the anomaly profile along the AF3 geotransect (Fig. 11), seems to indicate great structural and evolutional complexity in this area. Along this transect, a negative anomaly (-190 mGal) is shown, centered on the Yemen plateau noticeably north of the Sheba Ridge spreading center, and replaced southwards by a relative high extending over the whole Gulf of Aden. Also in this case, the presence of an asthenospheric rise is assumed in the gravity modelling, as already recognized by several Authors (Laughton and Tramontini, 1969; Tamsett, 1984; Stein and Cochran, 1985; Isaev, 1987; Tamsett and Searle, 1990).

Melting degrees are referred to a model in which the following parameters are assumed:

a) more than 100 km thick original continental lithosphere;

b) β factor $>>5$ (as expected in an oceanic basin);

c) Tp value of 1330 °C, which is slightly greater than that typical for oceanic ridges (1280°C), as already explained in sect. 3.3.

Table 4 — Density of the upwelling asthenosphere near the Ethiopian Rift obtained resulting from gravity modelling.

Depth range of the layer	Density	Depth range of the layer	Density
110-100 km	3.35 g/cm^3	70-60 km	3.23 g/cm^3
100-90 km	3.32 g/cm^3	60-50 km	3.29 g/cm^3
90-80 km	3.29 g/cm^3	50-40 km	3.3 g/cm^3
80-70 km	3.26 g/cm^3		

Choosing a geoterhm corresponding to the quoted Tp value, a melting trend, notably different from that of the Ethiopian rift, is to be expected. The petro-physical model indicates that partial melting starts only at a depth of 60 km and increases up to a value of 25% at a depth of 10 km, near the base of the oceanic crust. Consequently, the density decreases quickly upwards from 3.33 g/cm^3 at 60 km to 2.95 g/cm^3 at the base of the oceanic crust. The modelling suggests, on the other hand, a decrease in density controlled only by pressure from 110 to 60 km. Therefore, a gradually ascending LAB from more than 100 km under the Somali plate to 60 km under the coastline of Northern Somali was modelled, while northwards the depth of LAB strongly decreases, reaching 30 km under the boundary between the Somali continental crust and Gulf of Aden oceanic crust. The minimum depth (10 km) is finally reached near the oceanic ridge of the Gulf of Aden, corresponding to the relatively younger asthenospheric uprise. Toward the continental crust of the Arabian plate, the LAB depth increases more gently than in the southern part, reaching a depth of 35 km. This value remains constant under the Yemen plateau of the Arabian passive margin and starts to increase again 300 km north of the coastline. The Moho depth decreases strongly, passing from 35 km under the continental Somali crust to less than 10 km under the Gulf of Aden oceanic ridge, where it approximately coincides with the LAB. Northwards the Moho depth starts to increase again under the continental crust of the Arabian plate, reaching a depth of 35 km. The density values obtained for the upwelling asthenosphere are summarized in Table 5.

The results show a partial agreement with Isaev's (1987) gravity and magnetic interpretation and with the study by Laughton and Tramontini (1969) based on seismic profiles taken in the Gulf of Aden (profiles 6239 and 6235-6236).

Rather small Tp values, like those in agreement with the modelling (1330 °C), indicate that a thermal anomaly may also be caused by a tensional stress regime, due to external driving forces, in which the Moho's upward migration over the underlying uprising asthenosphere is "passive" (that is does not need the presence of a hot plume). In the case of the Gulf of Aden, the presence of transtensional stresses is clearly recorded by the lateral displacement of Sheba Ridge segments, and could be caused by northward drifting of the Arabian plate (Girdler and Styles, 1978) with a counterclockwise component of rotation around a pole located in the southeastern Mediterranean Sea (Girdler, 1983). This rotation could be related to the Indian plate's northward drift (McKenzie and Sclater, 1971) which coincides in time with the Gulf of Aden rifting, and leads to the hypothesis that when the Owen Fracture Zone became locked, the spreading began again on the Carlsberg Ridge, and the Arabian plate was torn off of the African plate and carried northward. But this hypothesis is put in question by the features of the slight seismic activity recorded in the northern branch of the Owen Fracture Zone (Banghar and Sykes, 1969).

The most remarkable features of our model are the large lateral extent of the upwelling mantle underlying the Sheba Ridge of the Gulf of Aden and, mainly, by its extremely asymmetrical morphology. This asymmetrical location (slightly northwards, near the southern passive margin of the Arabian plate) can be also observed directly on the large, long-wavelength negative anomaly. A similar feature was recognized by Vink (1984) with regard to Iceland, and by Dixon et al. (1989) for the Red Sea. A possible explanation of this peculiarity, suggested by Dixon et al. (1989), is that the tensional stress value necessary (200 MPa- 2 Kb) to induce a continental break-up is greater than that developed in a passive or active rifting; that is, rifting takes place

Table 5 — Density of the upwelling asthenosphere near the Gulf of Aden obtained resulting from gravity modelling Tp = 1330 °C; $\beta >> 5$.

Depth range of the layer	Density	Depth range of the layer	Density
110-100 km	3.37 g/cm^3	60-50 km	3.31 g/cm^3
100-90 km	3.36 g/cm^3	50-40 km	3.275 g/cm^3
90-80 km	3.35 g/cm^3	40-30 km	3.24 g/cm^3
80-70 km	3.34 g/cm^3	30-20 km	3.185 g/cm^3
70-60 km	3.33 g/cm^3	20-10 km	2.95 g/cm^3

only near a region of previous lithospheric weakness. In the case of the Gulf of Aden, a belt of E-W weakness has been recognized across the African plate, from the Gulf of Guinea to the present Gulf of Aden (Cornacchia and Dars, 1983) and does not coincide with the maximum stress area. Similarly, the tensional stresses in the Gulf of Aden along this weakness belt (located near the Afar hot spot but not necessarily related to it) could have played a capture role on the convective flow related to the nearby northern plume, causing an asymmetrical morphology. The coexistence of active processes (as occurred in the Afar Depression and the Ethiopian Rift Valley) with passive (sea floor spreading in the Gulf of Aden and the Red Sea) has already been hypothesized by Almond (1986).

The Owen Fracture Zone

With regard to the area between northeastern Somali and the Owen Fracture Zone, the gravity anomaly profile along geotransect AF5 (Fig. 12) shows a positive anomaly under Socotra Island, followed eastwards by a minimum area and by a gravity high under the intersection between the Owen Fracture Zone and the Sheba Ridge.

The gravity modelling, carried out along Geotransect AF5 (Fig. 12), confirms the presence of the continental Somali passive margin up to about 200 km ENE of Socotra Island (Beydoun, 1970; Beydoun and Bichan, 1970; Merla et al., 1973). As a result, the Moho under this region lies at about 35 km depth. The presence of continental crust in Socotra Island is also suggested by various evidences, like the shallow sea-bottom between Socotra and Cape Guardafui, and the eastward continuation of the Magnetic Quiet Zone north of Socotra.

The V3617 cruise carried out east of Socotra gives information only on the depth and nature of the oceanic basement. However, studying the evolution of the ocanic crust between the Error Ridge and the Sharbitat Ridge, Stein and Cochran (1985) recovered a model developed by McKenzie (1978), Royden et al. (1980) and Steckler (1981) which, by taking into account the vertical motion of a thinning lithosphere, allowed them to estimate:

a) a β factor higher than 5;

b) a thickness of crust in the range 2.5-3.5 km and of LID equal to 25 km.

These values have been also confirmed by heat flow measurements and seismic prospecting carried out during the VEMA 3617 (1980) cruise. In the present study, a β value >5 and a low Tp value (1280 °C) typical of oceanic crust have been chosen. An initial lithospheric thickness, before the start of lithospheric thinning, of about 100 km was assumed. Thus, an upward decreasing density trend like that shown in Fig. 9 was obtained.

Our model predicts a gently uprising Moho, reaching a depth of 20 km east of Socotra, where the thinned continental passive margin links with the oceanic crust. Eastwards, oceanic Moho rises up to 15 km under the Error Ridge, and to a depth less than 10 km below the connection between the Owen Fracture Zone and the Sheba Ridge. East of the Owen Fracture zone, the thickness of the Indian oceanic crust begins once again to increase gradually. The agreement between the results of our model and that adopted by Stein and Cochran (1985) seems good, except for the slight difference between the density values (2.8 g/cm^3 and 2.9 g/cm^3 respectively). The density distribution in the upwelling asthenosphere was assigned on the basis discussed in the previous cases; it is reported in Table 6.

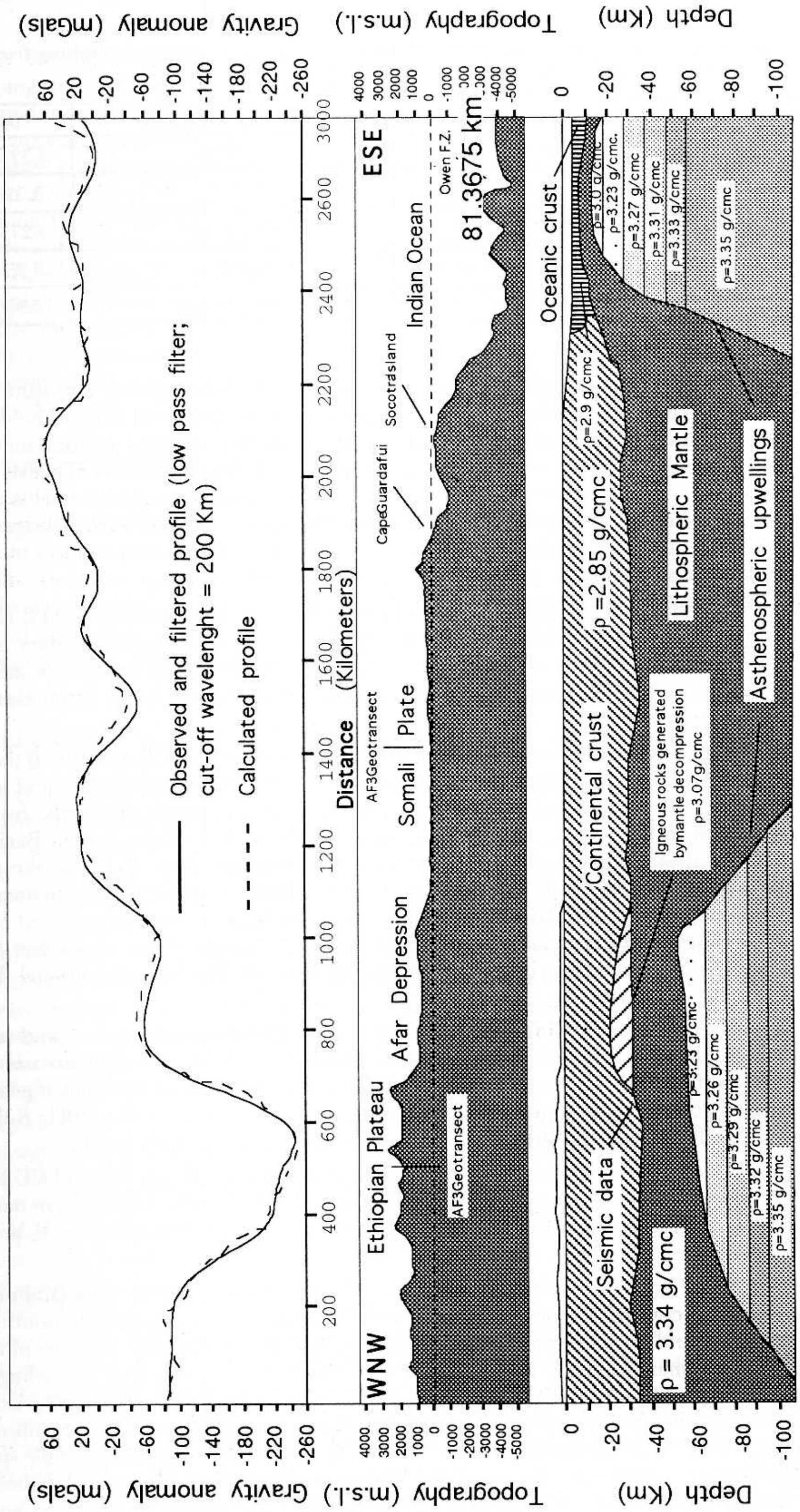

Fig. 12 — Gravity model along geotransect AF5.

Table 6 — Density of the uprising asthenosphere near the Owen F.Z. obtained resulting from gravity modelling Tp = 1280°C; $\beta >> 5$.

Depth range of the layer	Density	Depth range of the layer	Density
110-100 km	3.37 g/cm^3	60-50 km	3.325 g/cm^3
100-90 km	3.36 g/cm^3	50-40 km	3.31 g/cm^3
90-80 km	3.35 g/cm^3	40-30 km	3.275 g/cm^3
80-70 km	3.34 g/cm^3	30-20 km	3.23 g/cm^3
70-60 km	3.335 g/cm^3	20-10 km	2.95 g/cm^3

The results of the gravity interpretation show an asthenospheric uprise, 700-800 km wide at its base (100-110 km of depth), reaching the oceanic crust under the ridge. The estimated melting trend is similar to that of the Western Sheba Ridge and shows a gradual increase up to the thinned oceanic crust. The fit between observed and calculated data suggests that the Tp (1280 °C) and β values ($\beta >> 5$) chosen in the interpretation are correct. This confirms the passive nature of the oceanic rifting related to the eastern end of Sheba Ridge.

The Somali Basin

Gravity modelling was carried out in the Somali Basin along Geotransect AF2 (Fig. 10). The profile shows a positive anomaly under the continental-oceanic crust boundary along the eastern coastline of the Somali region in the Indian Ocean. Seawards, a negative anomaly is present, followed by gravity anomaly values gradually increasing up to the south-eastern end of the AF2 Geotransect.

The transition from continental to oceanic crust is located one hundred about off the Somali coastline in the Indian Ocean. Our model suggests a continental crustal thickness of about 30 km under the coastline. The oceanic crustal thickness decreases south-eastwards, from about 15 km to 5-7 km in the middle and deepest (over 5000 m) part of the Somali Basin. Then the Moho depth starts to gradually rise once again reaching less than 10 km at the endmost part of the profile. The suggested model is in agreement with the seismic results obtained during the R/V Vema (legs 3618 and 3619) and R/V Conrad (leg C1215) cruises reported by Coffin et al. (1986). These data indicate in fact the presence of oceanic crust about one hundred km beyond the Somali coastline in the Indian Ocean (Lat. 2° South) and a crustal thickness ranging from 12 to 15 km.

It is important to note that in order to obtain a good fit between measured and modelled gravity data, an upward migration of the LAB and an asthenospheric upwelling are not needed. Thus the Somali Basin does not seem to show a LID thinning, although on the basis of geophysical data (Francis, 1964; Francis et al., 1966; Lort et al., 1979; Scrutton et al., 1981; Rabinowitz et al., 1983; Coffin et al., 1986), it should be considered an oceanic basin.

This problem could be explained by N-S tectonic lineaments (Dohw F.Z., VLCC F.Z. and ARS F.Z.) which, initally only suspected (Schlich et al., 1972; Schlick, 1974), have been since clearly pointed out (Bunce and Molnar, 1977) on the basis of data obtained by the R.V. Atlantis II in April and May 1976.

The features of these fracture zones and the evidence from magnetic data (Rabinowitz et al., 1983) not only confirm the original position of Madagascar as being near the south-eastern Somali continent, and its subsequent southward drift, but also explain the absence of thinning in the lithospheric mantle, under the Wester Somali Basin. In fact, sea floor spreading started in the Somali Basin between 159 (Haq et al.; 1987) and 141 My B.P. (Besse and Courtillot, 1988) and finally stopped between 100 My B.P. (Bunce and Molnar, 1977; Scrutton et al., 1981) and 83 My B.P. (Besse and Courtillot, 1988) so indicating the very old age of the spreading phase. This is also confirmed by the great thicknesses of sedimentary cover which almost totally hide the fracture zones (Dohw F.Z., VLCC F.Z. and ARS F.Z.) related to the ocean spreading (Bunce and Molnar, 1977).

Thus, the low heat flow values, ranging from 43.8 mW/m^2 to 69.3 mW/m^2 (Sclater, 1966; Langseth and Taylor, 1967; Birch and Halunen, 1966) characterizing this area, can be explaned by a return of the local thermal regime to normal conditions. This could be related to the disappearance of the LAB rise; in fact, if the LAB represents the depth at which the upper mantle temperature reaches the solidus, a downward migration of the geotherms would cause also a sinking of the LAB towards normal depth values and a consequent lithosphere rethickening (McKenzie, 1978; Mongelli, 1991). Taking into account the age of the end of ocean spreading in the Western Somali Basin (between 100 My and 85 My), the actual depth of the LAB, as estimated for the gravity model described in Fig. 10, is in good agreement with the lithospheric rethickening (about 90 km) predicted by the Parker and Oldemburg (1973) model. The fossil nature of sea floor spreading in the Somali Basin, which probably represents one of the oldest parts of the Indian Ocean (Norton and Sclater, 1979; Besse and Courtillot, 1988) is confirmed by the results of gravity modelling (Fig. 10) which indicates the total absence of a rising LAB below the Western Somali Basin.

The Somali continental crust

The crustal thicknesses in the Somali region can be considered typical for a continental crust composed of a crystalline and methamorphic basement of pre-Cambrian age rejuvenated by tectonic and metamorphic events of Pan-African age (Dal Piaz and Sassi, 1986; Shackleton, 1986). This basement appears partially hidden by a Meso-Cainozoic sedimentary cover (Bosellini, 1989). In our models, the Moho is located at depths that generally change little and range from about 30 to 35 km. An exhaustive discussion on the modelling of the Somali plate crustal structure, based on gravity data interpretation, is reported in Rapolla et al. (1995).

CONCLUSIONS: ACTIVE AND PASSIVE RIFTING, A MEANS OF RECOGNITION

With regard to the aim of this research, the limits of interpretational ambiguity are considered here to have been strongly reduced. In fact, several gravity model variables were constrained and values assigned to them according both to data from geophysical prospection and to predictions of a good theoretical density model. For this aim, a petro-physical model (McKenzie and Bickle, 1988; White and McKenzie, 1989) was chosen. This model relates the thermal regime, and hence the change in degree of partial melting and the chemical features of the magma, to vertical motion of the earth's surface and to the degree of lithospheric thinning. By integrating this model with the density variation of a silicate melt suggested by Herzberg (1987), and according to pressure, temperature and chemical composition, the density trends of an asthenospheric plume have been inferred. The results were applied to a 2.5-D modelling along the geotransects. Only the lateral morphology of the mantle plume controlled the best fit between observed and modelled data.

Morphological changes in the asthenospheric upwelling, that is the different lateral extents of the mantle plume under the East Sheba Ridge (less than 800 km) and under the Ethiopian Plateau (over 1200 km), are probably not casual. A recent model (Keen, 1985) suggested on the basis of the rheological and thermodynamical aspects of the advective asthenospheric upwelling, gives a clue to distinquishing rifts related to active and to passive mantle upwelling. It predicts that a mantle plume due to a strong thermal anomaly, typical of active mechanisms, shows a much larger lateral extent than one related to passive rifting. In this last case, the lithospheric thinning is concentrated into a much smaller area. The Keen (1985) model, in particular, predicts that the lateral extent of lithospheric thinning, related to an active rifiting, is about 5 times the plume width at the base of a normal lithosphere. The results of our modelling lead to similar values; in fact, modelling of the asthenospheric upwelling under the Ethiopian Rift not only shows the features of active dynamics, from both the thermal and petrological points of view, but is also characterized by a very large lateral extent, and thus accords with the morphological features of an active mantle plume, as expected by Keen (1985).

The diameter of the lithospheric volume involved in the anomalous thermal regime, inferred from gravity modelling, extends here for over 1000 km; that is more than 5 times the diameter of the source hot spot, suggested to be about 200 km by White and McKenzie (1989).

A similar agreement, between the small lateral extent of the mantle rise near the Sheba Ridge, estimated using the same concepts, and the conclusions of the Keen model (1985), was found, indicating a local passive mechanism.

In conclusion, the present study may confirm the presence of active mantle dynamics as the cause of the Ethiopian Rift, and suggests a lithospheric thinning by at least 50% of the original thickness; it indicates, on the other hand, passive asthenospheric upwelling as generating the sea floor spreading in the Gulf of Aden, and a local thinning of the lithosphere to less than 10 km. The study has also indicated a possible asymmetry of the asthenospheric upwelling body below the spreading axis of the Gulf of Aden, implicitly suggesting asymmetric extensional mechanisms for the previous rifting stages. The presence of an old normal oceanic crust in the Somali Basin is confirmed, and the absence of mantle upwelling is suggested together with the fossil nature of the sea floor spreading. The modelling also confirms the continental nature of the crust beneath Socotra Island, the submerged passive margin of the north-eastern Somali plate, and the presence of an asthenospheric rise under the oceanic crust near the Owen Fracture Zone.

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