The West-African craton margin in eastern Senegal : a seismological study

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ABSTRACT. A vertical short period seismological array was operated for six months in eastern Senegal. Large *P* wave travel-time anomalies are in fairly good relation with the gravity and geological features. Two-dimensional inversion of the data shows the existence of a major vertical discontinuity extending from the surface to 150-200 km depth. The other heterogeneities are mainly located in the crust and related to specific segments of the regional geology : craton, Mauritanides and Senegalo-Mauritanian basin. The main discontinuity dipping to the east is interpreted as the trace of an old subduction slab. We propose the following geodynamical process to explain the formation of the Mauritanides orogenic belt : continental collision after opening of a back-arc marginal basin in late Precambrian and its closure until Devonian.

Key words : craton margin, two-dimensional inversion (2D-inversion), continental collision

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INTRODUCTION

During the last years, several seismological studies have clearly shown that structural differences extend deep in the lithosphere and perhaps more deeply, in the asthenosphere (see Aki (1982) for a review) between cratonic areas (stable since at least 1.500 My) and continental platforms which have been tectonized more recently ; we can mention, for example, Cara (1979) for the U.S.A. from Rayleigh waves dispersion, Romanowicz (1979, 1980) for the U.S.A. and Europe, Hovland *et al.* (1981), Taylor and Toksöz (1979) from *P*-wave travel-time residuals. Our purpose was to find out if such differences exist between the West-African craton and the Senegalo-Mauritanian basin. With this end in view the University of Leeds and the Office de la Recherche Scientifique et Technique d'Outre Mer (O.R.S.T.O.M.) operated a seismological array in eastern Senegal from October 1978 to April 1979. We selected this region on account of the presence of a permanent seismological station at Kedougou (KDS) and because a previous work (Dorbath C. and L., 1979) had shown large variations of mean *P*-residuals. Moreover geological data let us suppose that in this region the margin of the craton is less disturbed by the tectonic features of the Mauritanides than in the northern part which is characterized by large thrusts (Sougy, 1962 ; Lecorché, 1980). This array extended from the West-African craton which outcrops in the Kenieba inlier to the Senegalo Mauritanian sedimentary basin, cutting through the south end of the Mauritanides orogenic belt (figs. 1 and 2).

GEOLOGICAL SETTING

The main part of West Africa consists of the West-African craton (fig. 1), an old metamorphosed and granitized basement stable until about 1 700 My. To the west, this craton is fringed with tectonized and metamorphozed rocks, the main part of them forming the Mauritanides orogenic belt (Sougy, 1962). The direction of this belt is about north-south, and its length more than 1 000 km. Its most internal part is hidden by the tabular mesozoic to present sediments of the Senegalo Mauritanian basin.

The craton

The basement of the West-African craton outcrops along two large uplifts : the « dorsale de Man » in the south and the « dorsale de Reguibat » in the north, and in the two little zones of Kayes and Kenieba (Bessoles, 1977).

After 1.700 My the craton was crossed by long faults generally trending east-north-east. Among them, the fault of Kayes is interpreted as a major accident crossing all the craton (Simon *et al.*, 1981), and it was active during the whole Proterozoic and Paleozoic times. This basement is covered with detrital series ranging between upper proterozoïc and carboniferous, forming the Taoudeni basin. These sediments recorded slight deformations of the basement and show the influence of a far pan-african orogenesis. Among these sediments the late precambrian mixtite of glacial origin is an important stratigraphic level.

The Mauritanides orogenic belt (fig. 1)

Between the stable block and the orogenic belt *stricto sensu* the first deformation on a regional scale is an early bending of the basement (late-Precambrian) observed as well in Adrar (Trompette, 1973 ; Lecorché, 1980) as in the Koudougou region (Bassot, 1966). This bending is shown by the uncomformity of the late-Precambrian mixtite on the Proterozoic series and on the basement. This indicates that the thickness of eroded sediments is about 1 000 m.

The tectonic pattern allows Dia *et al.* (1979) to consider that the Mauritanides result in a polyphased deformation. This one is characterized north of the Kayes fault by at least two episodes of thrusting and a late sinistral strike-slip fault system (Le Page and Campredon, 1980). South of the Kayes fault the belt divided into two branches corresponding to two phases of deformation.

The seismological study described in this work was located in this segment. A cross-section (fig. 4) essentially due to Bassot (1969) shows from east-south-east to west-north-west :

1) Volcano-sedimentary formations, slightly metamorphised and folded during the Birrimian cycle (1.800 My). They are crossed by late-Birrimian granite (Bassot *et al.*, 1963) then by a basic atectonic magmatic series (Debat *et al.*, 1982) and in a few places by rhyolites (Bassot, 1966). It is the western margin of the outcropping craton.

2) A sedimentary series equivalent to the Taoudeni deposits and formed by :

— The late-Precambrian mixtite lying uncomfortably in east on the outcropping craton but interbedded in the west into a volcano-sedimentary basic complex.

— A green flyschoid series, very thick.

— A red detrital series, resting unconfortably upon the green series.

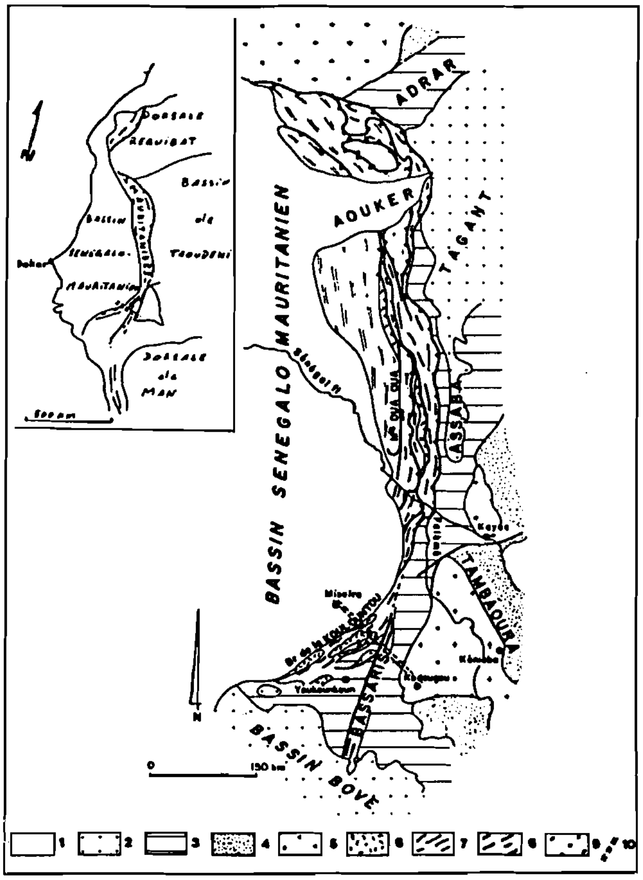

Figure 1

The main units of the Mauritanides (after Dia et al. 1979, *modified).*

1. *Post-Mesozoic sediments.*
2. *Ordovician to Devonian tabular sediments.*
3. *Upper Precambrian to Cambrian sediments tabular or gently folded.*
4. *Proterozoic tabular sediments.*
5. *Cratonic basement.*
6. *Acid-alkalin volcanism (Koulountou).*
7. *Pre-ordovician units of the Mauritanides later reactivated.*
8. *Post ordovician units.*
9. *Guidimakha granites.*
10. *Seismological array.*

Inlet : Structural pattern of western Africa.

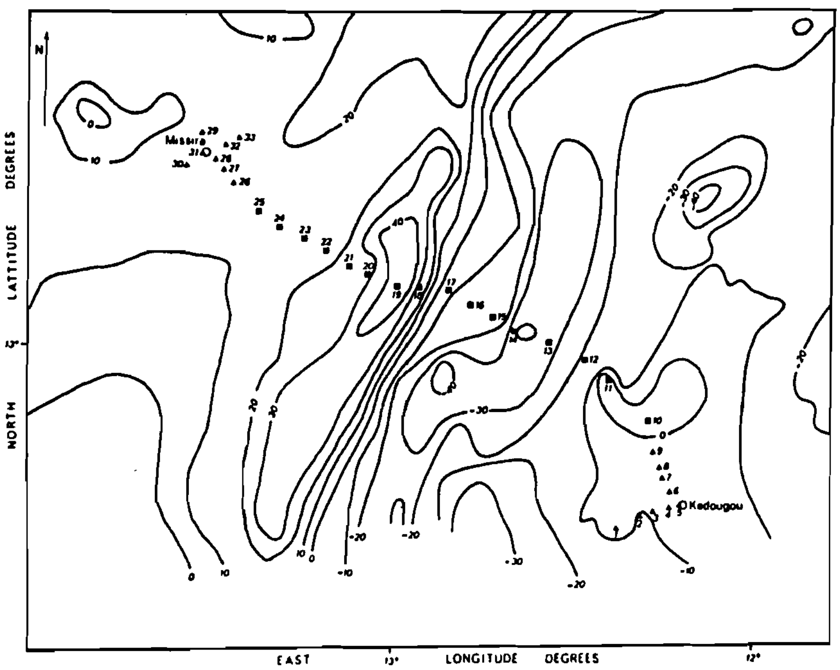

Figure 2

Seismological stations used in this study (permanent (6 months) station, temporary (1 month) station) superimposed on Bouguer anomaly contours.

The basis volcanism and the green series express the individualisation of a trough : the Faleme trough. These folded sediments constitute a large synclinorium eastward overfolded. It seems that some deformation of this synclinorium began prior to the deposit of the red detrital series.

3) Sericitoschists, chloritoschits and prasinites with few chips of serpentinites, affected by a polyphased tectonics (Chiron, 1964; Villeneuve, 1982). It is the Bassari ridge, a submeridian anticlinorium eastward overfolded, strongly deformed before the red detrital series deposit (Villeneuve, 1982). The Bassari ridge is limited in the west by a major submeridian fault (Bassari fault).

4) The Youkounkoun basin : it is a synclinorium, with, from top to bottom :

— Quartzite sandstones, probably from Ordovician, similar to those found at the base of the Bove basin, strongly folded after Silurian (Drot *et al.*, 1979).
— The red detrital series, very thick and folded under the quartzite sandstones.
— Red or green argilites, univalent with the green flyschoid series of the Faleme trough.
— Alcalin and calc-alcalin rhyolites and rhyolitic breccia, may be from late-Precambrian, very similar to the acid volcanic rocks observed in the west of the northern part of the belt.
— A basement made of gabbros transformed in amphibolites and granitized about 680 My ago (Bassot and Vachette, to be published).

5) The Koulountou ridge : it is formed with rhyolites and granite slices turning to sericitoschistes in the west. It is regarded as equivalent to the bottom series of the Youkounkoun basin and therefore as an anticlinorium. The flow clivage, at first vertical in the east, becomes less and less dipping in the west. This feature suggests that a tangential tectonics occurred west of Koulountou ridge; but these deformations were not clearly demonstrated in the outcropping part of the Koulountou ridge.

6) Horizontal detrital formations deposited between Mesozoic and Present. They constitute the Senegalo-Mauritanian basin that hides the western part or the Koulountou ridge.

Therefore the southern segment of the Mauritanides orogenic belt shows some important particularities. First the deformations are less severe than in the north. In this region the overturns to east or south-cast result at the very most in local scallings and one cannot actually speak of allochtonous units. On the other hand, both Caledonian and Hercynian tectonic stages are clearly distinct on the field. The Youkounkoun basin takes the part of a foreland towards the Hercynian structures of the Koulountou ridge. At last the outcropping basement of the craton in the Kenieba inlier which is limited northward by the Kayes fault, seems to have taken the part of a rigid block against which the Mauritanides structures have been moulded (Le Page, 1980). From this brief review one can point out that, at the edge of the stable zone which is considered as the craton by geologists, magmatism and sedimentation indicate the aperture of preorogenic troughs at the end of Precambrian. The basement, west of the Faleme trough, is intruded by acid volcanism which predates the closing of this trough. In middle Cambrian and later, the sedimentary formations record the progressive closing of the troughs. During this period the first structures occur in the external parts of the belt; the are reactived during the hercynian stage during which new thrusts appear in the internal part of the Mauritanides, north of the Kayes fault. Only some serpentine slices underlay the thrusted sheets. During this deformation, metamorphism remains very weak. On the other hand no actual ophiolitic sequence nor late orogenic granitisation were observed.

SEISMOLOGICAL STUDY

Two short period vertical (SPz) seismic arrays, each of about 20 km aperture, operated for the whole duration of the experiment : the nine stations Kedougou (KED) array on the craton and the eight stations Missira (MIS) array, 180 km to the north-west, on the Senegalo-Mauritanian basin. A third linear array of three or four SPz seismometers has been moved between these two permanent arrays, staying for one month in each of five successive positions, providing records at 10 km spacing along the profile. The coordinates and elevations of the permanent and temporary stations are given in table 1. Figure 2 shows the positions of the stations superimposed on the Bouguer anomaly map. Designed in that way, the seismological profile cuts through all known geological formations from the craton to the sedimentary basin and all gravity anomalies associated with the craton margin and the Mauritanides orogenic belt.

Table 1

Station	Latitude	Longitude	Elevation
1 IBL	12N 29.65	12W 22.46	168 m
2 BND	12N 31.83	12W 18.71	163 m
3 SIL	12N 32.59	12W 16.31	130 m
4 FAD	12N 33.22	12W 13.61	140 m
5 KED	12N 33.60	12W 11.81	124 m
6 CEN	12N 35.77	12W 13.43	164 m
7 NGA	12N 38.00	12W 14.81	134 m
8 TIN	12N 39.94	12W 15.40	153 m
9 MON	12N 42.73	12W 16.25	173 m
10 TIM	12N 47.48	12W 17.40	132 m
11 BAD	12N 54.11	12W 23.93	164 m
12 SIB	12N 57.59	12W 28.10	162 m
13 NIM	13N 00.41	12W 33.97	140 m
14 BEL	13N 02.23	12W 39.85	146 m
15 NIO	13N 04.72	12W 43.42	69 m
16 GFO	13N 06.66	12W 47.14	100 m
17 PLA	13N 08.90	12W 50.86	101 m
18 GTR	13N 09.79	12W 55.78	89 m
19 FAR	13N 09.77	12W 59.38	95 m
20 MUL	13N 11.69	13W 04.31	75 m
21 DDI	13N 13.09	13W 07.28	51 m
22 DAR	13N 15.65	13W 11.19	51 m
23 ABE	13N 17.62	13W 14.69	66 m
24 DIA	13N 19.63	13W 18.83	32 m
25 WAS	13N 22.33	13W 22.25	38 m
26 GBO	13N 26.91	13W 26.17	45 m
27 ARD	13N 29.16	13W 27.77	57 m
28 CPA	13N 30.84	13W 29.21	43 m
29 HAM	13N 35.16	13W 31.34	49 m
30 MAD	13N 29.80	13W 34.06	42 m
31 MIS	13N 31.80	13W 31.46	48 m
32 BAB	13N 33.32	13W 27.64	53 m
33 BAA	13N 34.24	13W 25.43	43 m

The seismological signal from each outstation of an array was amplified, modulated, radio-telemetered to a central station and recorded on a multichannel analogue tape. The internal time base of each recorder was controlled by means of time signals from B.B.C. World Service Transmissions. Then all the recorded events have been digitized at 50 samples per second and the relative arrival times automatically picked by waveform matching (see Briden *et al.*, 1982). The overall accuracy of the relative arrival times is estimated to be better than 0.05 s.

We used hypocenter determinations of the International Seismological Center and travel time residuals are calculated using the Jeffreys-Bullen tables. Only events, earthquakes or nuclear explosions, with epicentral distances greater than 30° recorded simultaneously by the three arrays and by more than half of the stations of each array have been retained. Events with epicentral distances between 137° and 150° have also been rejected to avoid possible misidentification among PKIKP, PKP and other core phases. Finally 119 events have been used, 92 with epicentral distances in the range 30°-100° and 27 with epicentral distances greater than 110°; the final data set includes 1 878 arrival times. Figure 3 presents the epicentres of these events as a linear fonction of distance and azimuth.

Mean station residuals and gravimetry

Given an event j recorded by a station i with an epicentral distance Δ_{ij}, the absolute residual is defined as :

$$R_{ij} = (T_{ij} - T_{0j}) - C_{ij}$$

where T_{ij} is the observed arrival time of P (PKIKP) wave at the station i, T_{0j} is the origine time of the event j, and C_{ij} is the computed travel-time for this hypocentre according to J.-B. tables. This term takes into account corrections for Earth ellipticity and station elevation. It is clear that absolute residual R_{ij} is affected by various errors and anomalies, including event mislocation and model biases.

The mean residual for event j is defined to be :

$$R_j = \left(\sum_i R_{ij}\right)/n_j$$

where n_j is the number of stations reporting P (PKIKP) arrival time for the event j. R_j includes travel-time errors due to mislocation and inhomogeneities in the source region and in the lower mantle which affect every ray from the focus to the array in the same way.

Relative residuals r_{ij} are calculated by reducing absolute residuals :

$$r_{ij} = R_{ij} - R_j$$

and at last mean station residual r_i is calculated for the station i :

$$r_i = \left(\sum_j r_{ij}\right)/n_i$$

where n_i is the number of events reported by station i.

Table 2 and figure 4 present the mean station residual for the 33 stations which were operated and the Bouguer anomalies (Creen and Rechenman, 1965). Negative residuals (i.e. high velocities) are observed on the craton, positive residuals (i.e. low velocities) on the external zone of the Mauritanides mainly formed of sedimentary, little or not metamorphosed formations : Faleme trough, Bassari ridge, Youkounkoun basin; then again negative residuals are found on the internal zone (Koulountou ridge) and at last the most positive residuals on the Senegalo-Mauritanian basin, west of

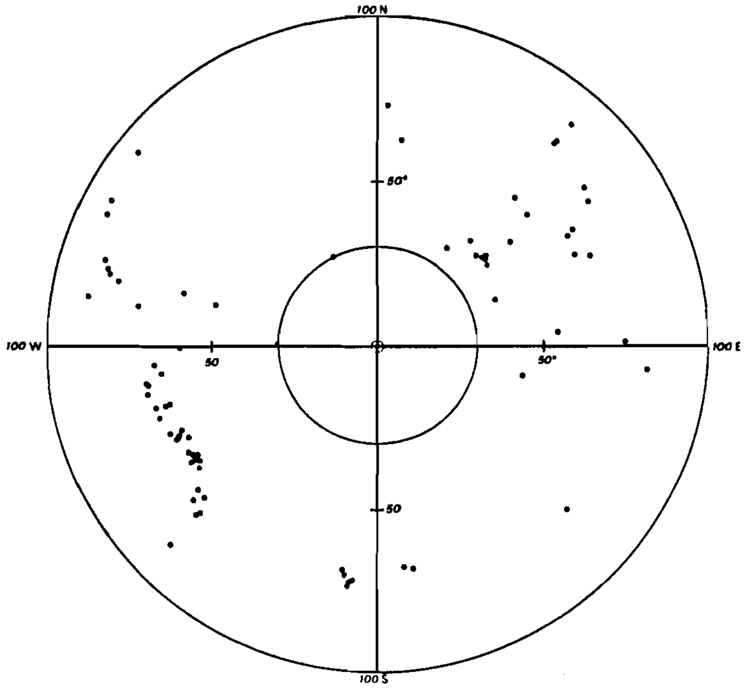

Figure 3
Epicenters of events used as P wave sources plotted as a linear fonction of distance and azimuth.

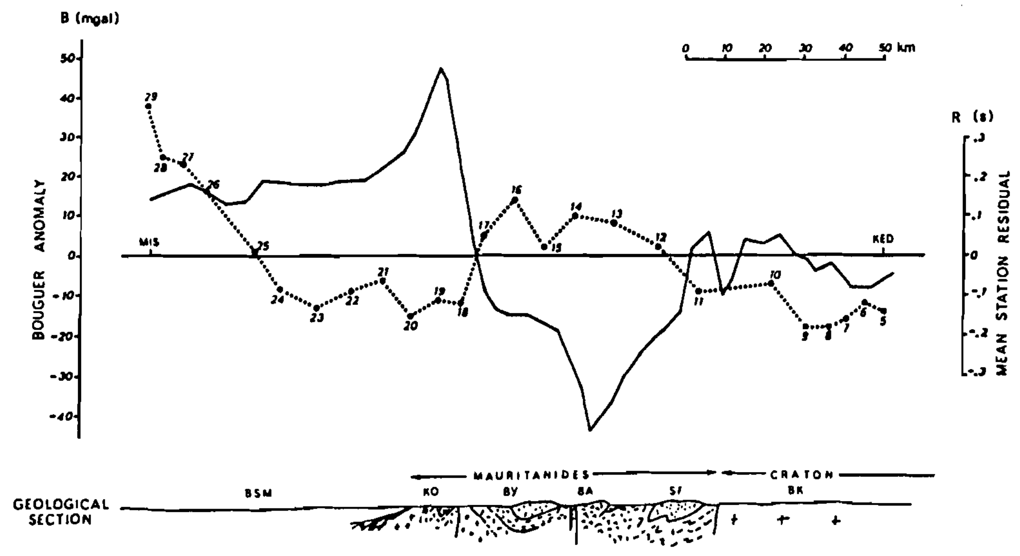

Figure 4

Mean station travel-time residuals for the seismological stations, Bouguer anomaly and geological section along the profile (after Bassot , 1966, Dia et al., 1979, *Villeneuve, 1982).*
BK *: Kenieba inlier;* SF *: Faleme trough;* BA *: Bassaris branch;* BSM *: Senegalo mauritanian basin;* BY *: Youkounkoun basin;* KO *: Koulountou ridge.*

station 26. On figure 5 the mean station residuals (calculated only with *P* waves) are plotted against the Bouguer anomaly. It appears clearly from this figure that three regions have to be distinguished : the craton with slightly negative Bouguer anomalies ($\simeq -5$ mgal) and the most negative residuals, the Senegalo-Mauritanian basin with Bouguer anomalies about + 15 mgal and the most positive residuals, and between these two regions the Mauritanides orogenic belt where the large variations of the Bouguer anomaly (from − 40 mgal to + 40 mgal) are linearly correlated with the mean station residuals.

The station 10, located on the craton but very close to its margin as seen by surface geology, marks the transition from the craton to the mobile belt; on the other side, station 26 marks geophysically the transition from the mobile belt to the Senegalo-Mauritanian basin although the Mauritanides formations do not outcrop between the stations 21 and 26 where they are covered by few tens of meters of mesozoic sediments (C.G.G., 1958). In consideration of their wavelengths, the variations of the Bouguer anomaly have probably a crustal origin (Guettat, 1981). The variation of the mean station residuals, calculated using events in wide ranges of azimuth and distances, principally reflects the variations in velocity under each station of the seismological array. Therefore since the 10 km spacing of the station is dense enough, we can make the hypothesis of a crustal origin to the variations of mean station residuals along the profile. Thus data from gravity and time residuals can be useful when setting up a model of the crust; we shall bring up this question later on. One can also note that the mean station residuals calculated only from PKIKP waves (table 2) are somewhat different with those calculated from *P* waves, the reason being that PKIKP rays have smaller angles of incidence than *P* rays (0° to 14° against 14° to 30°). Indeed the variations of mean station residuals calculated with only PKIKP waves reflect velocity variations in a thicker region.

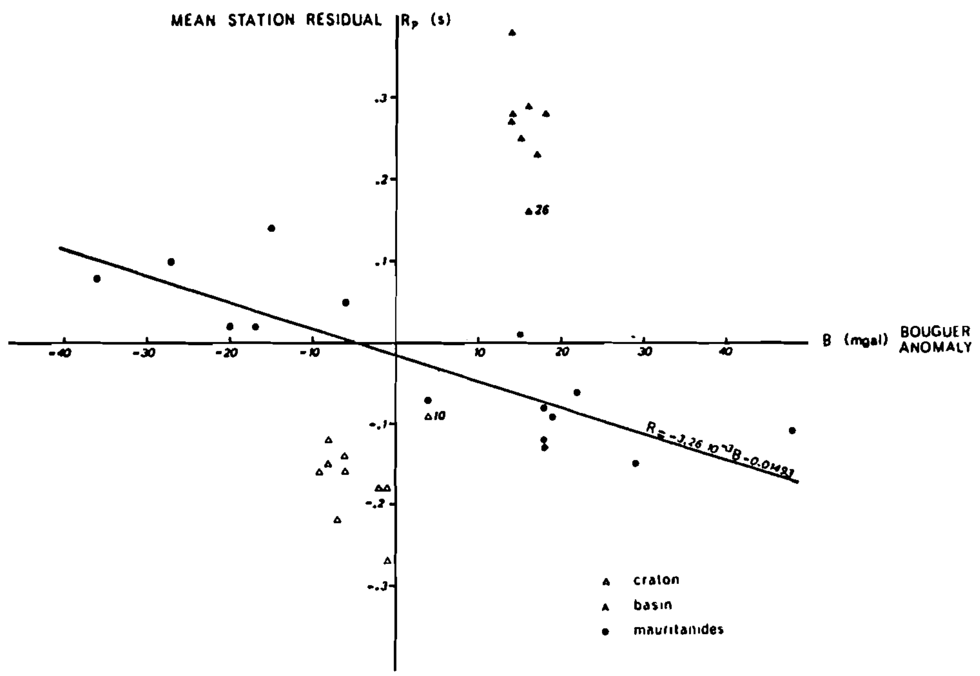

Figure 5

Mean station travel-time residuals for the seismological stations plotted against Bouguer anomaly.

Table 2

Station	N_p	R_p	N_{PKIKP}	R_{PKIKP}
1 IBL	66	− 0.27 ± 0.11	22	− 0.29
2 BND	82	− 0.22 ± 0.11	20	− 0.27
3 SIL	13	− 0.16 ± 0.16	3	− 0.25
4 FAD	82	− 0.15 ± 0.13	24	− 0.29
5 KED	76	− 0.14 ± 0.13	21	− 0.32
6 CEN	85	− 0.12 ± 0.12	23	− 0.27
7 NGA	37	− 0.16 ± 0.14	8	− 0.26
8 TIN	28	− 0.18 ± 0.09	9	− 0.29
9 MON	71	− 0.18 ± 0.14	19	− 0.27
10 TIM	10	− 0.07 ± 0.12	1	− 0.20
11 BAD	10	− 0.09 ± 0.18	1	− 0.11
12 SIB	10	0.02 ± 0.22	1	0.06
13 NIM	20	0.08 ± 0.22	7	0.26
14 BEL	19	0.10 ± 0.20	7	0.29
15 NIO	17	0.02 ± 0.16	5	0.17
16 GFO	17	0.14 ± 0.19	4	0.39
17 PLA	18	0.05 ± 0.17	5	0.28
18 GTR	17	− 0.12 ± 0.21	5	0.07
19 FAR	17	− 0.11 ± 0.14	4	− 0.09
20 MUL	17	− 0.15 ± 0.17	4	− 0.03
21 DDI	17	− 0.06 ± 0.17	3	− 0.02
22 DAR	19	− 0.09 ± 0.18	6	0.12
23 ABE	20	− 0.13 ± 0.16	7	0.05
24 DIA	19	− 0.08 ± 0.12	6	− 0.03
25 WAS	16	0.01 ± 0.13	6	− 0.05
26 GBO	71	0,16 ± 0.11	20	0.02
27 ARD	84	0.23 ± 0.11	24	0.13
28 CPA	57	0.25 ± 0.13	16	0.13
29 HAM	44	0.38 ± 0.15	12	0.21
30 MAD	83	0.27 ± 0.13	24	0.17
31 MIS	82	0.28 ± 0.13	23	0.15
32 BAB	79	0.28 ± 0.13	24	0.16
33 BAA	47	0.29 ± 0.12	12	0.18

Relative residuals inversion

On figure 6 relative residuals are plotted as a function of azimuth for several permanent stations. No clear effect is observed, but it can be noticed that, in the ranges of azimuth where the data are numerous enough, the dispersion of the data is quite large with respect to the expected dispersion due to arrival time errors. This dispersion comes mainly of the variations of relative residuals with distances, as it can be observed on figure 7 where relative residuals are plotted against distance in the two ranges of azimuth 40°-70° and 260°-300° for the same permanent stations. The almost linear variation of residuals with distance can be explained very simply by a model made of two blocks separated by a subvertical boundary, with higher velocities to the east; but there is no way to check the uniqueness of such a model. On the other hand, Briden *et al.* (1981) have shown that models of crustal structure which would account for all the observed anomalies have unacceptable features, too large lateral velocity and/or unacceptable Moho topography. Therefore we prefer to use an inverse method to construct a model adequate to explain the data. The method developped by Aki *et al.* (1977) has been followed.

This method can be summarized as follows : the space under the array of seismometers is divided into layers, each layer being characterized by its average P wave velocity. Then each layer is divided into blocks. In each block an unknown parameter is assigned, this parameter represents the difference between the true velocity and the velocity of the layer this block belongs to. For each ray it is possible to state an equation linking the velocity perturbations in the blocks that it samples to residuals.

We have therefore, in our case, an overdetermined system of equations. The problem is linearized by replacing velocity by slowness and then can be solved by an adequate inversion process. This method has been applied several times, particularly by Aki *et al.* (1977), Ellsworth (1977), Evans (1982) where general statements and examples of practical applications can be found.

As the array is only about 1.5° aperture and as we used only events with epicentral distances greater than 30°, large in comparison with the dimensions of the region to be modelized, the effects due to the Earth curvature can be neglected : all the layers and wave fronts are taken as plane. The initial distribution of velocity is Herrin's (1968). This initial model has a weak influence on the final solution, provided it is realistic.

In order to test a possible bias due to the mobile subarrays and a not uniform distribution of the sources in time, we checked that the mean station residual of the permanent stations does not change if one considers all the events or only events common with those recorded by the temporary network and so, for each of them. Moreover these mean station residuals do not correlate with the successive positions of the subarrays.

The main geological structures and the geophysical ones exhibit a clear linear orientation about 20° E ; on the other hand our linear model is orthogonal to these structures. Thus we have been led to construct a two-dimensional model, i.e. a model with prisms elongated

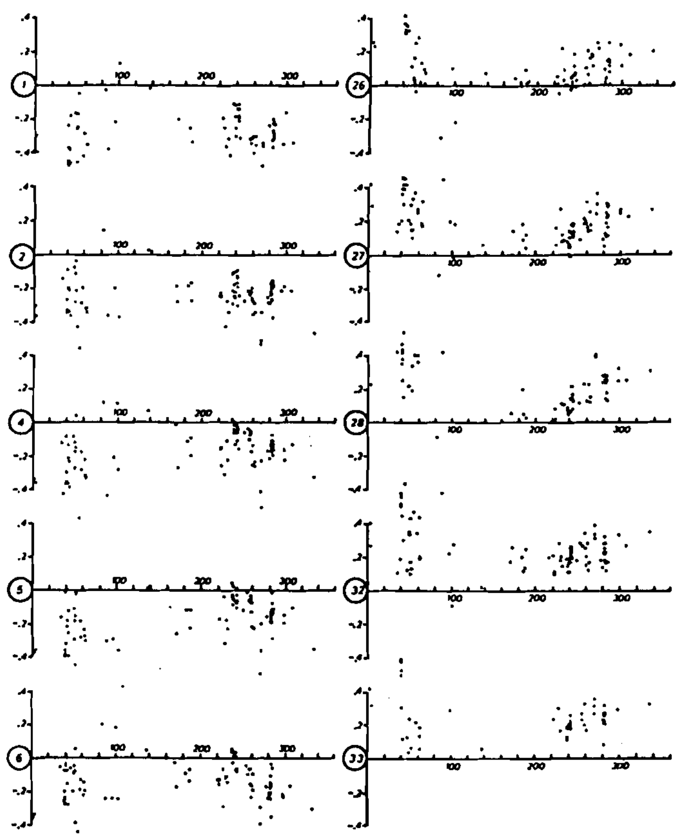

Figure 6
Relative residuals in some stations plotted as a function of azimuth to source.

in the direction parallel to the main structures. The number of the prisms in the first layer is at most the number of stations; in the other layers their sizes are taken as small as a good resolution allows. The model is presented in table 3. Its stability has been checked using the usual criteria : shifting of the boundaries, variation of the thickness of the layers and the size of the prisms... We have also varied the orientation of the prisms from 15° E to 30° E, without significant change. This model explains about 85 % of the data variance whereas we can reasonably assume that errors justify only 30 %. It is interesting to compare this result with a previous and similar work located on the Bangui magnetic anomaly (C. Dorbath *et al.*, 1982). In this later case a two-dimensional model failed to explain more than 60 % of the data variance and a three-dimensional model was required to reach the value obtained here with a two-dimensional inversion. This indicates that a two-dimensional model is good enough in this particular case. The solution is presented on figure 8. Figure 8*a* shows velocity perturbations in each block together with the resolution as defined by Aki *et al.* (1980). We note the overall good resolution (0.8). Figure 8*b* presents the same but smoothed solution. Standard error is about 0.8 % on each value. Only perturbations greater than standard error are dashed on this figure.

Geophysical model

The model is constrained mainly by *P* waves which are more numerous and sample it in a better way because their angle of incidence is greater than PKIKP's. Using this model we have computed the mean station residuals for subvertical PKIKP waves in the range of distance 160-180°. It is clear from figure 9 that the fit is remarkably good between the computed values and the observed ones. Thus the model explains the differences noticed previously between the station residuals calculated with the P data set and with the PKIKP data set.

Table 3

Model Center : Latitude 13N 3.83
Longitude 12W 52.37
Rotation Angle : – 20.0°

Layer N	Thickness km	Velocity km/s	$N_\perp$	$L_\perp$ km	$N_\parallel$	$L_\parallel$ km	Total depth km
1	15	6.00	1	400	20	10	15
2	25	6.75	1	400	12	20	40
3	40	8.07	1	400	14	20	80
4	40	8.11	1	400	16	20	120
5	40	8.17	1	400	18	20	160
6	40	8.27	1	400	20	20	200
7	40	8.27	1	400	20	20	240

where $N_\perp$ and $N_\parallel$ are the number of blocks respectively perpendicular to and along the profile.
and $L_\perp$ and $L_\parallel$ are the size of blocks perpendicular to and along the profile.

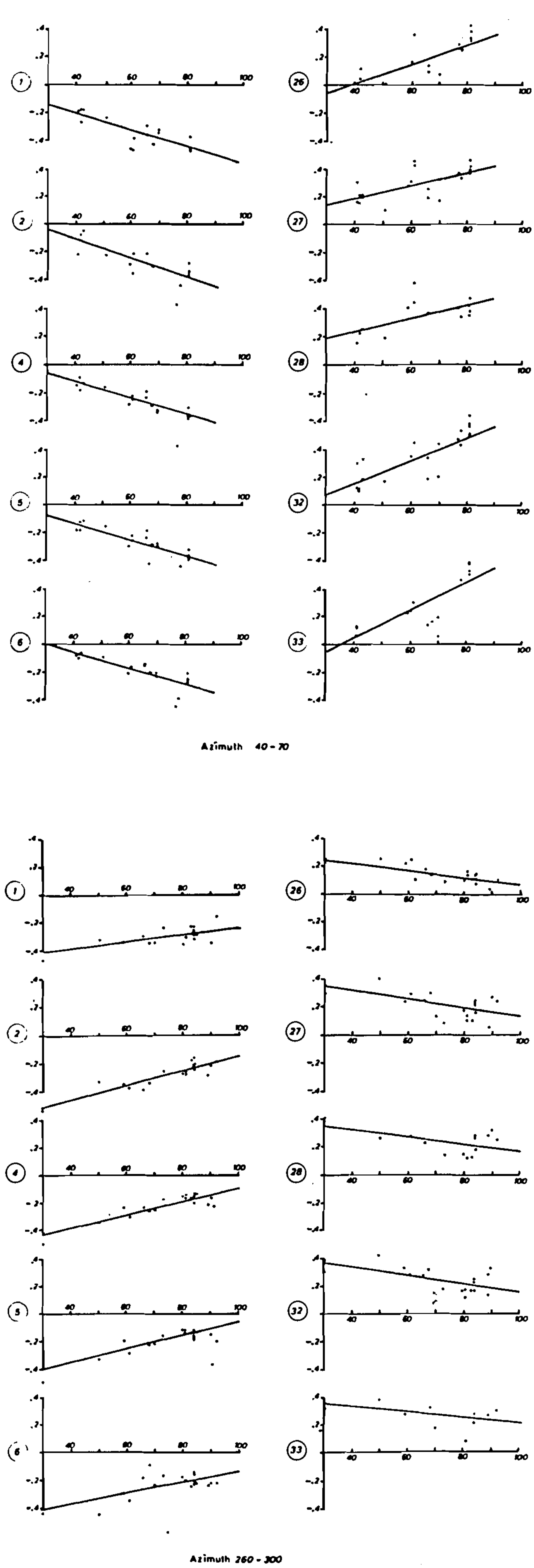

Figure 7 →
Relative residuals in the same stations plotted as a function of distance to the centre of the array :
a) azimuth range to the source = 40°-70°
b) azimuth range to the source = 260°-300°.

— *The crust*

The crust (layers 1 and 2, thickness 40 km) appears to be divided into four distinct zones (fig. 8) :

— To the east the craton is characterized by the highest velocities ; note the very good fit between the margin of the craton as seen on ground and the boundary of this high velocity zone.

— Just to the west of the craton extends a low velocity zone, about 60 km wide ; the low velocity takes up all the thickness of the crust. The perturbation reaches − 4 %. It fits the negative Bouguer anomaly and the outer zone of the Mauritanides belt.

— To the west of this low velocity zone, high positive perturbations are found again, less strongly marked than under the craton except near the surface. This zone coincides with the internal zone of the belt (Koulountou ridge).

— Finally the Senegalo-Mauritanian basin is characterized by low velocities (− 2.0 to − 2.5 %).

This regionalization is exactly the same as the one described when looking at mean station residuals, and the trendings of the perturbations are of course the same. This is another argument in favor of the hypothesis assuming that the main part of the mean station residual variations corresponds to the crust. In order to confirm this hypothesis we have computed, for a model with heterogeneities within crust only, the mean travel-time of *P* waves corresponding to a 70° epicentral distance with a uniform azimuthal distribution. Figure 9 shows the relative variations of these computed travel-times and the observed mean station residuals ; the fit is good, the maximum discrepancy is only 0.15 s to the west of the profile, on the Senegalo-Mauritanian basin. In the same way, using the relation of Nafe and Drake (1958) between velocity and density, we could be able to confirm, or not, the crustal origin of the gravity anomaly variations along the profile. To eliminate the edge effects due to the finite length of the model, we have extended it 500 km eastward with the same density as the craton and 500 km westward with the same density as the sedimentary basin. Note that :

— The top layer has, in our case, a density of 2.71 g/cm^3 instead of 2.67 g/cm^3 commonly used in the gravimetric calculations.

— The topography along the profile is flat and therefore the gravity anomalies inferred from the model are directly comparable with the Bouguer anomalies of Creen and Rechenmann (1965) deduced from field measurements using a mean density of 2.67 g/cm^3.

On figure 10 one can see the anomaly deduced from the model and the mean anomaly over the region in the direction parallel to the profile. The position of peaks and troughs and the amplitude are correctly restored, but not the regional trend. To reproduce this trend taking into account the velocity perturbations it is enough to make the crust thinner under the Senegalo-Mauritanian basin and to thicken it under the craton. With a 3 km thinner crust to the west and a 3 km thicker one to the east, the fit is fairly good. Introducing this Moho topography, we modify the initial model and the perturbations found at its extremities are only minimum ones, since an initial contrast (2 %) has to be added. To obtain a fairly correct solution an iteration would be necessary with this new initial model, but the uncertainties on both computed and observed anomalies (~ 5 mgal) make it needless. At last, for Guettat (1981), the negative Bouguer anomaly associated with the external zone of the Mauritanides belt results, in part, from a thickening of the crust in connection with a shortening which occurred during the growth of the belt. The velocities observed in this zone below 15 km are then explained by this thickening and by modifications in the crust during this process. Therefore the crustal density model deduced from the velocity model explains correctly the observed gravity feature, providing a 6 km thinning of the crust under the Senegalo-Mauritanian basin relative to the craton. A thickening of the crust under the external zone of the belt suggested

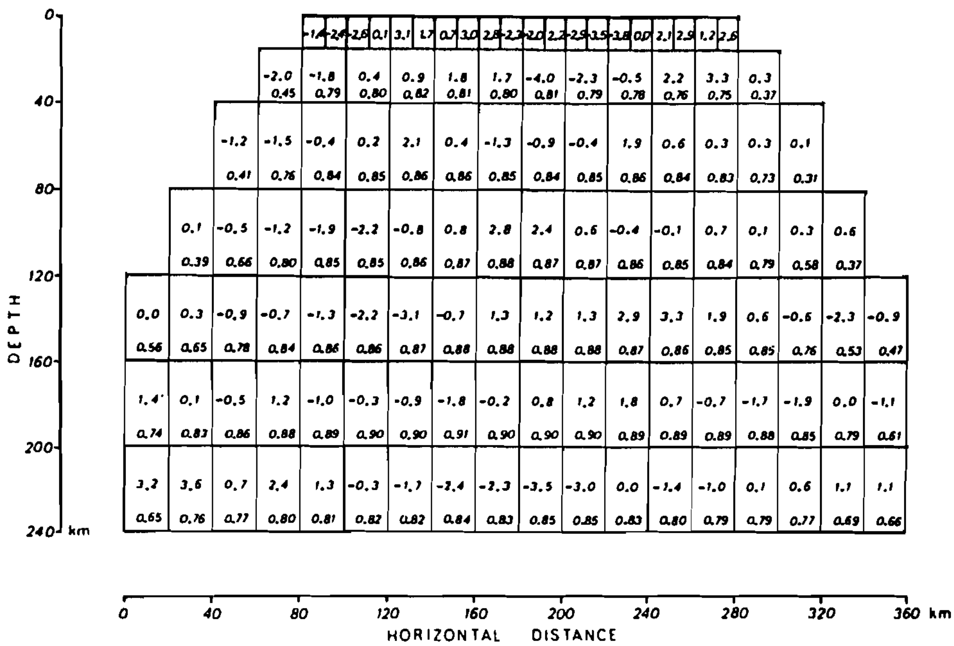

Figure 8
Velocity variations in the crust and upper-mantle beneath the array :
a) percent velocity variation in each block and diagonal element of resolution matrix.

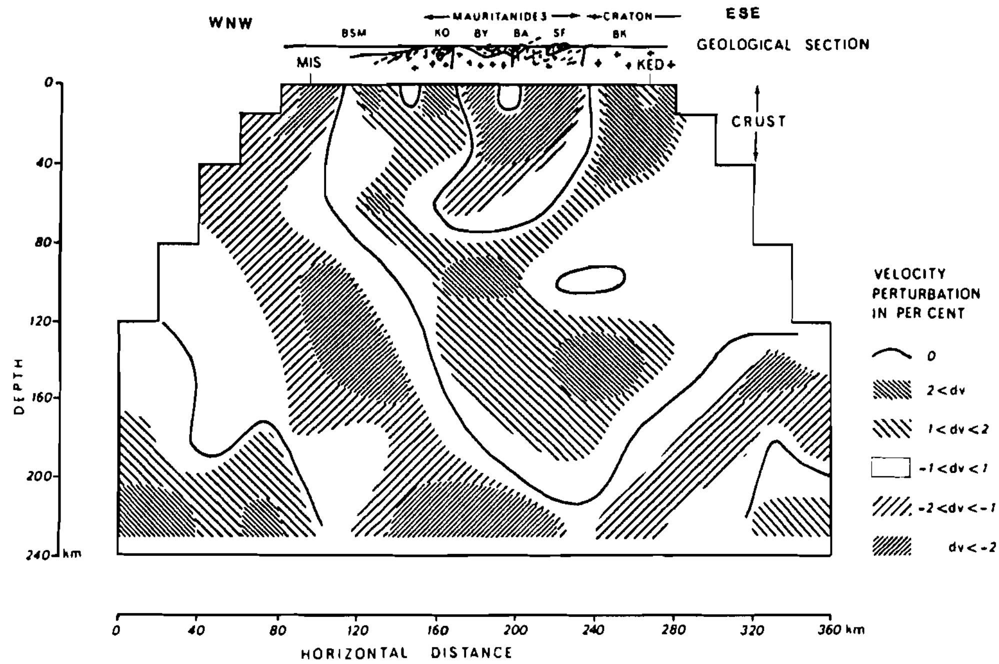

Figure 8

b) smoothed solution deduced from 10a. Only perturbations greater than standard error have been shaded.

by gravity studies is compatible with our inversion. Because of the poor vertical resolution we did not introduce such a perturbation of the crust thickness in our final inversion.

— *Upper mantle*

Below 40 km down to 150/200 km only two structures are obvious (fig. 8) : low velocities to the west under the Senegalo-Mauritanian basin, high ones to the east under Mauritanides belt and the craton. The maximum contrast is about 5 % but it decreases to the extremities. The boundary between theses two regions dips to the east with an angle of about 70°. Below 150 km this contrast is inverted : higher velocities are found under the Senegalo-Mauritanian basin, lower velocities under the craton but this result may be questioned because of some possible edge effects in the inversion. This general feature is similar to studies of lateral differences between older and younger regions in North America, Europe and Africa (Romanowicz, 1979 and 1980; Taylor and Toksöz, 1979; Hovland *et al.*, 1981; Dorbath and Montagner, 1982) : old structures have higher velocities down to about 200 km overlying lower velocities beneath. Our model, beneath crust, has also a feature similar to that proposed by Briden *et al.* (1981) but the main discontinuity dips to the east rather than being vertical or slightly westward as in Briden *et al.* In order to check this result concerning the direction of the slope and to eliminate possible bias due to the drawing of isanomalous lines, we decide to add a random value to the perturbations of each block to simulate the estimated error 0.8 % of the perturbation. Several trials have been made and they confirm our conclusion : the discontinuity does dip eastwards.

In summary the model obtained by inversion of travel-time anomalies explains also the gravity features, providing an adequate Moho topography according to the Nafe and Drake relationship between velocity and density. It is characterized by a major discontinuity in the crust and the lithosphere down to 150/200 km which will be called « Missira discontinuity ». This discontinuity separates a low velocity structure to the west form a high velocity structure to the east. It cuts the surface at about 120 km west of the outcropping margin of the craton, being overlapped by a mesozoic coverage. It begins as a vertical boundary in the crust, but below 40 km it dips to the east with an angle of about 70°. Beneath the eastern Mauritanides zone, a 60/70 km wide low velocity zone is found from the surface down to 50/60 km. This structure has its origin in a deep trough filled with volcano-sedimentary formations and a thickened and tectonized crust.

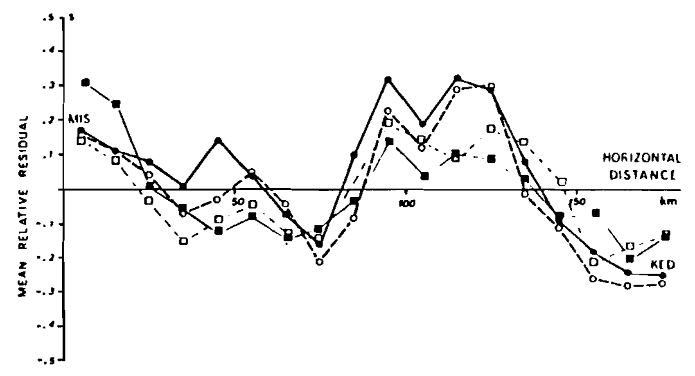

Figure 9

Mean PKIKP relative residuals in the distance range 160°-180°. (observed : filled square-computed : open square).

Observed mean station residual and computed mean relative delays in the crust with a 70° epicentral distance and an uniform azimuthal distribution of the rays.

(Observed : filled circle-computed : open circle.)

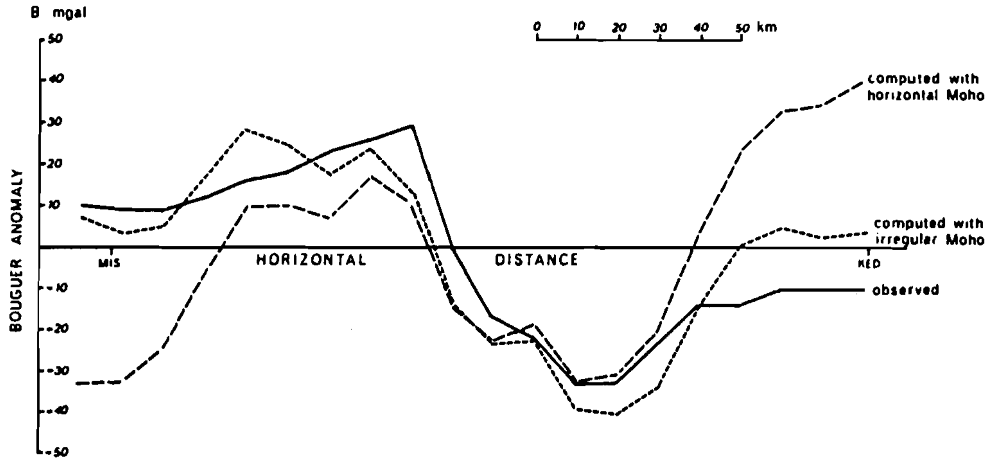

Figure 10
Comparison between the Bouguer anomalies observed in the region of the profile, computed with an horizontal Moho and computed with an irregular Moho.

DISCUSSION

With respect to its size and to the velocity difference on both sides, the Missira discontinuity is the main feature observed in the area. The other inhomogeneities are principally located in the crust and related to specific segments of the orogenic belt. The short length of the profile does not allow to determine the thickness of the lithosphere each side of this discontinuity, because the main part of the investigated volume consists of this central anomaly. This discontinuity separates two distinct structural blocks, the eastern one being the West-African craton. Theory of formation of orogenic belts at convergent plate margins leads us to make the hypothesis that the Missira discontinuity is the trace of a subduction zone which would have dipped eastward steeply. But collision could modify the dip of the slab. Thus the old subducted plate is now characterized by velocities lower than those under the craton. This is a similar result as that found by Taylor and Toksöz (1979) in the Northern Appalachians. The origin of such lower velocities is not clearly explained, but is not necessarily inconsistent with the thermal evolution of a subducted oceanic lithosphere (Toksöz *et al.*, 1971).

We deduce from the geological and geophysical observations the structural cross-section presented on figure 11. The basement, outcropping in the craton, sinks under the belt and reappears to the east of the Missira discontinuity. A synclinorial structure takes place between the two outcropping basement regions. Underneath this structure the crust is thickened and tectonized. To the west of the Missira discontinuity, the crust is thinner and has a lower velocity than in the craton but it is of continental nature (exploration cores and geophysical data). Moreover this continental basement was not submitted to important tectonic modifications after paleozoic (except the tholeitic permo-triasic volcanism linked to the atlantic aperture). It is therefore necessary to correlate the new geophysical observations and specially the Missira discontinuity with the geological features of the Mauritanides. From this point of view volcanism may be an important marker and in fact we can observe that the basic volcanism of the Faleme trough and the acid one of the Youkounkoun basin appear at the lower part of the green series at the end of precambrian. They are the evidence of a large distention phase at this time. Thus both calc-alcalin nature of the ryolithes and spilitic volcanism of the Faleme trough indicate an eastward subducting slab which we can see in the Missira geophysical discontinuity. That means that the Faleme trough represents a marginal sea at the western edge of the West-African craton. These data give us an indication about the precambrian time of the maximum subducting activity. The deposits of the green series show that the Faleme trough goes on subsiding; then volcanism has ceased. Thus we must admit that the closure of the marginal sea was beginning and that the subducting movement was slower. On another hand the main deformation in the Youkounkoun basin and the Koulountou ridge is later and that they appear after Silurian.

Thus the geophysical argument as well as the geological observations are in good agreement for admitting an eastward subducting plate at the end of the precambrian with a correlated formation of a marginal sea and of an island arc of continental nature following the process generally accepted now (see Cross and Pilger, 1982 for a review).

Then it is possible to propose a geodynamical process to explain the structures of the southern part of the Mauritanides. This one is summarized on figure 11. Figure 11A shows the opening stage of the marginal sea represented by the Faleme trough and the correlative acid and basic volcanisms. Figure 11B exposes the beginning of the closure. The Faleme trough goes on subsiding by cooling (Mc Kenzie, 1978) in this compressive contact. The green series syntectonic sedimentation may result in some unconformities which can explain the ones observed by Villeneuve (1982) in the Faleme trough, north of Guinea. During the closure, overthrusts and scallings occur in the western part of the basin resulting in the formation of the Bassari ridge (fig. 11C). In the latest stage the Youkounkoun-Koulountou pannel divides into fragments, generating the Youkounkoun basin and scallings in west (fig. 11D). This tectonic phase may be related with a late collision with a continental basement hidden under the Senegal-Mauritanian

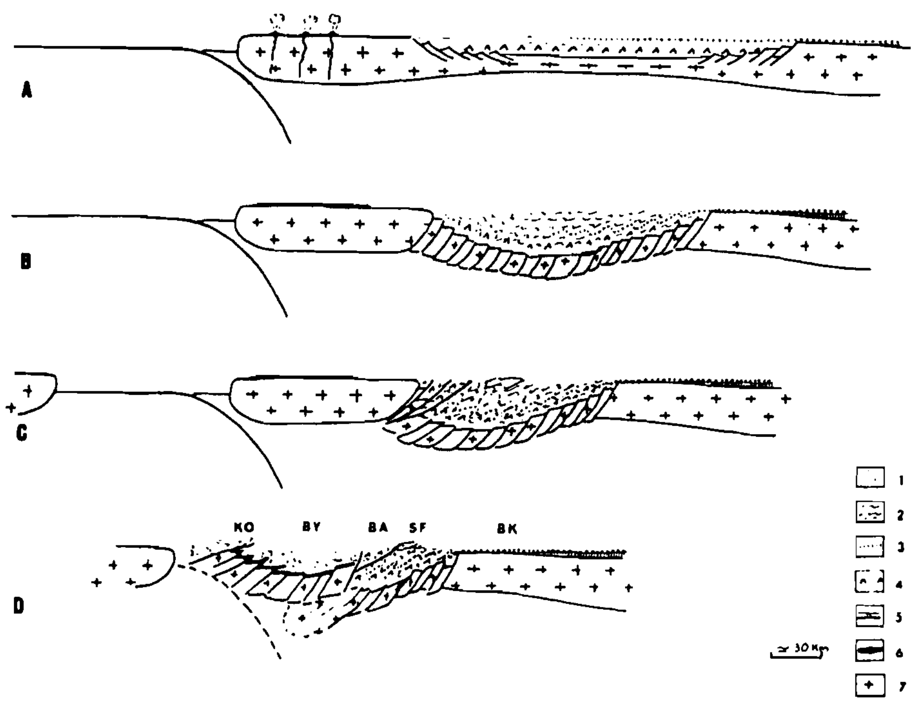

Figure 11
Geodynamical model for the southern part of the Mauritanides :
1. Red detrital series ; 2. green flyschoid sediments ; 3. upper precambrian mixtite ; 4. volcano sedimentary formations (Faleme trough) ; 5. proterozoic sediments ; 6. acid-alkalin volcanism ; 7. continental basement.
(For letters see figure 4. Only crust is drawn, but the whole lithosphere is interfering in the process).

basin sediments. It is this collision more that the subducting movement which may have induced the present structural pattern of the western part of the Mauritanides.

Further speculations may be added in this geodynamical model. The alcalinity increases with decreasing convergence rate (Uyeda, 1977). The Youkounkoun basin rhyolites would indicate that oceanic crust was subducting very slowly. That is confirmed by the fact that no andesitic volcanism is mentioned here. Moreover correlation between convergence rate and inclinaison angle of the sinking slab is generally accepted (see Cross and Pilger (1982) for a review) : slow convergence is associated with high-angle subduction. If confidence can be given to the observed inclination, ancient subduction would have taken place with slow convergence rate and high-angle Benioff zone.

Contribution I.P.G. n° 634.

CONCLUSION

The seismological study of eastern Senegal shows the existence of a major discontinuity from the surface to 150/200 km depth. It cuts the surface about 120 km at the west of the outcropping margin of the West-African craton. It separates two structural blocks, the east one, with higher velocities, being the structure of the craton. We propose here that the western one is an old subducted oceanic plate. Other heterogenities are found mainly in the crust and are related to specific segments of the Mauritanides orogenic belt. Seismological, gravity and geological data lead us to construct the structural cross-section shown on figure 11D. From this we infer that the Mauritanides orogenic belt results in the opening of a back-arc marginal basin in late-Precambrian and in its closure from that time to Devonian.

We have now to look if the geodynamical model proposed for this part of the orogenic belt can be extended to the whole belt. The bulge observed in the craton seems to be the first event due to the subduction. It is interesting to note that Toksöz and Hsui (1977) also found that a similar bulge occurs 200/300 km inside the continent prior to the opening of a back-arc marginal basin in their theoretical model. This bulge can be followed in eastern Senegal north of Kayes and in Adrar in north. On the other hand, in the internal part of the belt rhyolites similar to those found in Youkounkoun basin and Koulountou ridge are present all along the belt (Chiron, 1973 ; Le Page, 1978*a* ; Lecorché, 1980). Thus it is possible to imagine that some comparable geodynamical processes occurred along the whole folded zone. Geophysical studies should be undertaken in the northern region to confirm or not this generalization.

Finally the geodynamical history that we propose for the formation of the southern part of the Mauritanides belt is quite similar to that proposed by Hatcher (1978) and Iverson and Smithson (1982) for the Southern Appalachians, although in this later case the tectonic phenomena are more complex.

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