Crustal and upper mantle velocity structure of the Hoggar swell (Central Sahara, Algeria)

A. Ayadi [a], C. Dorbath [b,c,*], A. Lesquer [d], M. Bezzeghoud [a,e]

[a] *Department of ESS / CRAAG, BP 63 Bouzaréah, 16340 Algiers, Algeria*
[b] *EOST (UMR 7516), 5 rue René Descartes, 67084 Strasbourg Cedex, France*
[c] *IRD (ORSTOM), 213 rue La Fayette, 75480 Paris Cedex 10, France*
[d] *Institut des Sciences de la Terre, de l'Eau et de l'Espace (CNRS-UM2), CC 049, 34095 Montpellier Cedex 5, France*
[e] *Departamento de Fisica, Universidade de Evora, Apartado 94, 7001 Evora, Portugal*

Abstract

The Hoggar region is known as one of the most important swells in the African continent. Its altitude culminates at 2908 m in the Tahat hill (Atakor). The Hoggar and other massifs of central Africa (Aïr, Eghei, Tibesti, Darfur, Cameroon mount, ...) form a system of domal uplifts with similar scale, morphology and volcanic activity. The knowledge of the structure beneath the Hoggar swell will help us to understand the origin of continental swells. In order to get an image of the lithosphere in this region, we have performed a teleseismic field experiment. The 33 short-period seismic stations have been maintained for 2 1/2 month along a 700-km long NNW–SSW profile. This experiment crossed the Central Hoggar and extended northward into the In-Salah Sahara basin which is characterized by high heat flow values of deep origin. The high quality of the data recorded during this experiment allows us to perform a velocity inversion. The Hoggar appears to be characterized by lower mantle velocities. The anomalous zone extends from the upper lithosphere to the mantle. The weak velocity contrast is interpreted in agreement with gravity, geothermal and petrological data as due to extensive mantle modifications inherited from Cenozoic volcanic activity. It confirms that the Hoggar swell is not due to a large-scale uplift of hot asthenospheric materials but corresponds to a now cooled-off modified mantle. On the contrary, local low-velocity zones associated with the Atakor and Tahalra volcanic districts show that hot materials still exist at depths in relation with recent basaltic volcanism. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: Velocity structure; Hoggar swell; Algeria

1. Introduction

West Africa (Fig. 1a) mainly consists of two major tectonic provinces: the West African Craton (WAC), stable since 2000 Ma and the Pan-African belt which resulted from a collision tectonic process around 600 Ma (Black et al., 1979). The Pan-African

* Corresponding author. Institut de Physique du Globe, 5 rue René Descartes, 67084 Strasbourg Cedex, France. Tel.: +33-3-88-41-63-96; fax: +33-3-88-61-67-47; e-mail: cath@mapu.u-strasbg.fr
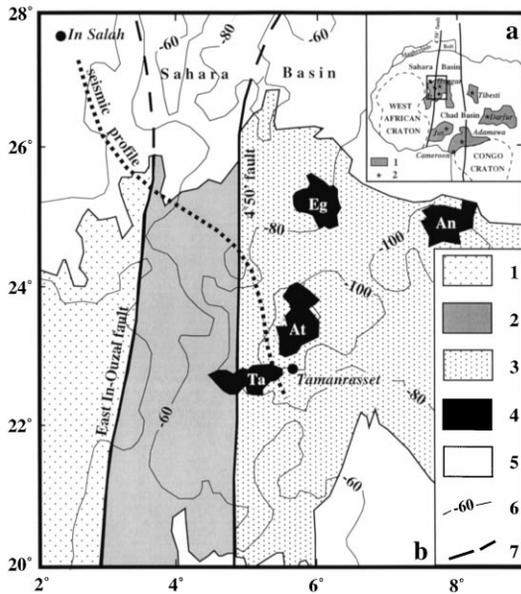
Fig. 1. (a) Map of North-west Africa with the location of the cratons, the NS megafaults and the seismic profile. (1) Uplifts of Precambrian basement. (2) Cenozoic volcanism. (b) Structural and gravity map of the Hoggar swell showing the location of the Lithoscope profile. (1) In-Ouzzal domain. (2) Pharusian belt. (3) Central Hoggar. (4) Cenozoic volcanism (An: Ad N'Ajjer, At: Atakor, Eg: Edjere, Ta: Tahalra). (5) Paleozoic sediments. (6) Bouguer anomaly contours in mgal (Lesquer et al., 1988). (7) Main faults.

domain east of the WAC is characterized by a system of broad Cenozoic volcanic uplifts (Tibesti, Darfur, Adamawa, Cameroon mount, Jos plateau) surrounded by Paleozoic to Mesozoic sedimentary basins. These domal uplifts present very similar patterns in scale and morphology as well as for gravity signatures (Brown and Girdler, 1980).

Situated between the Paleozoic and Mesozoic Sahara basins at north and the Tchad basin at south, the Hoggar shield (Fig. 1a) can be described as a broad (1000 km wide) SW–NE trending area of exposed Precambrian basement with a mean altitude of 500 m. The geological framework of this shield is inherited from the Pan-African orogeny (600 Ma). It is dominated by north–south elongated structural domains separated by north–south shear zone megasystems (Fig. 1b). Geophysical data and deep oil wells show that these Precambrian structures extend northward beneath the thick Paleozoic to Mesozoic sedi-

mentary cover of the Sahara basins (Takherist, 1991). A topographic swell — i.e., the Hoggar swell — is superimposed on this broad dome of Precambrian basement.

The highest massifs (> 2000 m), at the centre of the swell, correspond to the Cenozoic volcanic districts of Atakor and Adrar N'Ajjer.

A long-wavelength SW–NE negative Bouguer gravity anomaly (Fig. 1b) is associated with the Hoggar swell. The minimum (−120 mgal) corresponds to the Atakor volcanic district. By analogy with the East African rift system, Brown and Girdler (1980) have associated this anomaly with the thinning of the lithosphere. More recently, normal heat flow data measured on the southern part of the Hoggar shield and gravity data interpretation led
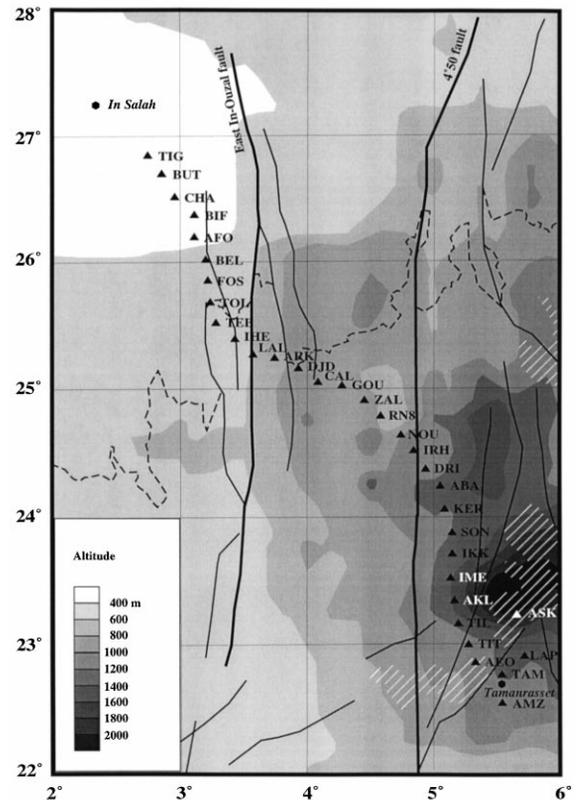

Fig. 2. Location of the seismic experiment together with the topography and the main structural discontinuities. The triangles show the position of the stations. The white hachured zones represent the volcanic provinces.
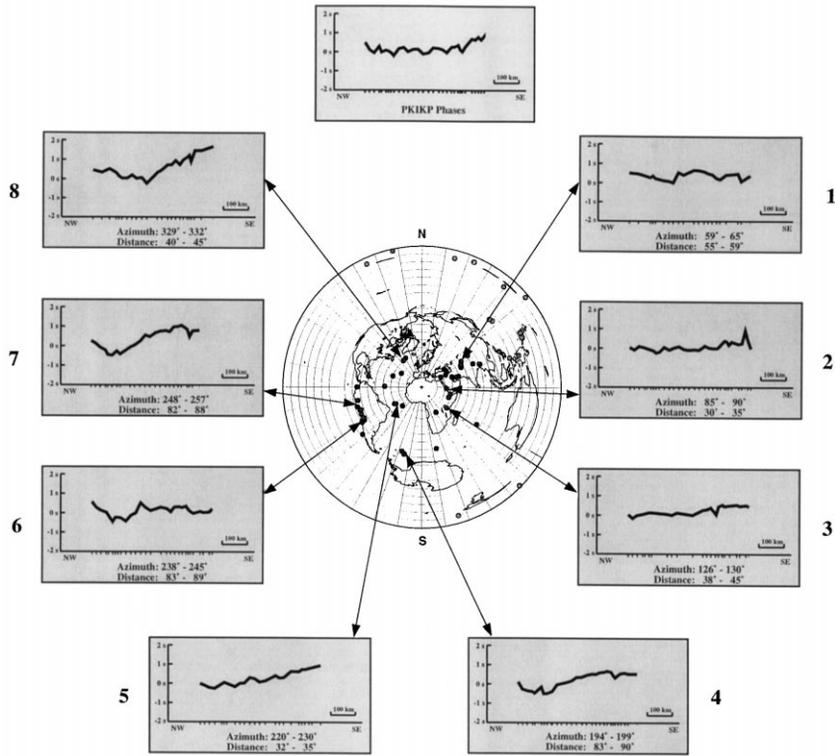
Fig. 3. Distance and azimuth distribution of the 95 events (grey dots) selected for the tomographic study. The world map is centred on the station ARK, which is taken as reference in the residual study. Average relative P residual curves are plotted around the map for eight source regions; the mean PKIKP residuals computed for 14 events are plotted in the top part of the figure.

Lesquer et al. (1989) to propose that the Hoggar swell is the isostatic response to a now-cooled altered upper mantle. They proposed that this anomalous zone is associated with an upper Cretaceous to Eocene hotspot type magmatic activity, which has led to extensive upper mantle modifications.

In contrast, the Sahara basins north of the Hoggar shield are characterized by high heat flow values.

Table 1
Initial P velocity model

Layer	Velocity (km/s)	Thickness (km)	Nb of NS blocks	Dimension (km)	Nb of EW blocks	Dimension (km)
1	5.90	10	33	cones	1	cones
2	6.40	24	20	30	10	45
3	8.00	40	22	30	12	45
4	8.05	40	20	40	10	65
5	8.10	50	22	40	10	65
6	8.15	50	20	50	10	75
7	8.50	60	20	50	12	75
8	8.70	60	20	60	10	95
9	9.00	60	18	70	10	105

Together with gravity field and surface waves velocities data, these high heat flow values have led to the proposal that the upper mantle beneath the Sahara basins, just north of the Hoggar shield, is anomalously hot and slow (Lesquer and Vasseur, 1992; Lesquer et al., 1990).

The surface wave large-scale resolution and the lack of heat flow data in the northern part of the Hoggar shield do not allow to define the southward extension of this anomalous mantle. To test these hypotheses and explore the structure of the crust and upper mantle beneath the Hoggar swell and the Sahara basins, a teleseismic field experiment was performed in 1993. Seismic events were recorded during 2 1/2 months at 33 seismic stations installed along a 700-km long NNW–SSE profile running from Tamanrasset at south to In-Salah at north (Fig. 2).

2. Geological and geophysical setting

From south to north, the seismic profile crosses three contrasted lithostructural and metamorphic domains separated by N–S-striking shear zones (Fig. 1b): the Central Hoggar, the Pharusian belt and the In-Ouzzal domain.

The Central Hoggar is limited westwards by the 4°50′ megafault. It is mainly composed of granulites
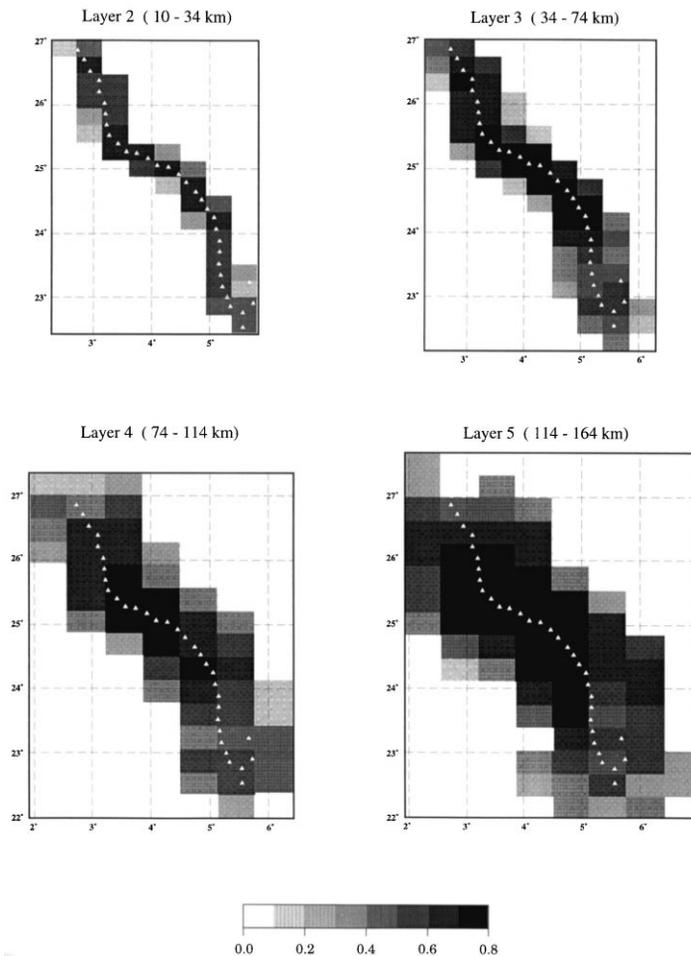
Fig. 4. Value of the diagonal elements of the resolution matrix corresponding to each block in the seven layers of the tomography. The darkest blocks are the best resolved in the inversion.
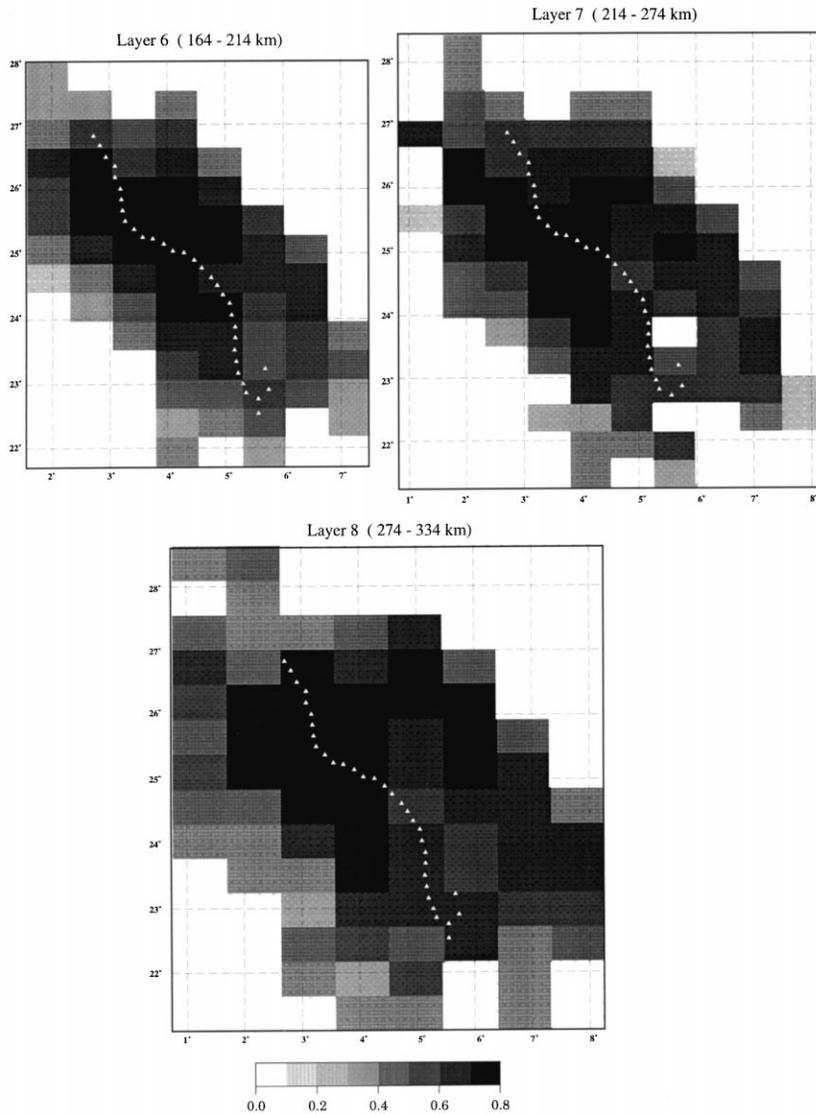
Fig. 4 (continued).

and gneisses Eburnean in age (2.1 Ga), reactivated and injected by abundant granitoids during the Pan-African orogeny. The 4°50′ shear zone, which crosses Africa from the Atlantic Ocean to the south to Mediterranean Sea to the north, is transected by the seismic profile between stations DRI and IRH (Fig. 2).

The Pharusian volcano-sedimentary belt corresponds to supra-crustal juvenile sequences: green schist facies, platform sedimentary rocks, mafic–ul-tramafic sills, calc–alkalic granitoids, volcanoclastic sequences.

The In-Ouzzal domain, which corresponds to a 2.1-Ga granulitic block slightly affected by the Pan-African orogeny, is interpreted as a Paleo-Protero-zoic fault-bounded lithospheric block (Caby, 1996). A line of positive gravity anomalies (Takherist, 1991) outlines the eastern limit of the In-Ouzzal domain. Locally, these anomalies can be correlated with gran-ulitic metagabbro and it has been proposed that these

belt-of-gravity anomalies outline a Pan-African cryptic suture zone (Takherist, 1991). The seismic profile crosses this gravity lineament north of 25°N, between stations ARK and FOS (Fig. 2).

The seismic stations north of station TOL are on the Paleozoic sediments of the Sahara sedimentary basin which overlap the Precambrian basement. In this area, the mean heat flow determined from oil wells data is more than 100 mW m^{-2}. These high heat flow values that are correlated to low gravity anomalies are interpreted as evidence for a high-temperature anomalous upper mantle (Lesquer et al., 1990).

Southward, the seismic profile crosses the western extremity of the Mio-Pliocene Atakor volcanic massif (which culminates at 2900 m near Assekrem (ASK) station) and the eastern extremity of the Plio-Quaternary Tahalgha volcanic massif. Both volcanisms of alkali affinity mark the late extension of the magmatic activity in the Hoggar which started early with the emplacement north-eastward in the present Amadror depression of Upper Eocene–Lower Oligocene lava and ring-shaped intrusion (Dautria and Lesquer, 1989; Aït Hamou and Dautria, 1994, 1997). The seismic profile transects the long wavelength gravity low which correlates with the Hoggar swell. Estimation of elastic plate thickness (Lesquer et al., 1988) allowed to propose that the Hoggar topographic load is supported by a low density body situated at around 50 km depth. A mining borehole near AMZ station at the southern end of the seismic profile exhibits a normal heat flow value of 63 mW m^{-2}.

3. Field experiment and data processing

Thirty-three temporary digital recorders of the French Lithoscope network, equipped with vertical short-period seismometers (Mark Product L4C, $T_o = 1$ Hz) were maintained from February 1993 to May 1993 (Fig. 2). In this region, very sparsely populated, few usable tracks exist, so the stations were located close to the main road. The profile was about 700 km long, and mean station spacing was 20 km. From south to north, the elevation along the seismic profile varies from 1500 to 500 m. Two more stations have been installed out of the profile, in the centre of the Atakor volcanic massif.

The time was given by GPS-controlled internal clocks. During the 2 1/2-month experiment, we have recorded up to 200 events which were reported in international bulletins. Only 95 teleseismic events with a clear P onset have been kept in the inversion study. The epicentres of these selected events are plotted as dots on a world map centred on the middle of the profile (Fig. 3). The good coverage in both distance and azimuth of the events allows us to perform a tomographic inversion. The final data set contains 1567 observations, 282 of which are PKP phases.

The absolute travel times have been computed using the National Earthquake Information Centre (NEIC) hypocentre's parameters and the travel time tables of Herrin (1968) for P and PKP phases. After station elevation correction, absolute travel time residuals were calculated. These absolute travel time residuals include contributions from hypocentres and origin time errors, and from the ray paths outside the study area. To remove these errors, we calculated relative travel time residuals by subtracting the residual of a reference station from the absolute residuals for each event. The reference station chosen is ARK (Fig. 2) situated on the basement near the limit of the Sahara basin, which has been operational throughout the whole field experiment.

4. Variation of the relative residuals

Because of the quasi-vertical incidence of the PKP phases, we observe no azimuthal variation of their residuals. Thus, we have plotted the mean PKP relative residuals along the profile for 14 events (Fig. 3). This curve gives a crude picture of the lateral velocity variations. The mean PKP residuals do not show any clear trend along the profile. Except on both extremities, the amplitude of the variations is within 0.5 s. At the southern end of the profile, we observe a small positive jump ($+0.5$ s) of the relative residuals, indicating the presence of lower velocities under the southernmost stations. The northern end of the profile also shows a higher residual, but only at the last station. The amplitude of the variation along the profile is very weak, about 1 s.

In Fig. 3, we present the variation along the profile of the average residuals of the P phases,

computed for eight source regions. These curves are plotted around the map at the approximate azimuth of the source region with respect to the centre of the profile. These curves present some common features, more or less pronounced depending on the azimuth: the highest relative residuals are at the south of the profile, the lowest ones at the stations located north of ARK, followed by an increase of the values at the northernmost stations. In detail, the amplitude of the variation along the profile is the smallest (about 0.5 s) on the curve corresponding to eastern and south-eastern source regions (profiles 2 and 3, Fig. 3), however, a weak increase of the residuals is observed in the south-eastern part of the profile. On the other hand, the plots corresponding to western and north-western source regions (profiles 7 and 8, Fig. 3) present the highest amplitude variations (up to 2 s), with a pronounced increase of the relative residuals in the southern part of the profiles. It should be noted that the two curves corresponding to sources regions perpendicular to the azimuth of the seismic profile (profiles 1 (azimuth 60°) and 6 (azimuth 240°), Fig. 3) show similar features, with a positive jump of the residuals south of ARK followed by a plateau. This similar behavior of opposite azimuth source regions allows us to suppose that we are not too far from the axial symmetry, which is a strong hypothesis imposed by the interpretation of a one-dimensional array.

The interpretation of the variations of the relative residuals in terms of a qualitative velocity model is not straightforward. However, we can state that there is a large zone of lower velocities under the Hoggar swell. A zone of higher velocities is situated north of ARK, the size of which has to be modest because it is not observed on every azimuth. Finally, we observe a zone of low velocities at the northern extremity of the profile, but much less pronounced than to the south. This coarse model will be refined by data inversion.

5. Data inversion and tomographic images

The relative travel time residuals have been used to perform a velocity inversion using the ACH technique (Aki et al., 1977) and to compute a model of the velocity perturbations in the studied region. The starting P velocity model (Table 1) we used is derived from the isotropic model for the average structure in Africa of Hadiouche et al. (1989). It includes a crustal thickness of 34 km, the same as in the mobilized zone model of Dorbath and Montagner (1983).

As the network was quasi-linear, we first performed a two-dimensional inversion. The resulting velocity model did not succeed in explaining the observed variations of the P residuals and reduced the initial data variance by only 57%. Thus, we have carried out a three-dimensional inversion. The volume under investigation was divided into layers, which were themselves divided into blocks except for the shallowest one (Table 1). In the 10-km thick upper layer we assigned a separate cone to each
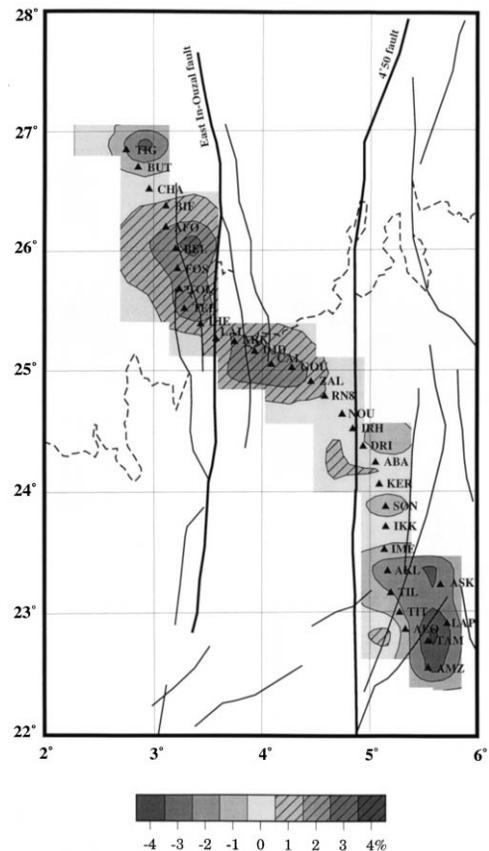

Fig. 5. Smoothed velocity perturbations (in %) in the crust, together with the main structural discontinuities. The regions with higher velocities than the reference model (positive perturbations) are hachured.

station, following the procedure of Evans and Achauer (1993). This method accounts better for the lateral heterogeneities in the upper crust where ray paths rarely overlap. The large aperture of our profile, 700 km, allows us to obtain a velocity model resolved down to a depth of 340 km. We tested several different models, with various layer and block partitioning and different general orientation. The chosen final model reduces the data variance by up to 90%. The value of the diagonal element of the resolution matrix blocks for the seven upper layers of our model are presented on Fig. 4. We only kept the blocks crossed by more than 10 rays. The central part of our model below the stations, crossed by numerous ray paths, is the best resolved.

The velocity perturbations obtained for the resolved blocks of our model are presented in map views on Figs. 5 and 6. The solution has been smoothed in order to remove the sharp changes between adjacent blocks.

5.1. The crust

To get a global image of the crust (Fig. 5), we performed the inversion with an initial velocity model, which included a single crustal layer 34 km thick in place of the two crustal layers of the model given in Table 1, the deeper layers remaining unchanged. This initial model allows us to get a smoothed and better resolved image of the crust. A
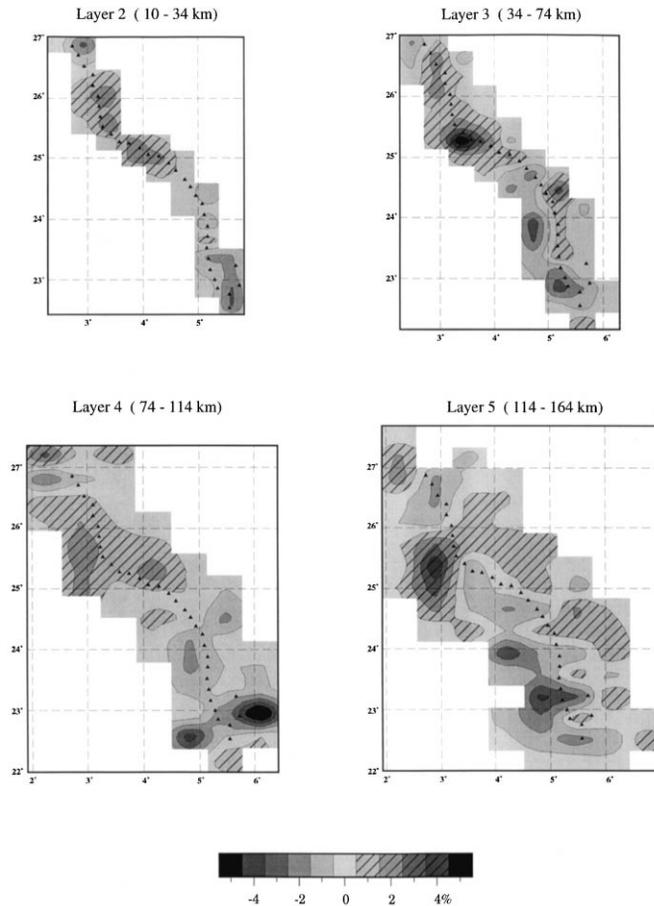

Fig. 6. Smoothed velocity perturbations (in %) in the eight layers below 10 km.
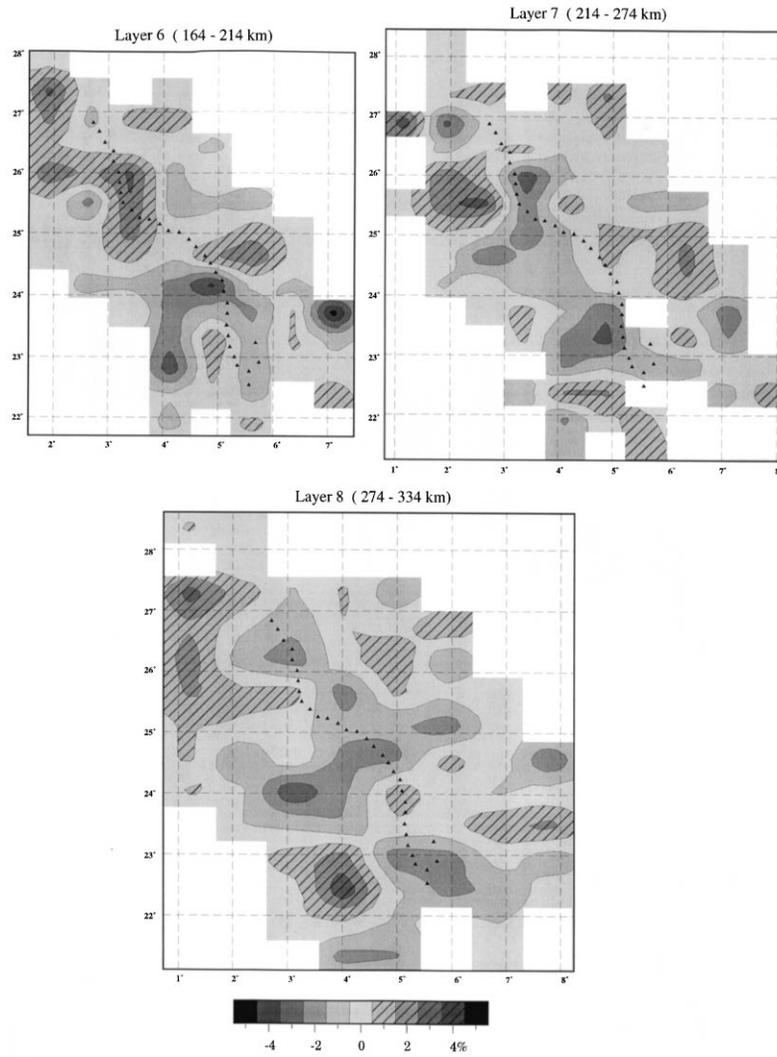
Fig. 6 (continued).

comparison with the image of the lower crust obtained by the first inversion (layer 2, Fig. 6) shows few differences.

Because of the linearity of the seismic profile, the image of the crust (Fig. 5) has no lateral extension and only displays north–south discontinuities. The southernmost part of the profile is characterized by lower velocities. The lowest values of the velocity perturbation, down to −4.5%, are observed near stations TAM and ASK. Between RN8 and BIF stations, we observe a zone of positive velocity perturbations with values up to +3.5% at the BEL station. The northern extremity of the profile is characterized by slight negative velocity perturbations.

The change from lower to higher velocities occurs at IRH station just on the 4°50′ megafault which separates the Central Hoggar domain from the Pharusian belt. As Central Hoggar is mainly composed of granulitic units characteristic of deep crust, it should consequently exhibit higher velocities than the Pharusian belt composed of slightly metamor-

phosed sedimentary and volcano-sedimentary rocks. Thus, the velocity pattern cannot be related only to changes in the nature of the Precambrian terrains.

South of IRH, the seismic profile crosses the western edge of the Hoggar swell, the velocity decreases with elevation (Fig. 2). The lowest values, which correspond to the Atakor volcanic high massifs (Fig. 1b), are well-controlled by the stations installed inside. Therefore, this velocity perturbation can be interpreted as the consequence of thermal perturbation linked to recent volcanic activity. Nevertheless, a relation with a thicker cratonic crust must not been rejected.

Between stations RN8 and LAI, the Pharusian belt is characterized by positive velocity anomalies which could be significative of a thinner crust. North of LAI, the higher velocities correspond to the Eburnean In-Ouzzal domain. This fault limited granulitic block extends northward beneath the Paleozoic platform sediments of the Sahara basins. The higher velocity values between IHE and AFO stations correspond to a positive gravity anomaly interpreted as a 5-km thick ultrabasic rocks massif (Takherist, 1991).

The negative anomaly, north of the profile, can be correlated with the Paleozoic sediments of the In-Salah sedimentary basin whose thickness is about 10 km under station TIG.

In conclusion, a simple interpretation of the velocity perturbations in the crust shows a clear consistency between the velocity perturbations and the geological domains.

5.2. The mantle

The velocity perturbations in the upper mantle (layers 3, 4 and 5, Fig. 6) present the same global pattern than in the crust: a negative anomaly to the south of the profile, a positive anomaly to the north. When going deeper (Fig. 6), the low-velocity zone associated with the Hoggar extends toward the west and the north; the velocity contrasts are weak, ranging from −3% to +3%. Correlatively, the high-velocity zone, that we observe in the crust west of the 4°50′ fault, progressively shortens and migrates toward the north. The highest velocities are observed just west of the East In-Ouzzal fault and seem to correlate at depth with the In-Ouzzal granulitic lithospheric compartment.

In the crust, the velocity pattern correlates with the geology, but the lower velocities in the Central Hoggar are not consistent with the petrological composition of the crust. The fact that the negative anomaly at south extends deeply in the mantle and correlates with the Hoggar swell is a strong argument for assuming that the observed velocity variations are not associated with changes in the upper mantle structure inherited from Precambrian history but essentially correspond to upper mantle modifications in relation with Cenozoic to recent magmatic events. This hypothesis is confirmed by the occurrence of the lowest velocities (down to −5.5%) in the fourth layer (74–114 km) just beneath the Atakor and Tahalra quaternary volcanic provinces: in these massifs, the volcanism is very young (from −20 to −3.5 Ma in the Atakor, from −3.5 Ma to present for the Tahalra) and its origin is very deep (about 100 km, Dautria and Girod, 1991).

6. Discussion

The major result of this study is the presence of a lower P wave velocity structure beneath the Hoggar at any depth, from the surface to the bottom of our model, that is more than 300 km depth. This structure extends toward the north-west with the depth; however, the location and the shape of the seismic profile do not allow to define the extension of the low-velocity zone and its exact relationship with the Hoggar uplift. The velocity contrast with the adjacent areas is modest, the amplitude of the velocity perturbation is more pronounced under the recent volcanic provinces of Atakor and Tahalra where it reaches −5% between 74 and 114 km depth.

When numerous gravimetric studies have been performed on the African volcanic uplifts, only one teleseismic experiment has been carried out on the Adamawa region (Cameroon) (Dorbath et al., 1986). Although generally compared with the other African Tertiary uplifts characterized by negative Bouguer anomaly accompanied by lithospheric thinning (Fairhead, 1979; Bergmingham et al., 1983), the Adamawa plateau is related to the Cameroon volcanic line and the Central African megashear zone. The actual plateau, up to 2000 m high, is made of a Precambrian basement, uplifted during terminal Cre-

taceous and Tertiary time, and intruded by an abundant Cretaceous to present-day volcanism. Therefore, it provides a good example for comparison with the Hoggar. The result of the teleseismic tomography is the presence of a lower velocity region under the plateau down to the bottom of the model, that is 190 km; the velocity contrasts are much weaker than for the Hoggar and do not exceed 2.5%. The spatial extension of this low-velocity upper mantle is controlled by pre-existing features.

Outside Africa, the geological evolution of the Massif Central (France) presents some similarity with the setting of the Hoggar. This Hercynian mountain has been uplifted during the Alpine orogeny, then volcanic activity is reported from the beginning of Cenozoic up to present-day time. A recent teleseismic tomography provides three-dimensional images of the velocity structure beneath the Massif Central down to 180 km (Granet et al., 1995). A zone of anomalously low-velocity (approximately -3%) upper mantle is observed beneath the graben system and the volcanic provinces, confirming and refining the diapiric model previously deduced from geophysical measurements and modelling. The authors interpret the upper mantle anomaly as due to active upwelling in the mantle. In the crust, low-velocity areas show a remarkable correlation with volcanic complexes and are considered to be essentially the remaining thermal signature of magma chambers or dykes associated with the volcanic activity.

The image of the upper mantle that we have found under the Hoggar is in many aspects consistent with the results obtained in the regions mentioned above. These three regions of reactivated and uplifted old basement are characterized by a moderately low-velocity upper mantle and a negative Bouguer anomaly. In the case of the Massif Central, the shape of the anomalous upper mantle correlates with the shape of the gravity anomaly and defines a bulge pattern. For the Adamawa, the array did not allow to define the lateral extension of the anomalous upper mantle. However, the elongated shape of the associated gravity anomaly following the Cameroon volcanic line leads to affect the same geometry to the low-velocity zone. Our linear profile is located on the western end of the Hoggar swell and does not extend southward enough to cross it entirely, so it is not possible to define the lateral and

southward extension of the anomalous upper mantle. The topography of the Hoggar massif, together with the elongated ENE–WSW shape of the low density body deduced from the negative gravity anomaly (Lesquer et al., 1988) suggest that the lower velocity zone extends mainly toward the east, where it could be more pronounced. This is supported by the presence of a well-marked negative velocity perturbation to the north-east of the ASK on the deeper layers (layers 6 to 8, Fig. 6).

As in the Massif Central, we observe in the Hoggar that the lower velocities are correlated at the surface with the recent volcanic provinces (Atakor and Tahalra). In these districts, the volcanic activity was paroxysmal during the Miocene and was continuing episodically up to the Quaternary. The mantelic origin of this volcanism is clearly imaged by the tomography which shows well-marked negative anomalies underneath the Atakor and the Tahalra.

The part of old structures, prior to the uplifting of the Hoggar swell, is not clear on our results. When the velocity perturbations could be related to the age of the substratum, the highest velocities should be observed beneath the Hoggar shield. The higher velocity zone, north of 24°N, corresponds to the Pan-African domain situated on the north-western edge of the uplift and therefore less affected by this event. The velocity perturbations are clearly controlled by the event that originated the uplift.

In any case, teleseismic tomography only provides relative velocity perturbations. We have to look at the results from other geophysical studies to proceed toward the absolute velocities. Dorbath and Dorbath (1984), studying the mean teleseismic travel time residuals in Africa, showed that the stations can be separated in two groups (craton and mobile zones); in each group, the station anomalies are correlated with the altitude. In their study, the travel time residual calculated at Tamanrasset seismic observatory (TAM), -0.46 s, places TAM on the line of the uplifted cratons. This result, together with the normal heat flow value measured in the region, lead Lesquer and Vasseur (1992) to propose that the mantle beneath the Hoggar swell is intermediate between a cratonic zone mantle and an active zone mantle. Study of xenoliths entrained by the basaltic eruptions have lead Lesquer et al. (1988) to propose that this low density mantle corresponds essentially to an

altered, now-cooled mantle resulting of modifications induced by gas, fluid and magma transfer from the deep mantle during Cenozoic (Dautria et al., 1988).

The negative velocity perturbation down to 300 km depth agrees with this hypothesis of extensive upper mantle modifications in connection with asthenospheric material transfer. Moreover, as the velocity contrast is weak, this anomalous mantle cannot be regarded as a present upwelling of the asthenosphere as proposed by Brown and Girdler (1980). These lower velocities can be interpreted, in agreement with previous geophysical data, as due to mantle modifications produced by asthenospheric material transfer in relation with the Late Mesozoic–Early Cenozoic events. Nevertheless, the lowest velocities associated with Mio–Plio–Quaternary volcanic districts could indicate that high-temperature mantelic zones still exist beneath the recent volcanic provinces and probably intrude the crust.

In conclusion, and in agreement with the conclusions of Aït Hamou and Dautria (1997), this P wave teleseismic tomography corroborates the assumption that the Hoggar mantle cannot be regarded as a present continental hotspot and that the swelling is probably due to the existence of an anomalous mantle resulting of the Cenozoic activity of a hotspot. This anomalous mantle should have migrated toward the ENE with the African Plate motion. As proposed by Aït Hamou and Dautria (1997), a displacement of the African Plate of near 650 km to the NNE in the past 35 Ma is probable and the hotspot would presently be situated 400 km WSW of Tamanrasset. It is interesting to notice (Fig. 6, layers 7 and 8) that as we go deeper in the mantle, the low-velocity zone appears to migrate toward the west.

The other target of this study was to explore the mantle beneath the south Sahara basin, where heat flow values greater than 100 mW m^{-2} were interpreted as significant of the presence of a hot anomalous mantle (Lesquer et al., 1990). The tomography model shows lower velocities in the mantle beneath the northernmost stations. The extension of the teleseismic profile does not allow to precisely analyze the correlation with the heat flow anomaly. However, the velocity model is consistent with other geophysical data and would confirm the existence of a low-density–high-temperature mantle under the Sahara sedimentary basins. The southern limit of this anomalous zone was impossible to define from heat flow data and surface wave velocities (Lesquer et al., 1990). Our study suggests that it should be situated near latitude 26°N (Fig. 6). It is noteworthy that this limit is quite the same for the negative gravity anomaly, which correlate with the high heat flow zone (Lesquer et al., 1990).

7. Conclusion

The results of this teleseismic tomography, in agreement with other geophysical data, confirm that the Hoggar dome can be related to the presence of an anomalous low-density–low-velocity mantle. We propose that this anomalous mantle is the consequence of the Cenozoic hotspot activity and has cooled off during the migration toward the NE of the African Plate. Is there a hot anomalous mantle beneath the Sahara sedimentary basins? The limited extension toward the north of the profile does not allow us to really confirm this hypothesis but it is obvious that if such a structure exists, it is limited to the Sahara basins and does not affect the Hoggar Precambrian shield.

Acknowledgements

References

Aït Hamou, F., Dautria, J.M., 1994. Le magmatisme Cénozoïque du Hoggar: une synthèse des données disponibles. Mise au point sur l'hypothèse d'un point chaud. Bull. Serv. Carte Geol. Alger. 5 (1), 49–68.

Aït Hamou, F., Dautria, J.M., 1997. Spatiotemporal and compositional variations of the Cenozoic Volcanism in the Ahaggar area: evidence for a hot spot in relation with the African Plate motion. Terra Nova, EUG9, Strasbourg, p. 9.

Aki, K., Christofferson, A., Husebye, E.S., 1977. Determination of the three-dimensional seismic structure of the lithosphere. J. Geophys. Res. 82, 277–296.

Bergmingham, P.M., Fairhead, J.D., Stuart, G.W., 1983. Gravity study of the Central African Rift System: a model of continental disruption. The Darfur domal uplift and associated Cainozoic volcanism. Tectonophysics 94, 205–222.

Black, R., Caby, R., Moussine-Pouchkine, A., Bayer, R., Bertrand, J.M.L., Boulier, A.M., Fabre, J., Lesquer, A., 1979. Evidence for Late Precambrian plate tectonics in West Africa. Nature 278, 223–227.

Brown, C., Girdler, R.W., 1980. Interpretation of African gravity and its implication for the breakup of the continents. J. Geophys. Res. 85 (B11), 6443–6455.

Caby, R., 1996. A review of the In-Ouzzal granulitic terrane (Tuareg shield, Algeria): its significance within the Pan-African Trans-Saharan belt. J. Metamorph. Geol. 14, 659–666.

Dautria, J.M., Girod, M., 1991. Relationships between magmatism and upper mantle heterogeneities as exemplified by the Hoggar volcanic area (Central Sahara, Southern Algeria). In: Kampuzu, A.B., Lubata, R.T. (Eds.), Magmatism in Extension and Structural Settings. Springer-Verlag.

Dautria, J.M., Lesquer, A., 1989. An example of the relationship between rift and dome: recent geodynamic evolution of the Hoggar swell and its nearby regions (Central Sahara, Southern Algeria and Eastern Niger). Tectonophysics 163, 45–61.

Dautria, J.M., Lesquer, A., Dupuy, C., Aït Hamou, F., Chikhaoui, M., 1988. Le Bombement du Hoggar dans le Sahara central represente t — il la trace d'un point chaud?, Programme et Résumés, Volcanisme intraplaque. Le Point Chaud de la Reunion.

Dorbath, C., Dorbath, L., 1984. Approche sismologique de la structure de la lithosphère en Afrique de l'Ouest. Thesis, U. Paris VI, 307 pp.

Dorbath, L., Montagner, J.P., 1983. Upper mantle heterogeneities in Africa deduced from Rayleigh wave dispersion. Phys. Earth Planet. Inter. 32, 218–225.

Dorbath, C., Dorbath, L., Fairhead, J.D., Stuart, G.W., 1986. A teleseismic delay time study across the Central African Shear Zone in the Adamawa region of Cameroon. Geophys. J. R. Astron. Soc. 86, 751–766.

Evans, J.R., Achauer, U., 1993. Teleseismic velocity tomography using the ACH method: theory and application to continental-scale studies. In: Iyer, H.M., Hirahara, K. (Eds.), Seismic Tomography. Chapman & Hall, London, pp. 319–360.

Fairhead, J.D., 1979. A gravity link between the domally uplifted cainozoic volcanic centres of north Africa and its similarity to the east African rift system anomaly. Earth Planet. Sci. Lett. 42, 109–113.

Granet, M., Stoll, G., Dorel, J., Achauer, U., Poupinet, G., Fuchs, K., 1995. New constraints on the Geodynamical Evolution of the Massif Central (France): a mantle plume hypothesis from teleseismic tomography. Geophys. J. Int. 121 (1), 33–48.

Hadiouche, O., Jobert, N., Montagner, J.P., 1989. Anisotropy of the African continent inferred from surface waves. Phys. Earth Planet. Inter. 58, 61–81.

Herrin, E., 1968. Seismological tables for P phases. Bull. Seismol. Soc. Am. 58, 1193–1242.

Lesquer, A., Vasseur, G., 1992. Heat-flow constraints on the West African lithosphere structure. Geophys. Res. Lett. 19 (6), 561–564.

Lesquer, A., Bourmatte, A., Dautria, J.M., 1988. Deep structure of the Hoggar domal uplift (Central Sahara, South Algeria) from gravity, thermal and petrological data. Tectonophysics 152, 71–87.

Lesquer, A., Bourmatte, A., Dautria, J.M., 1989. First heat flow determination from the Central Sahara: relationship with the Pan-African belt and Hoggar domal uplift. J. Afr. Earth Sci. 9 (1), 41–48.

Lesquer, A., Takherist, D., Dautria, J.M., Hadiouche, O., 1990. Geophysical and Petrological evidence for an ''anomalous'' upper mantle beneath the Sahara basins (Algeria). Earth Planet. Sci. Lett. 96, 407–418.

Takherist, 1991. Structure crustale, subsidence Mésozoïque et flux de chaleur dans les bassins nord-Sahariens (Algérie): apport de la gravimétrie et des données de puits. Doc. Trav. Centre Geol. Géophys., Montpellier, 29.