The Hoggar swell and volcanism, Tuareg shield, Central Sahara: Intraplate reactivation of Precambrian structures as a result of Alpine convergence

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Introduction

Hoggar (or Ahaggar) volcanism, in the Central Sahara, is Eocene to Quaternary in age (35 to nearly 0 Ma; *Aït-Hamou et al.*, 2000 and references therein). Hoggar is located within the African plate far from the recent East African rift system. Associated with a swell 1,000 km in diameter (Figure 1), Hoggar Cainozoic volcanism is classically considered to be a mantle plume product (*Sleep*, 1990) even though no thermal anomaly has been observed (*Lesquer et al.*, 1989). When considering available geological, geophysical and petrological data, such a model is hardly supported. When integrating the data with the structure of the Precambrian basement and the present geodynamic environment, an alternative model comes to mind. This comprises a source at the lithosphere/asthenosphere interface melted by adiabatic pressure release in response to stress applied by the Africa-Europe collision on pre-existing shear zones and fractures, within a semi-rigid block (metacraton).

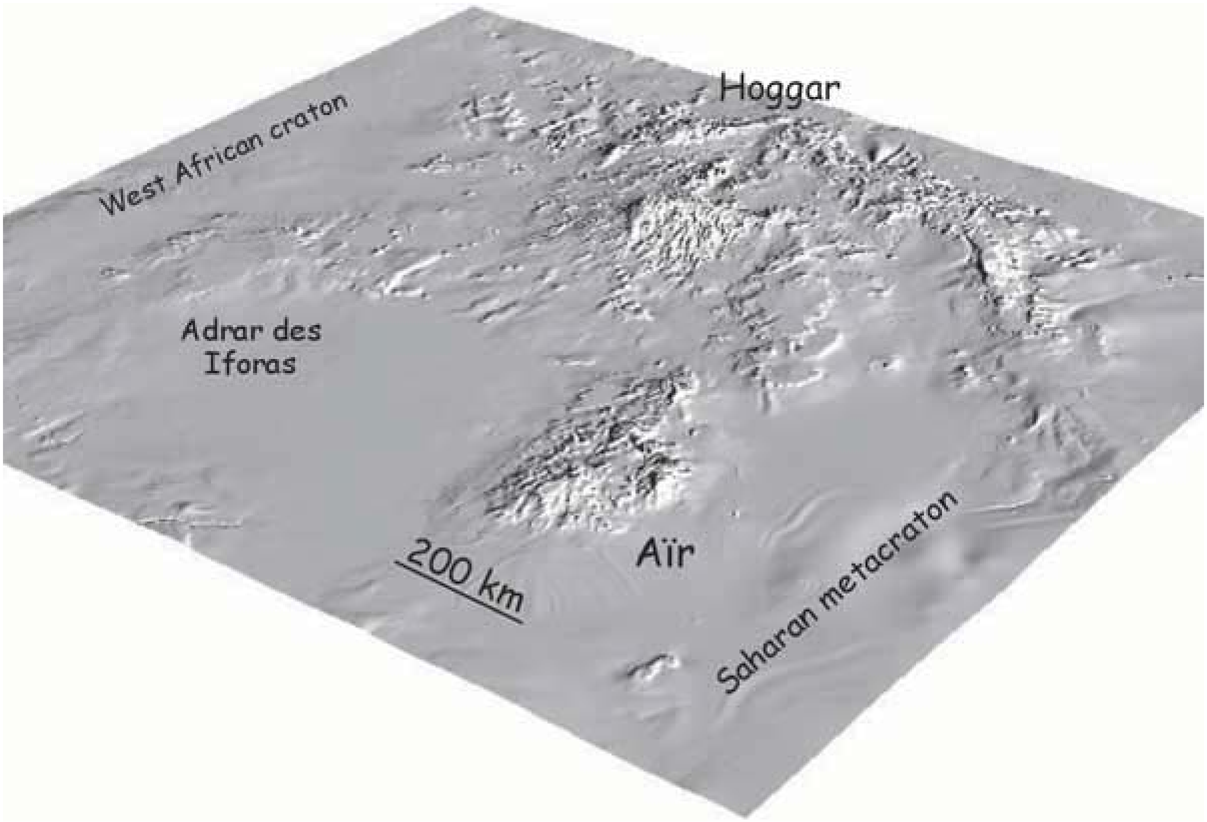


Figure 1: The Hoggar swell from GLOBE data, NOAA; processing by Ph. Trefois, MRAC.

The Precambrian basement

Hoggar forms the main part of the Tuareg shield, which is principally made of Archaean/ Palaeoproterozoic and Neoproterozoic terranes. These amalgamated during the late Neoproterozoic Pan-African orogeny (*Black et al.*, 1994) as a result of the convergence of the West African craton (WAC) and the Saharan (meta)craton (Figure 2).

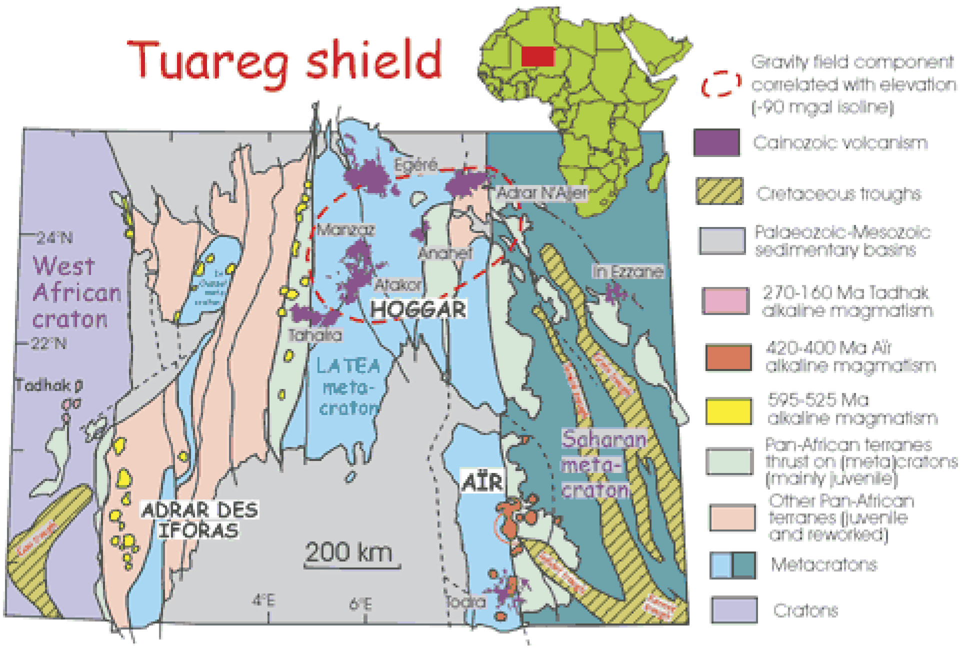


Figure 2: The Tuareg shield. Distinction of major types of terranes (after Black et al., 1994; Liégeois et al., 2003). Localities of alkaline magmatism: post-collisional (595-525 Ma; Ba et al., 1985; Liégeois & Black, 1987; Liégeois et al., 1998); Devonian in Aïr (Moreau et al., 1994); Permian-Jurassic in Tadhak (Liégeois et al., 1991) and Cainozoic volcanism (drawn from a satellite photograph – Orthorectified Landsat Thematic Mapper Mosaics as compressed color imagery in MrSIDTM file format from Lizardtech). Gravity anomaly from Lesquer et al. (1988), Cretaceous troughs from Genik (1993).

A first stage comprised the accretion of oceanic island arcs on these cratons and on microcratons during the period 900-680 Ma. Relics of these terranes, including ophiolites and eclogites, are preserved as thrust sheets on more rigid bodies (*Liégeois et al.*, 2003 and references therein).

The second stage was the regional northerly tectonic escape of the Tuareg terranes due to oblique collision with the WAC. It is characterised by spectacular N-S shear zones. During that stage, the metacratonization of the LATEA (Central Hoggar) microcontinent occurred, i.e. the squeezing of this rigid body, which was torn into several moving blocks. For a description of the metacraton concept, see *Abdelsalam et al.*, 2002. This induced sliding along N-S mega-shear zones and the intrusion of granitoid batholiths (615-580 Ma) with geochemical signatures indicative of the lower crust (Sr_i= 0.710; $epsilon_{Nd}$= -20). Linear lithospheric delamination beneath these mega-shear zones may occur under such circumstances, allowing a drastic increase in heat flow and melting of the crust. Post-collisional and anorogenic high-level alkaline plutons are aligned on the same mega-shear zones, particularly along (meta)craton margins (Figure 2; *Liégeois et al.*, 1998; *Azzouni-Sekkal et al.*, 2003).

Hoggar during the Palaeozoic and Mesozoic

At the beginning of the Phanerozoic, the Tuareg shield was entirely eroded and covered by Ordovician sandstones whose source region was to the south. The shield did not constitute an obstacle for the sedimentation flux (*Beuf et al.*, 1969).

The presence of rare Cretaceous continental remnants lying directly on top of Hoggar Precambrian indicates that the shield was already uplifted and slightly re-eroded before the Cretaceous. The palaeo-currents in Cretaceous sediments around the shield indicate that Hoggar was bypassed by the sediment flux from the north (*Faure*, 1985). Hoggar, as well as the Eglab and Tibesti massifs, was an island during the Cretaceous, being only covered by continental lakes (*Fabre*, 1976). This suggests that the swell already existed during the Cretaceous. The current presence of Cretaceous deposits at an altitude of 2,000-3,000 meters (Rognon; *Fabre*, 1976) indicate the importance of uplift during the Cainozoic, however. This Cretaceous uplift could be linked to the development of a series of Cretaceous elongate basins in West Africa (Figure 2; *Genik*, 1993; *Dautria & Lesquer*, 1989).

Cainozoic volcanism & plutonism

Recent volcanic activity began during the late Eocene (35-30 Ma; *Aït-Hamou et al.*, 2000) and lasted probably till the upper Palaeolithic or even Neolithic (*Lelubre*, 1952). It occurred in several districts (Figure 2): Tahalra (1,800 km^2), Atakor (2,150 km^2), Manzaz (1,500 km^2), Egéré (2,800 km^2), Anahef (400 km^2) and Adrar N'Ajjer (2,500 km^2), all in Central Hoggar, and also in Teria (100 km^2) to the NE, In Ezzane (800 km^2) to the east and Todra (1,050 km^2) to the south of Aïr. This volcanism forms high relief, often over 2,000 m in altitude, culminating in the Atakor (Mt. Tahat, which is 2,918 m high). The Hoggar swell reaches an altitude of 350-400 m on its margin and 1,000-1,500 in its centre. (The town of Tamanrasset is at an altitude of 1420 m). Some Precambrian basement inliers have been observed in Atakor at an altitude of up to 2,600 m.

Three main stages can be identified for Hoggar volcanism:

1. The oldest stage is Upper Eocene to Oligocene and is present only in the Anahef district. It corresponds to tholeiitic basalts of fissural origin (thickness of 600 m; 35-30 Ma; *Aït-Hamou et al.*, 2000), intruded by a dozen subvolcanic ring complexes (c. 29 Ma for the Achkal complex; *Maza et al.*, 1995) made of gabbros, diorites, monzonite, alkali-feldspar syenites and nepheline syenites. The largest is the Tellerteba ring complex, which is 8 x 5 km in size. These complexes are covered by alkaline rhyolites whose age (c. 24 Ma) suggests uplift of 0.4 mm/year (*Maza et al.*, 1995). The deeper levels currently exposed in the Anahef district are then linked to contemporaneous uplift of Hoggar.

2. The second stage is the most voluminous and is well represented in the Atakor district (*Girod*, 1971). It occurs between 20 and 12 Ma and between 7 and 4 Ma (*Rognon et al.*, 1983) and comprises 80% basalts and 18% trachytes and phonolites (Figure 3).

3. The latest stage is late Pliocene to late Quaternary (3 Ma to upper Palaeolithic). In Atakor, this volcanism only comprises basanites and nephelinites, which follow valleys and can cover upper Palaeolithic terraces. In the Tahalra district, and also probably in the lesser-known Manzaz, Egéré and Adrar N'Ajjer districts, most of the volcanic activity occurred between 3.5 and 2.5 Ma (*Aït-Hamou*, 2000). It comprises 95% alkali basalts (basanitic and hawaiitic flows). Some volcanoes are more recent, and some cover Neolithic terraces.

Figure 3. The Iharen phonolitic plug, near Tamanrasset.

The In Ezzane district is almost unknown, due to its remote location (Figure 4). The In Teria district comprises melilitic rocks rich in xenoliths (*Bossière & Megartsi*, 1982; *Dautria et al.*, 1992). In Aïr (Niger), the Todra district (*Black et al.*, 1967) began with about 30 trachytic and phonolitic volcanoes, many erupting very viscous lavas and trachytic tuffs. The Todra volcano itself lies at an altitude of 1,780 m. This stage was followed by the formation of about 130 basaltic volcanoes (Figure 5). They are generally regular cones and only one flow following valleys was generated. *Black et al.* (1967) estimated that some volcanoes cannot be more than a few centuries old. They remarked also that these volcanoes are located either on NW-SE oriented faults or on annular faults linked to Devonian ring complexes (Figure 6). The NW-SE faults belong to conjugate brittle faults that developed at the end of the Pan-African orogeny, due to the WAC indentation (*Ball*, 1980).

Figure 4. Satellite photograph of the In Ezzane district (Orthorectified Landsat Thematic Mapper Mosaics as compressed color imagery in MrSIDTM file format from Lizardtech). The lavas appear blue. The circular structures have not yet been studied. Click on image to enlarge.

Figure 5. Example of a small Aïr basaltic volcano, which generated one flow

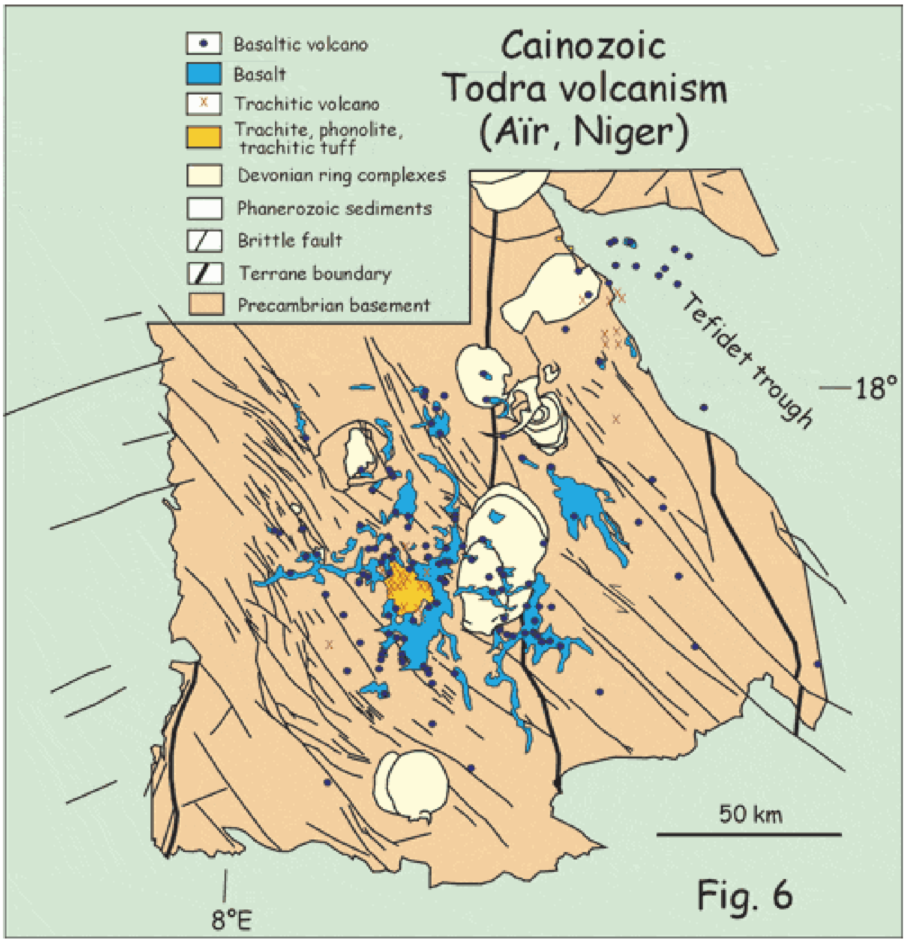


Figure 6. Map of southern Aïr showing volcanoes and lava flows and their link with the terrane boundaries, the late Pan-African brittle faults and the Devonian ring faults (from Black et al., 1967; 1985; Liégeois et al., 1994).

Geophysical data

Crough (1981) proposed that Central Hoggar is underlain by anomalously light mantle, on the basis of free air gravity data. He interpreted this as indicating a magmatic body at a depth of < 60 km below the surface. *Lesquer et al.* (1988) analysed the shape of the anomaly (Figure 2), which appears to be correlated with the basement elevation, and suggested that the body lies between 20 and 70 km depth and has a maximum thickness of 30 km. This association between the gravity anomaly and the Cainozoic volcanism led to the suggestion that a mantle plume underlies Hoggar (*Aït-Hamou & Dautria*, 1994; *Aït Hamou et al.*, 2000). However, heat flow measured inside the Hoggar swell is normal (50 mW/m^2) and there is no evidence for a thermally perturbed lithosphere (*Lesquer et al.*, 1989). On the contrary, a decrease in heat flow is observed with basement elevation, which is inconsistent with a thermal uplift. Only a minor thermal structure is possible below the centre of Atakor, where heat flow of 63 mW/m^2 has been measured (*Lesquer et al.*, 1989). The latter suggests that small high-temperature mantle bodies intrude the crust beneath the recent volcanic areas (*Ayadi et al.*, 2000).

The low-density zone proposed on the basis of gravity data thus cannot be interpreted as an upwelling of the lithosphere/asthenosphere boundary. In contrast, a large thermal anomaly occurs to the north, below the Saharan basins (*Lesquer et al.*, 1990; *Lesquer & Vasseur*, 1992). A low *P*-wave velocity structure has been detected beneath Central Hoggar, extending from the surface down to 300 km (*Ayadi et al.*, 2000). The velocity contrast with adjacent areas is modest, however, and the anomaly is strongest beneath Atakor and Tahalra where it reaches -5% between 74 and 114 km depth. Considering both the seismic and heat flow data, *Lesquer & Vasseur* (1992) proposed that the mantle beneath Central Hoggar is intermediate between cratonic mantle and activated mantle. If this anomaly is related to a hot body, the latter must have been emplaced before 60 Ma in order to have cooled off (*Lesquer et al.*, 1988).

The data from the xenolith included within the alkali basalts

Following the work of *Girod et al.* (1981) and *Dautria et al.* (1987), it may be concluded that amphibole-rich xenoliths from Hoggar alkali basalts indicate a mantle metasomatism event during or just before (max. 40 Ma) the generation of Hoggar magma between 1,000°C and 1,100°C and 15 to 18 kbar. Xenoliths and lavas have low $^{87}Sr/^{86}Sr$ isotopic ratios – between 0.70306 and 0.70344. Based on the xenoliths, the source is enriched by a factor of 7–9 chondritic in LREE and 2 in HREE. The degrees of partial melting vary from 1.2 to 2.1% for nephelinitic melts and 3.8% to 4.4% for basanitic melts, both leaving a garnetiferous residuum. There is a positive correlation between the degree of deformation of the xenoliths and the abundance of “metasomatic” minerals (principally amphibole), which indicates that the fluids were preferentially injected along strained zones. The wide regional distribution and abundance of such amphibole-rich rocks in the Hoggar upper mantle indicate that it is highly veined and hydrous along the shear zones. Such mantle metasomatization could be the cause of the present uplift of the Hoggar. In addition, the Hoggar basement uplift could be seen as an isostatic response to upper mantle density reduction controlled by magmatic events and associated metasomatism.

The isotopic data

Pb, Sr and Nd isotopic compositions of Hoggar volcanic rocks lie within the EM1 and HIMU components (*Allègre et al.*, 1981; *Dupuy et al.*, 1993; *Aït-Hamou et al.*, 2000), the influence of the latter seeming to increase with time. The main alkaline basalts have $^{87}Sr/^{86}Sr$ varying from 0.7030 to 0.7037, epsilon$_{Nd}$ from +3 to +7, $^{206}Pb/^{204}Pb$ from 19.3 to 20.5 and $^{207}Pb/^{204}Pb$ from 15.59 to 15.70, while the older (c.35 Ma) Taharaq tholeiitic basalts have $^{87}Sr/^{86}Sr$ varying from 0.7035 to 0.7045, epsilon$_{Nd}$ from -2.5 to +3, $^{206}Pb/^{204}Pb$ from 18.3 to 19.5 and $^{207}Pb/^{204}Pb$ from 15.54 to 15.60. The EM1 pole has been attributed to the Pan-African lithosphere and the HIMU component (with some DM) to a plume, itself recycling subducted Proterozoic oceanic lithosphere (*Aït-Hamou & Dautria*, 2000).

The geodynamics of the African plate during late Mesozoic and Cainozoic

During the Cretaceous, the African plate was subjected to an important rifting event. This led to the equatorial Atlantic opening and to the generation of intraplate rifts, including the Bénué rift (*Guiraud & Bosworth*, 1997, and references therein). In the Hoggar area, Cretaceous rifts developed to the SE, within the Saharan metacraton (Figure 2). They show sediment thicknesses of (*Genik*, 1993):

- Kermit trough – 10 km of Cretaceous (K) and 5 km of Cainozoic (C)
- Ténéré trough – 5 km of K and 2 km of C
- Tefidet trough – 3 km of K and 1 km of C
- Grein trough – 1.5 km of K, 1.5 km of Palaeogene and 2 km of Miocene/ Holocene.

This indicates contemporaneity of the Hoggar swell and the development of nearby troughs. During the Cretaceous and the Cainozoic, the African plate moved towards Europe, changing direction from NE to NW. The main tectonic events occurred at 112, 85, 65, 37, 20, 8 and 1 Ma at the plate margin, with effects within the African plate (*Guiraud & Bosworth*, 1997 and references therein). At 35 Ma Hoggar was located at 3°W and 18°N (using theGeomar-ODSN reconstruction facility - http://www.odsn.de/), i.e. 1,000 km to the SW, near the current position of Timbuktu.

An alternative to the plume hypothesis

The constraints on Hoggar volcanism are:

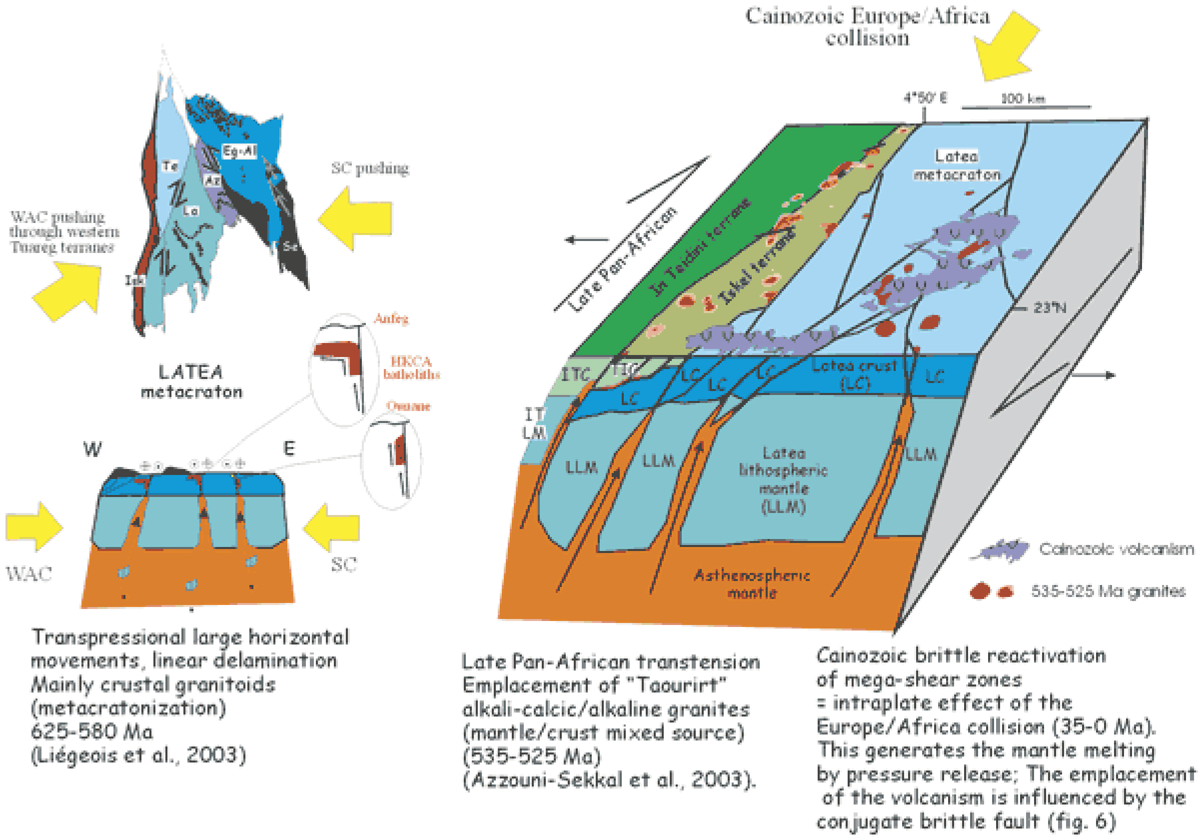
- If the negative gravity anomaly associated with the Hoggar swell is linked to a formerly hot body, the latter must be older than 60 Ma, to have had enough time to cool. It could be much older, however.
- Recent Hoggar volcanism is associated with a swell but not a thermal anomaly. A current large thermal anomaly is located to the north, below the Saharan basins, while the African plate has moved to the north-east. This thermal anomaly thus cannot be the present signature of a former Hoggar plume.
- Slightly higher heat flow has been measured very locally in the Atakor, suggestive of local intrusion of asthenospheric mantle into the Hoggar crust.
- As a whole, Central Hoggar has a structure intermediate between a craton and a mobile belt, which corresponds to the concept of a metacraton (a partially destabilized craton). Coupled seismic and heat flow data indicate also that the mantle beneath Central Hoggar is intermediate between cratonic mantle and active mantle.
- In Hoggar, strong xenolith deformation is observed, indicating the importance of modal mantle metasomatism. Fluids are preferentially injected along strained zones.
- The source of the Hoggar basalts is REE enriched. Their Sr-Nd-Pb isotopic ratios indicate an enriched mantle source, particularly at the beginning of the event. The involvment of old radiogenic lithosphere is very weak.
- The movement of Africa towards the north induced several tectonic phases (T)

the latest of which appears to have been followed by volcanism (V) in Hoggar at T= 37 Ma, V= 35-30 Ma; T= 20 Ma, V= 20-12 Ma; T=8 Ma, V= 7 – 2 Ma; and T= 1 Ma, V= most recent.

It may be added that the location of Hoggar, and more widely, Tuareg volcanism is linked to the rheological characteristics of the Pan-African terranes. It is located along shear zones within or at the boundaries of metacratons (i.e., semi-rigid blocks; Figure 2). It is also influenced by the late Pan-African brittle conjugate fault system (Figure 6). We could suggest the application of what has been demonstrated recently for the Bicol volcanic arc (*Pasquarè & Tibaldi*, 2003) where magma rose at depth along NW-striking transcurrent faults, but where in the uppermost crust magma rose along NE-striking fractures, parallel to the direction of greatest principal stress. Indeed, in the Hoggar-Tuareg shield, the general locations of the volcanic districts are related to the main shear zones while the precise locations of individual volcanoes are determined either by NE-SW or the NW-SE fractures. We know that the African plate moved, following the period considered, either to the NE (84-37 Ma; 21-8 Ma) or the NW (37-21 Ma; 8-0 Ma). More detailed study is needed to determine the details of this correlation. On the other hand, no trace of a plume is observed between the present location of Central Hoggar and its former position 35 Ma ago, c. 1,000 km to the WSW, near the current position of Timbuktu.

Hoggar volcanism may thus be the consequence of intraplate rejuvenation of Pan-African structures linked to the Africa-Europe collision. Based on the model proposed for the Pan-African (late Neoproterozoic) granitoids that intruded in same areas, I suggest that linear lithospheric delamination occurred along the Pan-African mega-shear zones located inside metacratons or at their margins (Figure 7). This event was more limited in intensity than at the end of the Pan-African orogeny. It allowed the rise of enriched material from the lithosphere/asthenosphere boundary (*Black & Liégeois*, 1993; *Anderson*, 1995) that can melt by adiabatic pressure release but this hot material was not abundant enough to melt the old lithosphere nor to generate a thermal anomaly, except very locally. (Minor involvement could be envisaged for the early tholeiitic basalts, generated by a larger degree of partial melting and characterized by a more enriched isotopic signature.) The role of the relatively shallow light body (20-70 km) is not yet clear as its age is not known. It is older than 60 Ma, and could be much older. Together with the mantle metasomatism linked to Hoggar volcanism and with the stress transmitted from plate boundaries, it could have contributed to the uplift and generation of the Hoggar swell. We must not forget, however, that the uplift observed in Aïr (Figure 1) is also important (the basement is elevated to up to 1,900 m) and is located far from this anomaly. It is, in contrast, located on the western margin of the Saharan metacraton, which is marked by one of the most important Pan-African shear zones in the Tuareg shield.

Figure 7. (Overleaf) Proposed alternative model for Hoggar. During the post-collisional Pan-African period, the LATEA craton was dissected by mega-shear zones (metacratonization). Major linear delamination induced asthenospheric upwelling able to melt the lower crust. this gave rise to mainly crustal granitoids (620-580 Ma; Liégeois et al., 2003 and references therein). At the end of the convergence, a similar event occurred but with less intensity, and gave rise to high-level alkali-calcic plutons of mixed mantle/crust origin (535-525 Ma; Azzouni-Sekkal et al., 2003 and references therein). This model proposed a similar scenario for the recent volcanism, that it is the intraplate consequence of Africa-Europe convergence. An even lower intensity event could cause the mainly mantle origin of Hoggar volcanism. See text for more details.



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