

The Hoggar swell and volcanism: Reactivation of the Precambrian Tuareg shield during Alpine convergence and West African Cenozoic volcanism

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Myth: A simplified picture, often illusory, that groups of humans elaborate or accept concerning a person or a fact and which plays a determining role in their behavior or their appreciation.

Translated from *Dictionnaire Le Robert* (1990)

ABSTRACT

We review the northwest African Cenozoic volcanic fields, including their regional geology. This provides a basis for understanding the relations between Hoggar volcanism and the Africa-Europe collision. Volcanic alignments are related to structural features, and no spatial age trend exists. In Hoggar, a close link is established between the volcanism and Pan-African structure. During the Mesozoic rifting period, the Hoggar area was already a topographic high well before any volcanism, which began at ca. 35 Ma, just after the initiation of the Africa-Europe collision at ca. 38 Ma. Hoggar volcanism continued episodically until now, as did the collision. We describe the Hoggar volcanic province based on available field, petrological, geochemical isotopic, and geophysical data, including data on gravimetry, heatflow, and seismic tomography. The latter suggests that northwestern African volcanism is linked to mantle structure down to 150 km but not deeper, implying a shallow mantle source. In Hoggar, lithospheric structures deduced from the seismic tomographic model and from geology are compatible when their respective resolutions are taken into account.

The considerations just stated cannot be reconciled with a plume model. We propose instead that intraplate stress induced by the Africa-Europe collision reactivated the Pan-African mega-shear zones mainly in metacratonic terranes, inducing linear

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lithospheric delamination, rapid asthenosphere upwelling, and melting due to pressure release. Edge-driven convection may contribute. The surface location of the volcanism is influenced by Paleozoic and Mesozoic brittle faults.

Keywords: Hoggar, northwestern Africa, Cenozoic volcanism, Pan-African structure, Africa-Europe collision

INTRODUCTION

Africa was mostly built during the Proterozoic. The Phanerozoic orogenies affected only the extreme northwestern and southern parts of the continent, and currently 90% of Africa is surrounded by spreading ridges. Notwithstanding, Africa has a high mean elevation, particularly its southern part (Doucouré and de Wit, 2003, and references therein). The link between volcanism and elevation is not simple: the southern part of the continent has the highest mean elevation (plateaus between 700 and 1000 m), but Cenozoic volcanism is scarce and old. In the Namaqualand-Bushmanland and Cape province of western South Africa, numerous small melilitite (with carbonatite- and kimberlite-type) plugs have ages of 77–54 Ma (Duncan et al., 1978; Moore and Verwoerd, 1985; Verwoerd et al., 1990). In Namibia, they are 50–48 Ma (Reid et al., 1990). Rare younger phonolite and melilitite occur in southwestern Namibia (37–35 Ma; Kröner, 1973). No more recent volcanism is associated with elevated southern Africa.

In the rest of Africa, Cenozoic volcanism can be subdivided in four groups: (1) the huge volcanic belt linked to the East African rift (800,000 km³; Ebinger and Sleep, 1998); (2) the Cameroon line; (3) provinces located within the Precambrian basement of northwestern Africa, which is sometimes linked to swells; and (4) provinces associated with the Alpine Atlas belt. In these four groups, volcanism took place from ca. 45 Ma until the present. This paper focuses on the Hoggar volcanic provinces and will encompass the two last groups (Fig. 1).

In Hoggar (or Ahaggar), recent volcanic activity is Upper Eocene to Quaternary in age (35 to nearly 0 Ma; Aït-Hamou et al., 2000, and references therein). Associated with a crustal swell 1000 km in diameter (Fig. 2), it is classically considered the product of a mantle plume (Sleep, 1990; Burke, 1996; Ebinger and Sleep, 1998; Aït-Hamou et al., 2000). However, no thermal anomaly has been detected (Lesquer et al., 1989), nor is uplift limited to Central Hoggar (Fig. 2), while a link between the location of Hoggar volcanic formations and Pan-African (late Neoproterozoic) geology is clear (Fig. 3). Looking at the whole of northwestern African Cenozoic volcanism, links can be discerned between the Alpine orogeny and the lithospheric morphology deduced from tomography. When considering the geological, geophysical, and petrological data available, particularly mantle xenoliths, the plume model is poorly supported.

Another model is suggested by the relationship between

structure of the Precambrian basement, lithospheric morphology, and the present geodynamic setting. In response to stress resulting from the Africa-Europe collision (Bailey, 1992), volcanism may be generated by adiabatic pressure release of uprising asthenosphere. Upwelling may be due on the one hand to the edge effect (King and Anderson, 1998) at the underlying lithospheric step. On the other hand, it may result from reactivation of pre-existing shear zones and fractures generated during the Pan-African (late Neoproterozoic) orogeny within a semi-rigid block (metacraton), inducing limited linear lithospheric delamination at the lithosphere-asthenosphere interface along these mega-shear zones (Liégeois et al., 2003). This may have induced upwelling of the asthenosphere immediately below. Therefore, taking account of the structure of the lithosphere and of the geology of the Precambrian basement is essential for understanding the location of Cenozoic Hoggar volcanism. Such ideas may be extended to the whole of northwestern Africa.

In order to set the scene for the ensuing discussion, the next section will describe in detail northwestern African volcanism. We have already presented a map of the volcanic fields in the area (Fig. 1). As far as we are aware, there is nowhere else a map showing all northwestern African Cenozoic volcanic fields, and no synopsis such as we give in Table 1. We have summarized much of the relevant literature, a great deal of which is obscure or in French. The details we present underpin our interpretation of northwestern African volcanism, in particular in relation to Hoggar.

GEOLOGICAL CONSTRAINTS

Cenozoic Volcanism in Northwestern Africa

The Hoggar swell is far from the only place in northwestern Africa where Cenozoic volcanic activity occurred (Fig. 1; Table 1). The largest province extends from Tibesti in Chad, where it is marked by a swell, through the Libyan plain toward Tripoli, where the main volcanic fields are Nuqay (or Eghei), Al Haruj, As Sawda, Al Hasawinah, and Garhyan (Fig. 1). Other provinces involve a series of much smaller volcanic areas associated with the Alpine Atlas belt. The Canary Islands can be included in this province (Fig. 1).

The Tibesti Volcanic Field. This region is important for understanding West African Cenozoic volcanism and therefore Hoggar volcanism. We present a relatively detailed description

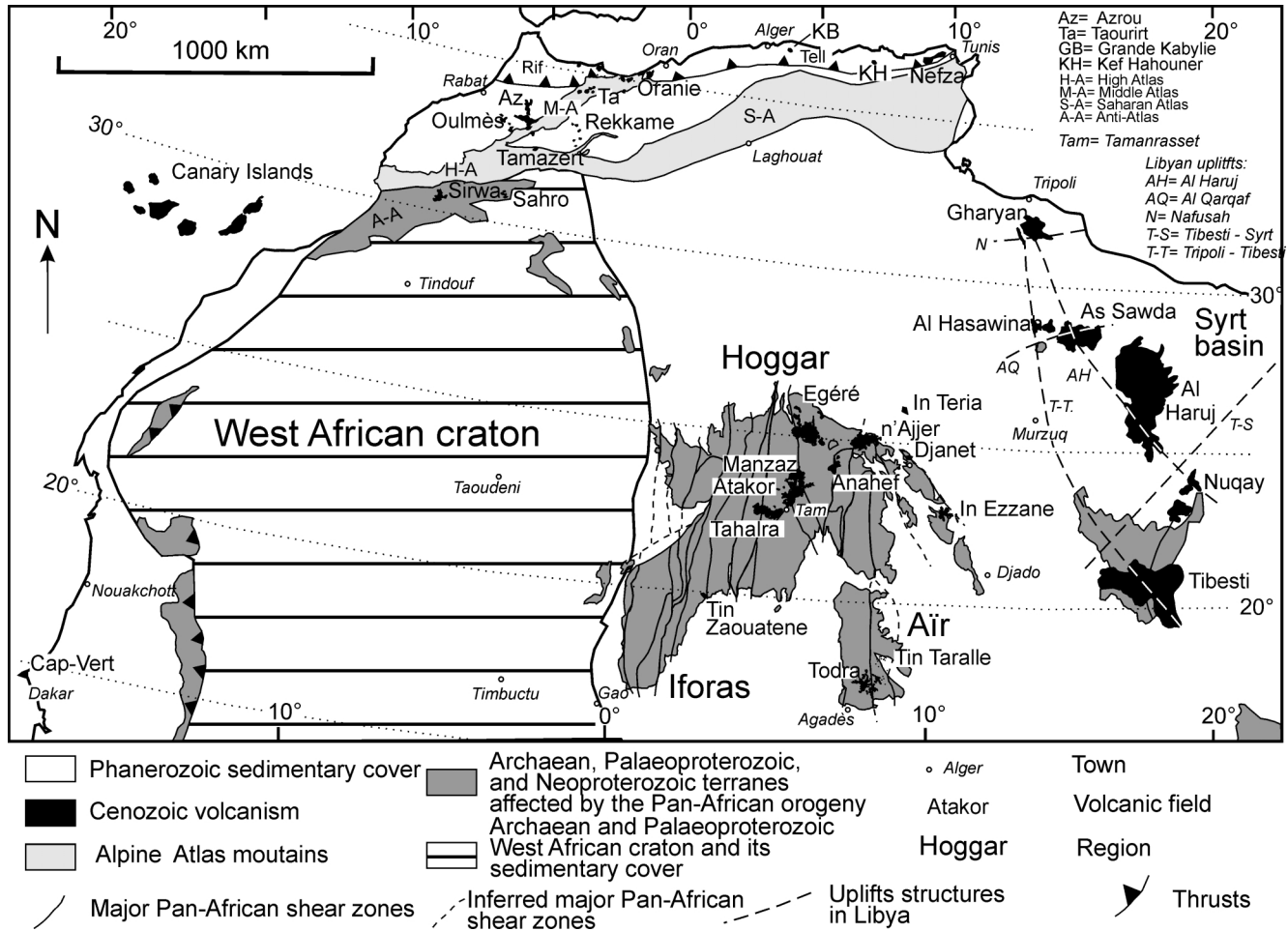


Figure 1. Cenozoic volcanism in northwest Africa. Map based on Fabre (1976); Black et al. (1994); Black et al. (1967); Saadi (1982). The volcanic fields in the Tuareg shield were drawn from a satellite photograph; Orthorectified Landsat Thematic Mapper Mosaics as compressed color imagery in MrSID™ file format from Lizardtech.

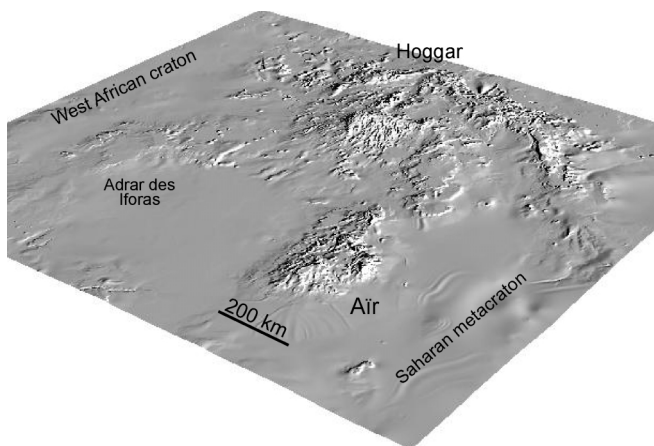


Figure 2. The Hoggar and Air swell. The uplift is more extensive than just the main Hoggar volcanic field (from GLOBE data, National Oceanic and Atmospheric Administration; processing by Philippe Trefois, Africa Museum).

as well as a map (Fig. 4). The Tibesti volcanic field covers an area of 30,000 km² with an estimated volume of 3100 km³ (Vincent, 1970) and lies on a swell whose Precambrian basement culminates at 2000 m above sea level. Volcanoes constitute the highest points, culminating with the Emi Koussi volcano at 3415 m. As a whole, the Tibesti volcanic field is located on a terrane boundary (El Makhrouf, 1988) separating Pan-African high-grade metamorphic rocks to the east from Neoproterozoic low-grade sediments to the west (Fig. 4).

The earliest Cenozoic volcanic series are younger than Middle Eocene sediments (Vincent, 1970). Tibesti volcanic activity has been intense since the Lower Miocene (Gourgaud and Vincent, 2004), and currently the Ehi Toussidé volcano can be considered dormant with fumarolic activity (Vincent, 1970). Ages between 9.7 and 0.3 Ma are cited with no detail given (Reynolds and Hall, 1976), while the large Emi Koussi shieldlike volcano (60–70 km in diameter) has been dated between 2.42 and 1.32 Ma (Gourgaud and Vincent, 2004). The earliest uplift is known to

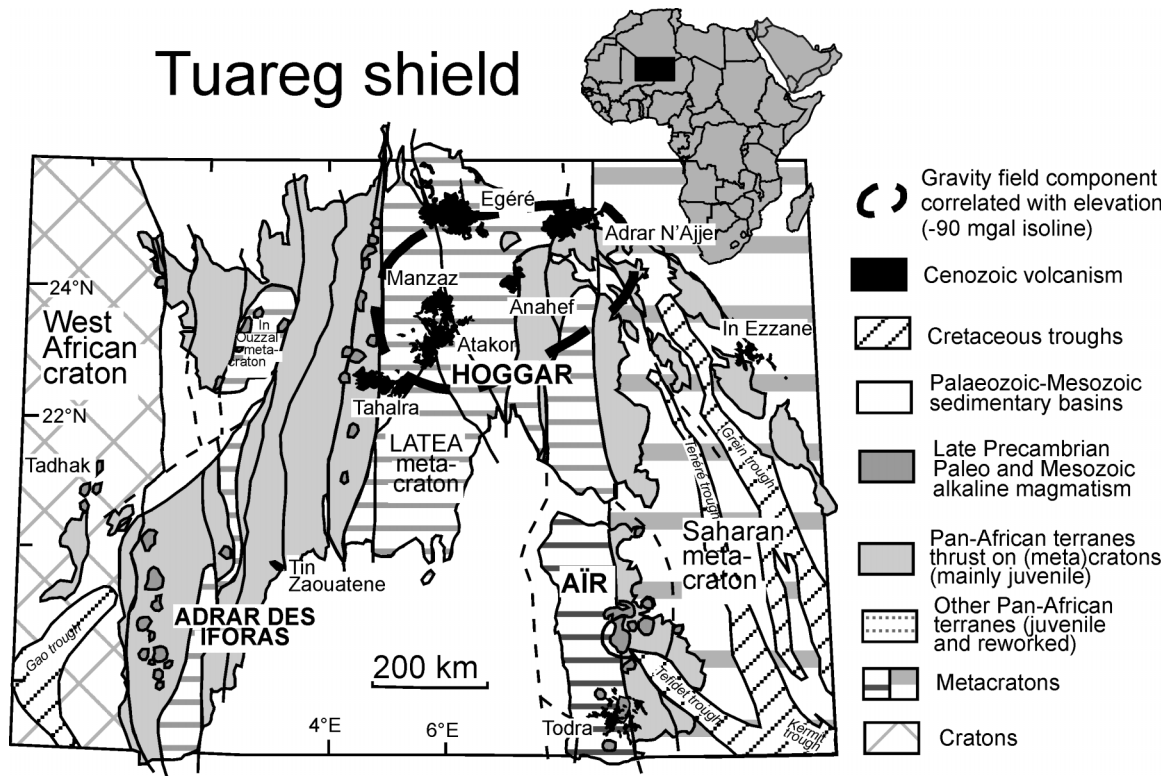


Figure 3. The Tuareg shield showing distinction of major types of terranes. After Black et al. (1994); Liégeois et al. (2003). Ages of alkaline magmatism are as follows: late Pan-African (595–525 Ma; Ba et al., 1985; Liégeois and Black, 1987; Liégeois et al., 1998); Devonian in Air (Moreau et al., 1994); Permian-Jurassic in Tadhak (Liégeois et al., 1991); and Cenozoic volcanism (drawn from a satellite photograph, with Orthorectified Landsat Thematic Mapper Mosaics as compressed color imagery in MrSID™ file format from Lizardtech). Gravity anomaly from Lesquer et al. (1988); Cretaceous troughs from Genik (1993).

be much older—pre-Cretaceous in age—because in the inner part of the Tibesti the thick Paleozoic sediments have been eroded and the erosion surface is covered by the Lower Cretaceous Nubian sandstones (Vincent, 1970; Fig. 4).

The first volcanic event produced Miocene composite flood lavas (plateau basalt) essentially made of alkali olivine basalts, but ignimbritic rhyolites may be locally abundant (Vincent, 1970). Spectacular trachytic and phonolitic plugs are associated with this phase.

The second event (Vincent, 1970) is represented by four large shield (Hawaiian-type) volcanoes (Oyoye, Tarso Yéga, Tarso Toon, Tiroko, all 40–60 km in diameter) and two large elevated domes (Dadoi and Abéki). The dominant rock type is labradorite-rich porphyritic tholeiitic to transitional basalt with a potassic tendency (mean $K_2O = 1.6\%$; Vincent, 1970) associated with alkaline rhyolites. The four large tholeiitic shield volcanoes are located on or very close to the terrane boundary (Fig. 4).

The third event is dominated by ignimbritic volcanoes with flattened cupolas centered on vast calderas (Vincent, 1970), among which the huge Emi Koussi volcano is also located on the terrane boundary (Fig. 4). This volcano is typically alkaline and exhibits two bimodal lava series, one silica-saturated

(basalt, trachyandesite, comenditic trachyte) and the other silica-undersaturated (basalt, phonolite) (Gourgaud and Vincent, 2004). Tarsa Voôn and Yirrigué, the other ignimbritic volcanoes, contain abundant peralkaline rhyolite (Vincent, 1970).

The fourth and latest event comprises Quaternary alkali basalt lava flows and trachy-andesitic volcanoes (Mousgou, Ehi Timi, Ehi Toussidé, Trou au Natron). This final event produced a very thin blanket of deposits representing only 2% of the volume of the Tibesti volcanic rocks.

The Large Libyan Volcanic Fields. In Libya, the volcanic fields are huge, covering more than 66,000 km² (Fig. 1), but correspond mostly to the first event in Tibesti, i.e., the deposition of the plateau basalts. They are located on uplifted areas delimited by NNW-SSE early Paleozoic and ENE-WSW Mesozoic structures (Klitzsch, 1970; Woller and Dediuk, 1980; Busweril and Oun, 1991).

To the south, Nuqay (also called Egheï) is close to the Tibesti massif and located on the terrane boundary depicted in Tibesti. It is composed of flood basalts that overlie Middle Eocene sediments (Lelubre, 1946) and are identical to the Tibesti plateau basalts (Vincent, 1970).

Al Haruj is the largest volcanic province in West Africa, its

TABLE 1: CENOZOIC VOLCANISM OF NORTHWEST AFRICA

Volcanic field	Province	Country	Age (Ma)	Estimated area (km ²)	Estimated volume (km ³)
Anahef	Hoggar	Algeria	35–24	400	200 (1000?)
Atakor	Hoggar	Algeria	20–12; 7–4; 3–0.01	2150	250
Djanet	Hoggar	Algeria	8.4–6.8	50	6
Egéré	Hoggar	Algeria	3.5–2.5	2800	325
In Ezzane	Hoggar	Algeria	Recent	500	60
Manzaz	Hoggar	Algeria	20–12; 7–4; 3–0.01	1500	175
N'Ajjer	Hoggar	Algeria	3.5–2.5	2500	290
Tahalra	Hoggar	Algeria	Pliocene to Holocene	1800	210
In Teria	Hoggar	Algeria	Holocene (?)	100	12
Tin Taralle	Aïr	Niger	28–20; 15–8; 4–0.7	50	6
Todra	Aïr	Niger	28–20; 15–8; 4–0.7	1050	120
Tin Zaouatene	Iforas	Mali	Holocene	50	6
Tibesti		Chad	9.7 to 0.3	30,000	3500
Nuqay		Libya	n.d.	8000	950
Al Haruj		Libya	? 6–2.2–0.4	45,000	5000
As Sawda		Libya	12.3–9.8	10,000	1200
Al Hasawinah		Libya	24.9; 15.7	1000	110
Gharyan/Nafusah		Libya	40; 11; 8–2	3000	350
Nefza-Tabarka	Tell	Tunisia	10.9–7	400	100
Kabylie-Algérois	Tell	Algeria	16.4–8.2	1000	500
Oranie	Tell	Algeria	11.7–0.8	2000	200
Azrou	Middle Atlas	Morocco	?35; 14.6–0.5	1500	175
Oulmès	Middle Atlas	Morocco	2.8; 0.9–0.31	600	70
Rekkame	Middle Atlas	Morocco	50; 32	100	12
Tamazert	Middle Atlas	Morocco	44–42	250	30
Taourirt	Middle Atlas	Morocco	57; 37	100	12
Sahro	Anti-Atlas	Morocco	Pliocene	100	12
Sirwa	Anti-Atlas	Morocco	11–9; 6.7	1000	120
Cap-Vert	Atlantic	Senegal	30; 21; 14.5–5.3; 1.5–0.57	100	12
Total				117,100 km ²	~14,000 km ³

area approximating 45,000 km² (Busweril and Wadsworth, 1980b). It was built up by two basaltic episodes, both post-Oligocene (Bellair et al., 1952). The first corresponds to the age of most of a huge plateau and the second to that of Pleistocene volcanic centers (Klitzsch, 1968). Available K-Ar ages range from 6 Ma to 0.4 Ma (Ade-Hall et al., 1974). Al Haruj lavas comprise homogeneous alkali basalt with CIPW (Cross-Iodlings-Pirsson-Washington)-normative nepheline (Busweril and Wadsworth, 1980b).

As Sawda extends to 6000 km², but probably initially covered an area of >10,000 km² (Busweril and Esson, 1991) located at the intersection of two major uplift structures (Fig. 1; Woller and Dediuk, 1980). The As Sawda volcanic rocks are exclusively basaltic, as is the case for the entire volcanic field linked to the Al Haruj uplift (Woller and Dediuk, 1980; Fig. 1). In addition to the flood basalt plateau it comprises, especially in the center of the province, a series of shield volcanoes that are aligned along two uplift directions (Woller and Dediuk, 1980). The main rock type is alkali olivine basalt with variations from olivine tholeiite to basanite (Woller and Dediuk, 1980). Stratigraphic evidence points to a post-Oligocene age (Klitzsch; 1970), which is in agreement with K-Ar ages in the 15.2–8.7 Ma range, providing evidence for increasing silica undersaturation

and a decreasing volume of lavas erupted with time (Busweril and Esson, 1991).

The Al Hasawinah volcanic field is tectonically related to Tertiary reactivation of the Tripoli-Tibesti and Al Qarqaf uplift lineations (Jurak, 1978; Busweril and Oun, 1991). It is composed of alkali olivine basalt, basanite, trachyte, and phonolite (Bordet et al., 1955; Jurak, 1978; Busweril and Oun, 1991) dated between 24.9 and 15.7 Ma (K-Ar; Jurak, 1978).

The Gharyan volcanic field shows more variation (Busweril and Wadsworth, 1980a). It comprises a large (3000 km²) but thin (a few tens of meters) old basaltic plateau (55–50 Ma), phonolite domes (ca. 40 Ma), and young volcanic centers (12–1 Ma). The plateau basalts are transitional between tholeiitic and alkaline. The phonolite domes are exclusively located to the north and northeast of the plateau (the Jabal Nefusah area; Bausch, 1978; Bausch and Meduna, 1991) and yield strongly silica-undersaturated and peralkaline agpaitic compositions (Busweril and Wadsworth, 1980a). The late volcanic centers are mainly basanitic but also comprise mugearite and hawaiiite. Lherzolitic xenoliths are locally abundant within the late volcanic centers. Their mineralogy is dominated by olivine with variable proportions of orthopyroxene, clinopyroxene, and spinel, and their chemistry points to a source depth of 45–60 km (Busweril and

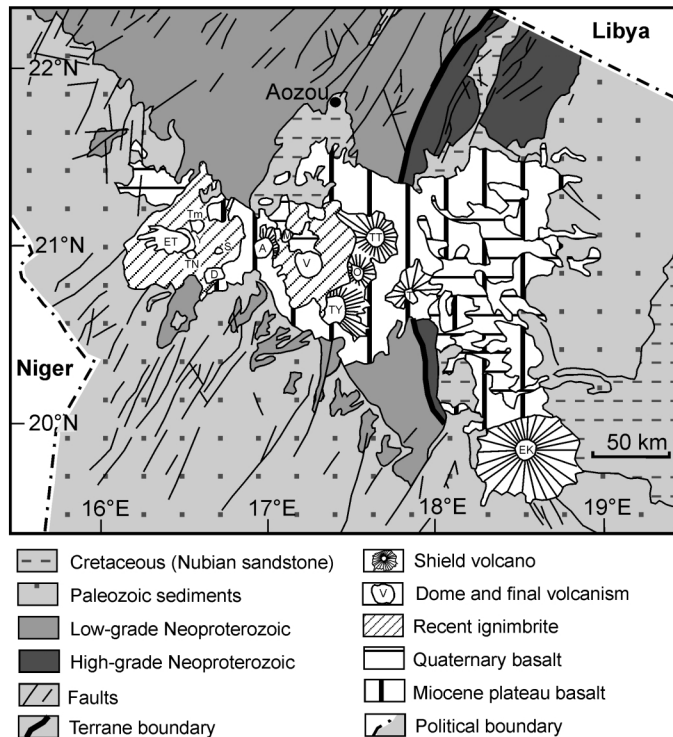


Figure 4. Geological map of the Tibesti area (redrawn from Vincent, 1970; terrane boundary from El Makhrouf, 1988).

Wadsworth, 1980a). Large age gaps between the three Gharyan groups seem to preclude genetic connections between them. These observations support a clear trend of volume decreasing with time (Busweril and Wadsworth, 1980a).

The Maghreb Province. From northern Tunisia to Morocco, the 1200 km-long linear Neogene to Quaternary magmatic belt of the Maghreb extends along the Atlas belt (Fig. 1). In addition to sparse Paleogene magmatic activity, the igneous centers of the Maghreb comprise two main types of post-collisional magmatism that differ in both their nature and their age of emplacement. High-K calc-alkaline to shoshonitic suites, associated with cordierite-bearing S-type granitoids (Fourcade et al., 2001), are mostly Miocene in age (16–8 Ma in eastern Algeria; 12–7 Ma in Oranie, Western Algeria, and Morocco; 11–7 Ma in northern Tunisia), while the alkaline series started ca. 8 Ma in Tunisia and Morocco and continued up to the Pleistocene in Morocco and Oranie (Coulon et al., 2002). The magmatic activity was not continuous, and volcanoes erupted in discrete episodes corresponding to abrupt changes in the kinematics of the European-African plate convergence (see, e.g., Rosenbaum et al., 2002a,b). The wide variety of rocks reflects the complex processes involved (Savelli, 2002).

Tunisia. In Tunisia, the Nefza region (Fig. 1) extends in an elongated area with a length of 50 km; comprises rhyolite, rhyodacite, and basalt plus pyroclastic rocks and granodioritic

bodies; and spans 10.9–7 Ma (Jallouli et al., 2003, and references therein).

Algeria. In Algeria, Cenozoic volcanism is mainly concentrated in the west in the Oranie region in the Tell thrust belt at the tip of the Middle Atlas mountain range (Fig. 1). It erupted between 12 and 7 Ma and between 4 and 0.4 Ma during extensional events following Alpine tectonic phases and involved the deposition of thick sedimentary series (Belanteur et al., 1995; Louni-Hacini et al., 1995; Coulon et al., 2002). Between 12 and 10 Ma, the volcanism produced high-K calc-alkaline to shoshonitic andesites between 10 and 7 Ma, then transitional basalts, and finally, after a quiescent period, alkaline basalts from 4 to 0.4 Ma. This magmatic evolution is interpreted as resulting from progressive evolution of a source in the lithospheric mantle previously enriched by subduction fluids to enriched asthenospheric mantle during a slab break-off period (Coulon et al., 2002). The Quaternary basalts brought mantle xenoliths to the surface (Zerka et al., 2002). Minor occurrences of metric rhyolitic tuff layers have been observed in the Grande Kabylie east of Algiers and date from Upper Aquitanian to Lower Burdigalian (20–18 Ma) (Rivière et al., 1977). Some Miocene volcanic layers (up to 100 m thick and 1300 m in length) of olivine trachyte and andesites have been described to the northeast of Constantine, at the Kef Hanouner (Raoult and Velde, 1971). Cenozoic granites are also present (e.g., Fourcade et al., 2001).

Morocco, Middle Atlas. Along the western margin of the Middle Atlas, three main volcanic fields are known: (1) the Taourirt field of Eocene dikes, sills, and breccia pipes, including camptonite (Mokhtari and Velde, 1988); (2) the large Azrou field formed by flows and volcanic centers of basalt but also of olivine nephelinite and ijolite (Harmand and Cantagrel, 1984, and references therein); and (3) the Plio-Quaternary Oulmès field (20 × 70 km in area), which comprises, in addition to subordinate basalt, strongly alkaline lavas (basanite, nephelinite, phonolite; Rachdi et al., 1985). The Rekkame volcanic field is located along the western margin of the Middle Atlas and comprises numerous small volcanoes made mainly of basalt, melanephelinite, and basanite, with three distinct ages (Rachdi et al., 1997): Eocene (ca. 40 Ma), Oligocene (ca. 30 Ma), and rarely Pleistocene (ca. 1.4 Ma). The first two are related to two major extensional episodes in the general regime of compressional collision between Africa and Europe (Rachdi et al., 1997).

Morocco, High Atlas. The High Atlas range comprises a smaller number of magmatic fields, but some are remarkable, such as the Tamazert complex, which comprises carbonatite (Agard, 1973). This is a 15 × 15 km Eocene elongated intrusion comprising micaceous pyroxenite, nepheline syenite, and magnesio-carbonatite (Mourtada et al., 1997), with dates of ca. 45 Ma and ca. 35 Ma. Isotopic data point to a metasomatized depleted mantle source (Bernard-Griffiths et al., 1991, and references therein). Numerous dikes including lamprophyres are associated with this complex (Bouabdli et al., 1988). Emplacement of the complex was related to the reactivation of Variscan transcurrent faults.

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The Sirwa (or Siroua) volcano is a major volcanic manifestation (25 km in diameter, elevation of 3304 m; Fig. 1). It is located on the South Atlas fault that marks the southern limit of the Alpine domain and correspondingly the northern boundary of the West African craton and *gedr* (Ennih and Liégeois, 2001). The Sirwa lavas (partly ignimbritic) and tuffs are mainly alkaline-peralkaline trachyte and phonolite with some mugearite and benmoreite and rare basic rocks (Berrahma and Hernandez, 1985). The radiometric ages available are in the 11–2 Ma range (Upper Miocene to Pliocene) and correlate with Alpine tectonic activity, both compressional and extensional (Berrahma and Delaloye, 1989, and references therein). Ti-Al-rich clinopyroxene megacrysts and gabbroic and pyroxenite xenoliths that are probably not cognate are found in ignimbrites; they crystallized at a pressure ^{3} 10 kbar and a temperature ^{3} 1160 °C (Bondi et al., 2002).

The Sahro (or Saghro) volcanic field, east of the Sirwa volcano, comprises a series of thin (max. 100 m) Pliocene phonolitic lavas and tuffs and a series of necks and diatremes (De Sitter et al., 1952). The lavas cover lacustrine Pliocene sediments, now at 1900–2000 m above sea level. The main event (production of olivine nephelinites) is dated between 9.6 and 7.5 Ma; a second event (production of pyroxene nephelinites rich in xenoliths, including carbonatitic ones) is dated at ca. 2.9 Ma (Ibhi et al., 2002).

The Atlas system can be regarded as a diffuse plate boundary in which the Atlas mountains are narrow deformable zones sandwiched between more rigid blocks (Gomez et al., 2000). These narrow zones are superimposed on early Mesozoic intra-continental rift structures (Piqué et al., 1998a) and have accommodated 17%–45% of total Africa-Eurasia plate convergence since the Early Miocene (Gomez et al., 2000). During the Mesozoic, Atlas sediments recorded sinistral extensional transcurrent movements in subsiding elongated areas (rifting) and, in post-Eocene times, north-south compressional movements contemporaneous with considerable uplift (Piqué et al., 1998a). Carbonatites, camptonite, and strongly undersaturated lavas are old (Table 1) and related to extension. Phases younger than Miocene related to the compressional event display evolution from calc-alkaline to alkaline compositions. The transition occurred mainly at the end of the Miocene (Messinian). The direction of maximum compressional stress changed from NE-SW to N-S, the latter reactivating, at the edges of the Maghreb indenter; there were also NW-SE faults in Tunisia with a sinistral sense of movement and NE-SW faults in Morocco with a dextral sense of movement, triggering the alkaline volcanic events (Piqué et al., 1998b). This evolution could be linked to slab break-off during the Miocene (Maury et al., 2000; Coulon et al., 2002). The fault zones are still active, as indicated by persistent seismicity.

The Canary Islands. The Canary Islands are located at the transition between African continental and oceanic crust. They were built up by abundant, prolonged (longer than 35 m.y.), and structurally controlled volcanic emissions (Araña and Ortiz, 1991; Carracedo, 1994). Aeromagnetic anomalies, for example,

show the influence of regional tectonics on the growth of Gran Canaria during its entire evolution (Blanco-Montenegro et al., 2003). It is tempting to relate this volcanism to High Atlas structure (Anguita and Hernán, 1975), considering the alignment of the Canary Islands with the Sirwa and Sahro Cenozoic volcanoes and even Middle Atlas volcanic units (Fig. 1). All these magmatic provinces share many characteristics (Anguita and Hernán, 2000), including the existence of carbonatites (dating to 23 Ma in Fuerteventura Island; Cantagrel et al., 1993). Such a link is likely, but discussion of this is beyond the scope of this paper. We refer the interested reader to the review paper of Anguita and Hernán (2000) and references therein.

The Cap-Vert Peninsula. Sporadic volcanic activity occurred along the Atlantic coast of West Africa. The largest zone is the Cap-Vert peninsula, where Dakar, the capital of Sénégal, is located (Fig. 1). The Sénégal basin is a typical passive margin, and the tectonic setting of the Cap-Vert peninsula corresponds to the transition from continental to oceanic crust, where north-south trending faults are regarded as old listric faults developed during the Mid-Jurassic rifting of the Atlantic Ocean. The alkaline volcanic activity began during the Oligocene, had its climax during the Miocene (three main episodes at 21 Ma, 15–10 Ma, and 8.5–5.3 Ma), and, after a renewal at 1.5 Ma, ended at ca. 600 Ka (Crévoila et al., 1994). The simultaneous igneous episodes of the Cap-Vert peninsula, the Cabo Verde archipelago, and the Canary Islands are ascribed to the reactivation of old fault zones related to the motion of the Africa plate, especially along its Mediterranean edge.

The Precambrian Basement of the Tuareg Shield

The Hoggar volcanic magmas penetrated the fractured Tuareg shield comprising the massifs of Hoggar in Algeria, Adrar des Iforas in Mali, and Aïr in Niger. The shield is principally made of Archean or Paleoproterozoic and Neoproterozoic terranes that amalgamated during the Late Neoproterozoic Pan-African orogeny as a result of convergence between the West African craton and the Saharan metacraton (Fig. 3; Black et al., 1994). For a description of the metacraton concept, see Abdelsalam et al. (2002) and Liégeois et al. (2003).

The first stage corresponded to accretion of oceanic island arcs on the (meta)cratons during the period 900–680 Ma. Relics of these terranes, including ophiolites and eclogites, are preserved as thrust sheets on top of more rigid bodies, such as the LATEA (Laouni–Azrou N’Fad–Tefedest–Egéré–Aleksod) microcontinent forming the Central Hoggar (Liégeois et al., 2003, and references therein) or eastern Aïr (Liégeois et al., 1994).

The second stage was marked by the regional northerly tectonic escape of the Tuareg terranes due to oblique collision of the Saharan metacraton with the West African craton. It is characterised by spectacular north-south shear zones (Fig. 3). During that stage, metacratonization of the LATEA microcontinent (Central Hoggar) occurred; i.e., this rigid body was squeezed, which caused it to be torn into several moving blocks, but kept

most of its rigid behavior (Liégeois et al., 2003). This induced transcurrent movements along north-south mega-shear zones and intrusion of granitoid batholiths (615–580 Ma) with geochemical signatures pointing to the lower crust (initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.710$; $\epsilon_{\text{Nd}} = -20$). Linear lithospheric delamination beneath the mega-shear zones could have occurred, allowing a drastic increase in heatflow and inducing crustal melting (Liégeois et al., 2003, and references therein). Postcollisional and anorogenic high-level alkaline plutons are aligned on the same mega-shear zones, particularly along (meta)craton margins (Fig. 3; Liégeois et al., 1998; Azzouni-Sekkal et al., 2003).

Hoggar during the Paleozoic and the 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 sedimentary flux (Beuf et al., 1971) during a long period lasting up until the Cretaceous.

Paleo-currents in Cretaceous sediments around the shield indicate that Hoggar was bypassed by the sedimentary flux from the north (Faure, 1985). During the Late Cenomanian-Turonian (95–90 Ma), a Trans-Saharan seaway separated the Hoggar and the Tibesti massifs linking the Tethys or Mediterranean sea and the (Proto)Atlantic Ocean (Kogbe, 1980). Another passed to the west of Hoggar during the Maastrichtian (Kogbe, 1980) and the Paleocene (Reyment and Reyment, 1980). This later transgression also occurred to the east but probably with a gap between latitudes 20°S and 25°S. Hoggar, as well as the Eglab and Tibesti massifs, was an island during the Cretaceous, covered only by continental lakes (Fabre, 1976). Only rare remnants of Cretaceous continental deposits rest directly on top of the Hoggar Precambrian basement. This indicates that the shield was already uplifted and slightly reeroded before the Cretaceous and therefore that the Hoggar swell already existed during the Cretaceous, along with other swells such as the Tibesti swell. Cretaceous uplifts were linked to the development of a series of Cretaceous elongate troughs in West Africa, which were well developed on both sides of Hoggar (Fig. 3; Dautria and Lesquer, 1989; Genik, 1993). The main vertical movements occurred during the Cenozoic, mostly after 30 Ma, though they began several tens of millions years earlier. The current occurrence of Cretaceous deposits at an altitude of 2000–3000 m above sea level (Rognon, 1967) emphasizes the dramatic nature of later Cenozoic uplift.

Cenozoic Hoggar Volcanism and Plutonism: 35 Ma to the Present

The Hoggar swell reaches an altitude of 350–400 m a.s.l. on its margin and of 1000–1500 a.s.l. in its center (the town of Tamanrasset is set at an altitude of 1420 m). Some Precambrian basement inliers have been observed in Atakor at an altitude of up to 2600 m a.s.l. Cenozoic to recent igneous activity occurred in several districts (Figs. 1 and 3): Tahalra (1800 km²), Atakor

(2150 km²), Manzaz (1500 km²), Egéré (2800 km²), Anahef (400 km²), and Adrar N'Ajjer (2500 km²), all within Central Hoggar, and also in Teria (100 km²) to the northeast, Djanet (50 km²) and In Ezzane (800 km²) to the east, Todra (1050 km²) and Tin Taralle (50 km²) in the south of Aïr, and Tin Zaouatene (50 km²) in the east of Adrar des Iforas. The major volcanic districts form high relief, often over 2000 m in altitude, culminating in Atakor (the 2918 m-high Mount Tahat). Igneous activity began in the Late Eocene or the Early Oligocene (35–30 Ma; Aït-Hamou et al., 2000) and lasted until the Neolithic (Lelubre, 1952) and later. We found historical pottery within basaltic lavas in the Tahifet area, and the Tuareg people relate ancestral stories suggesting that volcanoes were seen erupting. No equivalents of the volcanic formations emitted in the Maghreb area during the early extensional phase at 40–50 Ma have been found so far.

The volcanic rocks constitute a thin cover. The mean thickness is 100 m, with some exceptions such as the Anahef plateau basalt, which reaches a thickness of 700 m (Aït Hamou et al., 2000). The estimated volume of Atakor volcanism is ~250 km³ (Girod, 1971; Table 1), and the total volume of Tuareg Cenozoic volcanism is ~1650 km³. It could be as much as 2500 km³ if the estimate of 1000 km³ for the Anahef basalt before erosion that was made by Aït Hamou et al. (2000) is accepted.

Three main igneous episodes are identified in Hoggar:

1. The oldest Upper Eocene to Oligocene episode is exposed only in the Anahef district. It corresponds to fissural outpouring of tholeiitic flood basalt with a thickness up to 700 m (35–30 Ma; Aït-Hamou et al., 2000), intruded by a dozen subvolcanic ring complexes (ca. 29 Ma for the Achkal complex; Maza et al., 1995) comprising gabbro, diorite, monzonite, alkali feldspar syenite, and nepheline syenite. The largest is the 8 km × 5 km Tellertebe ring complex (35 Ma, Rossi et al., 1979). The ring complexes are covered by ca. 24 Ma alkaline rhyolites, indicating uplift rates of ~0.4 mm/yr (Maza et al., 1995). Exhumation of the deeper levels currently exposed in the Anahef district resulted from coeval uplift of Hoggar.
2. The second episode (Miocene, 20–12 Ma, and Mio-Pliocene, 8.5 and 4 Ma; Rognon et al., 1983) is the most voluminous at the present surface. It is well exposed in the Atakor district (Girod, 1971; Benhallou, 2000; Yahiaoui, 2003), where it is composed of 80% basalts and 18% trachyte and phonolite, the remaining comprising rhyolites. With its famous scenery, the Atakor massif is one of the largest volcanic districts of the province. The distribution of the various rock types is not random. Mafic rocks, abundant in the center of the massif, are scarce to the south, where only a few scarps are still exposed. The majority of phonolite domes and necks occurs in the Assekrem area (Fig. 5A), whereas trachyte (Fig. 5B) can be seen everywhere, with a marked enrichment in quartz to the south and southeast (the Tahifet area), where rhyolite is also exposed. From the field and petrological evidence, two magmatic groups are suggested: (1) a

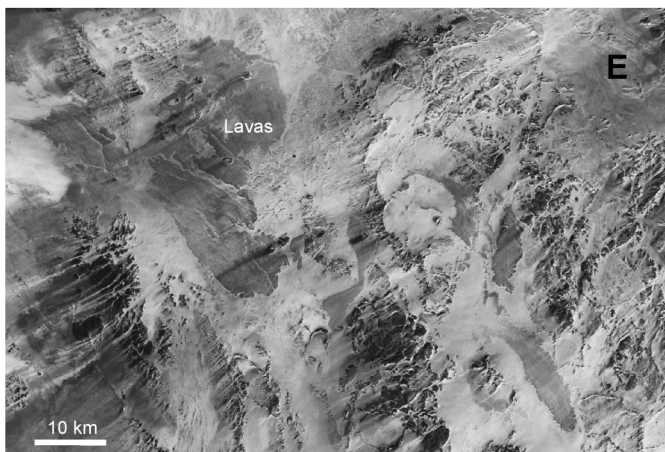
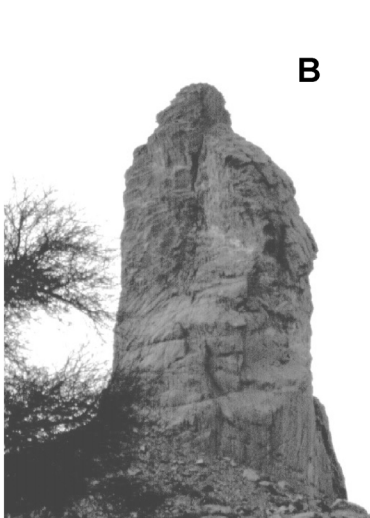
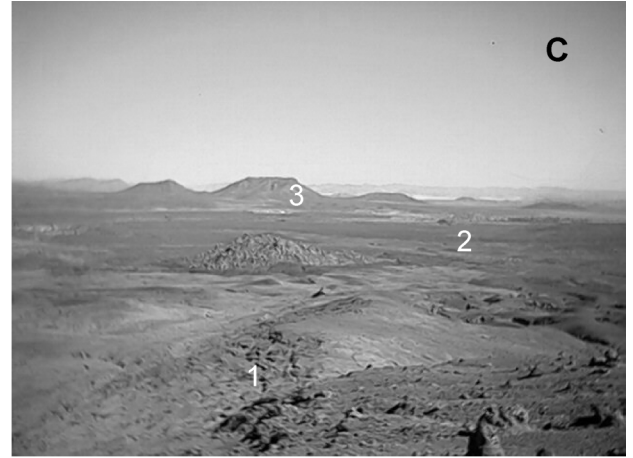
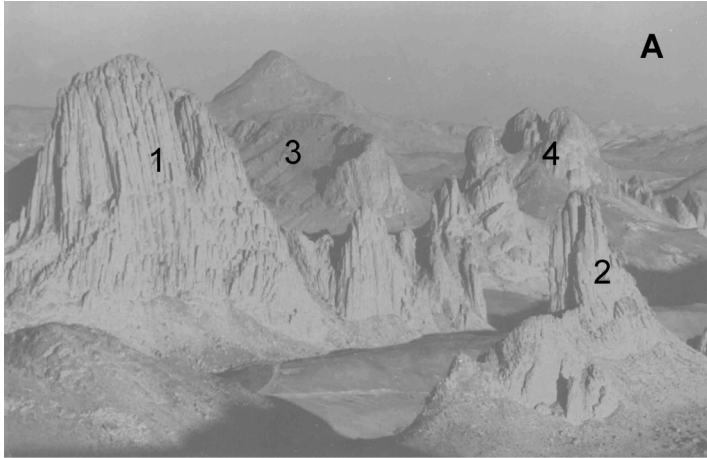


Figure 5. (A) View from the Assekrem plateau: (1) trachytic neck of Tezouièdje; (2) needle dome of Séouinane; (3) crater; (4) alkaline phonolitic domes of Tidjmayine. (B) Aegirine trachytic neck of Imros (region of Tahifet). (C) The Manzaz district: (1) volcanic ashes covering the granitic basement; (2) granitic basement inlier; (3) volcanoes with open craters. (D) Quaternary basanitic prismatic flow; oued = wadi Ségueika (Assekrem). (E) Satellite photograph of the In Ezzane district; Orthorectified Landsat Thematic Mapper Mosaics as compressed color imagery in MrSID™ file format from Lizardtech. The lavas appear dark gray. The circular structures have not yet been studied. (F) Example of a small Aïr basaltic volcano that generated a single flow. Except for E, photos taken by the authors.

mafic group composed of a suite of olivine basalt, hawaiite, and mugearite, the latter bearing kaersutite megacrysts, and (2) a felsic group comprising agpaitic phonolite, with feldspar and feldspathoid phenocrysts engulfed by aegirine-augite associated with aenigmatite, and a large variety of silicic rocks ranging from biotite trachyte to aegirine-augite—aegirine trachyte to aenigmatite arfvedsonite trachyte to quartz trachyte to rhyolite.

In the Manzaz district, the Miocene episode emitted very widespread vesicular basaltic flows covering or bypassing the Pan-African granitic basement. Late Miocene ages (8.4 and 6.8 Ma) have been obtained in the Djanet area (Morten et al., 1980).

3. The latest episode is Late Pliocene to late Quaternary (3 Ma to historic times). In the Atakor district, the volcanic suite is almost entirely represented by basanite. Basanite, often vesicular, is characterized by an abundance of analcite. Scarce nephelinite lavas flowed along the valleys (Fig. 5D) and can cover upper Paleolithic terraces. In the Tahalra district, a Pliocene to Holocene event built a series of small strombolian stratovolcanoes that generated basanite, hawaiite, and nephelinite lava flows (Dautria, 1988). In this district, the proportion of lavas is 80% basanite, 15% nephelinite, 2% hawaiite-mugearite, 1% trachyte, and 2% rhyolite. In the lesser-known Egéré and Adrar N'Ajjer districts, most of the volcanic activity occurred between 3.5 and 2.5 Ma (Aït-Hamou et al., 2000). The Manzaz district includes strombolian volcanic cones that emitted some lava flows. The cones, now greatly eroded, often have horseshoe shapes (Fig. 5C) and contain numerous peridotite and pyroxenite xenoliths. Abundant pyroclastites are crowded with megacrysts of amphibole, pyroxene, and feldspar, as in the case of the Oukcem double maar (Fig. 5C). The district contains 95% alkali basalt, basanite, and hawaiite lava flows. More recent volcanoes rest on Neolithic terraces.

No Cenozoic radial dike swarm is known in Hoggar or anywhere in northwestern Africa.

The In Ezzane district is almost unknown due to its remote location (Fig. 5E). The In Teria district comprises xenolith-rich melilitic rocks (Bossière and Megartsi, 1982; Dautria et al., 1992). In the Djanet area, a few basaltic flows occur. These contain unusual high-pressure megacrysts mainly composed of Tschermak's molecule-rich clinopyroxene and olivine as well as aluminous orthopyroxene and spinel (Morten et al., 1980).

Farther south, in Aïr (Niger), the Todra and Tin Taralle districts (Black et al., 1967) began with ~30 trachytic and phonolitic volcanoes, many having erupted very viscous lava flows and trachytic tuffs. The Todra volcano itself lies at an altitude of 1780 m. The eruption of this first volcanic group was followed by the formation of ~130 basaltic volcanoes, generally in the form of regular cones, each of which generated only one lava flow following valleys (Fig. 5F). Black et al. (1967) estimated that some volcanoes cannot be more than a few centuries old. Several hot

mineral sources are known (e.g., the Tafadek source at a temperature of 50 °C; Black et al., 1967).

EVIDENCE AGAINST A MANTLE PLUME MODEL

Structural Control of Volcanism

Black et al. (1967) remarked that the Aïr volcanoes are located either on faults oriented NW-SE or on annular faults linked to Devonian ring complexes (Fig. 6). The NW-SE faults belong to the conjugate brittle fault system (Fig. 7) that developed at the end of the Pan-African orogeny due to West African craton indentation (Ball, 1980). Quartz filling these faults in Aïr is older than Ordovician sandstones. Both directions of the conjugate system are present everywhere in the shield, but in a given area one of them is often predominant (Ball, 1980). The system was reactivated several times during the Phanerozoic. The main reactivation events occurred during (1) the Devonian, as seen in the emplacement of the Aïr ring complexes (Fig. 3, 6; Moreau et al., 1994); (2) the Cretaceous, as seen in the parallelism between the fault system and the Cretaceous troughs, which suggests that the latter were controlled by the former (Fig. 7; Ball, 1980; Dautria and Lesquer, 1989); and (3) the present, as seen in the current uplift, which is well exemplified by Aïr with its staircase relief molded on these faults (Vogt and Black, 1963).

In Atakor, the orientation of the dikes associated with the volcanic necks is that of the conjugate fault system (Girod, 1971). Special status is given to the NE-SW faults located to the southwest of the Central Hoggar volcanic field, which constitute the so-called Amded lineament (Dautria and Lesquer, 1989; Aït-Hamou and Dautria, 1994), mostly because it is parallel to the alignment created by the Tahalra, Atakor, Anahef, and N'Ajjer volcanic fields. The "Amded lineament" might even be related to the Libyan Al Qarqaf uplift (Fig. 1). The Hoggar, Tibesti, and Libyan volcanic fields are similarly controlled, because they are located at the intersection of major tectonic lineaments: the north-south or NW-SE being Pan-African and Paleozoic in age and the NE-SW Mesozoic in age. The latter are most probably linked to the early Mesozoic intracontinental rift structures in the Atlas region (Piqué et al., 1998a) described earlier, whose current mountain ranges feature the same orientations (Fig. 1).

Last, it should be noted that all the West African Cenozoic volcanic provinces formed in the same period (from ca. 40 Ma to the present; Table 1), i.e., since the initiation of the Europe-Africa collision. Also, the alignments of volcanic fields (Fig. 1) do not delineate any spatial age progression. Magmatism occurs repeatedly in a single place, as observed for the whole Phanerozoic (Bailey and Woolley, this volume).

Geophysical Data

On the basis of a free air gravity low, Crough (1981) proposed that Central Hoggar is underlain by anomalously light mantle involving a magmatic body at a depth of <60 km below

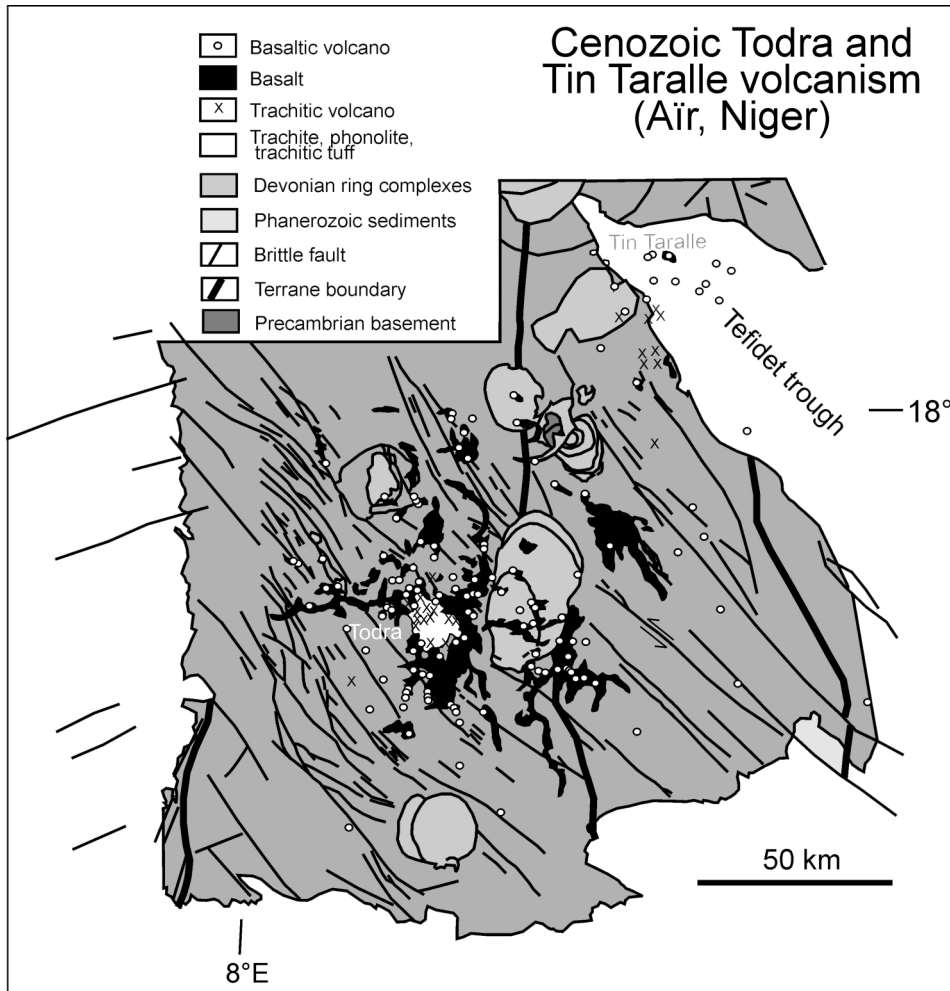


Figure 6. Map of southern Aïr (Todra and Tin Taralle volcanic fields) showing volcanoes and lava flows and their links with terrane boundaries, late Pan-African brittle faults, and Devonian ring faults (from Black et al., 1967, 1985; Liégeois et al., 1994).

the surface. Lesquer et al. (1988) analyzed the shape of the anomaly (Fig. 3), which appears to be correlated with basement elevation, and suggested that the body lies between 20 and 70 km depths and has a maximum thickness of 30 km. The association between a gravity anomaly and a Cenozoic igneous province led to the suggestion that a mantle plume underlies Hoggar (Aït-Hamou and Dautria, 1994; Aït-Hamou et al., 2000).

The heatflow measured inside the Hoggar swell is normal (50 mW/m^2), and there is no evidence for thermally disturbed lithosphere (Lesquer et al., 1989). On the contrary, heatflow is less at higher basement elevations, which is inconsistent with thermal uplift. Only a minor thermal anomaly can be envisaged below the center of Atakor, where heatflow of 63 mW/m^2 has been measured (Lesquer et al., 1989), suggesting that small high-temperature mantle bodies may intrude the crust beneath the recent volcanic areas (Ayadi et al., 2000). In contrast, a large thermal anomaly occurs to the north, below the Saharan basins (Lesquer et al., 1990; Lesquer and Vasseur, 1992). This led Lesquer et al. (1988) to suggest that the hot body was emplaced before 60 Ma and has cooled down completely. However, two questions arise: (1) Given the anomaly, the formerly hot shallow

body should have been very large. What kind of magmatic body could it be? (2) If this body has completely cooled, why does a low-density signature remain?

On the other hand, if a thermal anomaly exists deeper in the mantle, it will not dramatically increase surface heatflow if only conduction is modeled. For example, a body 250°C hotter than the surrounding mantle, at a depth of 120 km below the lithosphere and 100 km thick, would, after 30 m.y., increase the heatflow at the surface by only 0.2 mW/m^2 or, after 100 m.y., by 2 mW/m^2 (Carol Stein, 2004, written comm.). Thus, the absence of a heatflow anomaly at the surface cannot be cited as proof of the absence of a thermal anomaly in the mantle. In turn, this means that a thermal anomaly in the mantle cannot express itself, by conduction only, in the shallow lithosphere. The absence of a high-heatflow anomaly suggests a primarily dynamic or a compositional rather than thermal origin for the surface swell (De Laughter et al., this volume). This also suggests that the melting zones at the sources of the volcanism are of limited extent, because no extra heatflow due to convection (magma heat transfer) is detected. Considering both seismic and heatflow data, Lesquer and Vasseur (1992) proposed that the mantle beneath

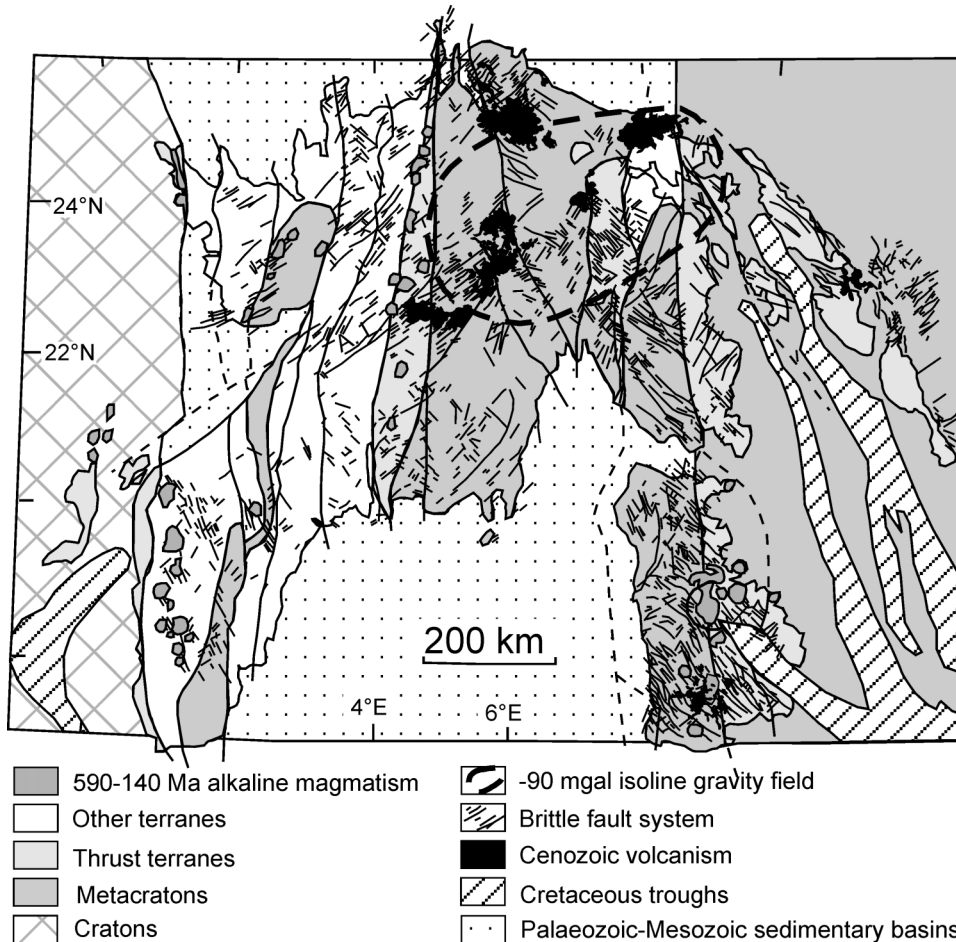


Figure 7. The brittle fault system in the Tuareg shield. Late Pan-African in age, it was reactivated several times during the Phanerozoic, including during the current swell uplift. Brittle fault system from Ball (1980), gravity anomaly from Lesquer et al. (1988), Cretaceous troughs from Genik (1993), terrane structure from Black et al. (1994) and Liégeois et al. (2003). A. L.—Ameded lineament.

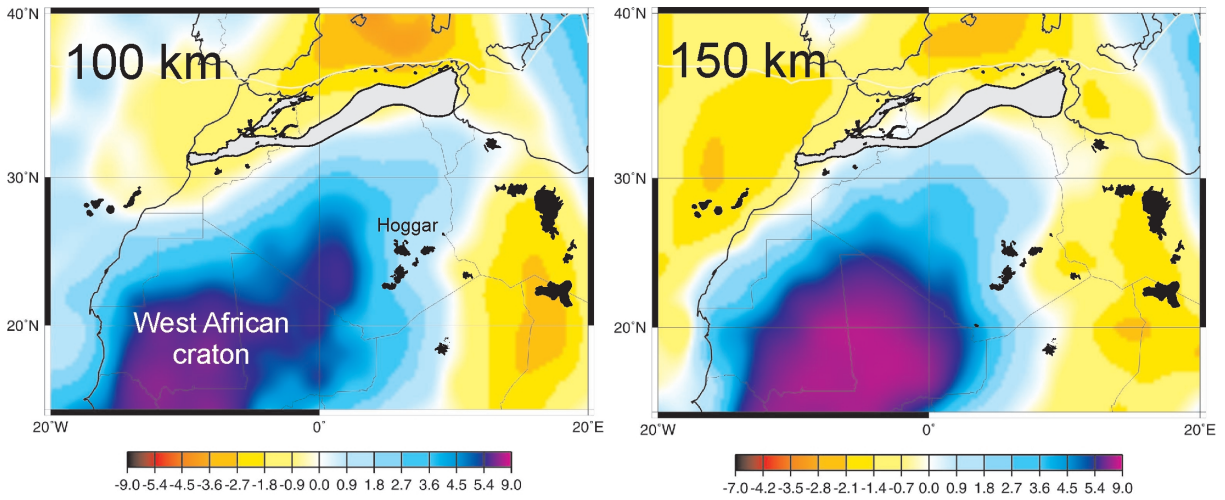
Central Hoggar is intermediate between cratonic mantle and activated mantle. All the preceding arguments call for additional constraints, and we now consider mantle tomography.

West African maps showing lithospheric structure were generated using the 3D shear wave velocity tomography model of N. Shapiro, which is based on surface wave diffraction data (<http://ciei.colorado.edu/~nshapiro/model/>). The data processing techniques are described by Ritzwoller et al. (2002) and Shapiro and Ritzwoller (2002).

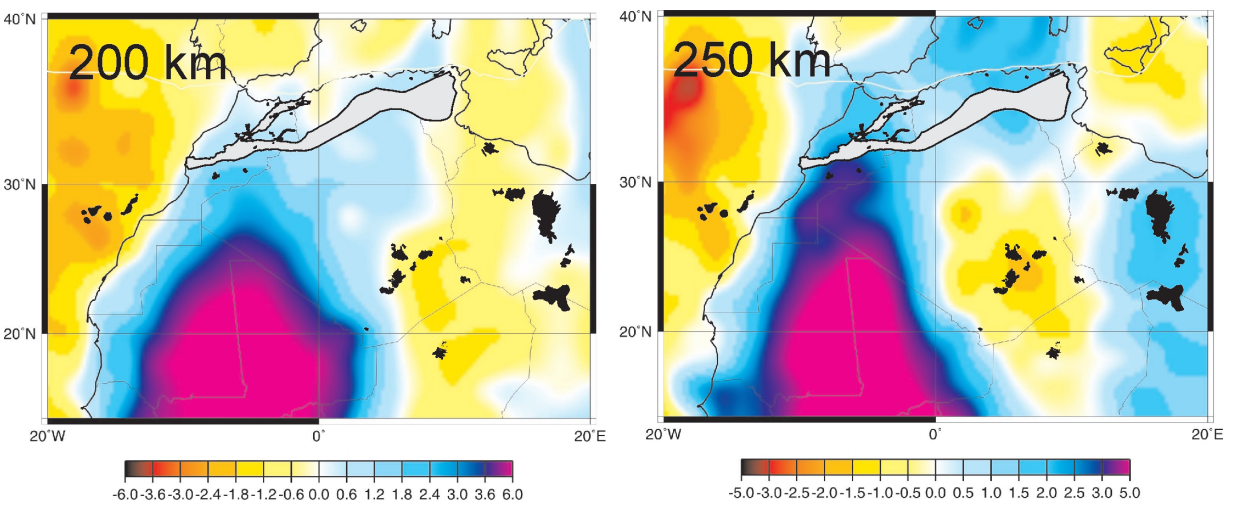
The Cenozoic volcanic fields and the Atlas Mountains are shown on the maps (Fig. 8A). The thick lithosphere of the West African craton is clear at all depths, particularly in the south, where very high shear velocities are still recognizable at even 250 km depth. The Atlas Mountains broadly correlated in the north with the zone of thick lithosphere, and this is seen particularly well at 150 km depth. This is consistent with the thinking that the Alpine orogeny had nearly no effect on Africa to the south of the Atlas Mountains and with the fact that this mountain range underlines the northern boundary of the West African craton (Ennih and Liégeois, 2001). It is noteworthy that there is no Cenozoic volcanism in the cratonic part of Africa (Black and Girod, 1970).

The northwestern African volcanic fields correlate with lithospheric structure. Three groups can be distinguished: (1) The Maghreb provinces, like the Atlas Mountains, follow the northern boundary of the cratonic lithosphere, especially to the west, where there is evidence for transcurrent movements (Piqué et al., 1998b). If low velocities in the tomographic images correspond

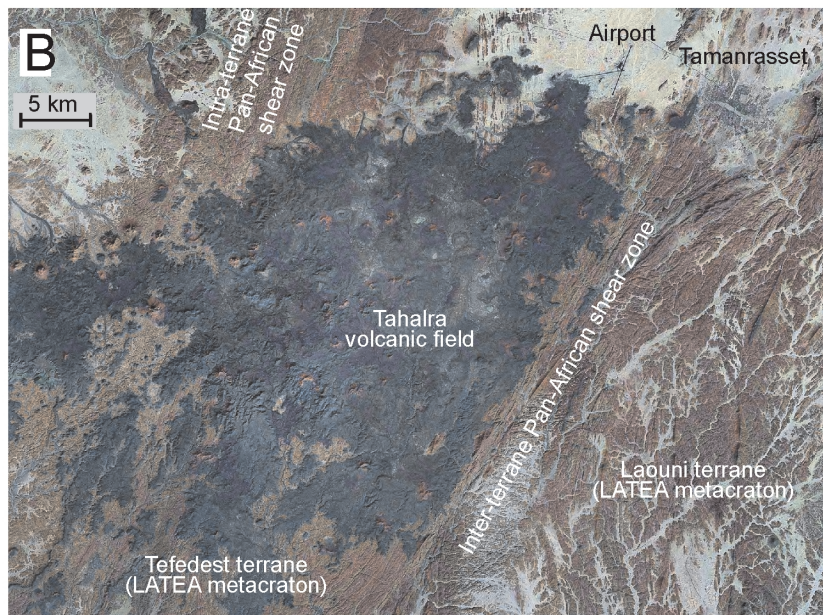
Figure 8. (A) West African maps showing the 3D shear wave velocity tomographic model based on surface wave diffraction tomography map generated from the model of N. Shapiro (<http://ciei.colorado.edu/~nshapiro/model/>). Surface wave diffraction tomography and data processing are described in Ritzwoller et al. (2002) and Shapiro and Ritzwoller (2002). Color scale shows the shear velocity as a percentage of perturbation relative to the reference velocity values, which are 4.483 km/s, 4.473 km/s, 4.494 km/s, and 4.562 km/s for depths of 100 km, 150 km, 200 km, and 250 km, respectively. (B) Satellite photograph of the Tahalra district southwest of Tamanrasset; Orthorectified Landsat Thematic Mapper Mosaics as compressed color imagery in MrSID™ file format from Lizardtech. The lavas appear dark blue and volcanic cones red due to several factors, including differential $\text{Fe}^{3+}/\text{Fe}^{2+}$ mobility. The linear eastern boundary of the Tahalra lava field follows the Pan-African shear zone separating the Laouini and Tefedest terranes, both belonging to the LATEA metacraton. This indicates that the latter was reactivated during the volcanic activity.



A: 3D shear velocity tomographic model



■ Cenozoic volcanic fields ■ Atlas mountain belt — Boundaries between states



to higher temperature, there is a link with a warmer mantle down to a depth of 150 km. (2) The Tibesti and Libyan volcanic fields are also located over a warmer mantle area at shallow depth (100 and 150 km), particularly the Tibesti, whereas at greater depth (250 km) the mantle is colder (as suggested by higher velocities) than the mean. (3) The Hoggar field is located on the eastern shoulder of the West African cratonic mantle, which extends down to 200 km, and a low-velocity seismic mantle exists at a depth of 250 km. For cross-sections and maps of the seismic structure of the entire mantle passing through Hoggar, see Ritsema (this volume). When we look at northwestern Africa as a whole, it is apparent that there is a correlation between low seismic velocities and Cenozoic volcanic fields down to a depth of 150 km but not deeper. This is in agreement with the melting depth of the source of such magmas, and would imply that deeper mantle velocity variations (e.g., those suggested at 250 km below Hoggar) have no direct influence on the volcanism.

The tomographic results also bear on the gravity anomaly in the Hoggar area (Fig. 3). The tomographic model indicates that there is relatively thick lithosphere below the Hoggar volcanic field. The residual gravity anomaly must thus be sourced beneath this. We suggest that the origin of the Hoggar gravity anomaly is the deep (>250 km) “thick low velocity (probably warmer) mantle region.” In that case, the small elastic thickness of Hoggar assumed by Lesquer et al. (1988) to imply a depth of 50 km for the top of the anomaly would be invalid. This proposition should be tested.

The location of the Hoggar volcanic field on the shoulder of the cratonic lithosphere, with colder mantle extending to a depth of 150 km, is unlikely to be a coincidence, and the correlation with the lithospheric structure inferred from geology (Fig. 3) should be considered. The latter suggests that there is rather thick lithospheric mantle under the metacratons and thinner lithosphere in between, not progressive thinning toward the east as shown by tomography. This apparent contradiction can be explained by considering the resolution of the tomographic model, especially in northwestern Africa, which is poorly instrumented by seismic stations. In this region, model resolution is limited and ~100 km-wide structures cannot be detected (N. Shapiro, 2004, written comm.). The tomographic model is thus very smooth. A schematic cross-section is presented in Figure 9 that is based on both the tomographic model and the geological constraints (Liégeois et al., 2003, and references therein). It illustrates how the thicker lithosphere of the metacratonic areas might be related to the progressive thinning of the lithosphere suggested by the tomographic model. A higher-resolution tomographic model is needed. There remains, however, a discrepancy between the tomographic and the geological models: the inferred Saharan metacraton, to the east of the Turaeg shield, is marked by low-velocity (warm?) shallow mantle, although at a depth of >150 km it is higher-velocity than the mean. This discrepancy cannot be solved at present due to poor geological knowledge of this area (Abdelsalam et al., 2002), including the easternmost Hoggar terranes, which are almost unknown. A global assessment of the lithospheric evolution of this

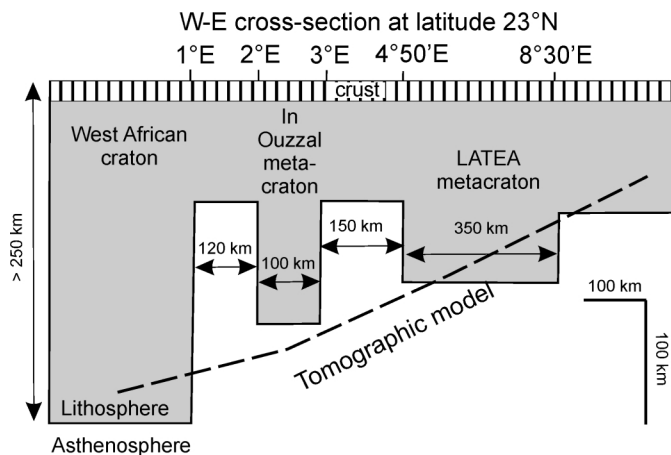


Figure 9. Model of lithospheric structure based on geological arguments and tomography that agrees with both kinds of data within the resolution limits.

area during the Phanerozoic is required, but this is beyond the scope of this paper. Here we stress that, even if the shallow mantle below the Saharan metacraton is low-velocity (warm?), lithospheric structures guide the volcanism in Tibesti, Libya, as in Hoggar, where reactivation of Pan-African interterrane shear zones during Cenozoic volcanic activity is demonstrated, for example, by the sharp linear shape of the Tahalra lava field along the shear zone (Fig. 8B).

The Alkali Basalts and Their Xenoliths

The Lavas. The volcanic suite of the Hoggar massif forms a typical alkaline series including silica-undersaturated, -saturated, and -oversaturated types. Rare earth element (REE) patterns of the mafic group are similar to patterns of ocean island basalts (OIB). Phonolite and trachyte patterns are typical of evolved alkaline and peralkaline rocks, with the amounts of light REE (LREE) up to 1000 times those of chondritic, amounts of heavy REE (HREE) ~30 times those of chondritic, and a strong negative Eu anomaly. Variations in the $(La/Yb)_N$ ratio show magmas evolving by stepwise fractional crystallization to phonolite, biotite trachyte, aegirine-augite to aegirine trachyte, aenigmatite arfvedsonite trachyte, quartz trachyte, and rhyolite (Dautria et al., 1988; Azzouni-Sekkal et al., 2004).

The Xenoliths. In addition to xenoliths coming from the Hoggar crustal basement (Leyreloup et al., 1982; Kornprobst et al., 1987; Pineau et al., 1987), Girod et al. (1981) and Dautria et al. (1987) have established that amphibole-rich xenoliths within Hoggar alkali basalts (Tahalra) are likely to have originated in a mantle metasomatic event that occurred during or just before the generation of Hoggar magmas (max. 40 Ma considering the isotopic signature and parent/daughter ratios) in the temperature range 1000 °C to 1100 °C under pressures of 1.5–1.8 GPa. In the Djanet area, higher values have been obtained for olivine, orthopyroxene, and clinopyroxene xenocrysts: a

minimum of 1150 ± 30 °C has been estimated using geothermometers, but mineral compositions suggest a crystallization temperature of 1400 °C at a pressure of 2.25 GPa (Morten et al., 1980). In Tahalra, both xenoliths and lavas yield low- $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios—between 0.70306 and 0.70344. Based on the xenoliths, the source is enriched by a factor of 7–9 times compared with chondritic LREE and by a factor of 2 in HREE. The degree of partial melting has varied from 1.2% to 2.1% for nephelinite magmas and 3.8% to 4.4% for basanite magmas, both leaving a garnet-rich residue. The positive correlation between the degree of deformation of the xenoliths and the abundance of the “metasomatic” minerals (principally amphibole) indicates that 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 and that its overall density there is fairly low. The Cenozoic increase of the uplift of the Hoggar basement could then be seen as the isostatic response to upper-mantle density reduction controlled by magmatic events and associated metasomatism (Dautria et al., 1987).

The Isotopic Data

The Pb, Sr, and Nd isotopic compositions of Hoggar volcanic rocks lie between the EM1 (enriched mantle) and HIMU (high m, i.e., high U/Pb mantle) end-members (Allègre et al., 1981; Dupuy et al., 1993; Aït-Hamou et al., 2000), the influence of the latter apparently increasing with time. The main alkaline basalts have $^{87}\text{Sr}/^{86}\text{Sr}$ varying from 0.7030 to 0.7037, ϵ_{Nd} from +3 to +7, $^{206}\text{Pb}/^{204}\text{Pb}$ from 19.3 to 20.5, and $^{207}\text{Pb}/^{204}\text{Pb}$ from 15.59 to 15.70. The older (ca. 35 Ma) Taharaq tholeiitic basalts have $^{87}\text{Sr}/^{86}\text{Sr}$ varying from 0.7035 to 0.7045, ϵ_{Nd} from –2.5 to +3, $^{206}\text{Pb}/^{204}\text{Pb}$ from 18.3 to 19.5, and $^{207}\text{Pb}/^{204}\text{Pb}$ from 15.54 to 15.60. Aït-Hamou et al. (2000) interpreted the isotopic dataset as indicating the presence of two end-members, namely the Pan-African lithosphere, corresponding to the EM1 end-member, and a plume, with some depleted mantle materials and a strong HIMU component, characterized by recycled subducted Proterozoic oceanic lithosphere.

The Geodynamics of the African Plate during the Late Mesozoic and the Cenozoic

During the Cenozoic, Africa approached Europe, but the rate of this movement is still debated. Some authors believe that the plate was nearly stationary (Burke and Wilson, 1972; Burke, 1996), while others favor movement of 20 ± 1 mm/yr for the last 19 m.y. and a similar speed or higher between 19 and 30 Ma (O'Connor et al., 1999). Using data from the Geomar-ODSN (Ocean Drilling System Network) reconstruction facility (<http://www.odsn.de/>), at 35 Ma Hoggar is estimated to have lain at 3°W and 18°N, i.e., 1000 km farther southwest. This implies that at 35 Ma the Hoggar was located near the current position of Timbuktu (Fig. 1). This indicates a mean speed of 28.5 mm/yr, which is in agreement with the estimate of O'Connor et al.

(1999). We do not favor the assumptions of Burke and Wilson (1972), which were challenged by Kröner (1973), who used reasoning similar to that presented here.

As described in the earlier section on the Cenozoic volcanism in northwestern Africa, a major rifting event affected the African plate during the Cretaceous, which led to opening of the equatorial and southern Atlantic Ocean and the generation of intraplate rifts, including the Atlas and Benué rifts. Africa-Europe collision began ca. 38 Ma (Hoernle et al., 1995). This and other peak periods of compression along the northern plate margin of Africa at 20, 8, and 1 Ma induced tectonic events within the continental plate (Guiraud and Bosworth, 1997, and references therein).

In the Hoggar area, Cretaceous rifts developed to the southeast within the Saharan metacraton and to the southwest within the West African craton (Fig. 3). They contain thick sediment piles (Genik, 1993): the Kermit trough is made up of a sequence of 10 km of Cretaceous and 5 km of Cenozoic sediment, the Ténéré trough of 5 km of Cretaceous and 2 km of Cenozoic sediment, the Tefidet trough of 3 km of Cretaceous and 1 km of Cenozoic sediment, and the Grein trough of 1.5 km of Cretaceous, 1.5 km of Paleogene, and 2 km of Miocene/Holocene sediment. By contrast, during the development of these nearby troughs, the Hoggar, Tibesti, and Libyan regions were elevated regions, prefiguring the current swells or highs.

AN ALTERNATIVE TO THE PLUME HYPOTHESIS

The constraints on Hoggar igneous activity are as follows:

- 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 down, and it could be much older. Alternatively, it could be tentatively linked to the deeper mantle, which may be warmer at a depth >250 km.
- Cenozoic Hoggar igneous activity is associated with a swell but not a thermal anomaly. A current large thermal anomaly is located to the north, below the Saharan basins, but it cannot be the present signature of a Hoggar plume because the Africa plate has moved to the northeast. Simple heat-flow modeling shows that a large, hot mantle body cannot significantly elevate the surface heatflow over periods of several tens of millions of years by conduction.
- Slightly higher heatflow has been measured very locally in the Atakor, suggesting local intrusion of asthenospheric mantle into the crust beneath Hoggar.
- No trace of a plume is observed between its postulated present location below Central Hoggar and the position it is predicted to have had at 35 Ma, ca. 1,000 km to the E-NE near the current position of Tripoli, on the Libyan Mediterranean coast. In contrast, large volumes of basalts were erupted in other swells or uplifts during the same period (35–0 Ma).
- Most of the Cenozoic northwestern African volcanic fields

were initiated at 35 Ma or younger, i.e., after the beginning of the Africa-Europe collision (ca. 38 Ma). Only carbonatites and strongly undersaturated massifs are older (e.g., Tamazert in Morocco) among rocks that have typical rift-related characteristics, corresponding to the last extension conditions resulting from the major Cretaceous rift period in northwestern Africa.

- Tomographic models indicate that the Cenozoic northwestern African volcanic fields can be correlated with the mantle structure down to 150 km but not at lower depth. This is not in agreement with the “primary” plumes (coming from the lowermost mantle boundary layer) postulated for Africa (Courtillot et al., 2003).
- On the whole, Central Hoggar (LATEA) has a geological structure intermediate between a craton and a mobile belt, and corresponds to the concept of a metacraton (a partially destabilized craton). The variation of lithospheric thickness deduced from Pan-African crustal structure can be reconciled with tomographic models. In particular, both indicate that the mantle beneath Central Hoggar is intermediate between cratonic mantle and active mantle, which has long been observed but not really understood.
- Strong deformation of mantle xenoliths is observed in Hoggar, illustrating that modal mantle metasomatism was important and that fluids were preferentially injected along strained zones.
- The source of Hoggar basalts is REE-enriched. Their Sr-Nd-Pb isotopic ratios indicate mostly a depleted mantle source, but enriched compared to mid-ocean ridge basalts source, particularly at the beginning of events. The involvement of old radiogenic lithosphere is minor.
- During the Cenozoic, the northward movement of Africa induced several tectonic phases (*T*), which appear to have been followed by volcanism (*V*) in Hoggar: *T* = 38 Ma (beginning of the Africa-Europe collision) and *V* = 35–30 Ma, *T* = 20 Ma and *V* = 20–12 Ma, *T* = 8 Ma and *V* = 7–2 Ma, and *T* = 1 Ma and *V* = recent.
- The Hoggar volcanic province and, more widely, the Tuareg volcanic province are controlled by the rheology of the Pan-African terranes. They are located along shear zones within or at the boundaries of metacratons (i.e., semi-rigid blocks; Fig. 3). They were also influenced by the late Pan-African brittle conjugate fault systems (Fig. 7). We propose that what has been demonstrated recently for the Bicol volcanic arc (Pasquaré and Tibaldi, 2003) can be applied to Hoggar. In the Bicol volcanic arc, magmas rose at depth along northwest-striking transcurrent faults, but in the uppermost crust they raised along northeast-striking fractures, parallel to the direction of greatest principal stress. In the Hoggar-Tuareg shield, in general the locations of volcanic districts are related to the main shear zones, while the precise locations of individual volcanoes are determined either by NE-SW or by NW-SE fractures. No radial dike swarm is known in Hoggar.

All the geophysical, geochemical, and geological constraints just described should be taken into account. We contend that they are not consistent with the current mantle plume model. We rather propose that Hoggar volcanism is the consequence of intraplate rejuvenation of Pan-African structures linked to the Africa-Europe collision, and more generally that intraplate volcanism is tectonically controlled and has a shallow source (Bonin and Lameyre, 1978; Lameyre et al., 1984; Black et al., 1985; Nni et al., 1986; Black and Liégeois, 1993). In the case of Hoggar, localized at a major change in lithospheric thicknesses (Figs. 8 and 9), the initiation of volcanism could have been additionally facilitated by edge-driven convection, which results where there is a strong lateral temperature gradient that induces small-scale convection and rapid movement of material through the melting zone (Anderson et al., 1992; King and Anderson, 1998; King and Ritsema, 2000).

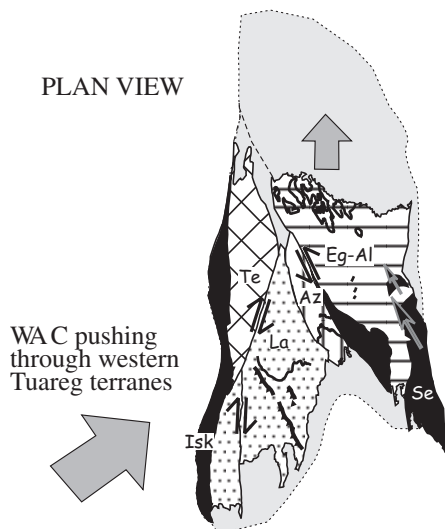
These considerations imply a shallow mantle source for the origin of the Cenozoic northwestern African volcanic fields. Their geochemical and isotopic compositions, comparable in many ways to that of OIB (e.g., Dautria et al., 1988, for Hoggar), can be easily explained with a shallow source when the degree of partial melting, the size of the melting regime, and the characteristic length scales of chemical mantle heterogeneity (“the statistical upper mantle assemblage”; Meibom and Anderson, 2004) are taken into account. The base of the lithospheric thermal boundary layer, by definition close to its melting point, may be enriched in various fusible components compared with typical OIB (Black and Liégeois, 1993).

Based on the model proposed for the Pan-African (Late Neoproterozoic) granitoids that were emplaced in the same areas, we suggest that linear lithospheric delamination occurred along the Pan-African mega-shear zones inside metacratons or at their margins (Fig. 10). Splitting of cratonic lithosphere (rather than



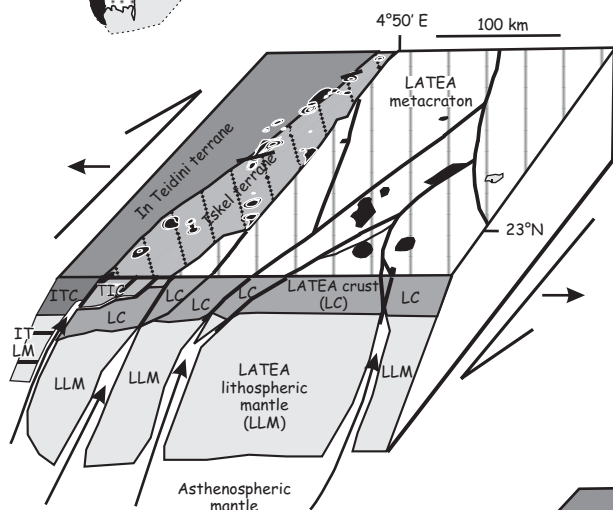
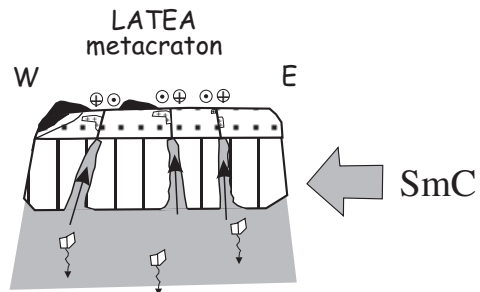
Figure 10. Proposed alternative model for the Hoggar magmatic episodes on a local scale. (A) During the post-collisional Pan-African period, the LATEA craton was dissected by transpressional mega-shear zones (metacratonization). Major linear delamination induced asthenospheric upwelling able to melt the lower crust. This gave rise to mainly crustal granitoids (620–580 Ma; Liégeois et al., 2003, and references therein). (B) At the very end of the Pan-African orogeny (535–520 Ma), a similar event occurred but with lower intensity and within a transtensional regime. It gave rise to high-level alkali-calcic plutons of mixed mantle and crust origin (535–520 Ma; Azzouni-Sekkal et al., 2003, and references therein). (C) This model suggests a similar scenario for recent volcanic activity, which may be an intraplate consequence of Africa-Europe convergence. The event, clearly intraplate, was more minor and gave rise to only minor delamination leading to mainly asthenospheric melt generation and the recent Hoggar volcanic formations. The ocean island basalt-type composition of Hoggar volcanic rocks is a consequence of the degree of partial melting, the size of the melting regime, and the characteristic length scales of mantle chemical heterogeneity (Meibom and Anderson, 2004). Asthenospheric upwelling is reinforced by edge-driven convection (King and Anderson, 1998) due to lateral temperature variations associated with basal lithospheric topography (not represented here; see Fig. 9).

PLAN VIEW



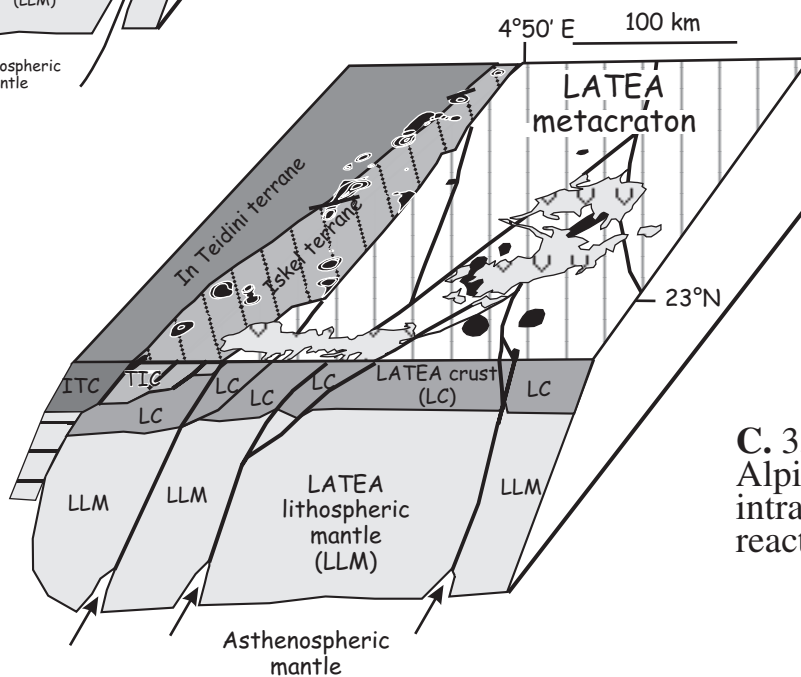
A. 625-580 Ma
metacratonisation

CROSS-SECTION



B. 535-520 Ma
Latest Pan-African
reactivation
("Taourirt" granitoids)

Cenozoic Africa-
Europe collision



C. 35- 0 Ma
Alpine
intraplate
reactivation

Thick lithosphere of
the In Ouzzal terrane and
of the West African craton

Edge-driven
convection

thinning) may allow asthenospheric adiabatic ascent from great depth and therefore extensive melting (Anderson et al., 1992).

During the Pan-African orogeny, the LATEA microcontinent was metacratonized, i.e., dissected by mega-shear zones along which linear delamination occurred, generating large amounts of granitoids (Fig. 10A; Liégeois et al., 2003). At the end of the Pan-African orogeny, reactivation of these shear zones induced emplacement of high-level alkali-calcic and alkaline plutons and ring complexes (Fig. 10B; Azzouni-Sekkal et al., 2003). We suggest that a similar process occurred during the Cenozoic (Fig. 10C), though it was more limited in intensity than that at the end of the Pan-African orogeny. It allowed the rise of enriched material from the lithosphere-asthenosphere boundary (Black and Liégeois, 1993; Anderson, 1995), perhaps encouraged by edge-driven convection (King and Anderson, 1998) and melting through adiabatic pressure release. The hot material was not abundant enough to melt the old lithosphere or to generate a thermal anomaly, except very locally. Only a minor amount of lithospheric upper mantle was involved in forming the early tholeiitic basalts. These were generated by large-degree partial melting and characterized by a more enriched isotopic signature than can be explained by major lithospheric involvement.

The Cenozoic volcanism was accompanied by uplift that led to the current swell. However, as shown earlier, the Hoggar area was already a major topographic high during the Cretaceous. This suggests that when the Africa plate was submitted to external stress, the Hoggar area, among others, was preferentially subjected to uplift. In Hoggar, maximum uplift occurred in the LATEA metacraton (Central Hoggar) and at the western boundary of the Saharan metacraton (Aïr). An explanation could be that preferentially uplifted areas are those that are semirigid (i.e., metacraton = "broken" craton). The two extremes are not favorable zones: cratons with their thick lithosphere are too stable, and mobile zones with their thin lithosphere would be able to deform, accommodating the stress and closing ways for magmas to surface. Such a model avoids invoking the coincidences and auxiliary hypotheses needed by the mantle plume model (Christiansen et al., 2002).

CONCLUSIONS

In northwestern Africa, early Cenozoic volcanism (>40 Ma) is rare, carbonatitic or strongly silica-undersaturated in composition, and limited to the Atlas region, which behaved at the time as a rift in the Tethys environment (Piqué et al., 1998a). Most of the volcanism and the largest volcanic fields predate the beginning of the Africa-Europe collision ca. 38 Ma (Hoernle et al., 1995). Geochronological data are too scarce to be decisive, but it appears that the discrete volcanic events that occurred in northwestern Africa followed successive compressive climaxes on the northern margin of Africa.

Cenozoic northwestern African volcanism is linked to tec-

tonic features: the Maghreb volcanism to the Atlas tectonics and the more intraplate volcanic fields (Hoggar, Tibesti, Libya) to the reactivation of earlier structures—north-south (Pan-African), NW-SE (Paleozoic), and NE-SW (Mesozoic). This intraplate tectonic activity occurred as early as the Cretaceous, generating rifts and uplifted regions, including Hoggar. The Cenozoic Africa-Europe collision accentuated preexisting swells.

The absence of heatflow anomalies (Lesquer et al., 1989) suggests that the zones of melting in the mantle are of limited extent and that heat transfer by magmas is low. Tomographic models suggest that the locations of the northwest African volcanic fields are correlated with warmer zones in the mantle down to 150 km, but not deeper. This calls for shallow mantle sources. The Hoggar is located in a zone in which the mantle, cold below the West African craton, becomes warmer toward the east. This temperature gradient suggested by tomography can be reconciled with the more complex lithospheric structure deduced from geology, the former having lower resolution.

We believe that all these constraints cannot be explained by the mantle plume model except by accepting many coincidences and adaptations. We suggest that the Africa-Europe collision generated intraplate stress that reactivated preexisting shear zones, causing a situation under which linear lithospheric delamination occurred. The consequent rising asthenospheric material, whose flow may have been reinforced by edge-driven convection, melted by adiabatic pressure release. The magma generated rose through the lithosphere along the shear zones. In Hoggar, the preferentially reactivated areas are cratonic boundaries and metacratons. The intraplate stress induced by the Africa-Europe collision enhanced the uplift of already high areas, generating swells. The semirigid character of these metacratons is probably favorable for both uplift and volcanism, which explains the association of both features.

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