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RESEARCH ARTICLE

The Cenozoic volcanic province of Tibesti (Sahara of Chad): major units, chronology, and structural features

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Abstract Using both field relationships and some absolute ages, the sequence of volcanic units in the Cenozoic Tibesti Volcanic Province (TVP) (Chad) is established as follows: (1) plateau volcanism, between at least 17 and 8 Ma, consisting of flood basalts and silicic lava plugs, with intercalated ignimbritic sheets in the upper basalt succession increasing in amount upwards. Ages decrease from NE to SW, following the migration of the small NW-SE flexures concentrating the feeding dike swarms; (2) Late Miocene large central composite volcanoes exhibiting diverse and original structures. Some of them (Tarso Toon, Ehi Oyé, and Tarso Yéga) are located along a major NNE fault, representing the main tectonic direction in Tibesti since Precambrian times; (3) construction of three large ignimbritic volcanoes, associated with significant updoming of the basement, ending with the collapse of large calderas: Voon (about 5–7 Ma), Emi Koussi (2.4–1.33 Ma), and Yirrigué (0.43 Ma); (4) basaltic activity, starting at about 5–7 Ma, and essentially consisting of cinder cones and associated lava flows (Tarso Tôh, Tarso Ahon, and Tarso Emi Chi); and (5) final volcanic activity represented by post-Yirrigué caldera activity in the Tarso Toussidé Volcanic Complex, and especially Ehi Toussidé (the only active volcano in Tibesti), plus Ehi Timi and Ehi Mousgou volcanoes, similar to Ehi Toussidé. The two tectonic directions controlling some volcanic features of the province correspond to the major old lithospheric structures delimiting the volcanic province, namely, the great NW-SE Tassilian flexure to the SW and a major NE-NNE fault zone to the E. Unusual conditions of uplift and erosion in the TVP enable exceptional exposure of the internal structure of its volcanoes.

Keywords Tibesti · Chad · Cenozoic · Volcanic units · Chronology · Tectonic control

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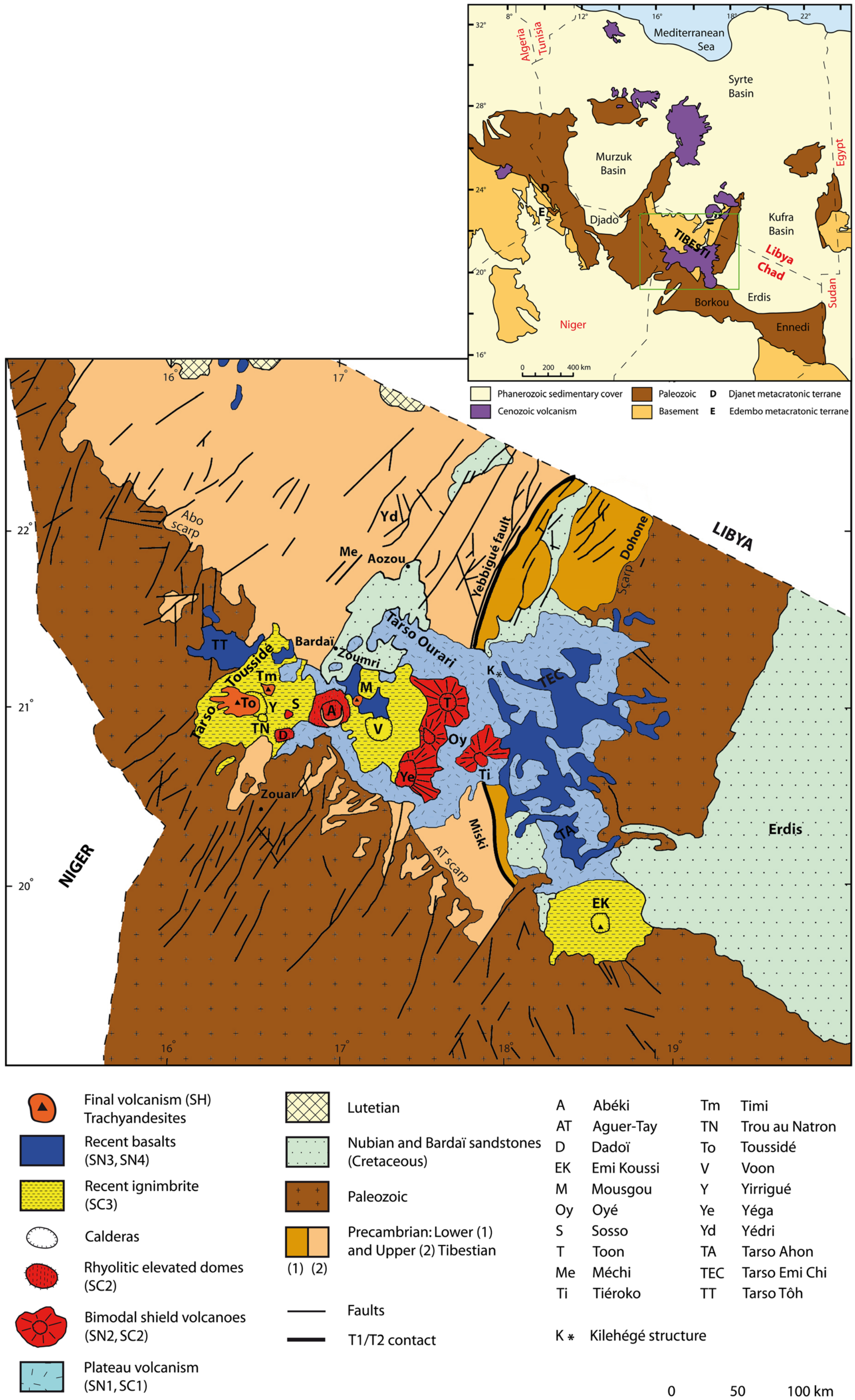
Introduction

The Tibesti massif is a mountainous area that straddles the extreme north of the Chad Republic and southern Libya (Fig. 1). In Chad, it spans a triangular area of approximately 100,000 km^2. Volcanic rocks occupy an area of about 30, 000 km^2 in the southern angle of this triangle, standing in marked topographic contrast to the low-lying surroundings. The Tibesti volcanic field is located at the southern end of a roughly NNW-SSE volcanic line stretching through Libya to the Mediterranean coast in the north (e.g., Conant and Goudarzi 1964) (see Fig. 1). Together, all these volcanic fields (about 100,000 km^2) represent the largest Cenozoic volcanic province in northwestern Africa.

All the high summits of Tibesti are volcanoes with elevations over 3000 m. However, the accumulation of volcanic products, interpreted to be between 5000 and 6000 km^3, is not responsible for the total topographic elevation of the area. Indeed, the surface of the underlying Precambrian and Palaeozoic basement is already high, reaching often more than 1500 m, and even locally 2000 m. Such associations of domal uplift and volcanism were also early outlined elsewhere on the

Fig. 1 Geological map of Tibesti Volcanic Province (modified after Gèze et al. 1959; Vincent 1970). *Insert map*: schematic regional geological map showing main geological units, including Edembo and Djanet metacratonic terranes of eastern Hoggar (modified after Hissene 1986, and Fezaa et al. 2010 for Djanet and Edembo terranes). *Green square* refers to main figure



African Plate (Gass 1972), especially in Hoggar and Aïr (e.g., Black and Girod 1970; Cantagrel and Karche 1983; Dautria and Lesquer 1989), and in Darfur and Adamaoua (e.g., Bermingham et al. 1983; Browne and Fairhead 1983; Dorbath et al. 1984). Hoggar, Tibesti, and Darfur were early associated with some of the most significant hotspots in the literature (e.g., Morgan 1972; Burke and Wilson 1976). However, more recently, it has been suggested that Cenozoic northwestern African volcanism (especially in Hoggar and Libya) is rather a consequence of the Africa-Europe collision, which generated intraplate stress, reactivating preexisting structural features (e.g., Liégeois et al. 2005; Azzouni-Sekkal et al. 2007; Beccaluva et al. 2008; Bardintzeff et al. 2012). Intraplate stress would have enhanced the uplift of already high areas and the magmas would have a shallow mantle origin.

Although Tibesti volcanism is supposed to be similar to volcanism of other African Cenozoic volcanic provinces, it is still poorly documented, both in Chad and Libya, as is the whole Libyan volcanism (e.g., Bardintzeff et al. 2012 and references therein). After the first exploration work of Nachtigal in 1869 (Nachtigal 1879) and of Tilho in 1915 (Tilho 1920), concluded by his topographic map at 1:1,000,000 scale, Dalloni (1934) published the first geological (including a sketch map at 1:2,000,000) and petrological data for Tibesti. Further exploratory geological survey and mapping of the Borkou-Ennedi-Tibesti area was performed at the end of the 1950s and during the 1960s (Wacrenier et al. 1958 with the first geological map at 1:1,000,000; Gèze et al. 1959), as well as a gravimetric study (Louis 1970). More detailed studies were later devoted to Western and Central Tibesti (Vincent 1960, 1963a, b, 1970).

Following these studies, Tibesti was practically inaccessible to scientists for a long time, for political reasons. Field work was again possible in some areas for a few years in the 1990s, allowing new observations in Western Tibesti and at Emi Koussi (Eastern Tibesti) (PM Vincent and A Beauvilain, unpublished data), with a petrological and geochemical study of this volcano (Gourgaud and Vincent 2004). Unfortunately, fieldwork is still largely hazardous in this country. As a result, this famous large intracontinental volcanic province remains the least known among African Cenozoic volcanic areas. Strangely, the very scarce and most recent literature about Tibesti (Guiraud et al. 2000; Permenter and Oppenheimer 2007) does not give much credit to the work of Vincent (1970), but this 1970 paper represents the only international contribution supplying field data from this largely inaccessible area.

This paper is focused on the description of volcanic units of the Tibesti Volcanic Province, their ages, relationships, and structural features. We use our own field data (both PM Vincent's data from the 1960s and more recent ones from PM Vincent and A Beauvilain) to establish the volcanic stratigraphy among these units, with some absolute ^{39}Ar-^{40}Ar age constraints. The resulting picture differs significantly from that given in the most recent papers on Tibesti, based exclusively on satellite image processing, reminding us of the limitations of such remote sensing methods when they are used without field control. The prominent role of NW and NE-NNE tectonic directions, characteristic of the Precambrian basement, in (1) the location of this volcanic province and (2) the development of some volcanic features inside the province itself, is also established.

Geological history and structural features of the Tibesti Volcanic Province (TVP)

The Precambrian basement underlying the Tibesti volcanic units is subdivided into two highly deformed series with a contact that was interpreted as a major unconformity (Wacrenier et al. 1958; Vincent 1963a). The oldest series, "Lower Tibestian" or T1, is exposed in the eastern part of the V-shaped plateau only (Figs. 1 and 2). It consists of high-grade metamorphic rocks intruded by different types of synorogenic granites. The youngest series, "Upper Tibestian" or T2, occupies a larger area and consists of low-grade metasedimentary rocks intruded by late and postorogenic granites of Late Neoproterozoic to Cambrian age (Rb-Sr ages on micas, Vachette 1964). Isoclinal folds with approximately N 20° axes, parallel to the T1/T2 contact, characterize this "Upper Tibestian" series. In the Miski graben, in SE Tibesti, this contact is NNW-SSE, becoming NW-SE before disappearing beneath the Emi Koussi volcanic units (Figs. 1 and 2).

Similar Precambrian units are also observed in Egheï (Libya), the northern extension of Tibesti. In this area, their contact is associated to a sharp aeromagnetic lineament and was interpreted as a crustal suture between the eastern part of a continental margin in the west and the western part of a mature island arc in the east (El Makhrouf 1988). The ages of these Precambrian units are scarcely known but are probably Neoproterozoic.

Interestingly, these "Lower Tibestian" and "Upper Tibestian" Precambrian units of contrasting metamorphic grade display similarities with the Neoproterozoic Edembo and Djanet Terranes, respectively, in Eastern Hoggar (Fezaa et al. 2010; Liégeois et al. 2013). These Hoggar terranes trending NW-SE (Fig. 1) are separated by a steeply dipping shear zone. Their contrasted behavior, during the late Ediacaran transpressive deformation, would reflect the presence of the Murzuq craton to the east, below the Murzuq basin (Fig. 1). Fezaa et al. (2010) and Liégeois et al. (2013) suggested that the "Upper Tibestian" could actually be the counterpart of the Djanet Terrane, on the southeastern side of the Murzuq craton. Both would then correspond to the reactivated boundary (or metacratonic margin) of the Murzuq craton with a thick continental lithosphere, whereas the Edembo Terrane (and "Lower Tibestian"?), much more strongly overprinted, behaved as a mobile belt, as a result of a thinner lithospheric mantle. This Murzukian (Ediacaran) intracontinental transpressive episode would be due to the indentation of the Murzuq craton toward the SW (Fezaa et al. 2010; Liégeois et al. 2013). In this hypothesis, the

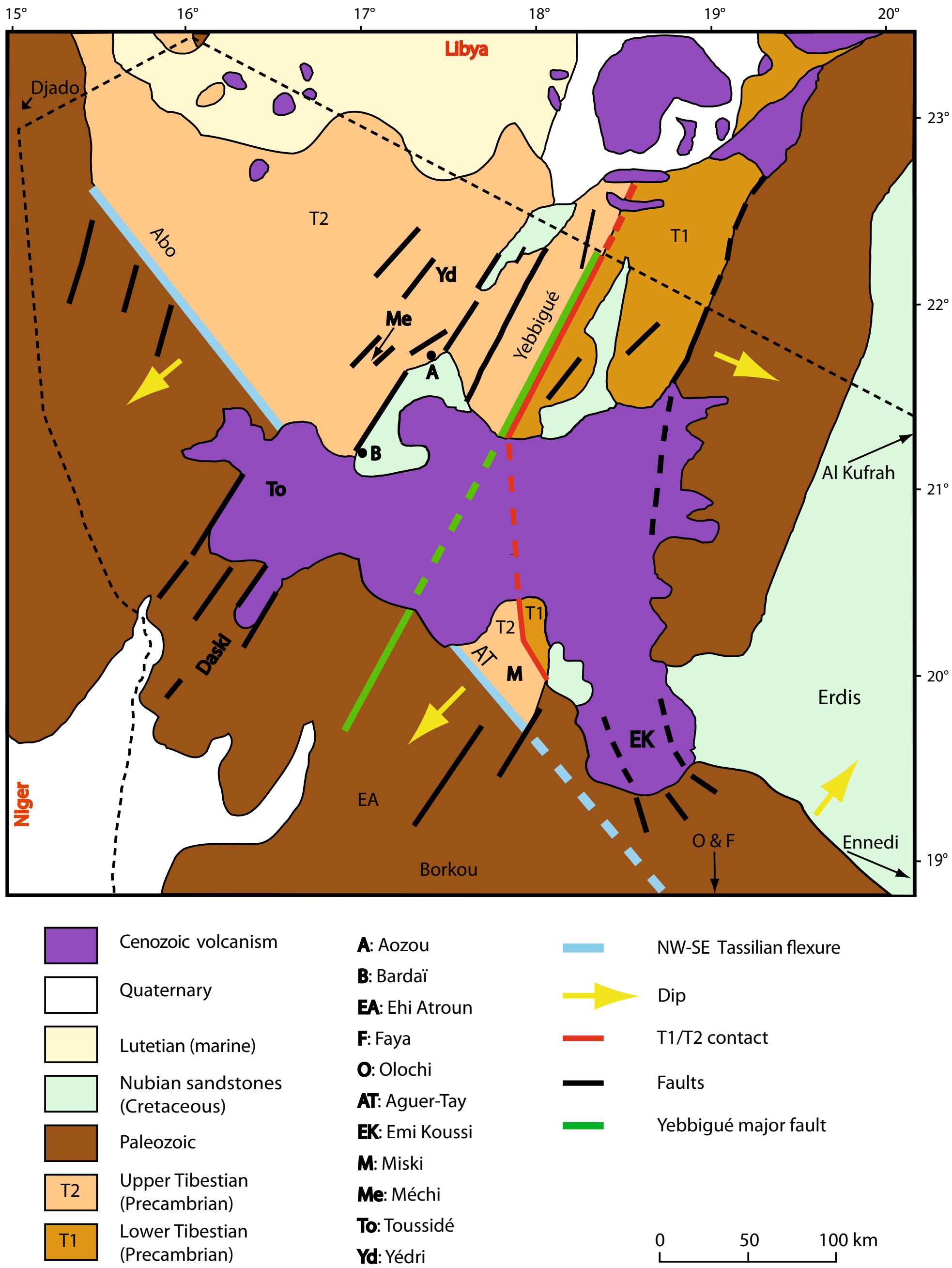


Fig. 2 Structural sketch map of Tibesti Volcanic Province, based on geological map (Fig. 1)

NNE T1/T2 contact in Tibesti would correspond to a shear zone also associated with this Murzukian episode. However, it can be expected that the situation in Tibesti is more complex. Indeed, it was recently proposed that Tibesti is bounded by additional cratons, Al Kufra to the east and Chad to the south (Liégeois et al. 2013).

In Tibesti, Palaeozoic sandstones rest unconformably on peneplained and weathered Precambrian units (Dalloni 1934; Bizard et al. 1955; Bonnet et al. 1955; Wacrenier et al. 1958). The Cambro-Ordovician deposits form escarpments, several hundred meters high, at the periphery of the plateau, except to the north (Figs. 1 and 2). This is beautifully illustrated by the great NW-SE flexure, or "Tassilian flexure," bounding the Tibesti highlands to the southwest (Fig. 2). Indeed, this major tectonic and morphological feature marks the contact between the Precambrian basement (Upper Tibestian) and Cambro-Ordovician sandstones and is very similar to the Tassilian cliffs surrounding Hoggar (Rognon 1970; Beuf et al. 1971). In the N and S of Hoggar, these cliffs surrounding the exposed Precambrian basement mark, over their respective lengths of 2000 and 1000 km, the lithological contact between this basement and the Cambro-Ordovician sandstones. Furthermore, the numerous faults bounding the Hoggar massif to the east, which also uplifted the Precambrian basement, especially the Edembo Terrane, are parallel to its contact with the Cambro-Ordovician sandstones (e.g., Rognon 1970; Fezaa et al. 2010). They also exhibit a NW-SE direction, attributed there to the Murzukian episode (Fezaa et al. 2010, see above). The NW-SE direction of the great flexure in Tibesti and of some major faults in the Murzuq Basin (Milesi et al. 2010) is probably also to be attributed to this Murzukian episode.

Inside the uplifted Tibestian block, younger nonfossiliferous continental sandstones rest directly on the Precambrian units. These "Nubian sandstones" (Figs. 1 and 2) were poorly dated to the Cretaceous period by Dadoxylon fossil wood (Dalloni 1934). However, some of these sandstones, the Bardaï sandstones (named after the closest village), more probably correspond to preserved outcrops of Late Palaeozoic to Early Mesozoic age (Deniel and Vincent, in preparation). Transgressive marine sediments were later deposited, during the Lutetian, in the far north of Tibesti.

Vincent (1970) suggested that the Tibesti basement is part of a vast Precambrian block which underwent movement over a long period of time, with the maximum uplift occurring after the Carboniferous. The denudation of the Precambrian basement would thus be the result of a Late Palaeozoic (probably Hercynian) uplift, also known in Libya as the Tibesti-Sirte uplift (Klitzsch 1966a). Faulting and uplift of Palaeozoic rocks in Hoggar have also long been attributed to Hercynian and Late Hercynian tectonic events (Conrad 1984). Subsequently, another major tectonic phase occurred in Tibesti, between Cretaceous (after the deposition of the "Nubian sandstones") and Early to Middle Eocene (pre-Lutetian), along NE to NNE trending faults developed in an antithetic Basin and Range Province style. This tectonic phase was interpreted as an extensional event related to a new swelling, reactivating Precambrian and Hercynian structures (Vincent 1963a). Based on thermochronological data, Rougier et al. (2013) proposed that the Tuareg shield underwent a widespread Eocene exhumation, prior to Cenozoic volcanic activity, outlining the importance of the Eocene uplift in the whole area. The complexity of uplift development has also recently been addressed in southern Libya (Ghienne et al. 2013; Carruba et al. 2014). This NE to NNE direction is actually predominant in Tibesti (Deniel and Vincent, in preparation), as also illustrated by the great Yebbigué fault (Figs. 1 and 2).

Stratigraphy and chronology of volcanic units in the Tibesti Volcanic Province

Three main volcanic units were defined in the TVP by Vincent (1970): plateau volcanism, central composite volcanoes, and ignimbritic shield sheets. Present-day volcanic activity is represented by the great solfatara field of Soborom, on the western flank of Tarso Voon, and the fumaroles of Ehi Toussidé (or Pic Toussidé) (Fig. 1).

The volcanic activity of the TVP has been intense since at least the Early Miocene. The earliest basalts of Egheï (Libya), close to the Chad border, overlie Lutetian marine deposits (Lelubre 1946). These basalts are identical to the earliest basalts of Tibesti but could be older.

Magmatism of Tibesti is mainly bimodal, with basalts sensu lato representing roughly 60 % in volume and felsic rocks, such as rhyolites, trachytes, and minor phonolites, approximately 40 %. Among these felsic products, large ignimbritic units are also observed (Fig. 3), whereas they have never been reported further north, in the huge volcanic fields of Libya. Intermediate lavas represent less than 1 % volume. They are only encountered in the final volcanic activity and mostly consist of trachyandesites, often exhibiting magma-mixing features (Vincent 1963a).

Plateau volcanism (SN1, SC1)

The older and irregular topography was flooded by the products of plateau volcanism which cover the largest area of the TVP (Fig. 4a, b). Alkaline olivine basalts (SN1 lavas in Fig. 3) first erupted. They look very similar to the "Plateau type" lavas of Mull (Hebrides) in the Paleogene volcanic province of western Scotland, first described by Bailey et al. (1924). The SN1 surface area and volume were estimated at about 10,000 km^2 and 1000 km^3, respectively, by Vincent (1969). They are associated with trachytic to phonolitic plugs (Fig. 4b); despite their small volume, these evolved rocks are important from the morphological point of view. They survived erosion of the associated basaltic lavas to be preserved as spines or even thick feeding dikes, only observable where higher degrees of erosion were reached. Besides, sheets of alkaline and peralkaline rhyolitic ignimbrite are intercalated in the upper half of the basaltic sequence, increasing in amount upwards (SC1 units in Fig. 3). The plumbing system of these old ignimbritic sheets, with no exposed vents, is unknown, except for the Kilehégé (or Kilinégué) ignimbrite which was fed by a radial dike swarm (see section "Outstanding volcanological features of Tibesti volcanism").

One of the Kilehégé ignimbrite feeder dikes was dated by ^{39}Ar-^{40}Ar at 17±0.2 Ma (N.O. Arnaud, personal communication, 10th August 1997). Only two other radiometric ages are available for the plateau volcanism. One is from a trachytic spine intersecting the western end of the Ourari trap, described later (Figs. 1 and 4) and known as Piton de la Balise. It gave a duplicated K-Ar age of about 8 Ma (8.4±1.4 and 7.9±0.9 Ma, RL Armstrong and PN Taylor, in Maley et al. 1970). The other K-Ar age of 10.2±0.5 Ma was obtained on a basalt from the area of Bardaï, SW of Piton de la Balise (von Jäkel 1982).

This first stage of Cenozoic migrating plateau volcanism covers the major part of the TVP and occurred over a long, but poorly known, time interval. It was built by successive eruptive pulses, separated by long repose intervals, as attested by

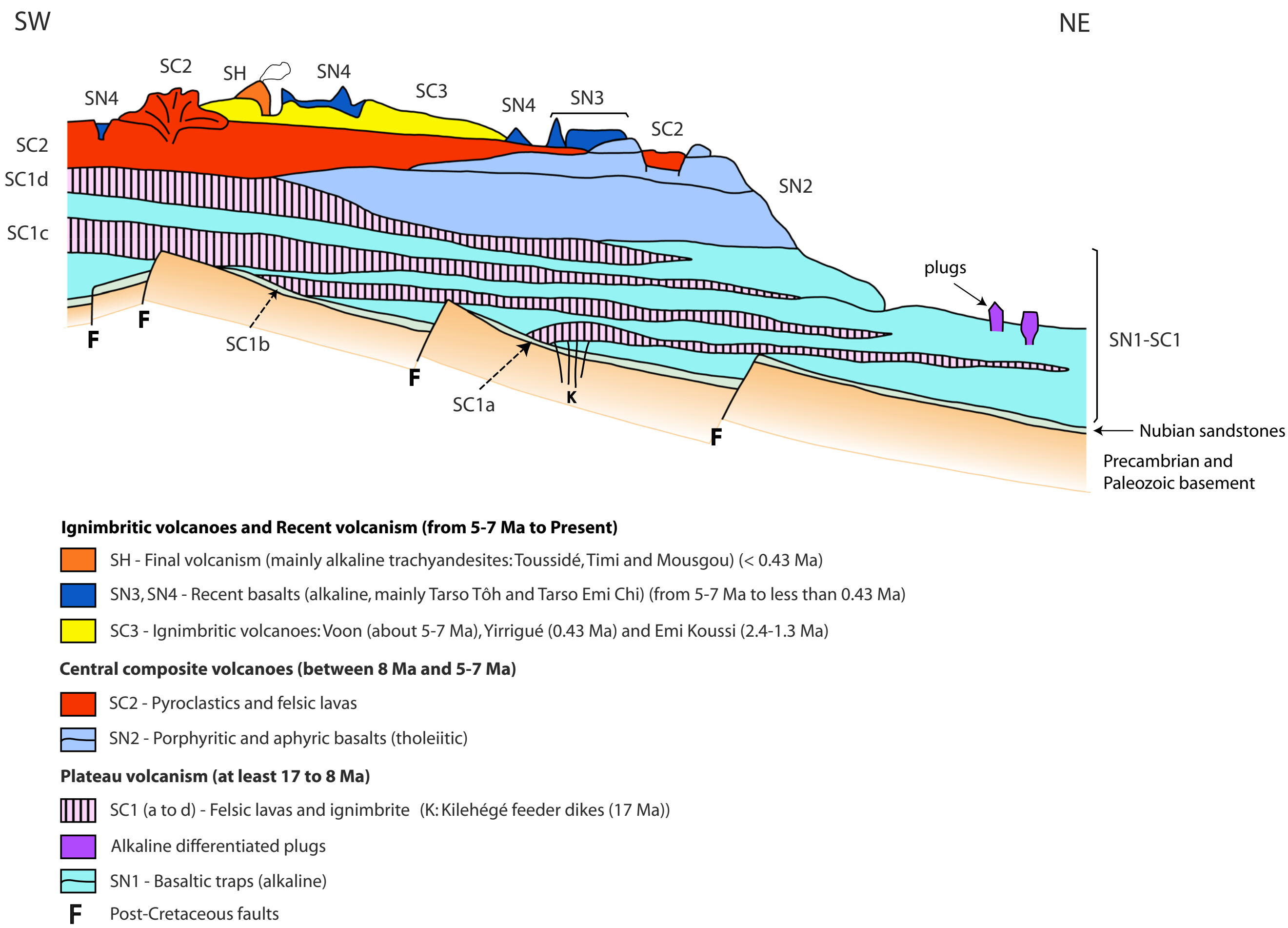


Fig. 3 Schematic and synthetic cross section through volcanic units of Tibesti (modified after Vincent 1963a, 1970), based on compilation of field data from various areas (mainly Western and Central Tibesti plus Emi Koussi), illustrating their chronology. For sake of clarity, both prevolcanic and volcanic units are not at scale. This cross section is about 300 km wide. SN and SC refer to the "Série noire" and "Série claire" of Vincent (1963a, 1970)

intercalated sedimentary deposits (e.g., Ourari elementary trap, Maley et al. 1970, and the section "Outstanding volcanological features of Tibesti volcanism"). This is reminiscent of the Hebridean Paleogene volcanic province of Scotland, especially in west-central and northern Skye, northwest Rum, and Mull (Staffa lava formation), where sedimentary units are intercalated in the lava sequence and were used to correlate the divisions between the lava formations (e.g., Anderson and Dunham 1966; Emeleus 1985; Williamson and Bell 1994, 2012).

Large central volcanoes

Different types of volcanoes are included in this group. The specific features, accounting for the terminology used here, will be described within the section "Outstanding volcanological features of Tibesti volcanism." These volcanoes are thought to be located along major faults, the best example being the 75 km long alignment of Toon, Oyé (or Oyoye), and Yéga volcanoes (Fig. 5), along the southern extent of the great NNE trending Yebbigué fault (Figs. 1 and 2).

Shield volcanoes (SN2, SC2)

These are a characteristic feature of the central part of the TVP (Fig. 4c). Each volcano was built in two distinct stages. A shield-like volcano, 20 to 40 km diameter, with 10 to 15° slopes, and a slightly convex profile was first built (Vincent 1970, SN2 in Fig. 3). Lavas associated with this first stage mainly consist of labradorite-rich porphyritic basalts with normative hypersthene and minor quartz and aphyric lavas. The tholeiitic (megaporphyritic and aphyric types) lavas are fairly similar to the two well-known "Central type" lavas of Mull's southeast caldera, initially recognized by Bailey et al. (1924).

Only pyroclastics and felsic lavas, mainly rhyolites, related to this tholeiitic series were emitted during the second stage of growth of these volcanoes (SC2 in Fig. 3). However, two distinct types of calderas characterize that stage: classic collapse calderas as for Toon and Yéga volcanoes and pseudocalderas associated with elevated domes, as for Tiéroko and Oyé (Vincent 1970; Vincent et al. 2004). There is no ring fault in the summit area of edifices belonging to this second group; the depression being

Fig. 4 Illustrations of main volcanic units in Tibesti Volcanic Province. **a**, **b** Plateau volcanism in Western Ourari. **a** Tilted Nubian sandstones (NS) in foreground covered by plateau basalts (SN1) on the left and trachytic plug in background (view from Pastis pass, about 47 km NE of Bardaï). **b** Western end of Tarso Ourari (view from Bardaï). Nubian sandstones (NS) in foreground (*on the left*), a NW-SE trending basaltic feeder dike (SN1) and various trachytic/phonolitic extrusions in background. **c** Northeastern part of Voon (SC3) caldera with Toon (SN2, SC2) shield volcano in background. **d** Emi Koussi (SC3) ignimbritic volcano, 3415 m a.s.l. (view from the SW foot of the edifice, March 1997). **e** Ehi Timi (SH) recent volcano, 3040 m a.s.l., with Yirrigué ignimbrite (SC3) in foreground (view from the S, just W of Petit Trou, April 1993)

caused by preferential erosion in the uplifted vent (Vincent 1970). They will be further described in the section "Outstanding volcanological features of Tibesti volcanism."

Rhyolitic elevated domes (SC2)

These are complex structures characterized by an association of extrusive and intrusive events (Vincent 1970). As with the edifices with pseudocalderas mentioned above, these elevated domes are characterized by the importance of vertical movements, related to underlying ring-dikes with laccolith behavior, which resulted in the uplift of their central part. However, the uplift was stronger for these pluto-volcanic structures and there is no obvious relation with a former basaltic shield. The laccolith may be directly observable, as in Dadoï, or is deduced from the local doming of the older volcanic or prevolcanic basement and the injection of cone sheets, as in Abéki (see section "Outstanding volcanological features of Tibesti volcanism").

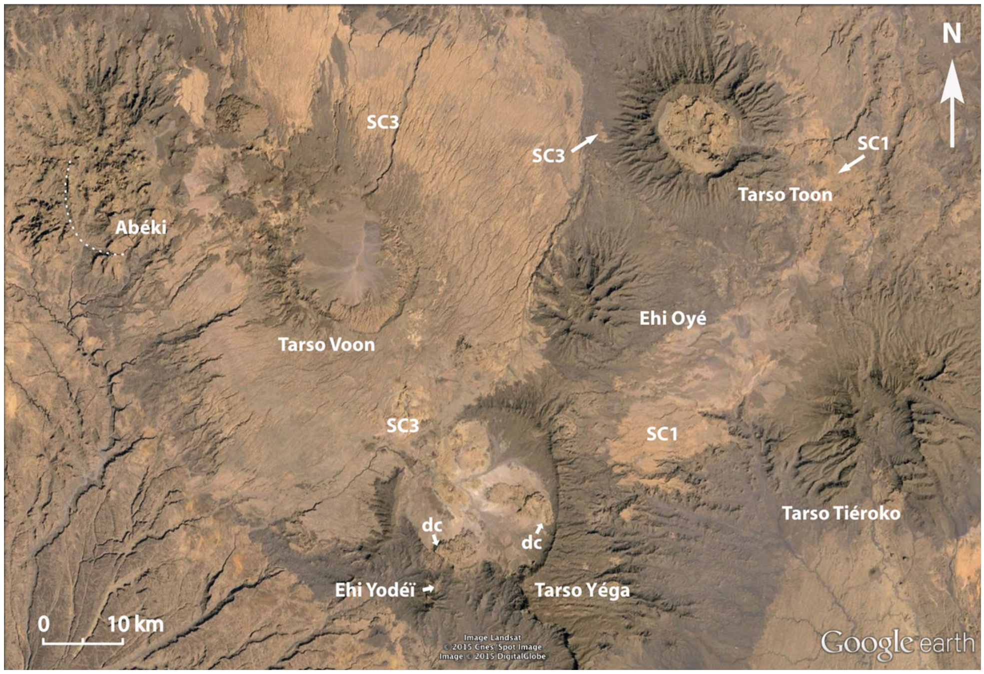


Fig. 5 Landsat images (31st August 2013) (Google Earth source) illustrating (1) diverse morphologies encountered among large central volcanoes* (about 8 to 5–7 Ma) of TVP and (2) Tarso Voon ignimbritic volcano (about 5–7 Ma). *Dashed white curve*: hemicircular depression in uplifted Precambrian schists in Abéki. *dc*: dome-coulées in Yéga caldera, *SC1*: plateau ignimbrite. *Toon, Yéga, Oyé, and Tiéroko are shield volcanoes (SN2, SC2), with classic collapse calderas for the first two and elevated domes and pseudocalderas for the last two. Abéki is a rhyolitic elevated dome (SC2). Yéga was invaded by younger Voon ignimbrite (*SC3*)

Lava domes and cumulo-domes (SC2)

Classic domes are also observed among these central volcanoes, such as Ehi Sosso (Fig. 1) consisting of a juxtaposition of large cumulo-domes. Rhyolites are vitreous at the surface and granophyric at depth. Viscous lava just spread laterally around the vent.

Ignimbritic volcanoes (SC3)

Three ignimbritic volcanoes with large calderas also occur in the TVP (Figs. 1 and 4c). From west to east, they are Yirrigué in the Tarso Toussidé volcanic complex, Tarso Voon and Emi Koussi (Fig. 4d). Contrary to the central composite volcanoes, these shield-like volcanoes are located away from major NNE-NE regional faults.

Yirrigué and Voon volcanoes mainly consist of peralkaline rhyolites (SC3 in Fig. 3). They are large, flat, slightly convex shields 70–100 km in diameter, with low angle slopes (2–3°). They were called ignimbritic “shield sheets” volcanoes by Vincent (1960, 1970). The Voon ignimbrite (SC3 in Fig. 5) covers an area of 2600 km^2 with an estimated volume of about 130 km^3. It flooded the lower slopes of the Tarso Toon volcano and entered into the eroded Tarso Yéga caldera (Figs. 1 and 5). Its eruption preceded the main excavation of the valleys, which occurred in the Late Miocene-Early Pliocene, between about 5 and 7 Ma (Swezey 2009 and references therein). The younger Yirrigué ignimbrite covers an area of 3200 km^2, including the western part of the Dadoï and Abéki structures, and has a volume of about 150 km^3. Its surface is almost flat (Fig. 4e), but distal tongues, following the valleys cut before its eruption, may have steeper slopes. The Yirrigué ignimbrite was dated by ^{39}Ar-^{40}Ar at 0.43±0.11 Ma (N.O. Arnaud, personal communication, 10th August 1997).

Emi Koussi is the biggest volcano of Tibesti and the highest summit (3415 m a.s.l.) in the Sahara desert (Fig. 4d). It was almost unknown (Gèze et al. 1959) before the field work of Vincent and Beauvilain in the 1990s and further petrological and geochemical investigations (Gourgaud and Vincent 2004). This shield-like volcano, 60–70 km in diameter, is classified as an ignimbritic volcano although it differs from the others due to several significant features. (1) Its slopes are less than 10° but clearly steeper than the 2–3° slopes typical of ignimbritic “shield sheets” volcanoes. (2) The different units of the ignimbrite, highly welded and with scarce *fiammes*, look very similar to lava flows, except in their upper and lower parts, and especially at their contact with the basement. Similar lava-like features are also observed in ignimbrites from other Continental Flood Basalt Provinces, such as Paraná-Etendeka and Karoo (e.g., Cleverly 1979; Milner et al. 1992). (3) The chemical composition of erupted products is also distinct, with the predominance of undersaturated magmas (Gourgaud and Vincent 2004).

Recent basalts (SN3, SN4)

These are alkaline undersaturated basalts (SN3 and SN4 in Figs. 1 and 3). In Central and Western Tibesti, the respective surface area and volume of SN3 were estimated at only 200 km^2 and 10 km^3, whereas for SN4, they were estimated at 1200 km^2 and about 15 km^3 (Vincent 1963a).

The SN3 basalts rest on the Voon ignimbrite and consist of untilted tabular basalts plus necks with vertical columnar jointing, as in the area of Bardaï. SN4 consists of numerous small (100–200 m high) basaltic volcanoes of Strombolian type and several maars. They may be scattered everywhere but are often grouped, the thin (except in valleys) lava flows making in this case an almost continuous “black tarso,” as Tarso Edri around Ehi Mousgou and Tarso Tôh, about 40 km north of Ehi Toussidé (Fig. 1). In these tarsos, volcanoes may be aligned, or not, along fractures of variable directions (NE-SW, NS, or NW-SE). Tarso Tôh consists of about 150 small volcanoes and rests, as many isolated SN4 volcanoes, on the Yirrigué ignimbrite. Note that these volcanoes are covered by explosive products of the post-Yirrigué Trou au Natron and Petit Trou (see the next section).

In Eastern Tibesti, this recent volcanism (SN3 and SN4) covers an even larger area between Emi Koussi and Tarso Emi Chi, often concealing older volcanic products (Fig. 1). Unfortunately, it is almost unknown.

Final volcanism (SH)

The final volcanic activity (Fig. 3) corresponds to the terminal hybrid series (SH) of Vincent (1963a) and is mainly concentrated in three volcanoes, each culminating over 3000 m high: Ehi Toussidé (or Pic Toussidé) and Ehi Timi in Western Tibesti and Ehi Mousgou in Central Tibesti (Fig. 1).

These volcanoes consist of large cones, 700–1000 m high and with 6–9 km basal diameter, high angle slopes, and convex flanks (Fig. 4e). The associated eruptive products, mostly potassic trachyandesites, cover an area of about 380 km^2 only, but their volume is estimated at about 40 km^3.

The initial Pelean stage is revealed by (1) trachytic lavas forming abrupt cliffs plus probable nuée ardente deposits on the southwestern flank of Ehi Timi and (2) the remnants of a Pelean dome still forming the central part and the two summital extrusions of Ehi Mousgou (Vincent 1963a). The second, effusive, stage of edification of these two volcanoes is documented by darker, mainly trachyandesitic, lava flows coating their flanks and making up the summit of Ehi Timi.

By contrast, Ehi Toussidé is totally coated by recent lava flows masking its initial products. However, the existence of an initial cone with much more viscous lavas is suggested by the steep slopes of the volcano, as for Ehi Timi and Ehi Mousgou. Ehi Toussidé represents the youngest volcanic activity (less than 100 ka) in the Tarso Toussidé volcanic complex (Deniel et al., in preparation). It largely flooded the Yirrigué caldera floor, after the formation of Trou au Natron (Doon) and Petit Trou (Doon Kinimi) calderas (Figs. 1 and 6). Ehi Toussidé is also the only active volcano in Tibesti, with fumaroles at its summit.

Outstanding volcanological features of Tibesti volcanism

Plateau volcanism

The elongated traps

The basaltic plateau consists of a suite of overlapping elementary composite traps, NW-SE trending, and called “traps oblongs” (Vincent 1969). The thickness of this trap series is largely variable, from a few meters to 240 m at the Kouaceur pass in northern Tibesti and more than 300 m in the NW-SE axis of Tarso Ourari, east of Bardaï. From NE to SW, successive traps extend further W, giving an “en échelon”

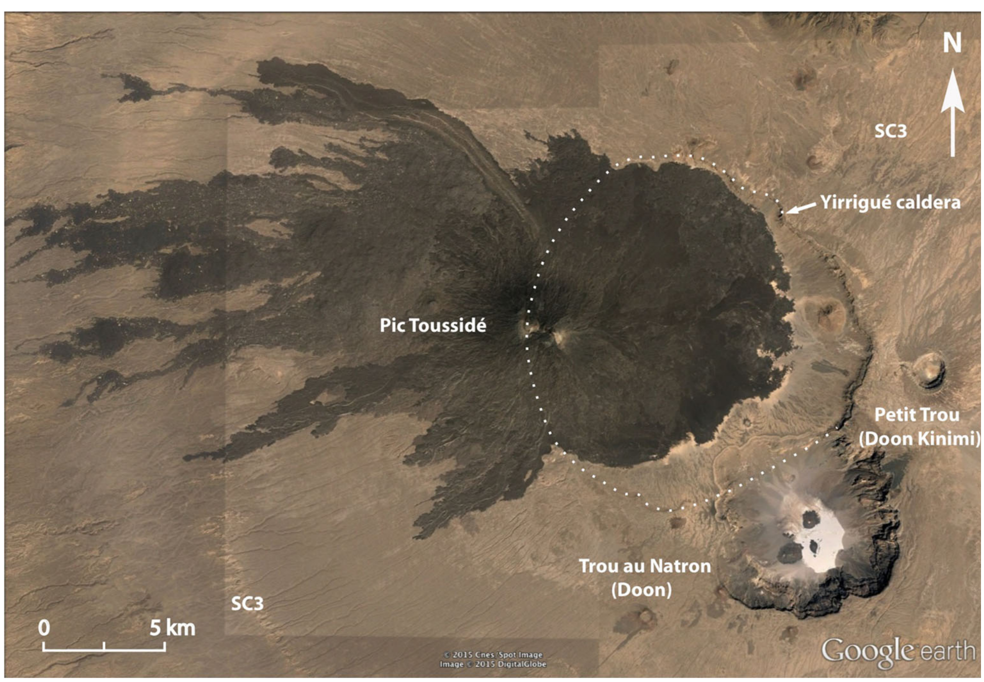


Fig. 6 SPOT images CNES (27th March 2003) (Google Earth source) illustrating relationships between various volcanic units of Tarso Toussidé Volcanic Complex (TTVC). *Dotted white curve*: outline of Yirrigué caldera ring fault where it is masked by Pic Toussidé or cut by Trou au Natron. Note that vents of both Trou au Natron (Doon) and Pic Toussidé are located on Yirrigué caldera ring fault. *SC3*: Yirrigué ignimbrite which flooded both caldera and surroundings

arrangement. The Tarso Ourari, NE of Bardaï, is the best known of these traps (Vincent 1969). Since basaltic flows are associated with ignimbritic sheets and lacustrine deposits in Eastern Ourari, their relative chronology can be established (Figs. 7 and 8). Field observations were also made in the northwestern end of the Ourari trap, over its 25 km width, but are probably still valid over its 70 km length up to the great Yebbigué fault, to the east. Dike swarms feeding basaltic flows are often observed, thanks to favorable erosion. They are located along the crest of NW-SE small flexures, where lava flows reach their maximum thickness (Fig. 7). The weak dip of lava flows on each side of these flexures indicates that this arrangement must be original. The dikes and associated flexures were indeed considered as contemporaneous (Vincent 1969; Maley et al. 1970). The dikes, generally 0.5–2.5 m thick, and overlapping flows are getting younger from NE to SW, following the migration of these flexures (Vincent 1969, 1970). This aging pattern, observed at the scale of the Ourari trap, seems to be valid for the whole Tibesti plateau volcanism but needs to be confirmed by radiometric dating. Indeed, to the NE of Ourari, the older age of the plateau basalts is suggested by their increasing degree of erosion up to the "dikes region" of Dalloni (1934). There, about 100 km NE of Ourari, only patches of basaltic flows are preserved, together with more or less eroded mafic dikes (Vincent 1969). Dikes of trachyte to phonolite, up to 8 m thick and corresponding to the "roots" of the plugs associated to the plateau basalts, are also observed in this "dikes region." To the SW of Tarso Ourari, outcrops of younger plateau basalts are scarce because of the widespread overlying products of Tarso Toussidé Volcanic Complex (Fig. 1).

The Kilehégé ignimbrite feeder dikes

The Kilehégé ignimbrite is exceptional among the old ignimbritic sheets associated with plateau basalts in that its plumbing system can be observed. This ignimbrite (SC1a), with an estimated volume of several 10 km^3, is intercalated between two flood basalt units (SN1a and SN1b in Figs. 7 and 8a) (Vincent 1963a). It was fed by a radial dike swarm, well exposed thanks to the associated, deeply eroded caldera (Vincent 1970; Ranvier 1998). Indeed, more than a hundred rhyolitic dikes, 3–15 m thick, are observed in the central part of the Kilehégé structure (10 × 8 km) (Vincent 1963c; Fig. 9), consisting mainly of the oldest SN1a basalts and forming a depression compared to the surrounding ignimbrite sheet. The dikes converge toward a focal zone located at the southern third of the structure, where the highest spines and several rhyolitic plugs occur (Fig. 9). Spines seem to result either from

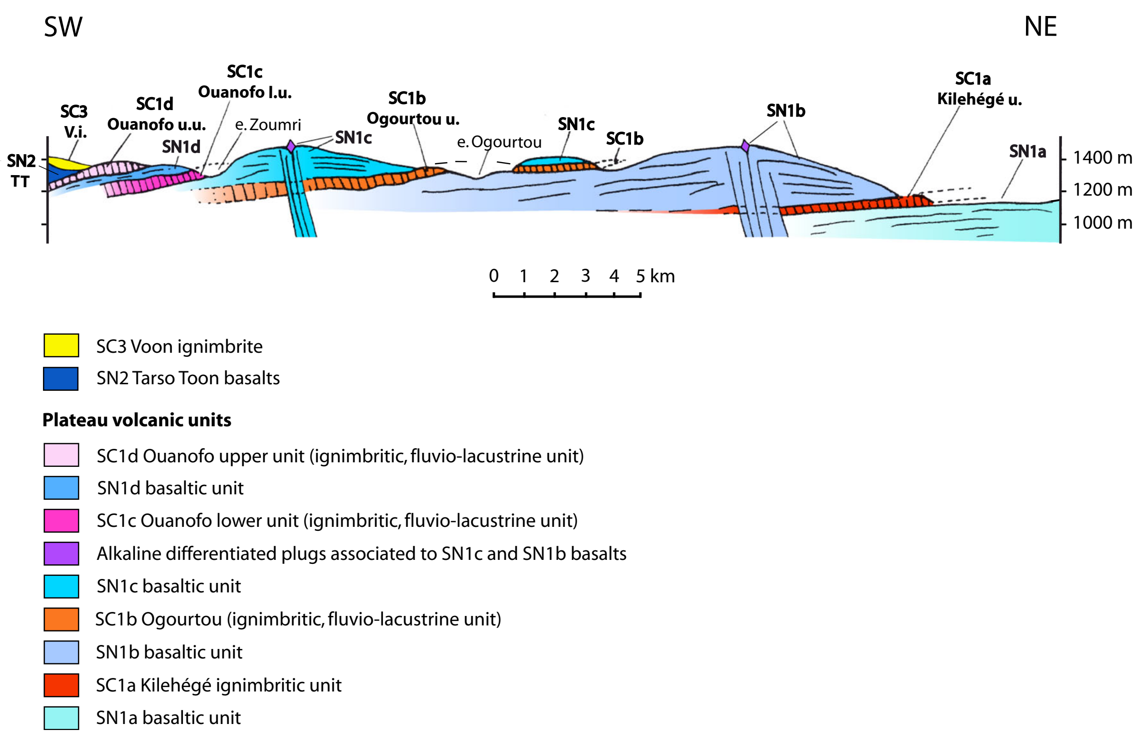


Fig. 7 NE-SW schematic cross section of Ourari trap (modified after Vincent 1969, 1970). Dike swarms feeding basaltic flows are situated along the crest of small flexures where lava flows reach their maximum thickness (SN1b and SN1c). These dikes and overlapping flows (SN1a to SN1d) are getting younger southwestward where they are covered by subsequent units: SN2 and SC3. This aging pattern observed at the scale of Ourari trap seems to be valid for whole plateau volcanism. SC1 units (*a* to *d*) refer to ignimbritic units (Kilehégé, Ogourtou, Ouanofo) ± related volcano-lacustrine deposits associated to SN1 basaltic traps (see Fig. 3 for details). *l.u.*: lower unit, *u.u.*: upper unit. SC1a (Kilehégé unit) is 17 Ma old. SN2 here refers to Tarso Toon (*TT*) which is between 8 and 5–7 Ma and SC3 to Voon ignimbrite (*V.i.*), about 5–7 Ma old. *e*: enneri

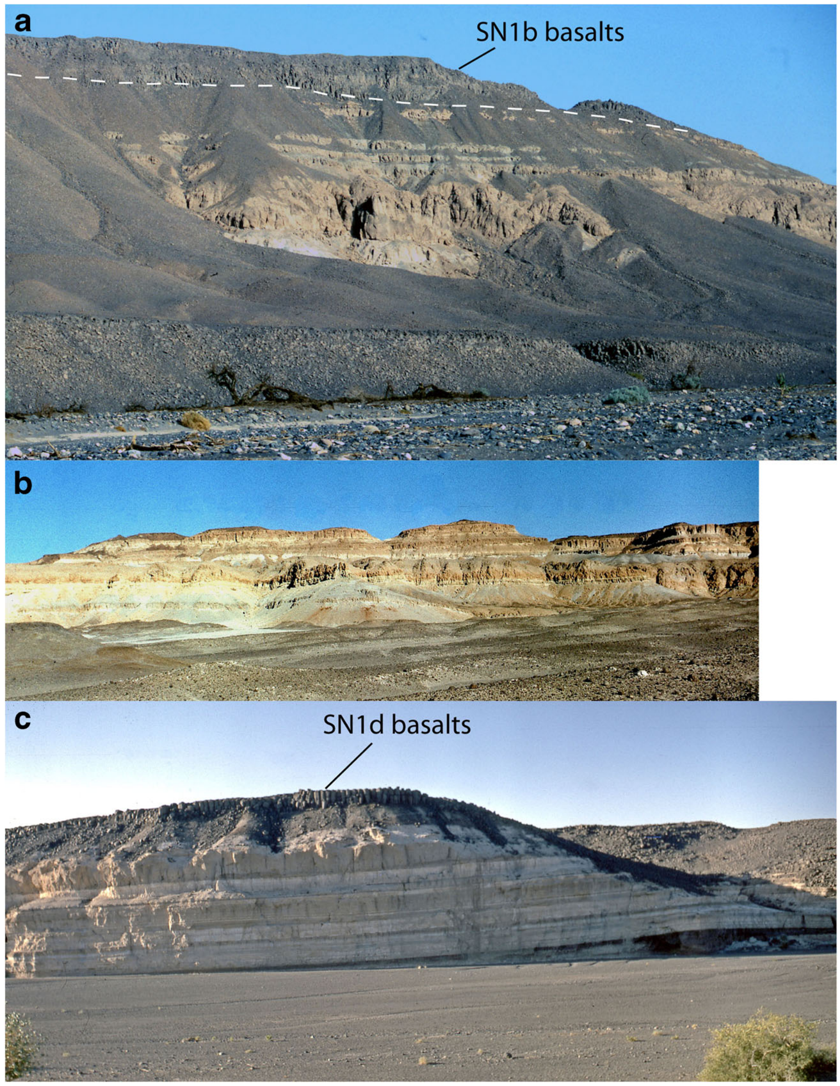


Fig. 8 Field photographs illustrating ignimbritic units occurring in Ourari trap (Fig. 7). **a** Kilehégé ignimbrite (SC1a), devoid of intercalated fluvio-lacustrine deposits, is overlain by columnar jointed SN1b basalts in enneri Iski, about 10 km NNW of Yebbi Souma. Its maximum thickness (75 m) is reached SW of Omchi (Maley et al. 1970). **b** Fluvio-lacustrine Ogourtou ignimbritic unit (SC1b) is about 80 m thick at Mohi Ma, in enneri Dohou (NNE of Toon volcano, about 19 km E of Ogourtou) (Maley et al. 1970). Its maximum thickness (110–120 m) is reached in enneri Yebbigué. **c** Fluvio-lacustrine lower Ouanofo ignimbritic unit (SC1c) is overlain by columnar jointed basalts (SN1d). It is about 14 m thick in enneri Zoumri, about 4 km E of Ouanofo (*this picture*), whereas it is 58 m thick 2.5 km W of Ouanofo (Maley et al. 1970)

Fig. 9 Field pictures of central part of Kilehégé structure. **a**, **b** Numerous rhyolitic feeder dikes of Kilehégé ignimbrite are associated with silicic plugs and spines. In **a**, scale is given by a person in the foreground

single dikes widening, or from the intersection of several dikes (Vincent 1963a).

The rhyolitic ignimbrite consists of a single unit (Vincent 1963c), devoid of intercalated fluvio-lacustrine deposits, unlike the younger ignimbritic sequences associated with Ourari plateau volcanism (Vincent 1969; Maley et al. 1970, Fig. 8). It is observable to the east, west, and north of this central basaltic core, with a weak but clearly external dip. It is observed continuously over 5 km toward the NW, whereas, elsewhere, it disappears under younger basaltic traps and later ignimbrites (SN1b and SC1b, respectively, in Fig. 7) (Vincent 1963a). The continuity of some dikes from the central part (SN1a) to the external part of the structure (Kilehégé ignimbrite: SC1a) is inferred from photo-interpretation only, but most dikes in SC1a trend in the same direction as dikes in SN1a (Ranvier 1998). Furthermore, the transition of a rhyolitic dike of the Kilehégé structure to the ignimbritic deposits was observed in the field, at the confluence of the Yebbigué and the Djiloa rivers, definitely establishing their contemporaneity (Vincent

1963a). A dike of ignimbrite was also observed in the field suggesting that the ignimbritic flow used these pathways to reach the surface. The conclusion of various field observations is that all these dikes are broadly contemporaneous, although not rigorously synchronous, and were the feeders of the Kilehégé ignimbrite, that erupted in a single stage. There is no field evidence of large circular faults delineating a caldera, but these might be masked by younger formations and the Kilehégé ignimbrite itself. However, a trap-door caldera is suggested and would be compatible with the centroclinal dips observed in the central part of the structure. The inferred caldera collapse must have occurred before the emission of the younger SN1b basalts which do not display an external dip, but probably immediately after the emplacement of the dikes. Based on analogic modeling, fractures would migrate from the surface to the reservoir during reservoir inflation, thereby allowing eruption (Ranvier 1998). The presence of numerous radial fractures would have permitted a rapid magma withdrawal from the crustal reservoir.

Note that, in general, the location of silicic eruptive centers in Large Igneous Provinces is poorly constrained, even in well-studied provinces such as Paraná-Etendeka or Afro-Arabia (e.g., Marsh et al. 2001; Ukstins Peate et al. 2005). Vent sites are indeed often buried by primary or redeposited products and may also be obscured by later tectonic and erosion (e.g., White et al. 2009 and references therein). The possibility of observing these Kilehégé ignimbrite feeder dikes is thus exceptional, by virtue of the deep incision of the Yebbigué river and its tributaries.

Large central volcanoes

Shield volcano calderas

Classic collapse calderas In Tibesti, the ring faults surrounding the subsided blocks of collapse calderas are always inward-dipping, becoming vertical with depth. However, it is known since Anderson (1937) that such cone-fractures result from an upward pressure, itself related to a surface doming. This is not a paradox if a two-stage mechanism is admitted. During the first stage, doming and formation of cone-fractures ends with eruption. In the second stage, collapse takes place into the empty space of the magma chamber, along one of these fractures acting as a normal fault, or along several fractures if it is a stepped caldera (Vincent 1970, 1989). The decreasing diameter of the caldera downwards explains the classical bowl-shaped geometry of the sinking volcanic deposits.

It seems that lateral stretching of the initial dome may also initiate a pre-eruptive caldera ("pré-caldera" in Vincent 1963a, 1989). This would explain why pyroclastics and felsic lavas may be confined within summit calderas, as beautifully illustrated in Tibesti by Toon volcano (Fig. 10). This caldera of the incremental subsidence type (Walker 1984) is characterized by small volume eruptions, the rhythm of which is possibly controlled by the weight of the truncated cone "stopper," acting as a safety valve (Vincent 1970).

Tarso Yéga is older than Tarso Toon and displays the largest caldera (about 250 km^2) of Tibesti. It is deeply eroded, especially on its southern flank where the caldera

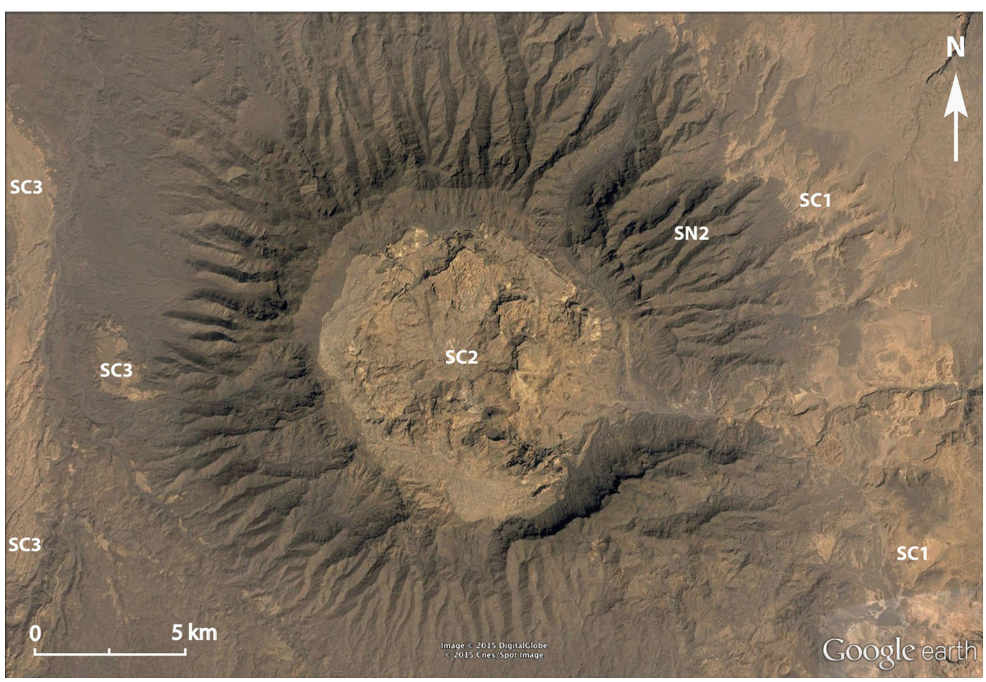


Fig. 10 SPOT images CNES (10th April 2013) (Google Earth source) of Tarso Toon shield volcano. Shield tholeitic lavas (SN2) rest over plateau ignimbrite (SC1) in the east and are covered by Voon ignimbrite (SC3) in the west. Note that Toon felsic products (SC2) occur only in the caldera

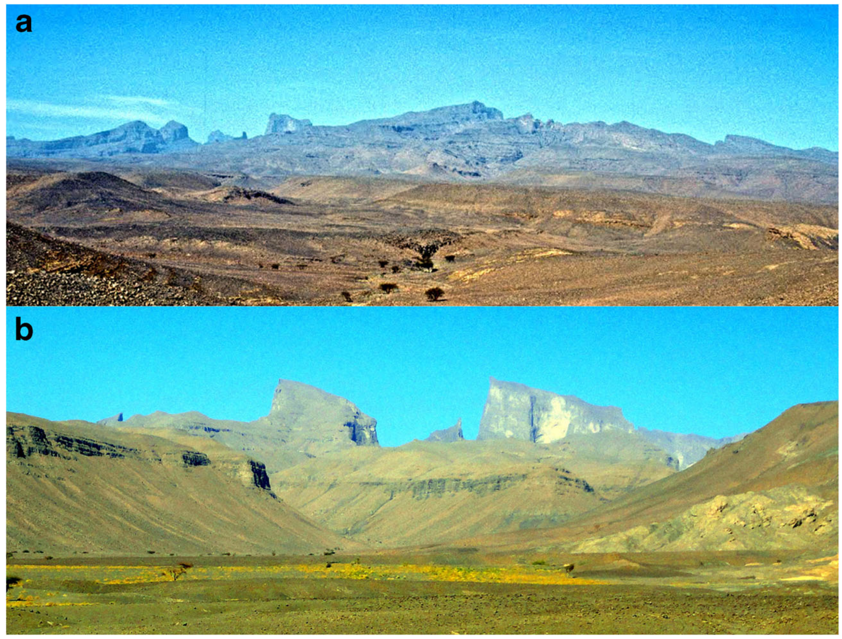


Fig. 11 Morphology of Tiéroko shield volcano, 2910 m a.s.l. **a** View from ENE, March 1997. **b** View from E, February 2012

wall was totally removed, and deformed by the subsequent intrusion of Ehi Yodéï about 3 km to the south (Vincent 1963a, Fig. 5). Two thick dome-coulées issued from its ring fault flowed toward the center of the caldera which was, after much erosion, invaded by the even younger Voon ignimbrite.

Elevated domes and pseudocalderas More structures consisting of elevated domes crowned by an irregular depression (pseudocaldera) characterize the second stage of construction of some basaltic shield volcanoes in the TVP (Vincent 1970; Vincent et al. 2004). Tiéroko and Oyé (Figs. 5 and 11) are examples of such volcanoes. Radiating vertical dikes are observed originating from the central area and cutting across the depression and the flanks. At Tiéroko, a central intrusion is clearly associated with these vertical radiating dikes (Vincent 1963a, 1970 and Fig. 12). In these volcanoes, felsic extrusions and intrusions in the central area resulted in (1) the uplift of earlier formations, including both the older volcanic and

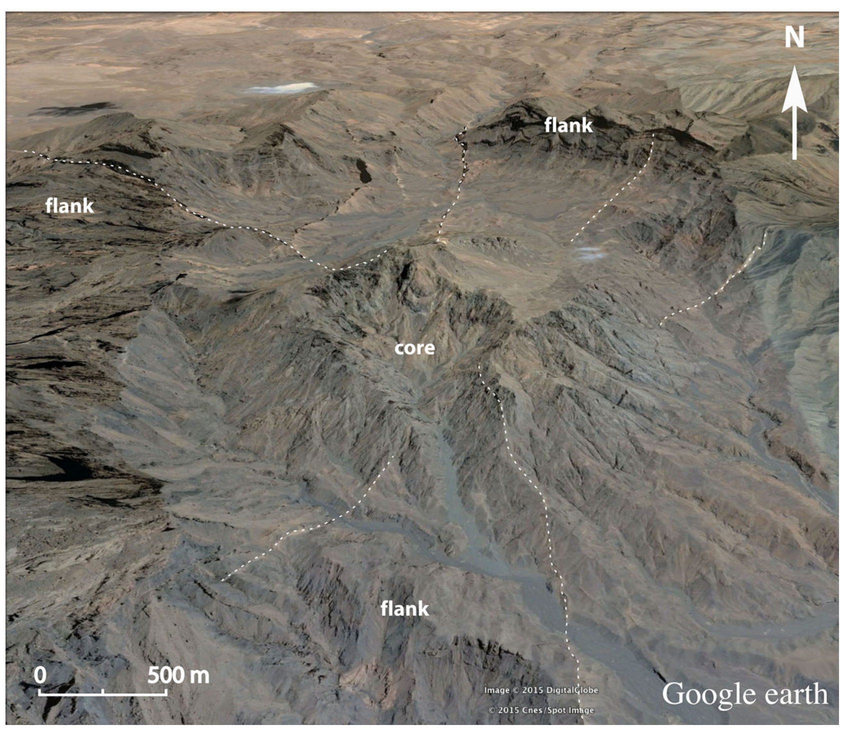


Fig. 12 SPOT images CNES (5th January 2013) (Google Earth source) illustrating main volcanostructural features of Tiéroko large central volcano. Radiating vertical dikes (*dashed white lines*) originate from the central intrusion and cut across the surrounding depression and flanks

prevolcanic basement, and (2) the injection of cone sheets. Furthermore, the dip of the flanks of these volcanoes increases abruptly toward the top of the edifices, from less than 10° in the lower slopes to more than 15° for Oyé and even 20° for Tiéroko (Vincent 1963a). This may indicate a restricted extension of the intrusion responsible for the uplift and which facilitated radial fractures. These pseudocalderas result from the preferential erosion of shattered and altered formations in the central area, above the roof of the laccolith. However, a tear of the intrusion roof itself, facilitated by radial fractures, may also have been involved.

Tiéroko and Oyé differ by the size of the central depression, 9×6 and 3×2 km, respectively (Fig. 5). Furthermore, the higher dip of the flanks at Tiéroko and the higher density of dikes compared to Oyé suggest that the central part of the volcano underwent a stronger uplift (Vincent 1963a). Unfortunately, field data are still insufficient to establish whether the upper basaltic lavas of the flanks belong to the same shield volcano, or are independent, as in rhyolitic elevated domes.

Rhyolitic elevated domes

Tarso Abéki (Fig. 1) is the best example of a large elevated dome in Tibesti (Vincent 1960). This hardly accessible asymmetric dome is 40 km wide with a central block forming a relief of 1200 m over an area of 17×11 km (Vincent 1963a, 1970). There is no caldera but the volcanic core is separated from the flanks by a hemicircular depression of uplifted Precambrian schists (Figs. 5 and 13). This basement high stands several hundred meters above its normal elevation (Fig. 14a). The core connects either continuously with the flanks or through a bordering fault or a flexure (Vincent 1960, 1963a). The dips of the flanks consisting of older material, including sandstones, are not original but result from doming (as von Buch (1836) wrongly believed for all central volcanoes!). These dips are inversely proportional to the width of the flanks (Fig. 13).

Over some 100 km^2, the core of the central block of Abéki displays the most complicated set of morphological features of Tibesti, with hundreds of spines and crests, often prismated, emerging from more eroded pyroclastic formations (Fig. 14c). Vertical, 300 m high cliffs are not exceptional and local relief in this central area may reach 700 m. Schematically, this volcanic core consists of a large cone of pyroclastic products (SC1 in Fig. 13), with (1) basement elements at the bottom, followed by (2) coarsely layered rhyolitic breccias (with cm to m size blocks) (Fig. 14b) with intercalated rhyolitic lavas in their upper part and, finally, (3) a rhyolitic lava flow, below the SC2 tuffs (Vincent 1963a). Furthermore, two sets of cone sheets feeding the younger thick lava flows and large extrusions (SC2 in Fig. 13 and Fig. 14d) are also observed in this central part. They are parallel to the peripheral set of dikes observed in Terkeï (NE flank of Abéki) and feeding the two rhyolitic needles of the top (Fig. 14e).

The flanks are covered by an impressive accumulation of (1) brecciated rhyolitic products including basement elements at the bottom, forming the base of SC1 and most of this unit on the SE flank (Fig. 13) and (2) thick rhyolitic lava flows ± tuffs (upper part of SC1 and SC2). The maximum thickness (at least 600 m) of these SC1-SC2 rhyolitic units is reached on the SW flank of Abéki (Adigouroumki, Fig. 14f). Basaltic lava flows intercalated with these rhyolites (Figs. 13 and 14d, e) are yet of unknown origin, but obviously came from the east of Abéki. This is

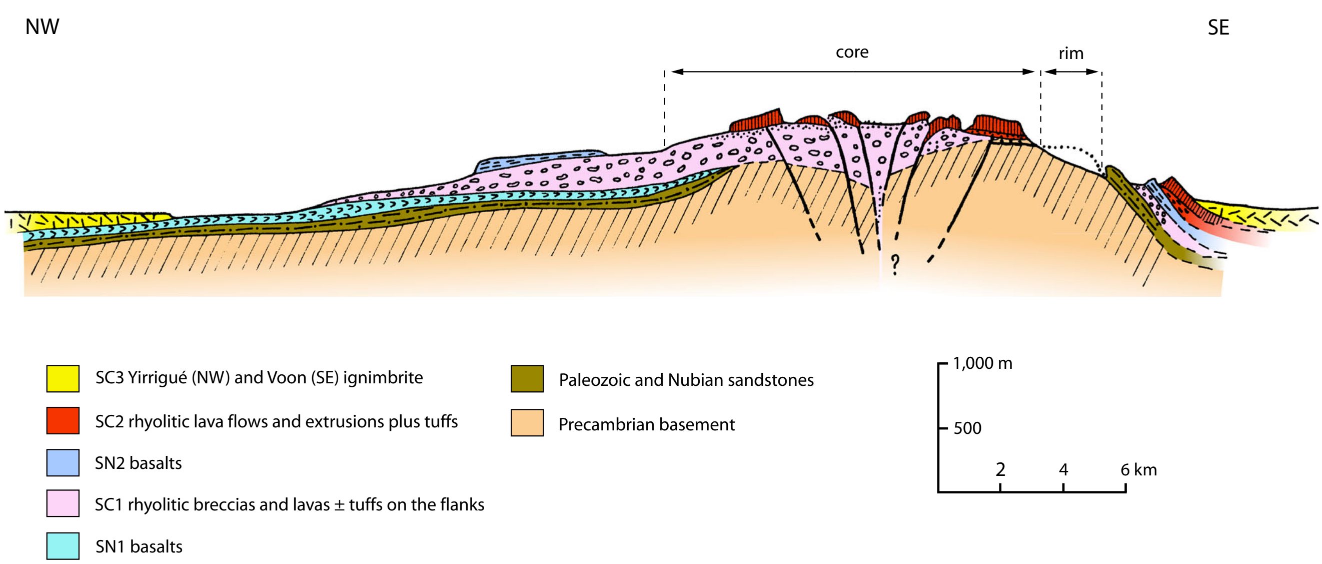


Fig. 13 NW-SE section of Abéki rhyolitic elevated dome. Note the strong asymmetry of flanks and uplifted Precambrian schists appearing in depression separating the core from the SE flank. *Continuous* and *dashed black lines* in the core delineate two sets of cone sheets plus peripheral set of dikes feeding younger lava flows and extrusions (SC2)

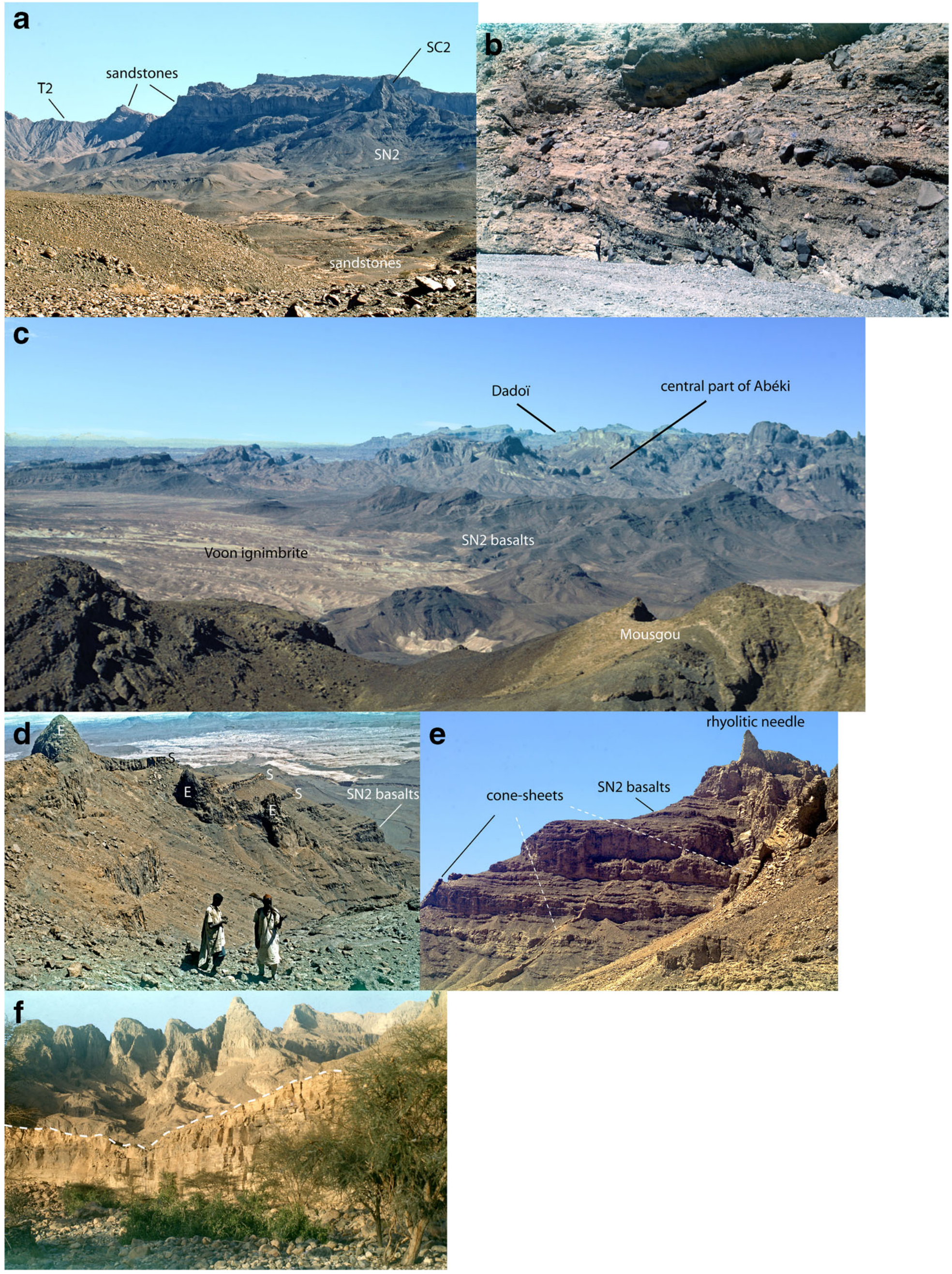


Fig. 14 Field photographs illustrating outstanding features of Abéki rhyolitic elevated dome. **a** Precambrian schists (T2) of the rim, *on the left*, with an outlier of uplifted sandstones (*middle background*) and rhyolites (SC2) of SSE flank overlying older basalts (SN2), *on the right*. Sandstones are observed in enneri (*foreground*) in their normal position, below basalts (southern part of Abéki). **b** Coarsely layered rhyolitic breccia (SC1, about 20 m thick) of initial cone. **c** Panoramic view from Mousgou over the central part of Abéki. Numerous spines and crests (SC2) emerge from more eroded pyroclastic formations (SC1). SN2 basalts and Voon ignimbrite (SC3) occur between Abéki and Mousgou. Dadoï is observed *in far background*. **d** Summital area of éhi Terkeï. A rhyolitic dike (SC2) crosscutting SN2 basalts fed three small extrusions (*E*) and associated sills (*S*) or flows. **e** Southern wall of éhi Terkeï (NNE flank of Abéki). SN2 basalts are crosscut by three rhyolitic cone sheets (SC2), the largest one *on the right* feeding the rhyolitic needle of top. **f** Thick ignimbritic tuff (*bottom level*) and rhyolitic lava flows (*upper level*) (SC1 and SC2 units) on SW flank of Abéki (Adigouroumki). Among 1000 m of volcanic products on this flank, 800 m consist of rhyolites and 200 m of basalts (Vincent 1963a)

because their maximum thickness (about 300 m) is observed in Terkeï, on the NE flank of Abéki (Fig. 14c), and is decreasing rapidly to the west, where they finally disappear. These basalts, younger than the Plateau basalts (SN1) but older than the Voon ignimbrite (SC3) (Fig. 13), are probably contemporaneous with shield volcano basalts (SN2). The dip of these flows was acquired during the above mentioned doming. Actually, the conformity of Palaeozoic sandstones, SC1 rhyolites, and SN2 basalts and the unconformity of SC2 rhyolites observed on the SE flank of Abéki (Fig. 13) suggest that doming occurred before the end of SC2 rhyolitic volcanism (Vincent 1963a).

The existence of three almost concentric sets of cone sheets feeding the tuffs and thick lava flows and extrusions of SC2, and the doming, imply the existence of a shallow intrusion. It is inferred that the partial collapse of a vast initial pyroclastic cone of debris (containing blocks of syenite and peralkaline rhyolites, e.g., Adigouroumki breccia on the SW flank), which was extending over the present areas of the core and flanks, induced the formation of an arched tensional fracture at depth in which magma was injected as a laccolith (Vincent 1963a, 1970; Vincent et al. 2004).

The more deeply eroded tarso Dadoï is located about 20 km WSW of Abéki (Figs. 1 and 14c), on the Daski horst. Although less studied than Abéki, Dadoï displays much similarities with it. However, the flanks are partially missing and the volcanic core is mainly occupied by a massive laccolith; breccias and cone sheets being subordinate. The laccolith is directly observable over 13×8 km and includes several Precambrian septa. It is made of alkaline microgranites and granophyres displaying glassy groundmass at the contact with the schists, in keeping with the shallow emplacement level. Based on field data, Vincent (1963a) estimated the maximum thickness of rocks above the intrusion roof as ca. 500 m. Note that about 600 m of rhyolitic lavas, probably mostly related to the western flank of Dadoï, are also beautifully exposed in the walls of the Trou au Natron caldera.

These unusual and very complex structures, built by intrusive, aerial extrusive and effusive events, still remain poorly studied, as are the edifices with pseudocalderas. However, a set of previously unknown "elevated domes," with impressive radial and concentric dike swarms, was also identified by Landsat imagery at the boundary between NE Tibesti and Libya (Vincent et al. 2004).

Large ignimbritic volcanoes

Relationships between volcanoes and basement

The uniqueness of Yirrigué and Voon volcanoes resides in their relatively high elevation, almost 1000 m, which is not related to the thickness of their eruptive products, reaching 100 m only in the walls of the Yirrigué caldera. The location of this caldera on the Daski horst can only account for part of the elevation too. Additional doming of the basement prior to the ignimbrite emplacement, and thus preceding their emission, is required to account for the remaining elevation. Incidentally, this is documented by the Voon ignimbrite resting unconformably on deformed, earlier volcanic units (Vincent 1963a). Such doming might reflect the past development of large crustal magma chambers.

Emi Koussi is located on a horst, as Yirrigué, and large uplift of the basement is also clearly evidenced. Along a SW transect, the highest exposure of Palaeozoic sandstones was observed at an altitude of 1200 m, that is 400 m higher than downslope, before disappearing under volcanic products. Gourgaud and Vincent (2004) suggested that the altitude of the basement upslope could even be higher. The thickness of eruptive products, observed in the walls of the two youngest calderas, is merely 300–400 m, here again insufficient to account for the observed elevation of the edifice.

These facts suggest that settings on horsts with significant basement uplift favor the accumulation of large volumes of magma in upper crustal magma chambers. Such accumulation of laccolithic bodies would result in regional updoming prior to any eruption.

Calderas

After the slow accumulation of magma in upper crustal magma chambers and inflation, a rapid eruptive tapping of the magma chamber would be followed by the collapse of a caldera. Although the vents of ignimbritic eruptions in Tibesti are only observed for Kilehégé, belonging to plateau volcanism, the same structures are likely to have been involved in younger ignimbritic volcanoes.

The complex summit caldera of Emi Koussi (11×15 km) results from the emission of three successive Quaternary (2.4–1.32 Ma) sequences, consisting of felsic products followed by small volume basaltic lavas (Gourgaud and Vincent 2004). The Voon and Yirrigué ash-flow calderas result from a single collapse and, as for Emi Koussi, the postcaldera activity occurred without floor resurgence. The older ignimbritic sheets intercalated in the plateau basalts, usually with no exposed vents, are also most likely related to this type of caldera.

The Yirrigué caldera (13×14 km) was later flooded by younger volcanic units of the TTVC (Fig. 6), so that its initial depth is unknown. The Voon caldera (18×11 km, Fig. 4c) is of the trap-door type, with a stronger collapse on its western side where arcuate folds are observed. Three anticlines emerge from a cover of recent basaltic lava flows (Vincent 1963a). The famous solfatara field of Soborom is located about 2 km west of this caldera, in the dissected core of one of these anticlines. Although the age of this structure is still uncertain, the anticline of Souradom (Fig. 15), located about 5 km southwest, and parallel to the western caldera cliff too, is clearly younger than the Voon ignimbrite. Here, folding cannot thus be due to the initial emplacement of a large magma body into the upper crust, as proposed for the famous concentric folds in southern Mull (e.g., Bailey et al. 1924; Bailey 1962; Skelhorn et al. 1969; Emeleus and Bell 2005). In this case, the lateral compression necessary for folding is better explained by the

Fig. 15 Field picture of periclinal closure of Souradom anticline, about 5 km southwest of the western caldera cliff of Voon ignimbritic volcano. Vertical layers correspond to Voon ignimbrite (SC3). Dip is decreasing very rapidly further away and SN3 basalts (*in the background*) are subhorizontal

collapse of a central block along an inward-dipping circular fault with a decreasing diameter downwards. Such collapse would occur after the ignimbrite eruption, as for classic collapse calderas in Tibesti.

Discussion

Chronological development of the TVP

Large discrepancies are observed between the chronology of volcanic units given in this paper and that proposed by Permenter and Oppenheimer (2007), based exclusively on their satellite image study.

The oldest volcanic activity in the TVP is clearly that of plateau volcanism (SN1) in the northern central part of the province, as early recognized by Dalloni (1934) and Vincent (1963a, 1970). The Kilehégé ignimbrite (SC1), here dated at 17 Ma, gives the oldest age available for the TVP. However, it is only a minimum age for this plateau volcanism as the ignimbrite was emplaced over even older plateau lavas which remain undated. These older volcanic units, although now largely eroded, extended further north (northeast of Tarso Ourari), as attested by the survival of basaltic and silicic dikes. These data suggest that the first phase of volcanic activity thus started before 17 Ma in the northern TVP and probably lasted more than 10 My. Indeed, the Tarso Ourari trap belonging to the plateau volcanism is associated to trachytic and phonolitic protrusions, one of them being dated at about 8 Ma. Besides, the focus of volcanic activity migrated during plateau volcanism. This migration, occurring southwestwards and perpendicular to the feeding NW-SE flexures, is apparently consistent with the NE motion of the African Plate at that time. However, available geochronological data are still insufficient to assess precisely the meaning of such migration.

The second stage of volcanic activity was mainly concentrated in the central TVP. It resulted in the growth of large central composite volcanoes (SN2, SC2). Although no radiometric data is available as yet, stratigraphic relationships suggest ages ranging from 8 to 5–7 Ma. Some of these volcanoes (Toon, Oyé, and Yéga) are also clearly associated with a major NNE fault, dating back from Precambrian and Hercynian times, which was reactivated in the Late Miocene. Toon and Yéga exhibit spectacular incremental collapse calderas, whereas other volcanoes exhibit more or less complicated structural features (Tiéroko, Abéki, and Dadoï) as well as more classic cumulo-domes (Sosso).

The aging of earliest volcanism toward the NE, documented along the 25 km width of Tarso Ourari and suggested by field data up to the "dikes region," and the location of large central volcanoes in the central TVP, are in clear opposition with the interpretation of Permenter and Oppenheimer (2007). Indeed, these authors proposed a reversed chronology between the first two stages of volcanic activity in the TVP and, thus, a migration of this activity toward the north. According to these authors, volcanism would have started in the central part of Tibesti, with the building of central composite volcanoes (our SN2-SC2) and, then, activity would have migrated to the north with Ourari plateau volcanism (our SN1-SC1) and to the east with Emi Koussi edification (SC3). In their model, the building of Emi Koussi volcano would have encompassed most of the time span of volcanic activity in the TVP. However, in spite of its large size, this volcano was essentially built over 1 Ma only (Gourgaud and Vincent 2004).

We already outlined the similarities between the plateau volcanism and central composite volcanoes of Tibesti and the Mull "Plateau" and "Central" groups defined in the Paleogene volcanic province of NW Scotland. This succession of plateau volcanism including ignimbrite and central type volcanoes is classically observed in other Continental Flood Basalt Provinces. In Ethiopia, for example, the Eocene-Oligocene flood basalts are overlain by Miocene lavas erupted from large central vent volcanoes (e.g., Merla et al. 1979; Berhe et al. 1987). On the Northwestern Ethiopian plateau, the thick Oligocene traps mainly consist of basalts, except on top of the plateau where ignimbritic tuffs, a few meters to 500 m thick, are observed (e.g., Pik et al. 1998). However, in southern Ethiopia, Eocene-Oligocene basalts are rare and the volcanic sequence is dominated by thick Eocene-Miocene acidic lavas and ignimbrite (e.g., Pik et al. 1998).

The third stage of activity in the TVP corresponds to the three large ignimbritic volcanoes (SC3) and their final collapse to produce large calderas. They consist of Emi Koussi in the southeast (2.4–1.32 Ma), Voon in the center, predating the main erosion of the valleys in the Late Miocene-Early Pliocene (probably between 5 and 7 Ma), and Yirrigué in the west (0.43 Ma).

The fourth phase (SN3, SN4) is represented everywhere on the plateau by basaltic activity. SN3 has a maximum age of 5–7 Ma because it is younger than the Voon ignimbrite, but no minimum age estimate can be proposed as relationships with the Yirrigué ignimbrite could not be observed. SN4 is younger than 0.43 Ma as it overlies Yirrigué ignimbrite, but older than Trou au Natron.

Finally, the most recent activity (SH) in the TVP is mainly located in the western part of the province. It is represented by (1) the post-Yirrigué caldera activity in the Tarso Toussidé Volcanic Complex, consisting of Trou au Natron and its small associated volcanoes, Petit Trou and Pic Toussidé volcano (Deniel et al., in preparation), and (2) Ehi Timi and Ehi Mousgou volcanoes. Although presently dormant, Ehi Toussidé is the only active volcano in Tibesti.

Again, this chronology is at odds with that proposed by Permenter and Oppenheimer (2007). These authors considered that the lava fields (SN3, SN4) predate the caldera collapses at Voon and Toussidé (Yirrigué) (SC3). They also ascribed the third caldera at Emi Koussi to the same, and terminal, phase of volcanic activity as Pic Toussidé. Although field data cannot bring constraints on this chronology, the two volcanoes being distant from each other, absolute ages do. Specifically, the youngest caldera at Emi Koussi was dated at 1.3–1.4 Ma, while Pic Toussidé is necessarily younger than 0.43 Ma as it is intersecting the Yirrigué caldera (Fig. 6). The only possibly Late Quaternary activity at Emi Koussi is represented by the very fresh basaltic scoria cones and lava flows of its late Strombolian volcanoes scattered both inside and outside the caldera.

Valuable information can certainly be extracted from satellite images in the survey of an unknown and/or inaccessible area. However, field data, when available, still remain the most productive means for determining the history of a volcanic province, as illustrated here. The old field data, obtained at the end of the 1950s and during the 1960s, supplied valuable information regarding age relations among the volcanoes of the TVP which should not be dismissed. The sequence of events discussed above, including new geochronological data along with new field observations on Western Tibesti and Emi Koussi areas, is fully consistent with these old field data. However, some chronological relationships (Figs. 5, 6, and 10) and volcanostructural features (Fig. 12) are beautifully illustrated by images from the satellite-borne SPOT sensors.

Tectonic control on volcanism location and features

Although it is beyond the scope of this paper to fully discuss the relationships between tectonic and magmatism in Tibesti, the tectonic control on the location of volcanism and of some of its features must be outlined here.

The TVP is located at the intersection of two major tectonic structures: the great NW-SE "Tassilian flexure" and the NE to NNE fault zone, delimiting the plateau to the SW and E, respectively (Fig. 2). As discussed above, both structural trends are probably related to the intracontinental transpressive Murzukian episode which occurred during the waning stages of the Pan-African orogeny, at about 575–555 Ma (Fezaa et al. 2010; Liégeois et al. 2013).

The NW direction was already considered as "predominant in the uplifts of the Tibesti and Ennedi" by Tilho (in Dalloni 1934). The NW-SE "Tassilian flexure" actually exhibits much similarity with the well-known Tassilian escarpments surrounding Hoggar, as discussed above. In Hoggar, most of the present contours of the Tassilian cliffs are located on tectonic structures corresponding to weakness lines of the basement, inherited from the precratonic geological history (e.g., Liégeois et al. 2013), that were variously reactivated, mainly during Caledonian and Hercynian events, but also during Cretaceous and Cenozoic times, including Quaternary (e.g., Rognon 1970; Beuf et al. 1971). It is suggested that the reactivation of old tectonic structures in Tibesti resulted in the development of the great NW-SE lithospheric flexure, mainly during Hercynian times.

Abundant field data indicate that the NE and NNE tectonic structures are much more frequent in Chad, Southern Libya, and Eastern Niger (e.g., Jacquemont et al. 1959; Jacqué 1962; Vincent 1963a, 1970; Klitzsch 1966a, b; Goudarzi 1980; Woller and Fediuk 1980; Ghienne et al. 2013 and references therein) than the NW-SE direction. In Tibesti, this major Precambrian direction was reactivated in Hercynian times, after the NW-SE flexure development, as testified by the Yebbigué fault following the Precambrian T1/T2 contact in the north and intersecting the NW-SE flexure in the south. NE trending faults on the plateau were also reactivated during Jurassic-Cretaceous times (Vincent 1963b) and probably even more recently. NE-NNE faults were still active in Tibesti after the Lutetian, at least until the Mio-Pliocene.

These two structural directions not only delineate the TVP but, further, control some volcanic features inside the province itself, providing evidence for their Cenozoic reactivation. Indeed, the basaltic plateau is made of a succession of overlapping NW-SE trending traps. Besides, the feeding dike swarms and the associated silicic plugs are located along the crest of small NW-SE flexures where lava flows reach their maximum thickness. This indicates that the NW trending faults controlled the emission of the older plateau lavas during the Miocene. In addition, some younger large central volcanoes such as Toon, Oyé (or Oyoye) and Yéga are aligned along the southern extension of the great NNE trending Yebbigué fault, outlining that this major fracture was also reactivated during the Mio-Pliocene.

Interestingly, the younger ignimbritic (ash-flow) volcanoes, are located away from major NNE regional faults but, instead, very close to the great NW-SE flexure (Fig. 1). Specific relationships are observed between these high-elevation volcanoes and their basement. The thickness of eruptive products at Voon, Yirrigué and Emi Koussi, as well as the location of the last two on local horsts, is insufficient to account for the observed elevation of these volcanoes. Additional doming of the basement prior to the ignimbrite emplacement is required to account for the remaining elevation. Settings on horsts with significant basement uplift seem to favor the accumulation of large volumes of magma in upper crustal magma chambers, resulting itself in regional updoming of the magma chamber roof before any eruption.

The localization of volcanism in Tibesti thus appears strongly controlled by lithospheric architecture, as for nearby Libyan, Hoggar, or Aïr volcanic fields (e.g., Goudarzi 1980; Woller and Fediuk 1980; Schäfer et al. 1980; Liégeois et al. 2005; Azzouni-Sekkal et al. 2007). Reactivation of lithospheric megastructures in relation to the Africa-Europe convergence and collision certainly played a major role in the formation of the TVP, as for Libyan and Hoggar volcanic fields. However, the possible relationships between the migration of magmatism in the TVP during plateau volcanism and the NE motion of the African plate cannot be fully evaluated as yet because of the lack of geochronological data.

Conclusions

The volcanic units of the Cenozoic Tibesti Volcanic Province (Chad) have been described and their chronology established, using both field relationships and some ^{39}Ar-^{40}Ar radiometric age constraints. The localization of the volcanism was strongly controlled by NW and NNE trending old lithospheric linear features.

Despite its large size and famous reputation as "hospot"-related, the TVP is still poorly documented. However, some specific features of this volcanism, although not yet fully understood, are worth outlining here.

1. Large volumes of ignimbritic products are encountered in the first and third stages of volcanic activity. Namely, alkaline and peralkaline rhyolitic ignimbritic sheets are intercalated in the upper part of plateau basalts and become more abundant upwards. Furthermore, there are three large ignimbritic volcanoes (Emi Koussi, Voon, and Yirrigué). These large volumes of ignimbritic products are a specificity of Tibesti, compared to neighboring Hoggar and Libyan provinces. There is no evidence of the existence of large ignimbritic edifices in these regions. Admittedly, ignimbritic products might have been largely removed by erosion owing to the older age of earlier volcanism there.
2. Alkaline plateau volcanism built by successive eruptive pulses separated by long repose intervals and tholeiitic central volcanoes, with megaporphyritic and aphyric lavas, are very similar to the classical Paleogene "Plateau" and "Central" volcanism in the Hebrides. In both areas, the location of magmatism and the orientation of the associated structures were largely controlled by tectonic features in the basement. However, a significant difference between the central volcanoes of these two provinces resides in the large diversity and originality of calderas, domes, and intrusions observed in Tibesti and their absence in the Hebrides. Indeed, despite clear evidence of surface volcanism in the Hebrides, only the eroded roots of central volcanoes were preserved after the erosion of a few thousand meters of material.

The exceptional conditions of exposure and uplift in Tibesti, combined with various degrees of erosion, clearly offer a unique opportunity of studying the structure of volcanoes at variable stages of their edification. This is the case for central volcanoes but also for some ignimbritic volcanoes, as illustrated by the Kilehégé ignimbrite feeder dikes. Unfortunately, this large and complex intracontinental volcanic province, although containing valuable keys for the understanding of other volcanoes worldwide, still remains very poorly known, as are the volcanic fields of Libya.

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Glossary

Ehi	Summit, peak
Enneri	Dry valley, occasionally with water, as "wadi"
Tarso	High plateau
Toussidé	"Which killed the local people (the Tou) by fire". ("Toubous" in French or "Tubu" in English is the name of the local population)

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