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Structural control on present-day topography of a basement massif: the Central and Eastern Anti-Atlas (Morocco)

J. GUIMERÀ [1] [*] M.L. ARBOLEYA [2] A. TEIXELL [2]

[1] **Departament de Geodinàmica i Geofísica, Facultat de Geologia, Universitat de Barcelona (UB)**
Martí i Franqués s.n. 08028 Barcelona, Spain. E-mail: joan.guimera@ub.edu

[2] **Departament de Geologia, Facultat de Ciències, Universitat Autònoma de Barcelona**
08193 Bellaterra, Spain

* Corresponding author

ABSTRACT

The Anti-Atlas basement massif extends South of the High Atlas, and, despite a very mild Cenozoic deformation, its altitude exceeds 1500m in large areas, reaching 3305m in Jbel Sirwa. Structural contours of the present elevation of a polygenic planation surface (the High Erosional surface) and of the base of Cretaceous and Neogene inliers have been performed to characterize the major tectonic structures. Gentle Cenozoic WSW-ENE- and N-S-trending folds, of 60 to100km wavelength, reactivate Variscan structures, being the major contributors to the local topography of the Anti-Atlas. Reactivated thrusts of decakilometric to kilometric-scale and E-W trend involving the Neogene rocks exhibit a steep attitude and a small displacement, but they also produce a marked topographic expression. The resulting Cenozoic horizontal shortening along N-S sections across the Anti-Atlas is about 1%. The position of the major anticlinal hinges determines the location of the fluvial divides of the Warzazat basin and the Anti-Atlas, and a structural depression on one of these hinges (Jbel Saghro anticline) allowed the formerly endorheic Warzazat basin to drain southwards. The first Cenozoic structures generating local topography are of pre-mid Miocene age (postdated by 6.7Ma volcanic rocks at the Jbel Saghro), whereas the youngest thrust movements postdate the Pliocene sedimentary and volcanic rocks (involving 2.1Ma volcanic rocks at Jbel Sirwa). In addition to these features, the mean elevation of the Anti-Atlas at the regional scale is also the result of a mantle thermal anomaly reported in previous works for the entire Atlas system.

KEYWORDS **Cenozoic contraction. Crustal folding. Uplift. Tectonic topography. Anti-Atlas. Morocco.**

INTRODUCTION

The Anti-Atlas of Morocco is a mountainous region located south of the Warzazat and Sous basins, which are the southern foreland basins of the High Atlas (Fig. 1). The present topographic elevation of the Anti-Atlas is over 1500m in extensive areas, exceeding 2000m in several places (Fig. 2): 3305m in the Jbel Sirwa, 2701m in the Amalou-n-Mançour (Jbel Saghro), and 2526m in the Imgand (Central Anti-Atlas). Being dominated by Proterozoic and Palaeozoic rocks, the Anti-Atlas contains occurrences of marine Cretaceous sediments indicating that such elevations are the signal of a Cenozoic uplift. However, much of the internal structure of the Anti-Atlas was formed during the

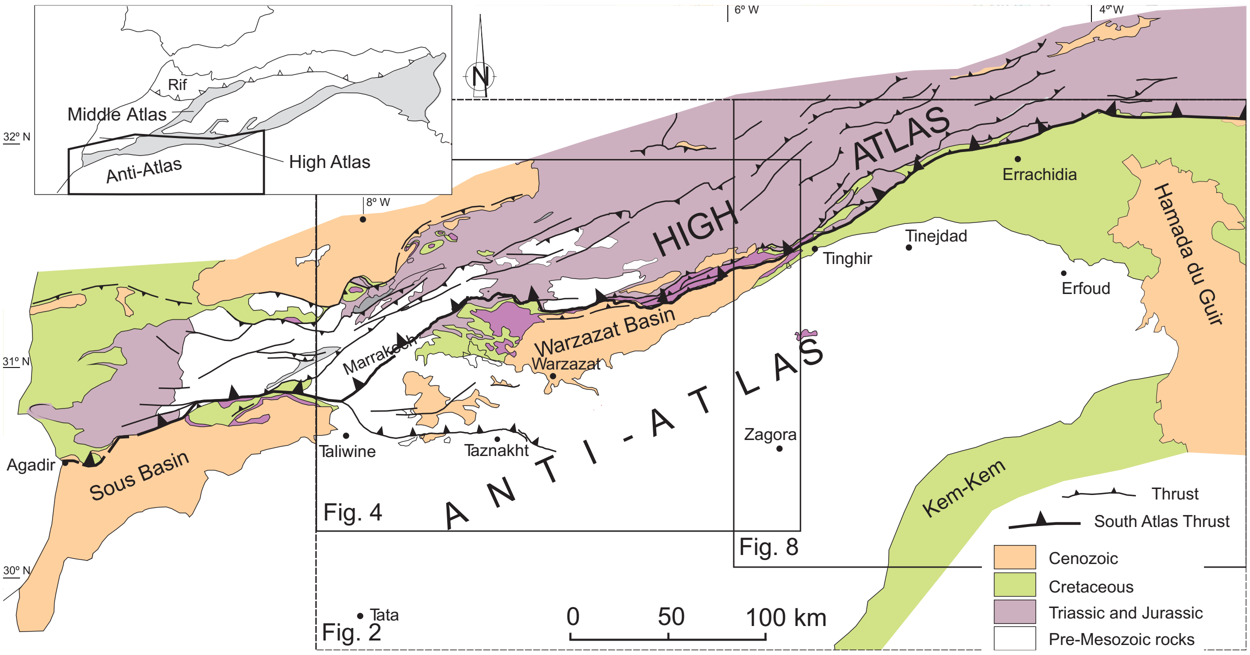

FIGURE 1 | **Geological map of the Anti-Atlas and the Central and Western High Atlas, simplified after Hollard (1985). Location of Figs. 2, 5 and 8 is shown.**

Panafrican and Variscan orogenies (late Proterozoic and late Palaeozoic, respectively) while the Cenozoic deformation is considered to be very mild (Choubert & Marçais, 1952; Hoepffner et al., 2005).

The high topography of the High Atlas thrust belt is the combination of crustal shortening and long-wavelength thinning of the lithosphere, affecting the mountain belt and the peripheral plains on a much wider scale (Teixell et al., 2003, 2005; Missenard et al., 2006). The differential topography between the High Atlas and the adjacent elevated plains results from tectonic shortening, manifested by well developed imbricate thrust systems. This is not so straightforward in the Anti-Atlas, where these systems are lacking. Choubert (1952) described the overall structure of the Anti-Atlas as a vast basement arch (*pli de fond*), which resulted from the remobilization of a previous fold of Variscan age, a view followed by Teixell et al. (2003), who interpreted it as a lithospheric-scale fold. A closer inspection evidences that topography is not homogeneous within the Anti-Atlas (Fig. 2) and, as pointed out by Choubert (1952), major faults do involve Mesozoic and Cenozoic rocks. The aim of this paper is to analyze the tectonic origin of topography on the basis of 1) the attitude of several surface markers that can be reconstructed from preserved erosional features and sedimentary inliers within the Anti-Atlas, and 2) the field analysis of the contractional structures that involve Neogene rocks and morphological surfaces. In any case, Post-paleozoic rocks are rare in the Anti-Atlas, and this study illustrates how a geomorphological analysis can provide grounds to the tectonic interpretation in regions where the sedimentary record is scarce.

GEOLOGICAL SETTING

Geologically, the Anti-Atlas is composed of a Proterozoic Pan-African basement overlain by post-Pan-African Uppermost Proterozoic and Palaeozoic rocks, deformed during the Variscan orogeny (Choubert, 1952; Choubert & Marçais, 1952). The Variscan deformation caused in the inversion of late Proterozoic and Palaeozoic intracratonic basins, remelting in a thick-skinned fold and-thrust-belt (Burkhard et al., 2006; Toto et al., 2008; Guimerà & Arboleya, 2008).

Directly overlying Precambrian and Palaeozoic rocks, a Cretaceous succession starts with pre-Cenomanian fluvial red beds, followed by Cenomanian-Turonian marine limestones and red-beds of Senonian age (Choubert, 1952). The pre-Cenomanian red beds, markedly unconformable on the pre-Mesozoic basement, cover a peneplain (Robert-Charrue, 2006) and are preserved in isolated outcrops, essentially on the northern and eastern side of the Anti-Atlas (Choubert, 1950; Choubert, 1955-56, Ferrandini et al., 1985) (Fig. 1). These pre-Cenomanian clastic sediments derive from the Sahara region, and were deposited by alluvial systems, whose regional slope was north-directed

(Guillocheau et al., 2007). The Cenozoic is largely confined to the Warzazat and Sous basins (Fig. 1), where it begins with Palaeocene to Middle Eocene marine limestones, followed by continental Eocene red beds (Choubert, 1950; Choubert, 1955-56; Gauthier, 1957). After a marked hiatus, the Aït Kandoula Formation (El Harfi et al., 2001) is composed by alluvial and lacustrine terrigenous and carbonate sediments of Middle to Upper Miocene age (Tesón, 2009). The Aït Kandoula Fm. constitutes the bulk of the infill of the Warzazat basin (ca. 1000m), and crops out in isolated exposures in the Sirwa massif, the western part of the Sous basin, and within the Anti-Atlas (Choubert, 1945; Gauthier, 1957; Görler et al., 1988), where it constitutes a valuable marker for Neogene deformation. It has not been documented whether the Anti-Atlas was ever completely covered by Mesozoic or Palaeogene sediments. Previous authors have assumed that most of it was exposed during these epochs (Choubert, 1952; Riser, 1988; Robert-Charrue, 2006; Malusà et al., 2007).

A characteristic feature of the Anti-Atlas is a widespread planation surface called the High Erosional Surface (HES) or Pre-hamadian Surface by Choubert, (1952). This author recognized this surface as polygenic in origin and attributed its development to an indeterminate period from the beginning of the Tertiary to the Aquitanian. Moreover, we have observed that North of the Sirwa region the HES derives from the exhumation of the pre-Cretaceous peneplain (Fig. 3 A). What can be ascertained is that this HES predates the sedimentation of the Aït Kandoula Fm. (mid-Miocene), which is either deposited on top of the HES (Sirwa massif and Draa valley; Fig. 3 A, B and C) or confined within paleovalleys downcut into it (northern part of the Anti-Atlas; Choubert, 1952; Görler et al., 1988; Fig. 3 C and D). Regardless of its origin, we have mapped this remarkably flat surface -which Choubert (1952) defined as a peneplain- as a single entity. This generalized planation surface is well preserved in large areas of the Anti-Atlas, over Precambrian and Palaeozoic rocks (Fig. 3 A, B, C and F), its presence implying a low relief region before the recent uplift of the Anti-Atlas.

Late Miocene to Pliocene phonolitic and trachytic volcanism (11 to 2Ma; Berrahma and Delaloye, 1989) occurred extensively in the Jbel Sirwa, where the remains of a major volcano reach 3305m. Miocene to Pliocene (10 to 2.8Ma) nephelinites, phonolites and alkaline basalts are found in the Jbel Saghro too (Berrahma et al., 1993).

Mesozoic and Neogene rocks around and inside the Anti-Atlas are cut by faults of decakilometric- to kilometric-scale and dominant E-W strike (Fig. 3 B and E, Fig. 4). The main fault is the Anti-Atlas Major Fault, inherited from the Panafrican orogeny, which vertically offsets the Neogene rocks some 500m (Choubert & Marçais, 1952). This fault dips steeply to the North and had a reverse slip during the Neogene (Missenard, 2006; Sébrier et al., 2006; Guimerà et al., 2006 and 2008) (Fig. 3 B; Fig. 5, section 2).

METHOD OF STRUCTURAL ANALYSIS

In view of the scarcity of post-Palaeozoic deformation markers, the method of analysis of the structures responsible for the uplift of the Anti-Atlas consists of structural contour mapping of reference surfaces. More than 900 points of elevation measurements were acquired from SRTM90 DEM (about 90m of pixel size, available at http://srtm.csi.cgiar.org) and our own field-based geological mapping on satellite orthoimages. After this, we have drawn georeferenced contour maps of the present-day elevation of the sub-Cretaceous unconformity, the HES and the base of the Neogene, using *Surfer* software with the Kriging gridding method. The results of contouring are presented in Figs. 4, 6 and 8.

For the sub-Cretaceous unconformity and the base of the Neogene, elevation measurements were acquired from geological maps. For the HES, elevation measurements were taken, after the recognition of this surface in the field (Fig. 3 A, B, C and F), from 3D visualizations of satellite orthoimages draped on the SRTM90 DEM.

Taking into account that the Aït Kandoula Fm. and the Neogene volcanic rocks are found both on top of the HES and filling paleovalleys downcut into it, the present differences in elevation of their base is the result of both the Neogene uplift and the filling of a paleorelief. This must be taken into account in the structural interpretation.

We have constructed generalized geological cross-sections to illustrate the attitude of the contoured markers and to define the Cenozoic deformation structure (Fig. 5).

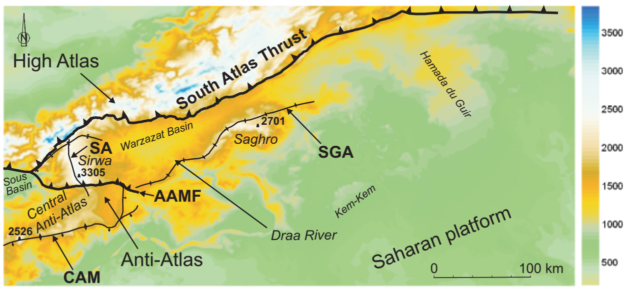

FIGURE 2 | **STRM90 digital elevation model of the area studied. The main geographic and tectonic features are shown. AAMF: Anti-Atlas Major Fault; CAM: Central Anti-Atlas Monocline; SA: Sirwa Anticline; SGA: Saghro Anticline. For location, see Fig. 1.**

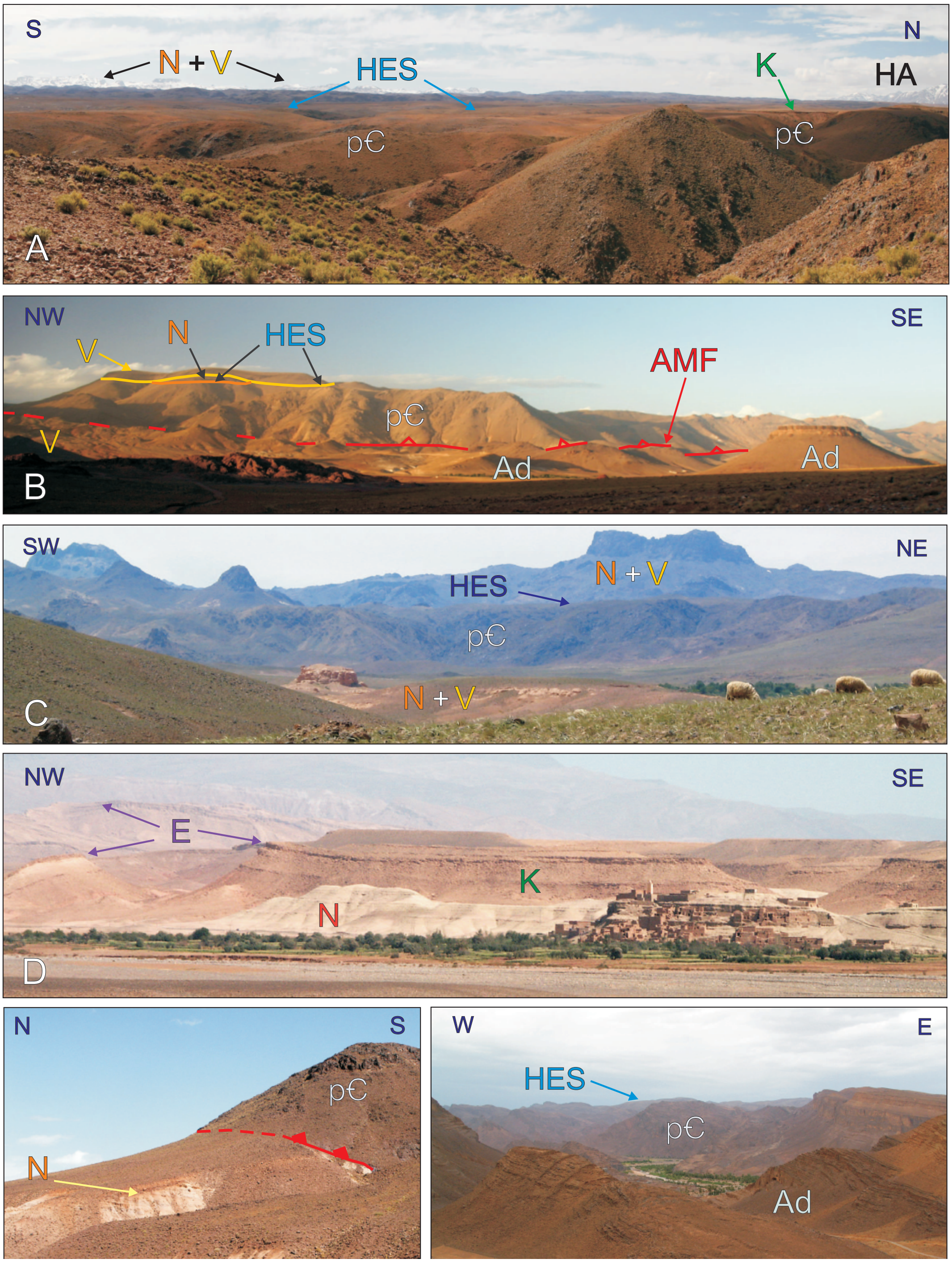

FIGURE 3 | **Field images of the area studied. p€: Precambrian. Ad: Adoudounian (latest Proterozoic). K: Cretaceous sediments. E: Eocene. N: Neogene sediments. V: Neogene volcanic rocks. HES: High Erosional Surface. AMF: Anti-Atlas Major Fault. HA: High Atlas. For location, see Fig. 4. A) View of the High Erosional Surface (HES) in the northeast Sirwa area, where it derives from the exhumation of a pre-Cretaceous peneplain. The erosion of the Cretaceous rocks uncovered a very flat erosional surface, which became a part of the HES. To the West, this surface is covered by the Neogene and volcanic rocks of the Sirwa (see Fig. 3 B). B) The Anti-Atlas Major Fault South of the Sirwa region. Neogene sedimentary and volcanic rocks have a vertical offset of 500m related to the reverse slip of the fault. In the uplifted hanging wall, the High Erosional Surface is preserved beneath Neogene rocks. C) East limb of the Sirwa NNW-SSE anticline. In the foreground, interbedded Neogene sedimentary and volcanic rocks fill a paleovalley downcut in the Precambrian rocks (in the middleground). On top of these Precambrian rocks, the HES is preserved and, over it, in the background, Neogene sedimentary and volcanic rocks, younger than the previous ones, lie on the HES. D) Western end of the Warzazat basin. Horizontal Neogene sedimentary rocks (Aït Kandoula Fm.) are entrenched in folded Eocene and Upper Cretaceous rocks. The town is Tizgzaouine. The High Atlas appears in the background. E) South boundary of the Hmam basin, inside the Anti-Atlas. Neogene sedimentary rocks are tilted 5º 9 to the S and thrusted by Precambrian rocks. F) The HES developed on top of Precambrian rocks around the area where the Draa river crosses the Anti-Atlas relief. The Draa, flowing south, is downcut about 200m.**

CENOZOIC STRUCTURE AND RELIEF

The combination of field observations and the analysis of the contour maps have enabled us to describe the major Cenozoic structural elements of the Central and Eastern Anti-Atlas. In addition to previously reported features, such as the Anti-Atlas Major Fault, we have identified a set of large-scale folds, huge tilted blocks, and major geomorphic features.

Main folds

The Saghro anticline

South of the Warzazat basin, the contoured surfaces define a large-scale WSW-ENE-trending major anticline which coincides with the relief of the Jbel Saghro (Figs. 4 and 5A). The northern limb of the anticline is defined by the NNW-directed slope of the HES and the base of Neogene North of the Jbel Saghro (Fig. 4), whereas the southern limb is evidenced by the S-directed slope of the HES (2400 to 1600m) to the SE of the Jbel Saghro (Fig. 4). Therefore, both limbs are well defined and the fold hinge is located around the ridge crest of the Jbel Saghro. In addition, two isolated buttes (called N and S Tourt) of Neogene lacustrine limestones and marls are preserved almost 30km south of the fold hinge, at 1100m in altitude (easternmost side of Fig. 4) (Choubert, 1959); according to Riser (1988), these rocks are Miocene in age. Near the base of the South butte, we have found a thin level (0.5m) of interbedded, very altered volcanic rocks. The Saghro anticline is 200 km-long and has a 70km-wavelength. Its gentle attitude (4° dip of its southern limb: 800m in 11.4km), coinciding with the tighter Variscan Sagrho anticline of Choubert (1952) and representing, therefore, a reactivation of the late Palaeozoic structure.

The hinge of the Cenozoic fold of the Saghro anticline plunges to the ENE, as deduced from the slope of the HES in the eastern termination of the Jbel Saghro (Figs. 4 and 5 C). The continuity of the Palaeozoic succession in the southern limb of the Saghro anticline, well established in geological maps (Choubert, 1959; Faure-Muret and

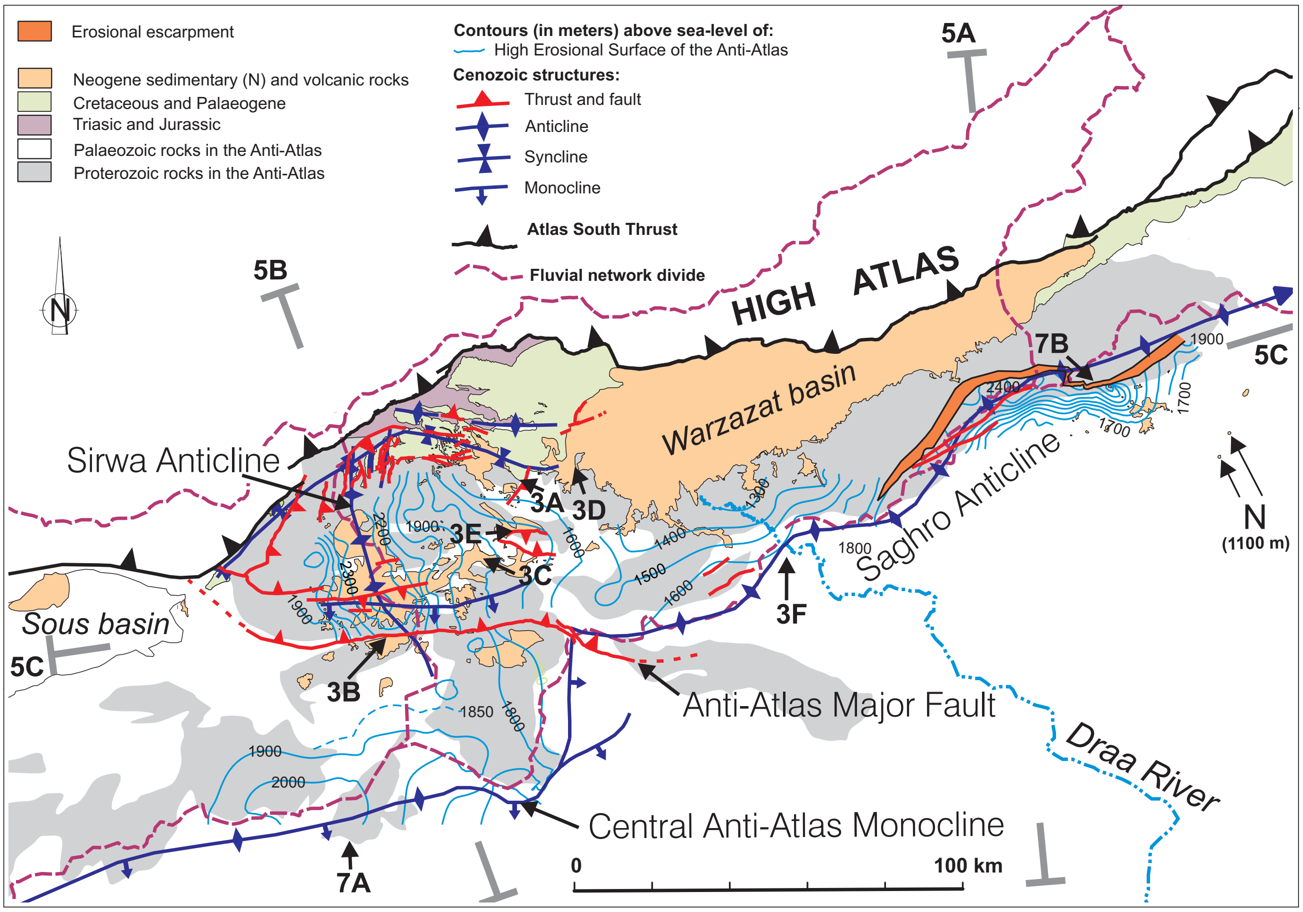

FIGURE 4 | **Map displaying the major Cenozoic structures of the study region. Also, the contours of the High Erosional Surface are displayed. Geological contacts are taken from Hollard (1985). N: Neogene. The location of pictures of Figs. 3 and 7 and of sections of Fig. 5 is shown. For location, see Fig. 1.**

Choubert, 1974-77), indicates that no major thrusts are exposed at the surface (Fig. 5 C).

The Sirwa anticline

In the Sirwa massif, the contours of all the reference surfaces define a broad NNW-SSE-oriented anticline, some 100km in wavelength, which coincides with the topographic high between the present-day Warzazat and Sous basins (Figs. 4 and 6). The fold amplitude, as defined by the contours of the HES (covered by the Neogene sedimentary and volcanic rocks), is close to 1000m. In the eastern limb, the HES and the base of the Neogene rise from 1300m in the western Warzazat basin, to 2300m in the fold hinge (Figs. 4 and 6). In the western limb, the HES drops from 2300m to 1800m, defining a west-directed slope. Moreover, 30km to the west of the last Neogene outcrop of the Sirwa massif, one small outcrop of Neogene sedimentary rocks is preserved at an elevation of 725m (Figs. 5C and 6) (Choubert, 1955-56). This outcrop is actually the eastern tip of the Sous basin, and shows lacustrine limestones dipping 5º-10º to the W. The 30km-wide area with no Neogene rocks preserved is the result of the recent erosion by the deeply incised (reaching 1000m of incision) Assif-n-Tfnout valley.

Cretaceous rocks north of the Sirwa massif are also folded with a similar pattern (Fig. 6). At a smaller scale, these rocks are also affected by hectometric- to kilometric-scale thrusts and folds oriented N-S and E-W (Fig. 6).

Central Anti-Atlas monocline

A gentle WSW-ENE-trending and SSE-vergent monocline coincides with the south-facing topographic step of the Central Anti-Atlas (Figs., 4, 5 B and 7 A). As in the Jbel Saghro anticline, Palaeozoic succession is continuous across the monocline, being not disrupted at the surface by any major thrust (Choubert, 1950-56; Choubert, 1959).

Main thrusts

Thrusts of arguable Cenozoic age are scarce in the Anti-Atlas. The previously described Anti-Atlas Major Fault,

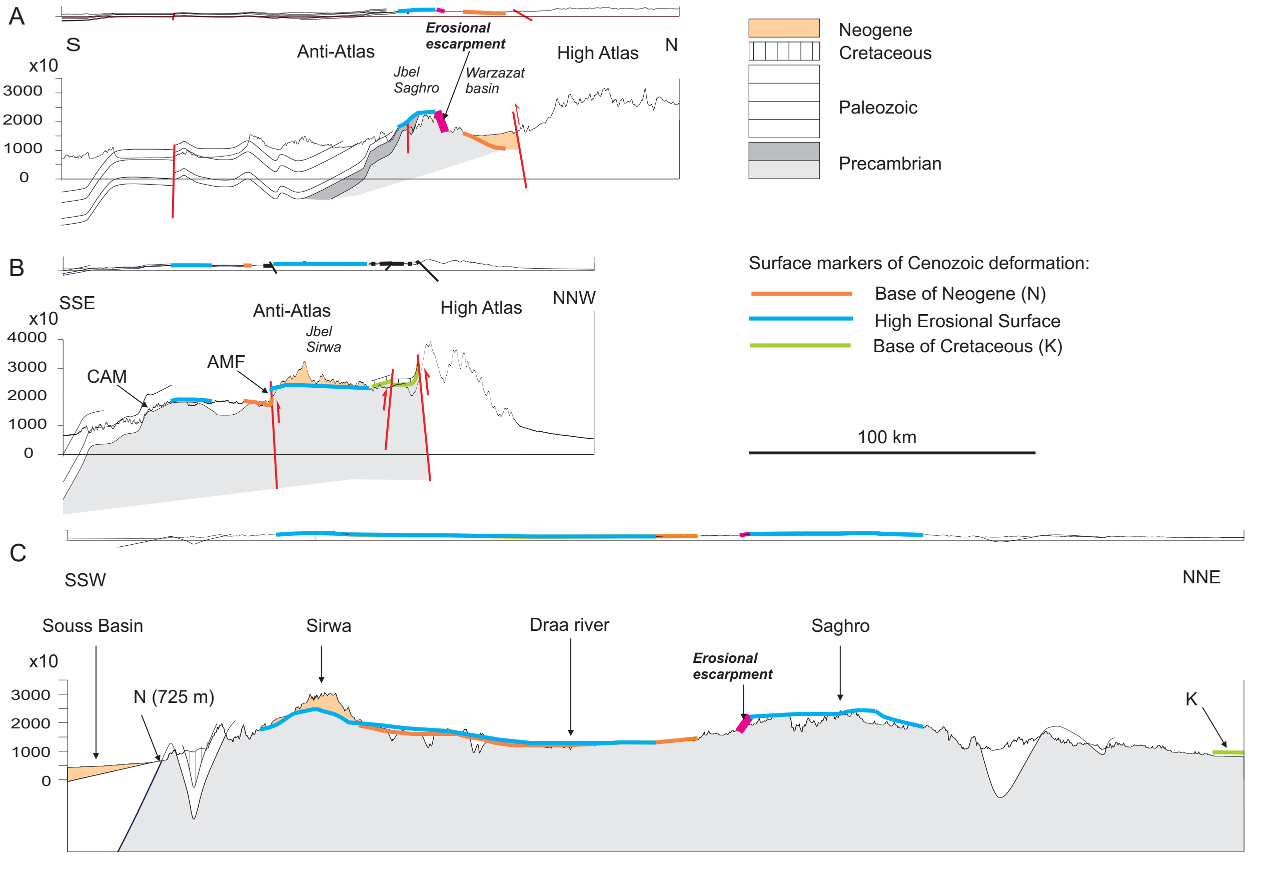

FIGURE 5 | **Geological cross-sections of the Central and Eastern Anti-Atlas. See Figs. 4 and 8 for location. Above is the section at the real scale, and below, ten times vertically exaggerated sections are shown (dips are distorted according to the exaggeration applied). In section A, the base of the Neogene was taken from Faure-Muret et al. (1994). K: Cretaceous sediments, QB: Quaternary basalts. AMF: Anti-Atlas Major Fault. CAM: Central Anti-Atlas Monocline.**

a South-vergent thrust bordering the Sirwa massif, appears clearly expressed by a step of some 500m in the Neogene contours (Figs. 4, 5 C and 6). This step is still well expressed in the present topography (Fig. 3 B), as reported by Choubert (1952) and Missenard (2006). Other thrusts in the area, smaller in dimensions, are the north-directed Hmam thrust, which bounds to the south the Neogene Hmam basin (Fig. 3 E and Fig. 6) and the Qal'at thrust, also north-directed, which cuts the Sirwa volcano (Fig. 6). The trace of these is 15 to 20km long, and their vertical slip is of the order of tens of metres. The dip of the thrust surfaces is 65-70º to the S in both cases. The actual dip of the Anti-Atlas Major Fault is difficult to determine, but it also appears to be very steep. Therefore the shortening associated to all these structures is very limited. Volcanic rocks of the Sirwa massif as young as 2.1Ma (Berrahma and Delaloye, 1989) are affected by the reverse slip of the Anti-Atlas Major Fault.

The Jbel Saghro escarpment

On the northern slope of the Jbel Saghro, the HES is truncated by a large North-facing erosional escarpment up to 700m-high and over 75km-long with a NE-SW trend (Fig. 4, 5 A and 7 B). Its development is previous to the volcanic rocks mapped by Faure-Muret & Choubert, (1974-77) at its foot, at Ikniwn (2000m), and its top, at the Amalou-n-Mançour mountain (2700m) (Fig. 7 B). These volcanic rocks have a late Miocene K-Ar radiometric age of 6.7Ma and 6.6Ma respectively (Berrahma et al., 1993). This erosional feature marks the southern outcrop boundary of the Aït Kandoula Fm. in the Warzazat basin (Figs. 4 and 5 A). The occurrence of coarse grained conglomerates of Anti-Atlas rhyolites and andesites in the southern Warzazat basin (Gauthier, 1957; Schmidt, 1992) is consistent with the existence of a marked relief in the Saghro area and with the development of the erosional escarpment.

The eastern end of the Anti-Atlas and the Hamada du Guir

Cretaceous rocks north and south of the eastern termination of the Anti-Atlas define a large and very gentle anticline, the limbs of which dip 2º-3º to the N and S.

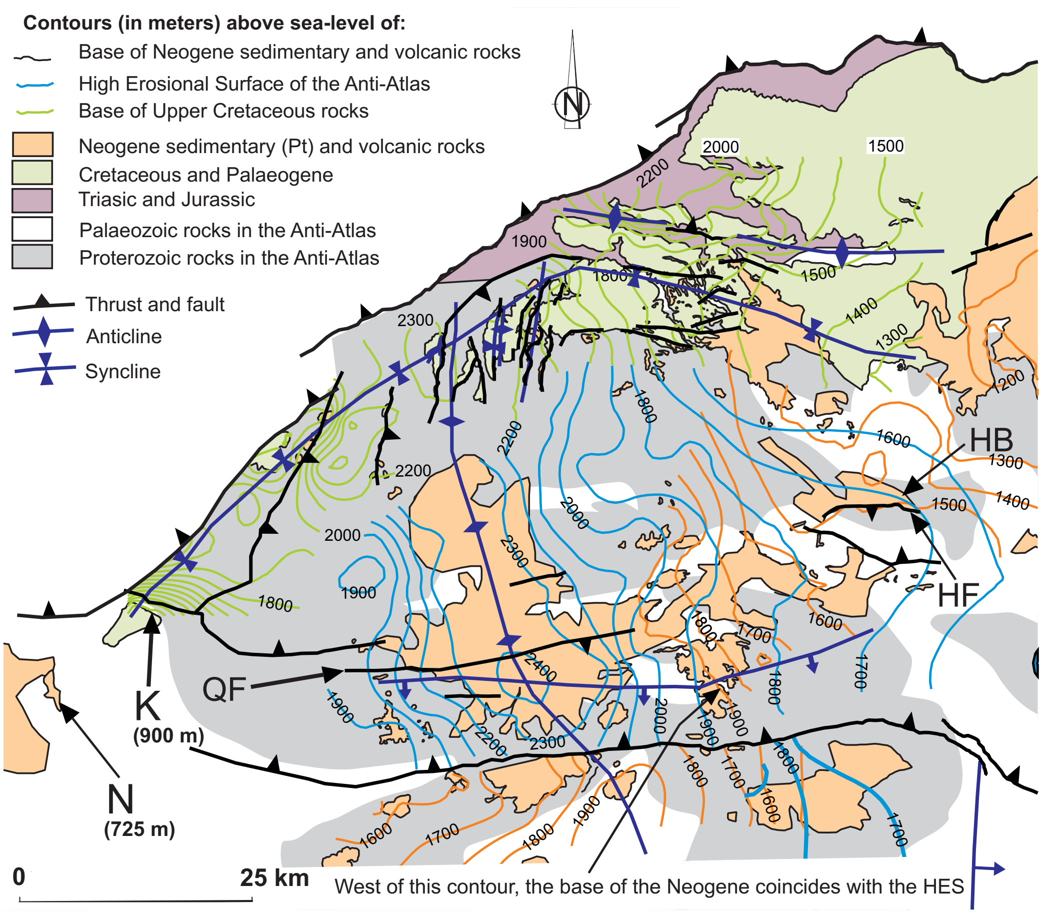

FIGURE 6 | **Map of the Sirwa massif and surrounding areas. Contours of the High Erosional Surface, the base of Cretaceous and the base of Neogene (in the areas where it is entrenched into the HES) are displayed. Geological contacts are taken from Hollard (1985). K: Cretaceous; N: Neogene; HB: Hmam basin; HF: Hmam Fault. QF: Qal'at Fault. For location, compare to Fig. 4.**

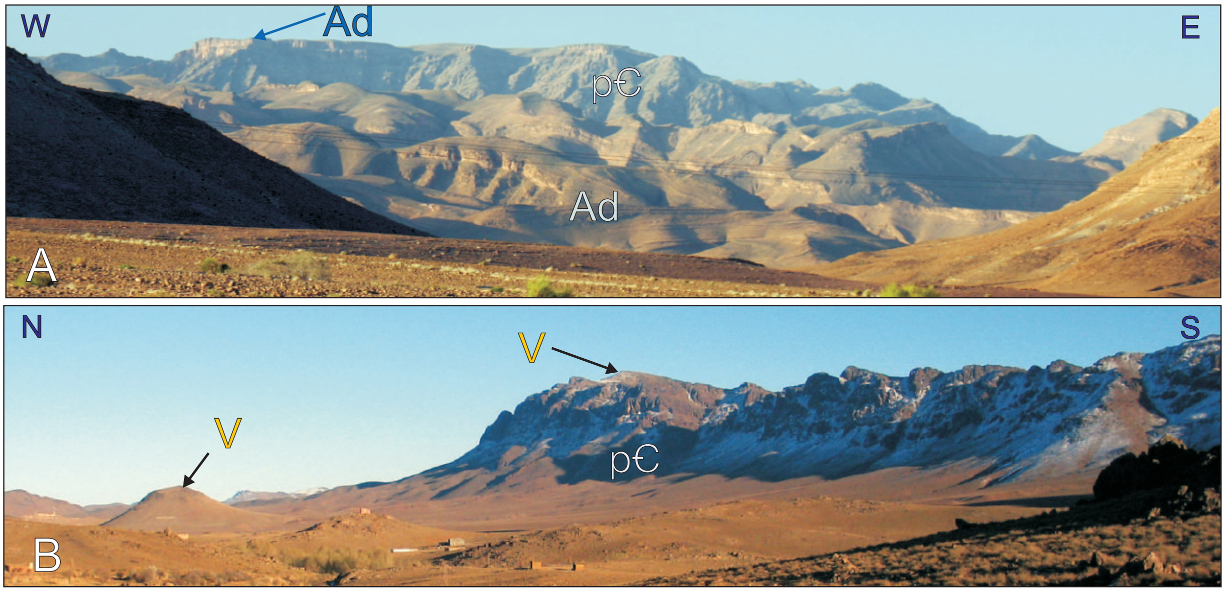

FIGURE 7 | **Field images of the area studied. p€: Precambrian. Ad: Adoudounian (latest Proterozoic). V: Neogene volcanic rocks. For location, see Fig. 4. A) The Central Anti-Atlas Monocline. In the background, Adoudounian rocks lie horizontal onto the Panafrican basement rocks. In the middleground, slightly folded Adoudounian rocks show a mean dip towards the observer, constituting the monocline limb. The tectonic relief produced by the monocline is 600m. B) North-facing erosional escarpment of the northern slopes of the Jbel Saghro, as seen from the NW. Miocene volcanic rocks lie on top of the escarpment (in the Amalou-n-Mançour mountain) and at its base (in the village of Ikniwn, shown in the photograph). See Figs. 5 and 6 A for location of the escarpment**.

Moreover, in the southern limb, this dip is expressed by the structural contours (Fig. 8).

The contouring of the Neogene strata of the Hamada du Guir evidences a continuous slope descending to the S, from 1100m to below 600m (Fig. 8). Consequently, the Neogene of the Hamada is not affected by the gentle anticline previously described.

INTERPRETATION AND DISCUSSION

The dimensions of the Cenozoic folds of the Anti-Atlas (hundreds of kilometres in length, and 60-100km in wavelength), suggest that they may involve the whole crust or most of it. The Cenozoic folding of the Saghro anticline and, possibly the Central Anti-Atlas Monocline, actually represent the reactivation or tightening of pre-existing Variscan anticlines, as originally proposed by Choubert (1952). The origin of the Sirwa N-S anticline is more debatable. Missenard et al. (2006) interpreted the difference in elevation between the basement of the Sirwa volcano (where Proterozoic rocks exceed 2300m) and the Warzazat basin floor as the result of the combination of a mantle thermal anomaly and a hypothetical mid-crustal magma injection. However, the wavelength of the lithospheric thinning beneath the Atlas domain (about 400km according to Teixell et al., 2005 and Fullea et al., 2007) exceeds the extent of the Sirwa massif and the western Warzazat basin, so we discard the mantle anomaly as responsible for the difference in elevation between both units. The available data do not allow to ascertain whether the fold is certainly due to a magmatic chamber or, alternatively, to a component of W-E (constrictional) shortening during the Cenozoic orogeny.

The steep attitude of thrusts involving Mesozoic or Cenozoic rocks within the Anti-Atlas is due to the fact that they reactivate older faults, probably post-Panafrican extensional faults. The history of these faults is long, as they were already reactivated as Variscan thrusts before the Cenozoic deformation (Burkhard et al., 2006; Toto et al., 2008; Guimerà and Arboleya, 2008).

The Cenozoic N-S horizontal shortening produced associated to the structures represented in cross-sections A and B (Fig. 5) is about 1% (E-W line-length shortening across the Sirwa antiform is of even smaller magnitude). This value was obtained by restoring the folds affecting the different surfaces of reference and comparing them to the present horizontal length of their trace in the cross-sections. Moreover, the heave (horizontal component of the slip, perpendicularly to the fault strike) of the thrusts was added to obtain the final horizontal shortening.

With regard to the topographic expression of the tectonic structures, it appears that anticlines give way to elevated areas, in contrast to less elevated regions in intervening synclines or structural depressions. This indicates that the local relief of the Anti-Atlas is strongly controlled by individual deformation structures, likely supported by the crustal strength, as suggested by their small amplitude in relation to their wavelength. The position of the major anticlinal hinges determines the location of the fluvial divides of the Warzazat basin and the Anti-Atlas (Fig. 4), as well as the present-day outlet of the Warzazat basin: the Draa river occupies a structural low of the Saghro anticline hinge, as demonstrated by the fact that the HES reaches 2500m at Jbel Saghro and 2300m at the Jbel Sirwa, while it is at 1400m in the intervening structural depression (Figs. 2, 4 and 5C).

In addition to the local topography, we must recall that the total elevation of the Anti-Atlas massif is also due to a mantle thermal anomaly put forward by different authors, on the basis of potential-field modelling (notably, of a large-scale positive geoid anomaly), the scarce orogenic thickening, and the widespread alkaline magmatism (Teixell et al., 2003, 2005; Missenard et al., 2006). A component of forebulge uplift in relation to the High Atlas thrust loads cannot be discarded, although its signal will be difficult to separate from that of the thermal uplift and of the local tectonic structures.

East of the Saghro anticline, the slope to the south of Cretaceous and Neogene and at the Hamada du Guir (Fig. 8) is not primary. In the case of the Cretaceous rocks, the regional slope of the alluvial systems during its sedimentation was to the North (Guillocheau et al., 2007), while in the case of the Neogene rocks, the limestone that

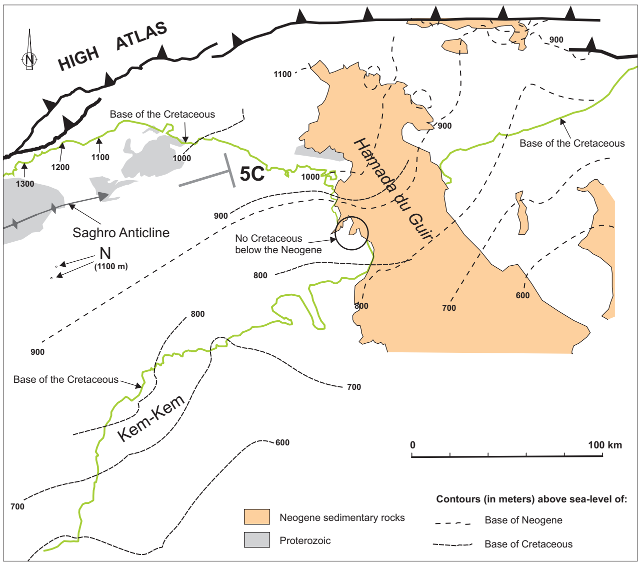

FIGURE 8 | **Map of the region east of the Jbel Saghro with contacts taken from Hollard (1985). The contours of the base of the Cretaceous and Neogene sediments are shown. N: Neogene. Location of cross-section in Fig. 6 C is shown. For location, see Fig. 1.**

caps the Hamada was deposited in a lacustrine environment (Lavocat, 1954; Thiry and Ben Brahim, 1997) defining a paleohorizontal (Babault et al., 2008). This southward slope is also the result of the thermal anomaly, which produced the elevation of a wide region extending from the Middle Atlas to the Anti-Atlas (Babault et al., 2008).

In the Sirwa anticline, two stages of growth can be deduced. The first one occurred prior to the deposit of the Aït Kandoula Fm. (Middle to Upper Miocene) and to the Neogene volcanism. This folding stage created a positive relief separating the Warzazat and Sous basins. Paleovalleys were entrenched in this relief, although we cannot tell neither their age nor whether they were remotely connected to the sea or formed part of an internally drained system. Later on these paleovalleys were filled up by the endorheic Aït Kandoula Fm. sediments until the latter overflowed the HES, communicating then the Warzazat basin with the eastern part of the present-day Sous basin. This is suggested by the continuity of Neogene outcrops described earlier (Fig. 6). This hypothesis necessitates a threshold separating the easternmost Sous basin from the areas to the west, should we follow Görler et al.'s (1988) view of the Neogene Warzazat basin as endorheic. The Lower Miocene erosion episode postulated by Missenard et al. (2008) and Balestrieri et al. (2009) in the Marrakech High Atlas after apatite fission tracks could be contemporaneous with this first stage of growth.

During the second folding stage, the continued growth of the Sirwa anticline deformed the Neogene, as deduced from the structural contour pattern (Fig. 6), thus disconnecting again the present-day Sous basin from the Warzazat basin, leading to the present configuration.

As for the Jbel Saghro anticline, constrains on its age can be obtained from the erosional escarpment described in its North face. This erosional feature implies that the relief of the Saghro anticline existed during the sedimentation of the Aït Kandoula Fm., which is accumulated at the escarpment foot. This, together with the presence of late Miocene (6.6-6.7Ma) volcanic rocks both on top and at the foot of the erosional escarpment indicates that the anticline and its subsequent erosion formed before the late Miocene, probably in early to mid Miocene times.

The existence of late Cretaceous limestones of similar facies on both sides of the Anti-Atlas and below the Hamada du Guir, dipping away from the Precambrian and Palaeozoic core (Fig. 8), indicates that the Cretaceous marine sediments once covered the majority, if not all, of the Anti-Atlas region. A Palaeocene-Eocene marine limestone, characteristic of the High Atlas and its forelands, gently truncates north-dipping Cretaceous sediments beneath the Warzazat basin and laps on the Precambrian basement of the basin (Tesón, 2009). This indicates that the Anti-Atlas was already a positive area in the Palaeogene and was not covered by these marine sediments. We cannot precise when the Saghro anticline developed a critical relief and was eroded into the escarpment, but in the adjacent High Atlas mountains, Tesón (2009) described a major hiatus marked by significant deformation and erosion from late Eocene to early-mid Miocene times.

On the other hand, the existence of Neogene isolated buttes at low elevation South of the Saghro massif suggests that they were probably in connection with the Warzazat basin around the periclinal termination of the Saghro anticline. The continuous Aït Kandoula exposure at the Hamada du Guir also points in this direction.

CONCLUSIONS

The Anti-Atlas basement massif is an example of a mountainous area where different spatial scales of topography result from different geodynamic processes. Mantle-related thermal uplift controlled the mean elevation at large wavelengths (hundreds of kilometres), while crustal-scale folding and thrusting controlled topography at the scale of tens of kilometres. Crustal structures have been revealed by structural contour mapping, consisting in very gentle Cenozoic folds of 60 to 100km-wavelength. These structures show a very strong correlation with the present-day topography. A structural depression along the hinge of one of these folds, the Saghro anticline, allowed the recent incision of the Draa river gorge, thus opening to the sea the formerly endorheic Warzazat basin. This study illustrates how the geomorphological analysis of surface markers can support the tectonic characterization, in particular in regions where the sedimentary record is scarce as the basement massif of the Anti-Atlas.

A first stage of growth of topography-forming structures (pre-mid Miocene) is evidenced by the erosional escarpment in the northern limb of the Jbel Saghro anticline (developed prior to the 6.7Ma-old volcanic rocks) and by the entrenchment of the fluvial network (later on filled by Neogene sediments) in the eastern limb of the Sirwa anticline. A second stage of growth postdates the Aït Kandoula Fm. and the Miocene to Pliocene volcanic rocks in the Sirwa area, where rocks as young as 2.1Ma are deformed by the Anti-Atlas Major Fault and other reverse faults.

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REFERENCES

Babault, J., Teixell A., Arboleya, M.L., Charroud, M., 2008. A late Cenozoic age for long-wavelength surface uplift of the Atlas Mountains of Morocco. Terra Nova, 20, 102-107.

Balestrieri, M.L., Moratti, G., Bigazzi, G., Algouti, A., 2009. Neogene exhumation of the Marrakech High Atlas (Morocco) recorded by apatite fission-track analysis. Terra Nova, 21, 75-82.

Berrahma, M., Delaloye, M., 1989. Données géochronologiques nouvelles sur le massif volcanique du Siroua (Anti-Atlas Morocco). Journal of African Earth Sciences, 9, 651-656.

Berrahma, M., Delaloye, M., Faure-Muret, A., Rachdi, H.E.N., 1993. Premières données géochronologiques sur le volcanisme alcalin du Jbel Saghro, Anti-Atlas, Maroc. Journal of African Earth Sciences, 17, 333-341.

Burkhard, M., Caritg, S., Helg, U., Robert-Charrue, C., Soulaimani, A., 2006. Tectonics of the Anti-Atlas of Morocco. Comptes Rendus Geoscience, 338, 11-24.

Choubert, G., 1945. Note préliminaire sur le Pontien au Maroc (Essai de synthèse orogénique du Maroc Atlasique). Bulletin de la Société Géologique de France, série 5, 16, 677-764.

Choubert, G., 1950. Carte géologique du Maroc 1:500 000, Feuille Hammada du Guir. Service Géologique du Maroc.

Choubert, G., 1952. Histoire Géologique du domaine de l'Anti-Atlas. Géologie du Maroc, Notes et Mémoires du Service Géologique du Maroc, 100, 77-194.

Choubert, G., 1955-56. Carte géologique du Maroc 1:500 000, Feuille Marrakech. Service Géologique du Maroc.

Choubert, G., 1959. Carte géologique du Maroc 1:500 000, Feuille Ouarzazate. Service Géologique du Maroc.

Choubert, G., Marçais, J., 1952. Le domaine de l'Anti-Atlas. Géologie du Maroc, fascicule I, Notes et Mémoires du Service Géologique du Maroc, 100, 13-27.

El Harfi, A., Lang, J., Salomon, J., Chellai, E.H., 2001. Cenozoic sedimentary dynamics of the Ouarzazate foreland basin (Central High Atlas Mountains, Morocco). International Journal of Earth Sciences, 90, 393-411.

Faure-Muret, A., Choubert, G. (eds.), 1974-77. Carte Géologique du Maroc 1:200 000, feuille 161: Jbel Saghro-Dadès. Notes et Mémoires du Service Géologique du Maroc, 161.

Faure-Muret, A., Morel, J.L., Dahmani, M. (eds.), 1994. Carte néotectonique du Maroc. Notes et Mémoires du Service Géologique du Maroc, 368.

Ferrandini, M., Philip, J., Babinot, J.-F., Ferrandini, J., Tronchetti, G., 1985. La plate-forme carbonatée du Cénomano-Turonien de la région d'Erfoud-Errachidia (Sud-Est Marocain. stratigraphie et paléoenvironnements). Bulletin de la Société Géologique de France, 8 (I 4), 559-564.

Fullea, J., Fernàndez, M., Zeyen, H., Vergés, J., 2007. A rapid method to map the crustal and lithospheric thickness using elevation, geoid anomaly and thermal analysis. Application to the Gibraltar Arc System, Atlas Mountains and adjacent zones. Tectonophysics, 430, 97-117.

Hoepfner, C., Soulaimani, A., Piqué, A., 2005. The Moroccan Hercynides. Journal of the African Earth Sciences, 43, 144-165.

Hollard, H. (ed.), 1985. Carte géologique du Maroc 1:1.000.000. Éditions du Service Géologique du Maroc.

Gauthier, H., 1957. Contribution à l'étude géologique des formations post-liasiques des bassins du Dadès et du Haut-Taudra (Maroc meridional). Comptes Rendus de l'Académie des sciences Paris, 236, 99-106.

Görler, K, Helmdach, F., Gaemers, P., Heissig, K., Hinsch, W., Maedler, K., Schwarzhans, W., Zucht, M., 1988. The uplift of the central High Atlas as deduced from Neogene continental sediments of the Ouarzazate province, Morocco. Berlin, Springer-Verlag, Lecture Notes in Earth Sciences, 15, 361-404.

Guillocheau, F., Rolland, N., Colin, J.P., Robin, C., Rouby, D., Tiercelin, J.J., Dauteuil, O., 2007. Palaeogeography of Africa through Meso-Cenozoic times: A focus on the continental domain evolution. Workshop TOPOAFRICA, Evolution of the African topography over the last 250 My. Rennes, November 13-16, 2007, Mémoire de Géosciences, Rennes, 7, 33-35. Website: Http://www.geosciences.univ-rennes1.fr/IMG/pdf/TopoAfrica.pdf

Guimerà, J., Arboleya, M.L., Missenard, Y., Leturmy, P., Frizon de Lamotte, D., Teixell, A., 2006. Tertiary structure of the Jbel Sirwa region (Anti-Atlas Mountains, Morocco). Geophysical Research Abstracts, 8, 07376.

Guimerà, J., Arboleya, M.L., 2008. Post-Panafrican late Proterozoic basins and their influence on the Variscan contractional structures (Central Anti-Atlas, Morocco). Geo-Temas, 10, 449-452.

Guimerà, J., Arboleya, M.L., Teixell, A., 2008. Structure and evolution of topography during the Cenozoic in the central and eastern Anti-Atlas (Morocco). Geo-Temas, 10, 361-364.

Lavocat, R., 1954. Reconnaissance géologique dans les hammadas des confins algéro-marocains du sud. Notes et Mémoires du Service Géologique du Maroc, 116.

Malusà, M.G., Polino, R., Feroni, A.C., Ellero, A., Ottria, G., Baidder, L., Musumeci, G., 2007. Post-Variscan tectonics in eastern Anti-Atlas (Morocco). Terra Nova, 19, 481-489.

Missenard, Y., 2006. Le relief des Atlas marocains: contribution des processus asthenospheriques et du raccourcissement crustal, aspects chronologiques. Doctoral Thesis. Cergy-Pontoise, Université de Cergy-Pontoise, 236pp.

Missenard, Y., Zeyen, H., Frizon de Lamotte, D., Leturmy, P., 2006. Crustal versus asthenospheric origin of relief of the Atlas Mountains of Morocco. Journal of Geophysical Research, 111, 1-13. B03401, doi:10.1029/2005JB003708.

Missenard, Y., Saddiqi, O., Barbarand, J., Leturmy, P., Ruiz, G., El Haimer, F.-Z., de Lamotte, D.F., 2008. Cenozoic denudation in the Marrakech High Atlas, Morocco: Insight from apatite fission-track thermochronology. Terra Nova, 20, 221-228.

Riser, J., 1988. Le Jbel Sarhro et sa retombée saharienne (sud-est marocain). Étude géomorphologique. Notes et Mémoires du Service Géologique du Maroc, 317.

Robert-Charrue, C., 2006. Géologie structurale de l'Anti-Atlas oriental, Maroc. Doctoral Thesis. Neuchatel, University of Neuchatel, 98pp.

Sébrier, M., Siame, B., Zouine, M., Winter, T., Missenard, Y., Leturmy, P., 2006. Active tectonics in the Moroccan High Atlas. Comptes Rendus Geosciences, 338, 65-79.

Schmidt, K.H., 1992. The tectonic history of the Pre-Saharan depression (Morocco) -a geomorphological interpretation. Geolologische Rundschau, 81, 211-219.

Tesón, E., 2009. Estructura y cronología de la deformación en el borde Sur del Alto Atlas de Marruecos a partir del registro tectono-sedimentario del la cuenca de antepaís de Ouarzazate. Doctoral Thesis. Bellaterra (Cerdanyola del Vallès), Universitat Autònoma de Barcelona, 217pp.

Teixell, A., Arboleya, M.L., Julivert, M., Charroud, M., 2003. Tectonic shortening and topography in the central High Atlas (Morocco). Tectonics, 22, 1051. doi: 10.1029/2002TC001460.

Teixell, A., Ayarza, P., Zeyen, H., Fernàndez, M., Arboleya, M.L., 2005. Effects of mantle upwelling in a compressional setting: the Atlas Mountains of Morocco. Terra Nova, 17, 456-461.

Thiry, M., Ben Brahim, M., 1997. Silicifications de nappe dans les formations carbonatees tertiaires du piedmont atlasique (Hamada du Guir, Maroc). Geodinamica Acta, 10, 12-29.

Toto, E.A., Kaabouben, F., Zouhri, L., Belarbi, M., Benammi, M., Hafid, M., Boutib, L., 2008. Geological evolution and structural style of the Palaeozoic Tafilalt sub-basin, eastern Anti-Atlas (Morocco, North Africa). Geolologische Journal, 43, 59-73.

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