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The two main steps of the Atlas building and geodynamics of the western Mediterranean

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Abstract. The Atlas system (Morocco, Algeria, and Tunisia) constitutes an important morphologic barrier fringing the Sahara platform. Its structural style changes along strike from a thick-skinned style in Morocco to a thin-skinned one in Algeria and Tunisia. The position relative to the Tell-Rif system is also different in eastern Algeria and Tunisia where the two systems are adjacent and in western Algeria and Morocco where they are separated by large rigid cores (Moroccan Meseta and Algerian High Plateaux). New data, as well as a reappraisal of available data, show that the Atlas build up occurred everywhere during two main phases of late Eocene and Pleistocene-lower Quaternary age, respectively. These phases are clearly distinct and do not represent end points of a progressive deformation. An additional Tortonian event exists in the eastern region where the Tell-Rif is thrusting directly over the Atlas. From Oligocene to middle Miocene the development of the Tell-Rif accretionary prism is coeval to subduction rollback of Maghrebian Tethys lithosphere and related to the opening of the western Mediterranean Sea. For kinematic and chronological reasons this process cannot account for the two specific steps of the Atlas building. They are better explained assuming that they record two jolts in the convergence of Africa with respect to Europe and correspond roughly to the initiation and the cessation of the subduction processes active in the western Mediterranean region.

1. Introduction

The geodynamics of the western Mediterranean region is still a matter of debate. The central question of these discussions is why and how, during the Miocene, extensional basins (namely, Alboran Sea, Algero-Provençal Basin, Tyrrhenian Sea and Valencia Trough) initiated and developed within the convergent domain situated between the Africa and Europe plates (Figure1). In the published syntheses the geological context in which these basins developed is directly derived from a knowledge of the "Alpine" chains fringing directly the western Mediterranean Sea: Betics in Southern Spain, Rif in Morocco, Tell in Algeria and Tunisia, and Apennines in Italy (Figure 1) [*Doglioni et al.*, 1999; *Jolivet et al.*, 1999, and references therein]. On the contrary, information coming from "intracontinental" chains located away from the above mentioned internal cores, such as the Iberian range and the Pyrenees in Europe or the Atlas and Saoura-Ougarta chain in North Africa, is generally neglected (Figure 1).

Recent studies carried out in the Atlas of Morocco, Algeria, and Tunisia allow a better understanding of these mountains and should be used to highlight the geodynamics of the western Mediterranean region. The aim of this paper is to present these new data and interpretations and, more generally, to give an "African point of view" on the Africa-Europe collision zone.

We will first present briefly the geology of the three main structural domains forming North Africa: the Sahara Platform, the Atlas Mountains, and the Tell-Rif orogenic system. Then we will examine the geometry and kinematics of the Atlas Mountains along strike from Agadir (Morocco) to Tunis (Tunisia). In the discussion we will integrate the kinematics of the Atlas within the plate kinematics framework of the western Mediterranean region. In particular, we will show that the two main phases of Atlas building (late Eocene and Pleistocene-lower Quaternary) can be correlated to the beginning and the end of the formation of the western Mediterranean Sea.

2. Geological Setting of North Africa

North Africa can be divided into three main structural domains (Figure 2a): The Sahara domain, the Atlas Mountains (including, in our acceptation, the Meseta-High Plateaux rigid cores), and the Tell-Rif system fringing the Mediterranean Sea. From a topographic point of view the Atlas domain forms a major morphologic barrier between the Sahara domain and the western Mediterranean Sea (Plate 1). The highest peak (Jebel Toubkal) is situated in Morocco and tops 4165 m. From this point and laterally the altitude of the Atlas decreases rapidly westward to the Atlantic and gently eastward to the Hodna basin. From the Hodna to the Gulf of Sirt the same lateral asymmetry is observed: the highest peak (Jebel Chelia, 2328 m) is located very close to the Hodna (Figure 3). Compared to the Atlas, the Tell-Rif appears as a domain of relatively moderate altitude: maximum 2500 m in the central Rif and Great Kabylie. In the western Sahara, the Anti-Atlas region forms an elevated arch (up to 2000 m) situated immediately in front of the Atlas Mountains (Plate 1 and Figure 3).

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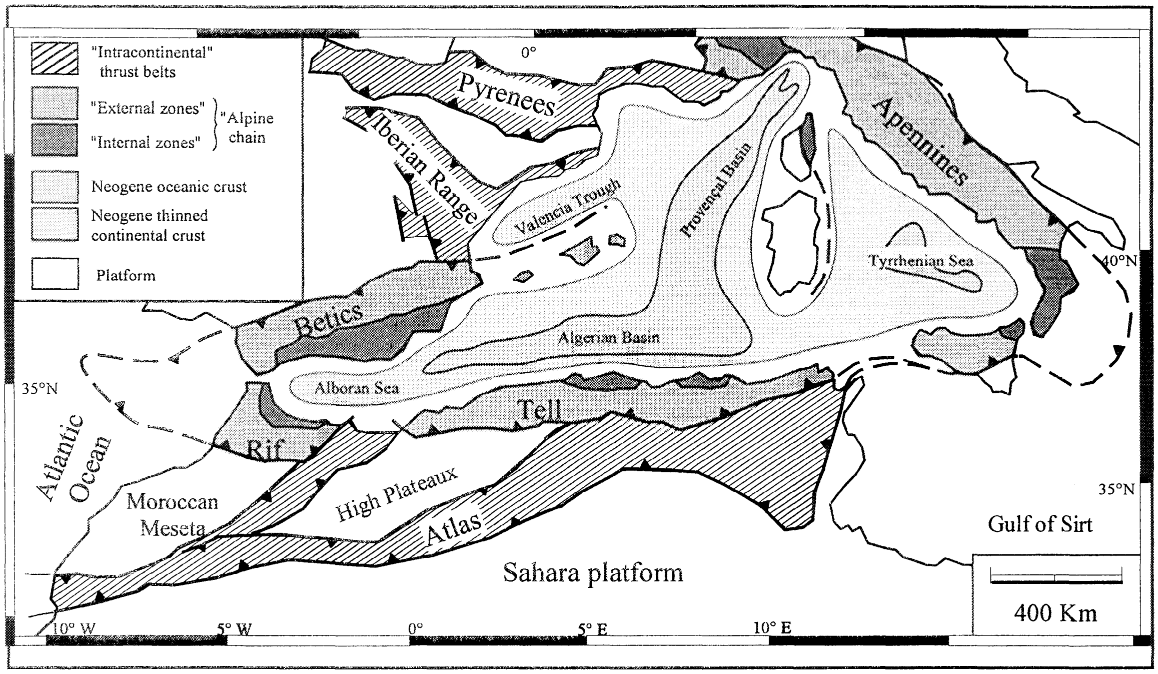

Figure 1. Schematic map of the western Mediterranean region showing the main orogens and extensional basins (modified from *Doglioni et al.* [1999]), s.l., *sensu lato.*

2.1. The Sahara Domain

Except for the late Eocene reactivation of the inverted Saoura-Ougarta basin [*Ziegler*, 1988] (Figure 2a), the Sahara domain is characterized by the lack of important Meso-Cenozoic deformations. The Precambrian African basement is exposed in the Anti-Atlas domain and in the Riguibate and Hoggar massifs. The basement is buried beneath Paleozoic basins (namely Tindouf, Bechar) distributed south and east of the Anti-Atlas and north of the Riguibate and Hoggar massifs (Figure 4).

A thick Meso-Cenozoic basin (up to 4500 m), the so-called "Triassic Basin" [*Boudjema*, 1987], occupies the eastern half of the domain (Figure 4). At the base of the Mesozoic series the "Hercynian unconformity" is an erosional and structural unconformity dividing the sedimentary cover into two "Super Cycles": Paleozoic (i.e. gondwanian) and Meso-Cenozoic (i.e., Atlantic and Tethyan), respectively [*Boote et al.*, 1998].

2.2. The Atlas Mountains

The Atlas Mountains (i.e, Middle and High Atlas in Morocco, Saharan Atlas in western Algeria, Aurès in eastern Algeria and Tunisian Atlas in Tunisia, Figure 2a) were uplifted during the Cenozoic. The faulted and folded Meso-Cenozoic cover pertains to different basins: the Western High Atlas belongs to the Atlantic margin whereas the Middle Atlas and the eastern High Atlas are situated on the sites of Tethyan basins (i.e., connected eastward or northward to the Tethys ocean, Figure 4). In Algeria and Tunisia the Atlas basin constitutes the northward extent of the "Triassic basin". Consequently, if in Morocco the South Atlas Front follows more or less the southern boundary of the Mesozoic Atlas basin, in Algeria it cuts across the basin (Figure 4).

The two rigid cores (Moroccan Meseta and Algerian High Plateaux), situated between the Atlas and the Tell-Rif systems, separated by the Middle Atlas (Figure 2a), are characterized by a thin or even absent Meso-Cenozoic cover overlying more or less metamorphosed Paleozoic strata (Figure 4). The basement of the Atlas system is well known in Morocco where it is made up of Paleozoic rocks deformed during the Variscan orogeny. In Morocco the Variscan belt is stretched along N45°-N70° shear zones [*Piqué and Michard,* 1989]. In Algeria, deformed Paleozoic rocks are known as blocks within diapirs of Triassic evaporites, suggesting that the Variscan domain extends below the Saharan Atlas. So, in western Algeria it is assumed by *Piqué et al.* [1998a] that the southern boundary of the Variscan orogeny parallels the South Atlas Front. In Tunisia the Paleozoic substratum of the Atlas is unknown. However, it is likely that an erosional unconformity, related to moderate Late Carboniferous movements, exists as in the Sahara domain [*Boote et al.*, 1998].

There is a general agreement that the Atlas Mountains developed along zones of crustal weakness inherited from rifting episodes associated with the opening of both Atlantic and Tethyan oceans during Late Triassic to Early Liassic times [*du Dresnay*, 1975; *Stets and Wurster*, 1977; *Mattauer et al.*, 1977, *Laville and Petit*, 1984; *Winterer and Hinz*, 1984; *Andrieux et al.*, 1989; *Aït Ouali*, 1991; *Stets*, 1992]. The fault pattern at the base of these basins is dominated by NNE-SSE trending normal faults along the Atlantic shoreline and within the Middle Atlas and by ENE-WSW trending faults along the Atlas main alignment (Figure 2b). From the end of the Liassic up to the late Mesozoic regional subsidence affected the Atlantic margin [*Le Roy*, 1997], the Central and eastern Maghreb in Algeria [*Vially et al.*, 1994], and Tunisia [*Burollet and Ellouz*, 1986]. In the eastern part of the Central High Atlas (Morocco), basic intrusions (gabbroic bodies) of Upper Jurassic age are

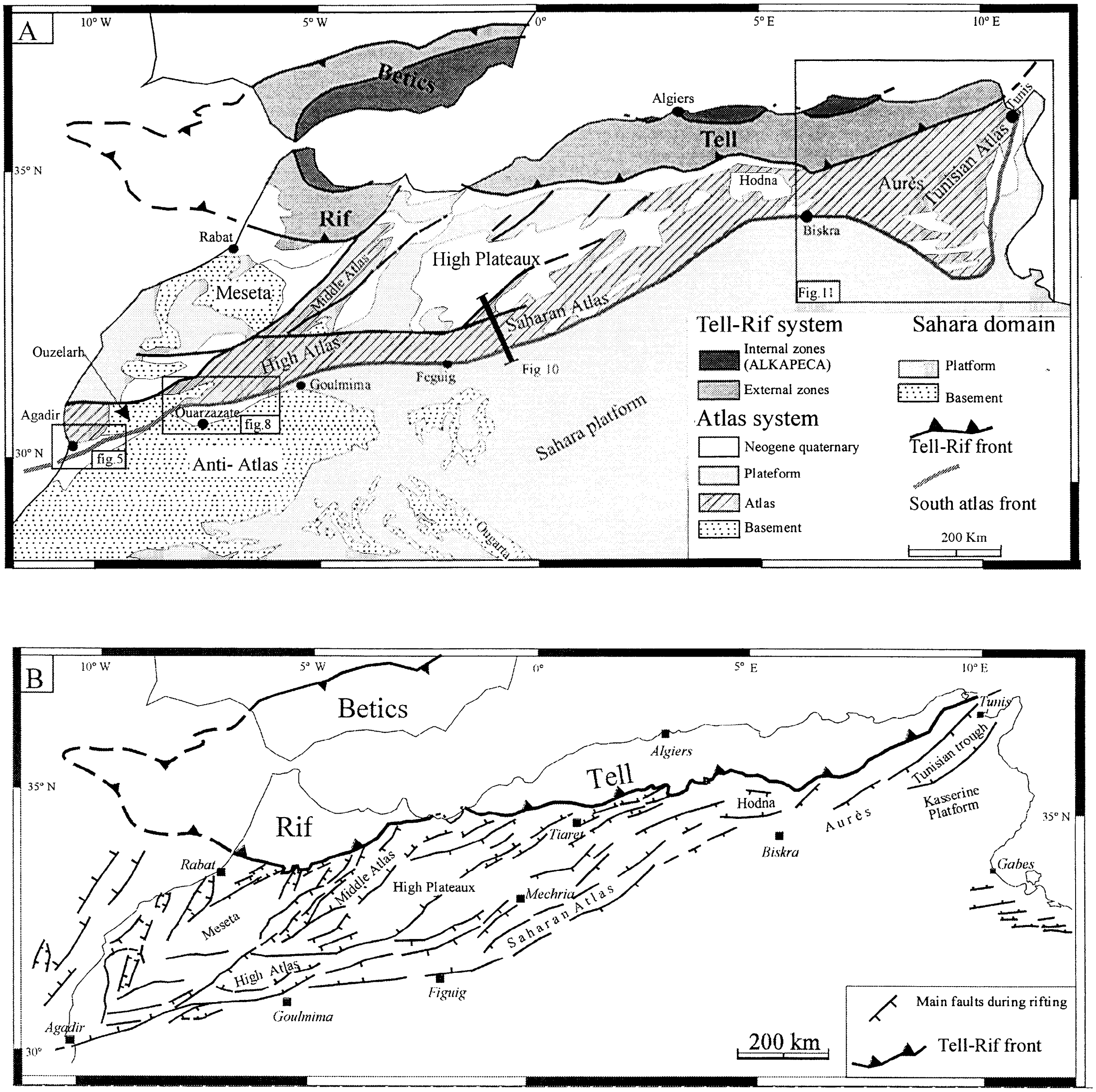

Figure 2. (a) Structural map of North Africa illustrating the main structural domains: Sahara Platform and Atlas and Tell-Rif systems. (b) Faults network linked to the Atlantic-Tethyan rifting episode in the Atlas system. Sources of data are the following: Morocco [*Laville and Piqué*, 1991]; Algeria (SONATRACH database, 1999); Tunisia [*Lucazeau and Ben Dhia*, 1989; *Research Group for Lithospheric Structure in Tunisia,* 1992].

interpreted as resulting from left-lateral transcurrent movements [*Laville and Harmand*, 1982; *Laville*, 1985; *Piqué and Laville*, 1995], but their origin as well as their mode of emplacement are hardly discussed. Apparent Upper Jurassic normal faults observed on seismic profiles from the Saharan Atlas (Algeria) are also interpreted as resulting from transtensional displacements [*Aït Ouali*, 1991]. Tectonic inversion of Mesozoic basins, leading to the building of the Atlas Mountains, took place during the Cenozoic up to the lower Quaternary, as will be detailed in section 3.

2.3. The Tell-Rif System

The Tell-Rif system consists of severely inverted basins, which are presently integrated into nappes structures. To the west the front of the Rif swings through a 180° arcs and connects to the Betics. To the east the front of the Tell swings through a 90° angle and joins the Apennines front (Figure 1).

The Tell-Rif system is considered as an Alpine-type orogen [*Durand-Delga and Fonboté*, 1980]. This means an orogen resulting from opening and closing of an ocean, followed by a

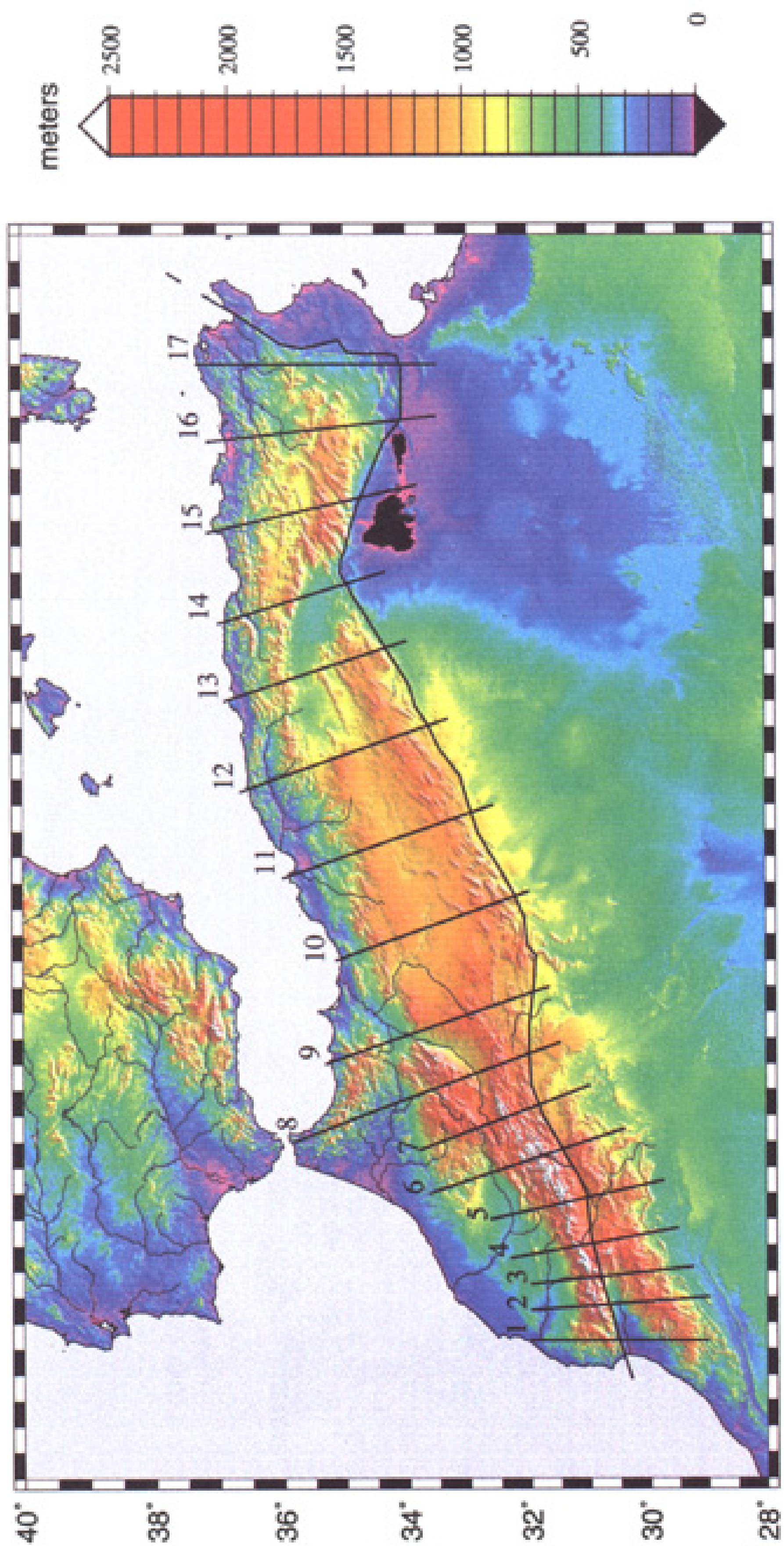

Plate 1. Topography of North Africa (source of data is U.S. Geological Survey EROS data center).

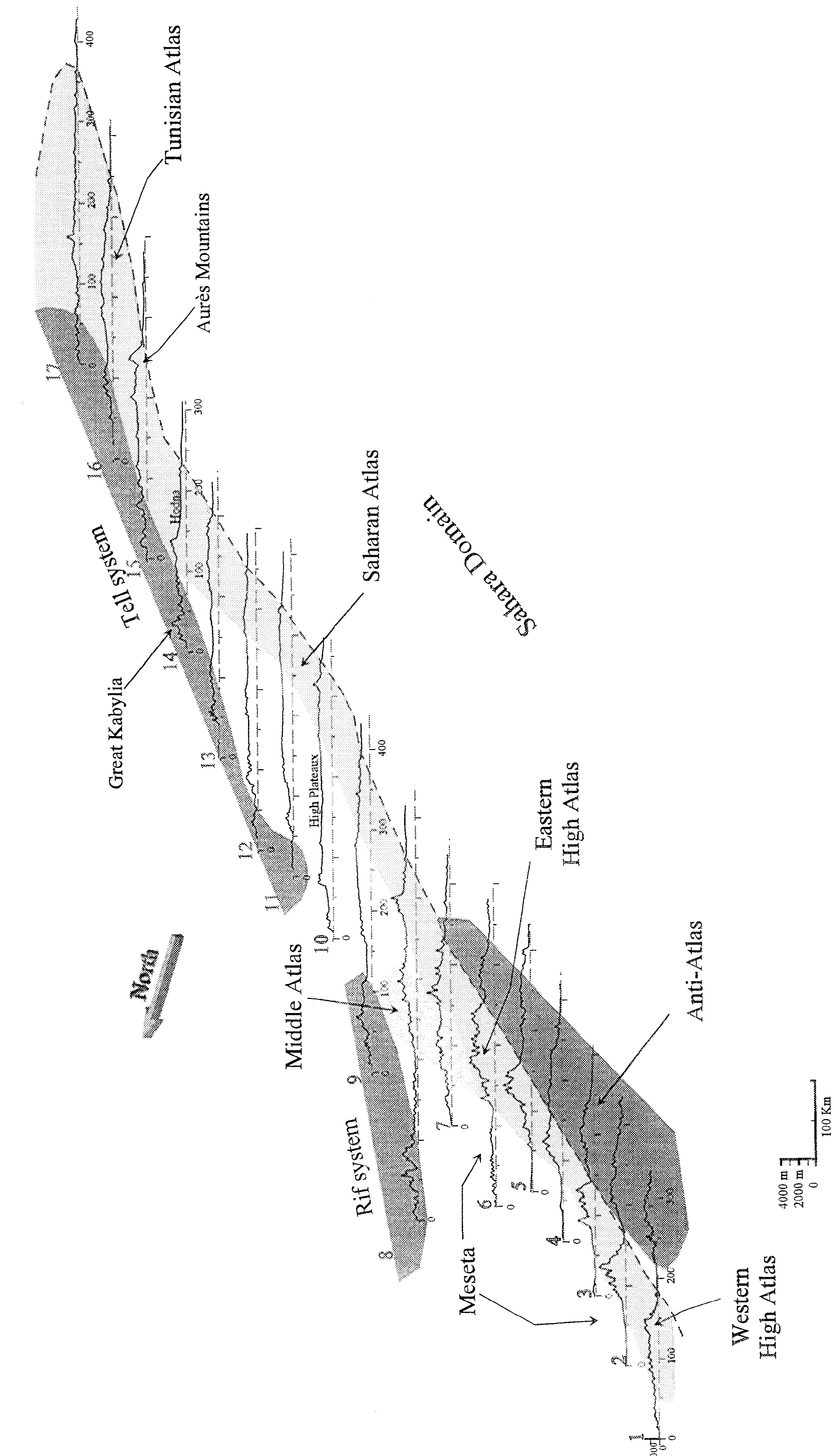

Figure 3. Cross sections through North Africa illustrating the topographic variation across the Atlas and Tell-Rif systems.

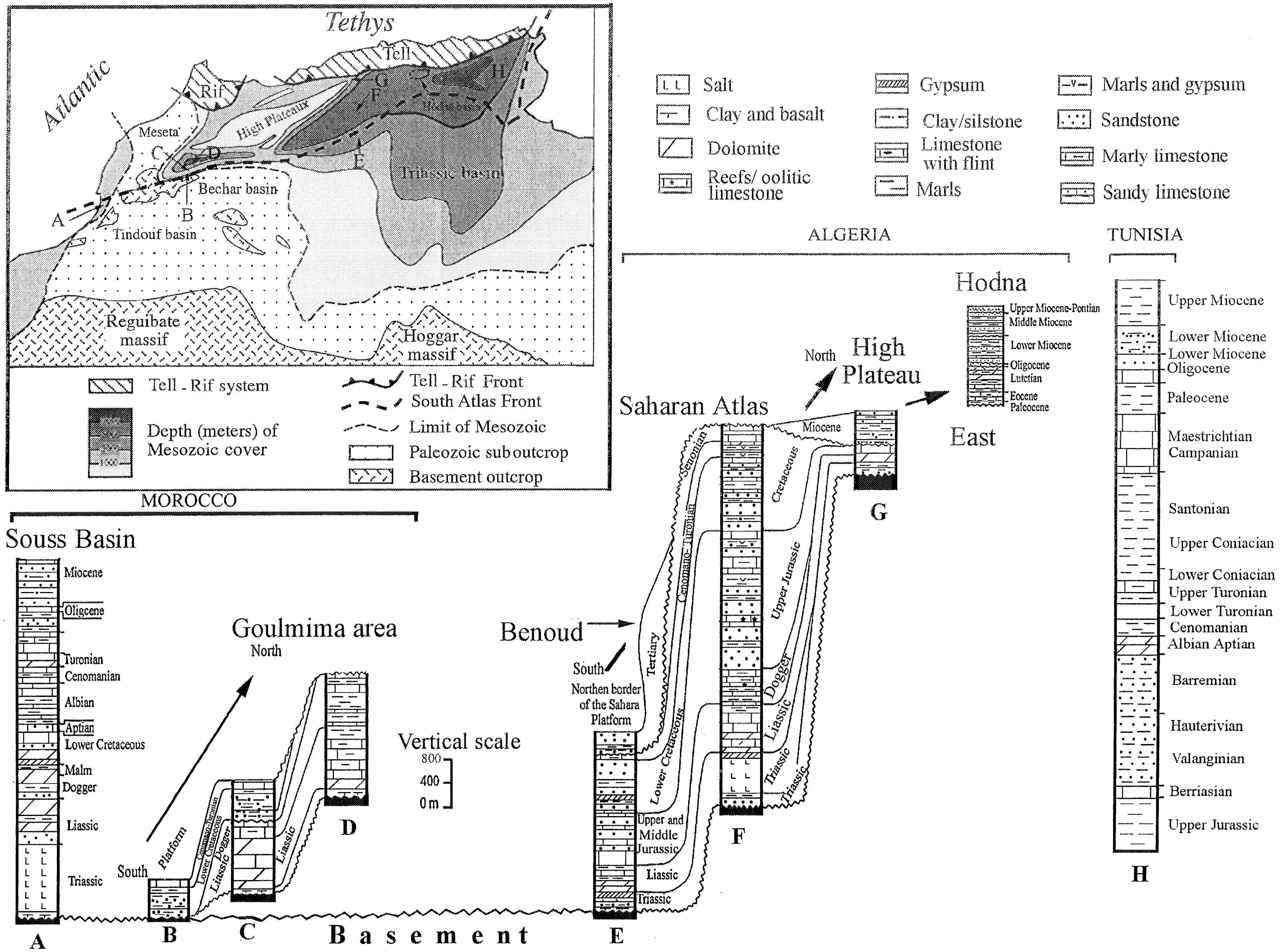

Figure 4. Distribution of the Mesozoic basins within the Sahara and Atlas domains illustrated by some generalised lithostratigraphic columns. Source of data are the following: Souss [*Mustaphi*, 1997]; eastern High Atlas [*Saint Bezar et al.*, 1998]; Algeria (SONATRACH database); Tunisia [*Ben Ferjani et al.*, 1990].

continent-continent collision. We will see that this classical interpretation can be revisited. In any case, the present-day external zones of the Tell-Rif are considered as the former southern margin of the Tethys Ocean: the "flyschs" domain as a remnant of ocean floor deposits and the cover of the present day internal zones ("Dorsale Calcaire") as remnants of the northern Tethys margin [*Durand-Delga and Fontboté*, 1980; *Bouillin*, 1986; *Cattanéo et al.*, 1999]. The internal zones, sometimes called ALKAPECA domain (for Alboran, Kabylies, Peloritan, and Calabria, [*Bouillin et al.*, 1986]), are only fragments of a stack of nappe complexes, including peridotite flakes known in the Betics (Spain), the internal Rif (Morocco), as well as the Kabylies (Algeria) (Figure 2a). The major stages of the Tell-Rif tectonic history are roughly the same as the Atlas ones: Triassic-Liassic rifting (with high stretching factors), a phase of quiescence during the middle and late Mesozoic, and thrusting initiation by the late Eocene [*Favre*, 1995].

The sequence of thrusting is quite complex and involves the development of out-of-sequence thrusting at the boundaries between the main zones [*Frizon de Lamotte et al.*, 1991; *Morley*, 1992a]. Compression within the external zones is more or less coeval to extension within the Alboran and Algerian basins, which was initiated in early (?)/middle Miocene times, followed by thermal subsidence and sedimentation during late Miocene up to Pleistocene [*Morley*, 1992b; *Watts et al.*, 1993; *Comas et al.*, 1997; *Comas et al.*, 1999; *Doglioni et al.*, 1999].

The Neogene tectonic transport direction, related to nappes emplacement, is different from one place to another in relation with the bending of the system. It is westward in the Betics and Rif [*Frizon de Lamotte*, 1987; *Frizon de Lamotte et al.*, 1991; *Morley*, 1992b; *Guézou et al.*, 1992; *Vissers et al.*, 1995; *Lonergan and White*, 1997], mainly southward (or south-eastward) in the Tell [*Vila*, 1980] and eastward in Calabria-Sicily [*Doglioni et al.*, 1999]. By the end of the Pliocene, on the contrary, the whole domain suffers a quite constant NW-SE shortening responsible for inversion structures within the Alboran Sea [*Watts et al.*, 1993; *Comas et al.*, 1999] as well as, in Algeria, within the Cheliff onshore basin [*Meghraoui et al.*, 1996].

3. Geometry and Kinematics of the South Atlas Front: Consequences for the Interpretation of the Atlas Mountains

Although the Atlas Mountains are continuous, their architecture changes drastically along strike: in Morocco the Atlas system is interpreted as a typical intra-continental chain [*Mattauer et al.* 1977; *Durand-Delga and Fontboté*, 1980; *Laville*, 1985; *Giese and Jacobshagen*, 1992; *Errarhaoui*, 1997] whereas in Tunisia it corresponds to the foreland fold-thrust belt of the Tell [*Rouvier*, 1977; *Snoke et al.*, 1988; *Burollet*, 1991; *Morgan et al.*, 1998; *Tricart et al.*, 1994]. The main reason for these different interpretations is that as we have shown, rigid blocks (Moroccan Meseta and Algerian High Plateaux) are sandwiched between the two systems in the western regions (Figure 2a and Plate 1).

The basement-involved style of the Moroccan High Atlas, where the basement frequently outcrops, is unquestionable. In our view, this "basement" of the High Atlas includes not only the Panafrican substratum but also the Paleozoic rocks, which are here affected by the Variscan orogeny and locally intruded by granites. In the eastern regions the question of the thin- versus thick-skinned style of deformation is still under debate. Traditionally, and probably by reference to the Moroccan Atlas, structural models relate the Saharan Atlas, Aurès, and Tunisian Atlas deformation to basement fault reactivating [*Laffitte, 1939a, b; Guiraud*, 1975; *Ouali et al.*, 1987; *Kazi-Tani*, 1986; *Boccaletti et al.*, 1988; *Snoke et al.*, 1988; *Vially et al.*, 1994]. This point of view has been illustrated recently by *Morgan et al.* [1998] and by *Hlaiem* [1998]. However, other publications have emphasized the thin-skinned nature of compressive deformation in the Aurès and Tunisian and the major role played by a detachment located within Triassic evaporites [*Frizon de Lamotte et al.*, 1990; *Ghandriche*, 1991; *Creuzot et al.*, 1992; 1993; *Outtani et al.*, 1995; *Addoum*, 1995; *Anderson*, 1996; *Ouali and Mercier*, 1997, *Bracène et al.*, 1998; *Frizon de Lamotte et al.*, 1998]. This question, which concerns the mode of the control exercised by basement faults on the development of cover structures, is important and must be discussed in detail.

Basin inversion is a well-known mechanism that occurs under a range of kinematic settings [*Ziegler et al.*, 1998]. We will examine two end-member situations. In the first case the same basement faults suffer two successive movements: a normal slip during the rifting stage and a reverse slip (or strike slip) during the inversion. If this mechanism is general, as postulated by the supporters of the thick-skinned models, we can expect a quite difficult propagation of the deformation front toward the foreland. In the second case, on the contrary, the lower detachment ignores the staircase geometry of the basement and generates hanging wall short cuts leading to the development of thin-skinned structures. If the detachment is efficient (evaporite), then we can expect a fast and easy propagation of the thrust front until the occurrence of a blockage due to a tectonic or sedimentologic disruption of the décollement level. Such a blockage subsequently leads to out-of-sequence structures superimposed on the structures developed during the propagation of the décollement. As we will see in sections 3.3 and 3.4, such a tectonic scenario, supporting the thin-skinned model, is recorded in eastern Algeria and in Tunisia.

On the other hand, an argument frequently used to support the thick-skinned model is the observation of congruence between the location of normal faults and thrust faults. In our opinion this does not represent evidence supporting the models because the normal faults located in the cover can be transported very far from their initial position. Additionally physical modelling of fold-thrust structures initiated in a sand box affected by early normal faults show that the ramps climbing from the basal décollement are preferably located at the intersection between the décollement and the previous normal faults [*Ballard et al.*, 1987; *Wilschko and Eastman*, 1988; *Colletta et al.*, 1991; *Vially et al.*, 1994]. Consequently, one must consider that the congruence mentioned above is not circumstantial but is insufficient to validate thick-skinned models (*Martin and Mercier*, 1996).

The aim of the section 3 is to deal with these alternate tectonic models to understand why the architecture of the same orogen changes laterally. We will focus on the South Atlas region where structures are well exposed and the timing of

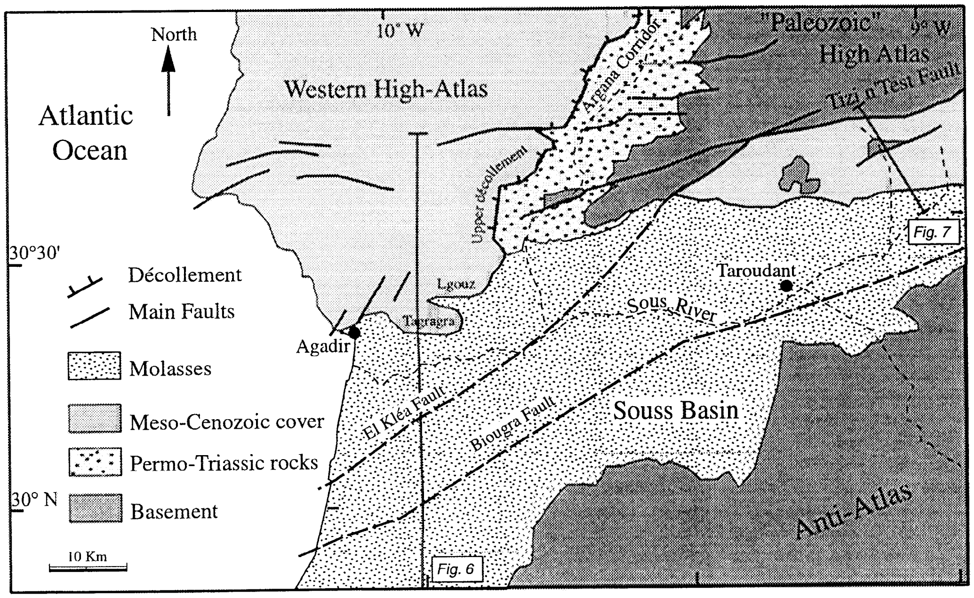

Figure 5. Structural map of the Western High Atlas (modified from *Choubert* [1959]).

deformation can be defined more precisely. The South Atlas Front (SAF) is a major thrust front running continuously from Agadir (Morocco) to Tunis (Tunisia) [*Bracène et al.*, 1998; *Frizon de Lamotte et al.*, 1998; *Saint Bezar et al.*, 1998, and references therein] (Figure 2). To the east the SAF connects the southern Apennines through Sicily and the Thyrrenian Arc. To the west the SAF disappears in the Atlantic Ocean. A connection with some of the structures present in the Canary Islands has been advocated [*Robertson and Stillman*, 1979] but this assumption is still under discussion [*Le Roy*, 1997]. We will examine the geometry and kinematics of the SAF from the west to the east in four keyareas.

3.1. The Souss Segment

The Souss segment is characterized by the presence of a Mio-Pliocene foreland basin (the Souss basin) fringing the mountain front. However, its geometry changes from the western region (Agadir) toward the east (Taroudant) (Figure 5).

North of Agadir, Mesozoic sediments forming the passive margin of the Atlantic Ocean are involved in folds forming the "western High Atlas," characterized by a mean altitude of ~1.000 m (Figure 4). The main structures are E-W folds and thrusts [*Ambroggi*, 1963] that are bounded by NNE-SSW strike-slip faults acting as tear faults [*Outtani*, 1996]. The structural style of this area is illustrated by the N-S cross section cutting through the Tagragra and Lgouz anticlines (Figure 6).

The El Kléa anticline is buried under the Souss basin. The fold, imaged by a seismic line, mimics a drag fold developed along a Triassic 45°N normal fault inherited from the Atlantic rifting (Mustaphi, 1997). The zero point, situated approximately at the top of the Dogger strata, displays the two successive movements (normal and reverse) taking place along this fault. The slip accommodated in the fold is mainly transferred from the front of the Tagragra anticline (Figure 6). Downward the fault soles out into an intracrustal reflector, interpreted as representing an intracrustal décollement. South of the El Kléa fault, the Biougra fault is also a major Triassic fault branched at depth on the same décollement, but no inversion occurred during the build-up of the Atlas Mountains.

The subsurface geometry of the Tagragra anticline is well constrained (seismic profiles and boreholes). It fits very well a fault bend fold [*Outtani*, 1996; *Mustaphi*, 1997]. As a consequence of this modelling, the fold geometry has been derived in considering décollements situated in the evaporites of Upper Triassic- Lower Jurassic age (upper flat) and at the top of the Paleozoic substratum (lower flat), respectively. The slip transmitted forward onto the upper flat is accommodated by the El Kléa anticline.

The Lgouz anticline is interpreted as a detachment fold built on the shallower décollement described above (Figure 6). This décollement, which is itself folded by a basement culmination (Figure6), is exposed in the field along the western boundary of the Argana corridor (Figure 5), a large transverse zone occupied by Triassic rocks. We assume consequently a complete tectonic uncoupling between a substratum (including basement, Paleozoic strata, and Triassic red beds) and the overlying cover. The structural style of the substratum, inferred from observation made in the Argana corridor, is dominated by steep faults inherited from the rifting episode and more or less subsequently inverted. It is likely that these faults are branched at depth on the intra-crustal reflector observed at the lower tip of the El Kléa fault. Along this section the amount of shortening (calculated using the length of Jurassic beds) is ~10%.

The region situated north and northeast of Taroudant is characterized by the lack of Triassic/Jurassic strata. From the Souss basin to the north, one can observe a zone of foothills formed by a couple of large asymmetric anticlines associated with a narrow syncline. Paleozoic beds (of Silurian age) truncated by an erosional and structural unconformity and the overlying Cretaceous to Eocene strata are involved in these folds. To the north, Mesozoic and Cenozoic beds truncated and

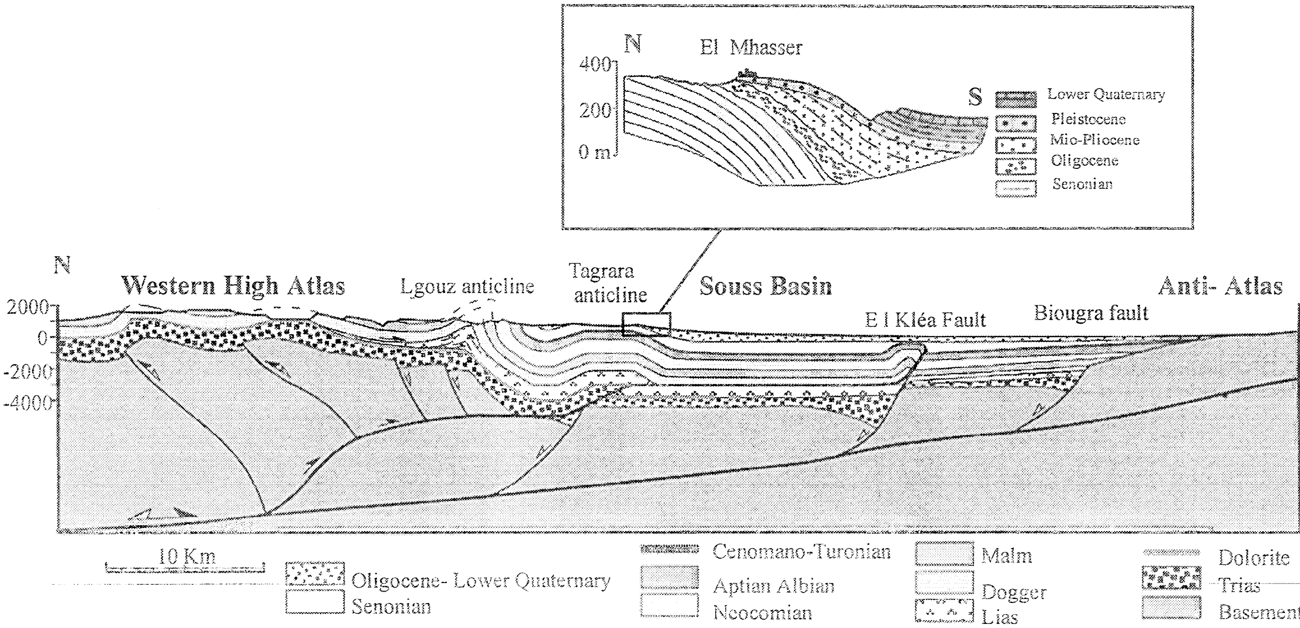

Figure 6. Generalized cross-section through the Souss basin and the Western High Atlas (modified and completed from *Outtani* [1996] and *Mustaphi* [1997]; sources of data are field work and Office National de Recherche et Exploration Petrolières (ONAREP) seismic profiles). See explanations in the text and location in Figure 5. Insert shows the geometry of Tertiary and Quaternary series along the front of the Tagragra anticline (modified from *Ambroggi* [1963]).

overthrusted by Paleozoic rock, containing the Variscan Tichka granite, form the western part of the "Paleozoic High Atlas" (Figures 5 and 7). An abrupt difference in altitude is evident between the Taroudant foothills where the mean altitude is 1000m and the Paleozoic High Atlas, which reaches 3500m at the Tichka peak.

Within each fold of the foothills, one can observe variations in the thickness of the Lower to Middle Cretaceous strata between the forelimb where they are thin (or even absent) and the back limb where the succession is complete and the thickness maximum. This could suggest that the anticlines were initiated during Upper Cretaceous times and then strongly reactivated during the building of the High Atlas. This gives additional evidence supporting the assertion that a first compression occurred in the Upper Cretaceous (see review by *Guiraud* [1998]). Classically, such structures are linked to the beginning of the convergence between the Africa and Eurasia plates. However, in our experience the effects of this first inversion remain weak.

The fault zone along which Meso-Cenozoic and Paleozoic rocks are juxtaposed is known as the Tizi N'Test Fault Zone (TNTFZ) (Figure 5). This major fault zone is a polyphase structure, active at least since the upper Paleozoic. It has been the site of major right-lateral movement during the late Paleozoic, extension during Triassic, and then thrusting during the building of the Atlas system [*Proust et al.*, 1977].

To the west the TNTFZ is connected to the El Kléa fault (see above), whereas to the east it cuts through the Central High Atlas. Therefore we suggest that the TNTFZ is branched on the intracrustal décollement flooring the High Atlas.

At least two superposed décollements are needed to explain the structures of the Souss region. The shallower one, which is located within Upper Triassic-Lower Liassic rocks, is restricted to the region west of the Argana corridor. The deeper one is

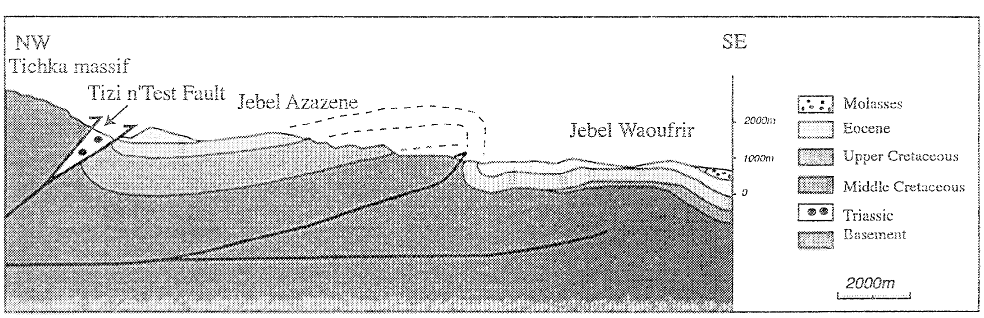

Figure 7. Cross section NE of Taroudant. See explanations in the text and location in Figure 5.

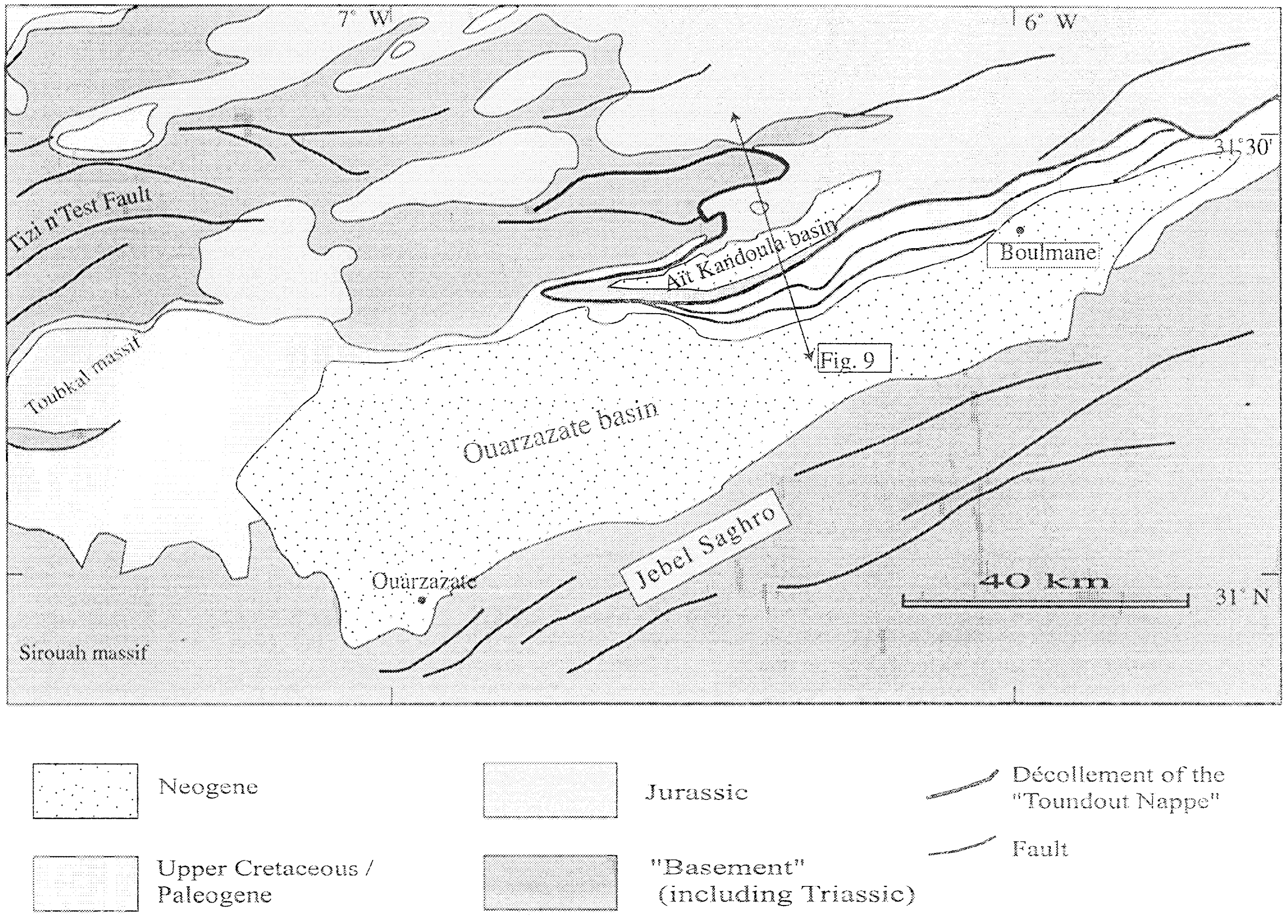

Figure 8. Schematic structural map of the Central High Atlas (modified from *Choubert* [1959]).

intracrustal. It is imaged below the Souss basin and extends southward below the Anti-Atlas. It is possible that this deep décollement plays a role in the recent uplift of the Anti-Atlas [*Giese and Jacobshagen*, 1992].

The timing of the deformation can be deduced from some tectonostratigraphic evidence. The Oligocene conglomerate underlining the base of the Souss basin molasses rests unconformably on Upper Cretaceous (Agadir) and Eocene (Taroudant) rocks. This suggests that an important tectonic phase took place between the Eocene and the deposit of the Oligocene conglomerate [*Ambroggi*, 1963]. Offshore, the part of the Atlantic margin extending the Atlas domain westward, exhibits the same unconformity underlying a major hiatus [*Le Roy*, 1997].

The molasses of the Souss basin are folded. Growth strata, observed along the forelimb of the Tagragra anticline, show that deformation was active during the Pleistocene and the lower Quaternary (Figure 6). Along the coastal line, Quaternary marine terrace elevation changes significantly near the Agadir fold. The 1960 Agadir earthquake (*Ms* 5.9) is also likely related to this structure which, consequently, must be considered as still active [Meghraoui *et al.*, 1998].

The partition between the structures linked to the upper Eocene event and those related to the Pleistocene-lower Quaternary cannot be easily made. However, we stress the recognition of two different phases separated by a break during the Oligocene-Miocene period.

3.2. The Central High Atlas Segment

From the easternmost part of the Souss basin to the westernmost part of the Ouarzazate basin, the Mesozoic cover is not preserved and the South Atlas Front is concealed within the basement. Its precise trajectory is not known but it necessarily cuts across the "Ouzellarh promontory" [*Choubert*, 1959] made up of Panafrican basement and isolates the Toubkal massif, which belongs to the High Atlas, from the Sirouah massif, which belongs to the Anti-Atlas (Figure 2a).

Eastward, the South Atlas Front underlines the northern boundary of the Ouarzazate foreland basin (Figure 8). The stratigraphy of the Ouarzazate basin can be divided into two major tectonostratigraphic units: a carapace-like cover and molasses. This carapace is divided into three main formations: red Upper Jurassic-Lower Cretaceous sandstone and siltstone, Cenomanian-Turonian marine limestone, Upper Cretaceous continental siltstone passing upward in Eocene carbonates. Syntectonic molasses of Oligocene to Pliocene ages overlie these sediments [*Fraissinet et al.*, 1988; *Görler et al.*, 1988]. South of

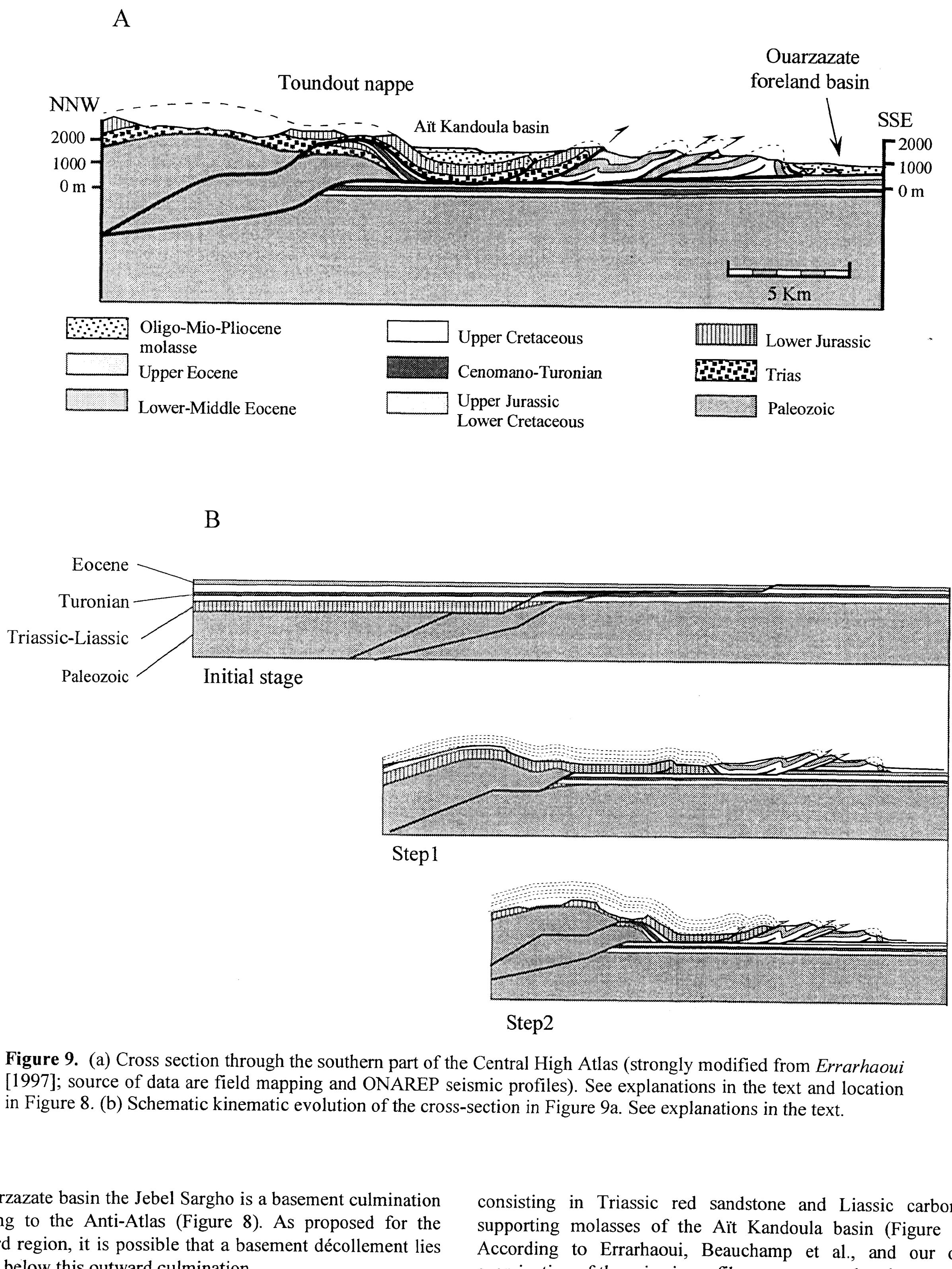

Figure 9. (a) Cross section through the southern part of the Central High Atlas (strongly modified from *Errarhaoui* [1997]; source of data are field mapping and ONAREP seismic profiles). See explanations in the text and location in Figure 8. (b) Schematic kinematic evolution of the cross-section in Figure 9a. See explanations in the text.

the Ouarzazate basin the Jebel Sargho is a basement culmination belonging to the Anti-Atlas (Figure 8). As proposed for the westward region, it is possible that a basement décollement lies at depth below this outward culmination.

Errarhaoui [1997] and *Beauchamp et al.* [1999], using subsurface data, show a frontal tectonic repetition of the Upper Cretaceous-Eocene sequence along the active margin of the Ouarzazate basin (Figure 9a). This duplication of the stratigraphic pile is overthrusted by the "Toundout nappe" [*Roch*, 1939; *Laville et al.*, 1977; *Fraissinet et al.*, 1988], consisting in Triassic red sandstone and Liassic carbonate supporting molasses of the Aït Kandoula basin (Figure 9a). According to Errarhaoui, Beauchamp et al., and our own examination of the seismic profiles, we assume that the molasses are not completely involved in the frontal duplication of the platform.

We propose that the "Toundout nappe" developed firstly as a fold bend fault with an upper flat located within Upper Cretaceous series (Figure 9b). At the same time, the slip transmitted forward was accommodated by the duplication of the

Upper Cretaceous-Eocene sequence. Subsequently, the development of a basement duplex led to the tilting observed in the rear zone of the "Toundout nappe." During this second step, the floor thrust of the duplex reached the surface as an out-of-sequence thrust fault (Figure 9b).

The amount of shortening is ~60% for the part of the section situated between the front of the Toundout nappe and the Ouarzazate basin (Figure9a). However, according to *Beauchamp et al.* [1999], we must keep in mind that deformation of the cover is concentrated along the borders of the High Atlas. For a complete section of the High Atlas in this area these authors propose a total shortening of 36%.

The timing of the deformation can be deduced from the analysis of the age and regional distribution of the molasses [*Fraissinet et al.*, 1988; *Görler et al.*, 1988; *Errarhaoui*, 1997). In the "Aït Kandoula" piggyback basin overlying the Toundout nappe, lower Miocene conglomerates rest directly on the Liassic carbonates, pointing to an important erosional period before their deposition. In the Ouarzazate basin, on the contrary, Oligocene molasses lie conformably on the upper Eocene [*Beauchamp et al.*, 1999]. Consequently, it is likely that at that time the Ouarzazate basin was relatively far away from the deformation front. For its part, the Toundout nappe homeland was probably much closer to the front. Although growth strata exist within the lower-middle Miocene sediments of the Ouarzazate and Aït Kandoula basins, we do not have any evidence for large-scale displacements during this period. On the contrary, the Upper Pliocene-Pleistocene, marked by the development of coarse conglomerates, appears as the main period during which an out-of-sequence basement duplex developed, leading to the second uplift of the Atlas.

The analysis of the Ouarzazate segment shows the occurrence of two distinct stages in the development of the Atlas system, corresponding to two episodes of rapid uplift of the axial zone of the High Atlas [*Görler et al.*, 1988]. Furthermore, the analysis shows that after a first step of activity (likely during late Eocene and Oligocene) and a phase of relative quiescence characterized by widespread lake deposits at the end of the Miocene [*Görler et al.*, 1988], the main deformation took place during the Pleistocene and the lower Quaternary.

3.3. The Eastern High Atlas and the Saharan Atlas

The Goulmima area (Figure 2a), studied recently by *Saint Bezar et al.* [1998], can be considered as representative of the eastern High Atlas. The main feature of this zone is that the southernmost thrust of the High Atlas remains blind below frontal tip line folds. Within the foreland, thin Mesozoic-Cenozoic series form a horizontal tableland overlying Paleozoic rocks or the Precambrian basement. An important observation is the lack of a Neogene foreland basin, suggesting that eroded material was transported laterally westward to the Ouarzazate basin and/or eastward to the front of the Saharan Atlas (Algeria).

From surface and subsurface data it has been demonstrated that the frontal fold is superimposed on a reactivated Triassic-Liassic normal fault, forming the southern boundary of the Atlas basin [*Jossen and Filali-Moutei*, 1992; *Saint Bezar et al.*, 1998]. The externalmost anticline is a thin-skinned feature, related to a décollement located within Triassic rocks. Kinematic modelling shows that through time the thrust tip lines propagated progressively southward. In detail, however, the sequence of thrusting is quite complicated with occurrences of out-of-sequence thrusting as in the Ouarzazate area [*Saint Bezar et al.*, 1998]. Compared to the westernmost regions, the geometry of the South Atlas Front in the eastern High Atlas is very simple. The frontal thrust is branched northward on a deep décollement, which, near the South Atlas Front, cuts down into the basement. A prolongation of this intracrustal décollement south of the front is not required.

Because of the lack of Tertiary deposits, an accurate timing of deformation cannot be established. However, the calculations of scarp retreat of the tableland bounding the SAF shows that scarp back wearing began in the late Eocene [*Schmidt*, 1992]. This transition from depositional to erosional activity is likely related to the beginning of the inversion process [*Schmidt*, 1992]. It seems, consequently, that the timing, established in the Ouarzazate region, is also valid for the eastern High Atlas.

The trend of the deformation front, which is roughly E-W from Agadir to Figuig, is striking ENE-WSW between Figuig and Biskra along the Saharan Atlas (Figure 2). Flanked by the High Plateaux and the Sahara Platform, the Saharan Atlas exhibits a moderate relief (maximum 2136 m) which decreases eastward toward the Hodna basin (Plate 1 and Figure3). As in the Eastern High Atlas segment, the frontal anticline is almost everywhere a tip line fold [*Bracène et al.*, 1998]. However, seismic lines crossing the deformation front show that, in this region, the frontal fold is not superimposed on a major basement fault (Figure 10). Moreover, on these lines one can observe that folded Cretaceous rocks overlie a décollement surface that climbs up from Triassic beds. Below this décollement, sedimentary rocks remain flat lying; thus the structural style is thin-skinned.

This geometry is very different from the one observed along the northern boundary of the Sahara Atlas. *Vially et al.* [1994] show seismic profiles crossing the North Atlas Front (i.e., the boundary between the Sahara Atlas and the High Plateaux). Here

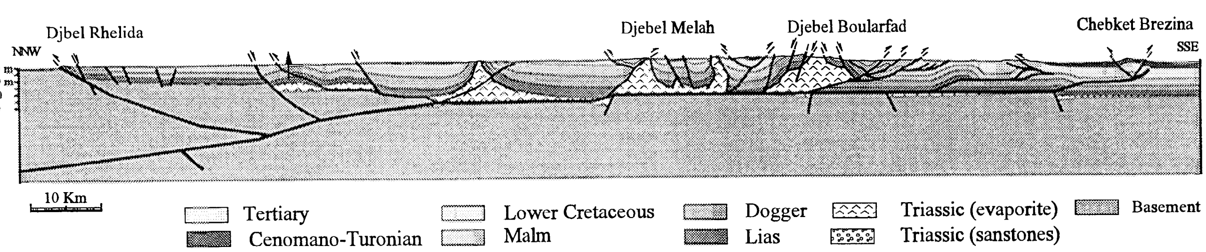

Figure 10. Cross section through the Saharan Atlas from the Sahara Platform to the High Plateaux domain. Source of data are field work and SONATRACH subsurface database. See location in Figure 2.

a basement fault of Triassic age is clearly imaged by subsurface data. Inversion of the Saharan Atlas basin does not lead to the reactivating of this fault but to the development of a new fault branched on the Triassic décollement.

The comparison between the two borders of the Sahara Atlas led *Bracène et al.* [1998] to propose for the Sahara Atlas basin an initial geometry close to that of a half graben, assuming that the major Triassic and Liassic normal faults were situated along the present-day High Plateaux-Atlas boundary.

During the Tertiary inversion the High Plateaux acted as a rigid buttress that was translated southward on an intracrustal décollement. This lower décollement joins the Triassic-Liassic décollement flooring the Atlas near the High Plateaux-Atlas boundary. On our section (Figure 10), we have chosen to draw the connection at the deepest point of the basin. South of this branching point, we consider that the basement is not involved in the Atlas folds. Nevertheless, we observe a change in folding style between the southern front characterized by ramp-flat geometry and the Saharan belt itself where detachment folds with salt intrusions in their cores dominate [*Vially et al.*, 1994; *Bracène et al.*, 1998]. The thin-skinned shortening ratio is of ~20% inferring 20 km of rigid translation for the High Plateaux.

Kinematic modelling of some structures of the Saharan Atlas by *Bracène et al.*, [1998] argues for a polyphase deformation history. However, because of the absence of Paleogene and Neogene deposits within the Saharan Atlas, it is impossible to discuss the effects of the two successive tectonic events observed in the western region. Within the Sahara Platform and in the Hodna basin (Figure 3), situated at the eastern triple junction between the Tell Atlas, the High Plateaux, and the Saharan Atlas, subsurface data show unconformities between Eocene and Oligocene and between Pliocene and Pleistocene strata [*Vially et al.*, 1994; *Mekireche et al.*, 1998]. They likely can be related to the major phases of the Atlas building. In northern part of the Hodna basin, an additional compressional event is isolated within the late Miocene [*Mekireche et al.*, 1998]. We will see below that such a "Tortonian event," related to the emplacement of the Tell front, characterizes the easternmost segment.

3.4. The Biskra-Tunis Segment

In this zone, which is separated from the Saharan Atlas by the Hodna basin, the High Plateaux do no longer exist, and the Atlas (i.e., Aurès and Tunisian Atlas, Figure 2) is in contact with the Tell. To the south and the east it is flanked by the Sahara and Pelagian platforms, respectively (Figure 11). The Aurès Mountains peak reaches 2328 m at Djebel Chelia (Figure 11b). From this point the relief decreases toward the platforms as well as toward the Tell (Plate 1 and Figure 3).

Even if the High Plateaux domain is absent, paleogeographic studies show that a platform existed between the Atlas basin and the Tell passive margin. This platform, referred to as the "neritic Constantine unit" [*Vila*, 1980] in Algeria and the "folded foreland" in Tunisia [*Wildi*, 1983], is now allochthonous and integrated in the Tell thrust belt [*Vila*, 1980]. Immediately south of this shelf, the Atlas basin was particularly deep (up to 10,000 m of sediments) in the Aurès and Tunisia Trough (Figure 4).

As in the Saharan Atlas, this strongly subsiding basin was connected to the south with the Sahara Platform and to the east with the Pelagian platform. At the western border of the Pelagian platform, the N-S axis, which represents the deformation front of the Tunisian Atlas, is associated with normal faults that developed during a Mid-Jurassic-Early Cretaceous rifting episode (Figure 11a). The whole Eo-Cretaceous extensional tectonics is sealed by the marine transgression that started in the middle Albian and reached its maximum in the upper Cenomanian. This rifting, unknown in the western area, is responsible for the development of NW-SE to N-S trending normal faults that parallel those of the Sirt basin north of Libya and its northward prolongation in the Pelagian Sea [*Burollet and Ellouz*, 1986].

These inherited tectonic patterns are related to two successive rifting episodes. During the Lower Jurassic (Tethyan rifting cycle), NNE-SSW trending faults developed. During the Lower Cretaceous (Sirt rifting cycle), NW-SE to N-S trending faults developed [*Ben Ferjani et al.*, 1990; *Delteil et al.*, 1991]. This complex rifting history explains the observed thickness variation in Jurassic and Early Cretaceous sediments.

Local and diachronous unconformities, observed in Upper Cretaceous to Paleocene series, are interpreted as resulting from diapir emplacement [*Ben Ferjani et al.*, 1990; *Perthuisot et al.*, 1999] or discrete compressional [*Touati*, 1985; *Boccaletti et al.*, 1988] or extensional [*Gourmelen*, 1984; *Ouali*, 1985] events. In any case, it is clear that these movements occurred within the basins before its true inversion, which occurred later.

It is established that the building of the Aurès and Tunisia Atlas results from successive and distinct tectonic events recognized decades ago [*Laffitte, 1939a, b; Castany*, 1949; *Burollet*, 1956]. Two events are generally distinguished on the basis of fold trends and observed unconformities. In the Aurès a first event, responsible for the development of NE-SW trending folds (the "Atlas" folds), occurred between the Middle Eocene and the Aquitanian [*Laffitte, 1939a, b; Guiraud*, 1975; *Ghandriche*, 1991; *Frizon de Lamotte et al.*, 1998]. In the Tunisian Atlas., the same name ("Atlas event") refers to an event which is Tortonian in age and which is responsible for folds exhibiting also mean NE-SW trends [*Soyer and Tricart*, 1989; *Ben Ferjani et al.*, 1990, *Tricart et al.*, 1994].

In the Tunisia trough, prolonging the Aurès eastward, *Snoke et al.* [1988] shown an important unconformity between molasses and folded Eocene beds. The molasses are not precisely dated, but the authors present convincing arguments allowing their attribution to the Lower Miocene. South of the Tunisian trough, only some smooth structures can be related to this event [*Bramaud et al.*, 1976; *Perthuisot*, 1978]. Consequently, we will consider that a late Eocene event occurred everywhere along the studied segment. However, it is restricted to the inner part of the Atlas, corresponding to the deeper basin (Aurès and Tunisia trough, Figure 11d)

In Tunisia, the effects of the Tortonian event extend far south of the late Eocene deformation front: the Tortonian deformation front coincides with two major inherited faults, the Gafsa fault and the N-S axis [*Outtani et al.*, 1994; *Tricart et al.*, 1994](Figure 11d). Folds trends mainly NE-SW, but close to the front, they parallel the Gafsa and N-S axis faults [*Yaïch*, 1986; *Creuzot et al.*, 1993]. In the core of the Aurès, as well as in the Hodna basin, newly formed E-W thrust faults reactivate or cut through late Eocene NE-SW folds [*Ghandriche*, 1991; *Addoum*, 1995; *Mercier et al.*, 1995; *Frizon de Lamotte et al.*, 1998;

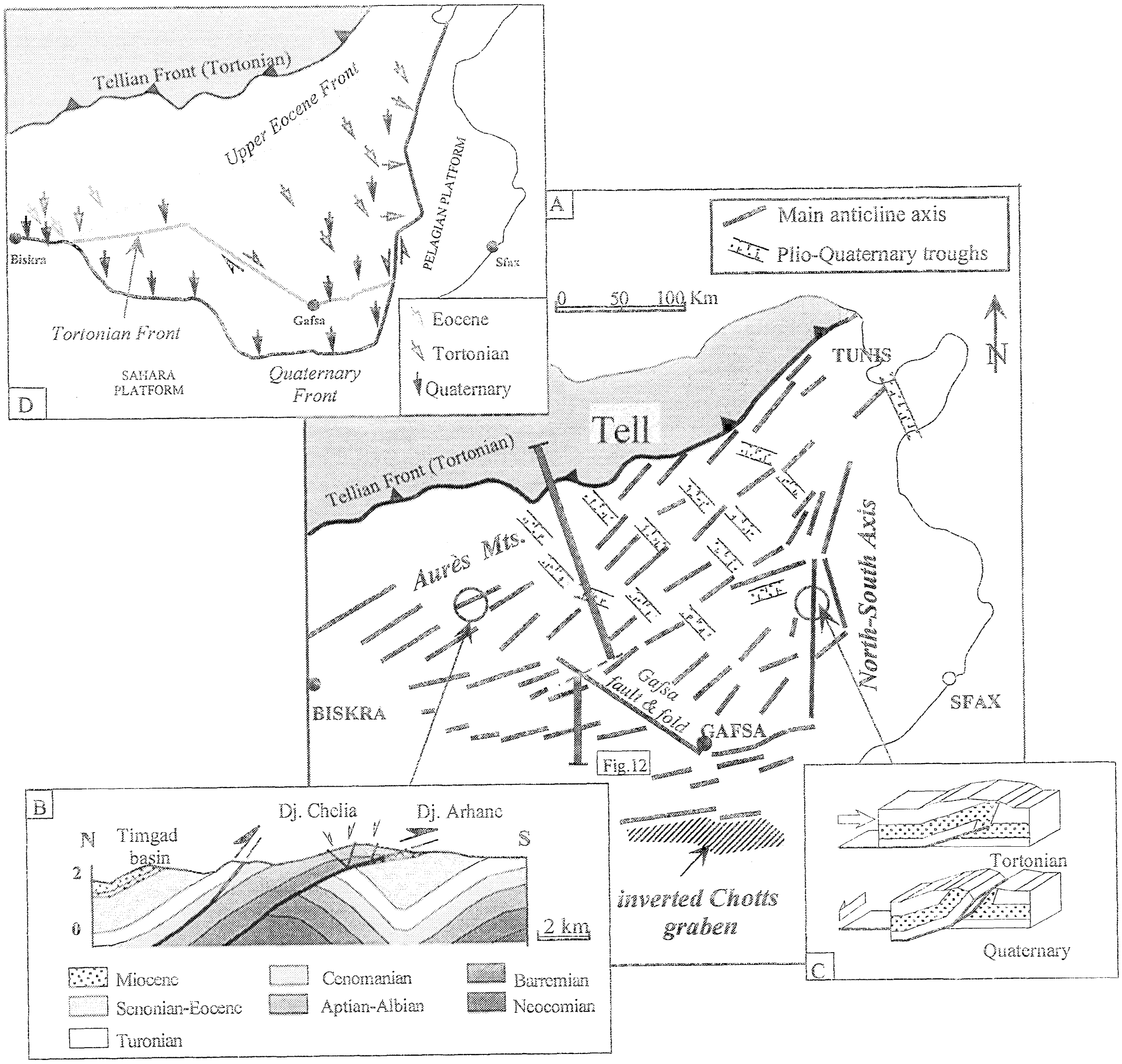

Figure 11. (a) Schematic map of the Aurès and Tunisian Atlas. The three successive deformation fronts of upper Eocene, Tortonian and, Pleistocene-lower Quaternary ages are underlined. (b) A late Eocene fold cut out by a Tortonian thrust fault in the Aurès Mountains (modified from *Ghandriche* [1991]). (c) A Tortonian fault propagation fold cut out by a lower Quaternary strike slip fault along the N-S axis (modified from *Creuzot et al.* [1993]). (d) The available kinematic indicators referring to the late Eocene, Tortonian, and Pleistocene lower Quaternary fronts.

Mekireche et al., 1998](Figure 11a). Pleistocene conglomerates rest unconformably on these new structures [*Ghandriche*, 1991].Therefore it is possible to correlate them to the Tortonian structures known in Tunisia, and we can assume that a curved fold-thrust belt was formed during the Tortonian between Biskra and Tunis (Figure 11b). According to *Tricart et al.*[1994] and *Casero and Roure* [1994], we acknowledge an offshore extension of this fold belt through the Sicily Channel. In contrast, there is no evidence that this fold belt extended westward. We emphasize the coincidence of the Tortonian fold-belt with the zone where the Tell directly overthrusts the Atlas (Figure 11 and 12).

South of the Aurès as well as south of the Gafsa fault in Tunisia, the present-day deformation front involves Plio-Pleistocene conglomerates [*Laffitte, 1939a, b; Guiraud*, 1975; *Zargouni*, 1985; *Frizon de Lamotte et al.*, 1990; *Ghandriche*, 1991; *Tricart et al.*, 1994, *Addoum*, 1995; *Outtani et al.*, 1995; *Frizon de Lamotte et al.*, 1998]. After a new phase of quiescence, during the Messinian and the Pliocene, the deformation front advanced to its present-day position during a late Plio-Pleistocene event. The Tortonian and recent fronts are superimposed on each other west of Biskra and along the N-S axis up to Tunis but are well separated in the central area (Figure 11c). Reactivating of the front and its southward propagation

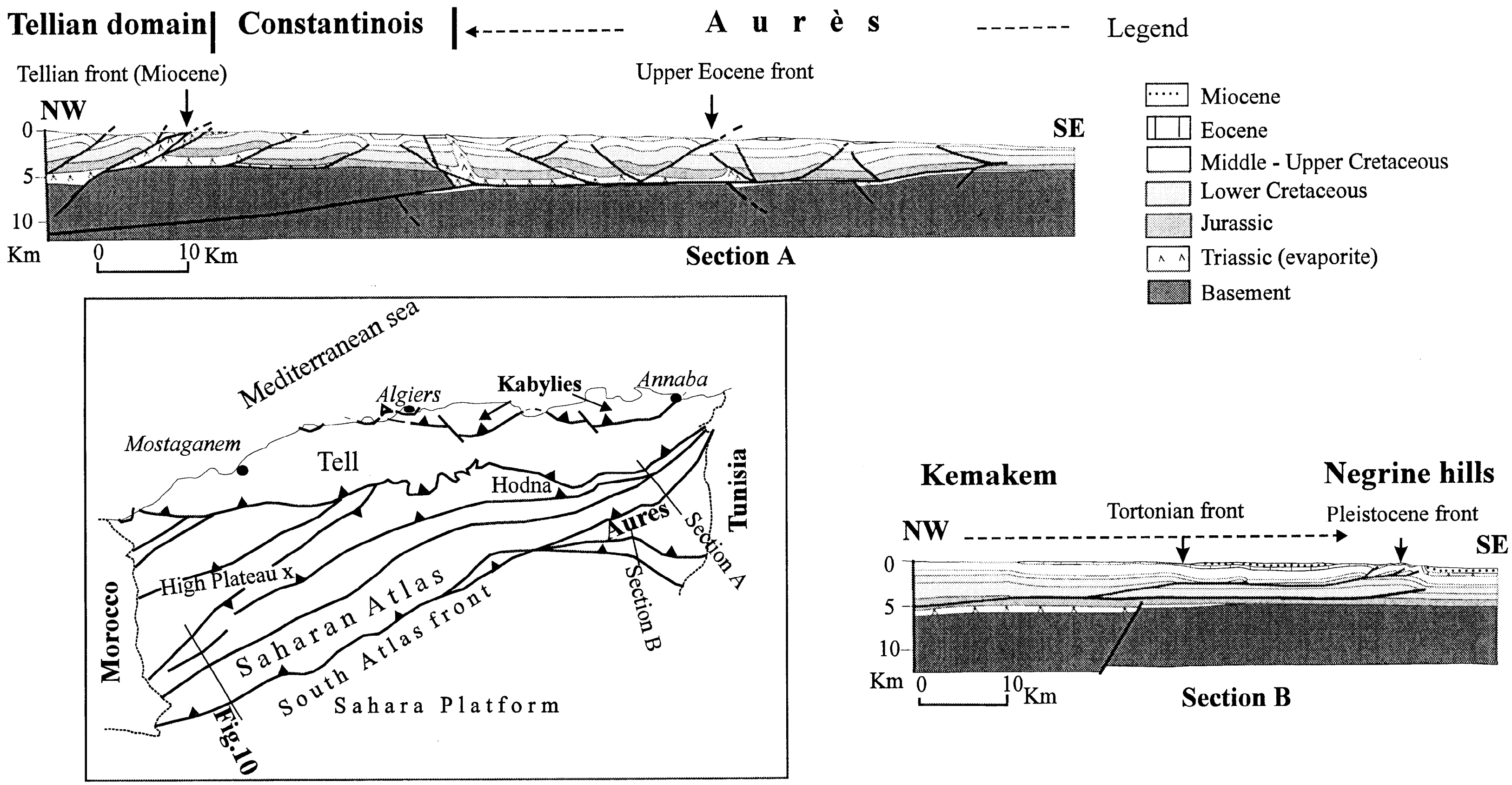

Figure 12. Cross section through the Aurès Mountains (see location in Figure 11). Sources of data are *Addoum* [1995], *Outtani et al.*, [1995], and SONATRACH subsuface database.

involved upramping of the basal décollement from Triassic evaporites into intra-Cretaceous levels [*Outtani et al.*, 1995](Figure 12). Eastward, the N-S axis acted as a lateral ramp (Figure 11b). To the south the "chaîne des Chotts" is interpreted as an inverted Jurassic graben. As proposed for the Anti-Atlas, it is likely that an intrabasement décollement was activated during the inversion of the Chotts graben [*Outtani et al.*, 1995; *Frizon de Lamotte et al.*, 1998].

Contrary to the complex pattern of the Tortonian folds and thrusts, the Plio-Pleistocene structures exhibit E-W trends, suggesting quite constant N-S shortening and a weak influence of the inherited framework (Figure 11d). This is in agreement with the position of the front within the platform (Figure12).

The Biskra-Tunis segment is characterized by three distinct deformation fronts of late Eocene, Tortonian and Plio-Pleistocene age, respectively (Figure 11 and 12). In addition, an extensional event, not discussed in this paper, took place before the Tortonian folding (see review by *Bouaziz et al.*, [1998]). The first and last compressional events are known elsewhere in the Atlas Mountains and can be considered as attributes of the Atlas system. The second is restricted to the Biskra-Tunis segment. We emphasize its coincidence with the final nappes emplacement along the Tell front [*Vially et al.*, 1994; *Tricart et al.*, 1994]. Correspondingly, the Biskra-Tunis segment exhibits superimposed structures which have different geodynamic significance. An interpretation of this segment as a foreland fold-thrust belt is only relevant for the upper Miocene period.

Because of the superimposition of different tectonic events, shortening ratios are difficult to establish. *Outtani et al.* [1995] give a maximum translation of 4 km for the part of the domain that is situated south of the Aurès between the Tortonian and Quaternary fronts. For the same event, displacements amount to 15-20 km along the N-S axis [*Creuzot et al.*, 1993]. By comparison, the shortening achieved during the late Eocene and Tortonian events is probably smaller, because it is mainly associated with the development of detachment or fault propagation folds without breakthrough thrusting (Figure 12). A total shortening ratio of ~30% seems reasonable.

3.5. Discussion

Whereas the Moroccan High Atlas is a thick-skinned thrust belt, the Saharan Atlas, the Aurès, and the Tunisian Atlas must be considered as thin-skinned belts with Triassic-lower Liassic evaporites providing a basal décollement. The transition between the two structural styles cannot be clearly defined but occurs near the Morocco-Algeria boundary in a zone where the orientation of the belts changes from an E-W to a NE-SW strike (Figure 2 and Plate 1).

Our assumption that an intracrustal detachement level was (and likely remains) active below the Anti-Atlas implies that the southern limit of Neogene basement thrusting generally does not coincide with the South Atlas Front. In Morocco it is located somewhere south of the Atlas whereas in Algeria and Tunisia it follows firstly the High Plateaux and then the Constantine Unit in eastern Algeria. This line corresponds also, more or less, to the southern limit of the Variscan orogeny as defined by *Piqué et al.* [1998a].

Neglecting the minor inversion that occurred at the end of the Cretaceous, the first clearly expressed compressive event occurred in the Atlas after the Lutetian and before the Burdigalian. This so-called "Atlas event" of authors working in Algeria [*Laffitte, 1939a, b; Guiraud*, 1975; *Ghandriche*, 1991; *Addoum*, 1995], is well documented in the Aurès and in the Hodna basin (see above) and is generally evident throughout the Saharan Atlas where the major folds trend NE-SW as in the Aurès. It probably also extends into the Tunisian trough. In Morocco the Souss and Ouarzazate regions give evidence supporting the existence of an important late Eocene event; unfortunately, because of the importance of subsequent deformations, we have very few indications on the geometry of the related structures. In the Middle Atlas (Morocco), *Charrière* [1990] distinguished two Tertiary folding and thrusting events. The first one, which we propose to correlate with the "Atlas event," occurred after the middle Eocene and is sealed by upper Miocene sediments. The timing of this event is not exactly known. However, a late Eocene age (35 Ma) agrees with all available data.

The second major tectonic event characterizing the Atlas system occurred during the Pleistocene and the lower Quaternary. It is responsible for the uplift of the entire Atlas system and for the development of important thrust-related structures distributed all over the domain.

An additional event of Tortonian age is due to the interference between Atlas and Tell systems. It is confined to the Biskra-Tunis segment where the two systems are adjacent.

4. The Atlas System in the General Context of Western Mediterranean Geodynamics

In this section we address the significance of the tectonic events recognized within the Atlas system and their relationship with the main evolutionary stages of the western Mediterranean. In Figures 13 and 14 we present a kinematic model for the evolution of the western Mediterranean region along a transect from the Sahara Platform to the Balearic Islands. This model is based on the crustal kinematic model presented by *Vergés and Sàbat* [1999] and has been complemented for Africa by data presented above.

Convergence between Africa and Europe began in the late Cretaceous [e.g., *Le Pichon et al.*, 1988; *Dewey et al.*, 1989; *Stampfli et al.*, 1991; *Dercourt et al.* 1993; *Ricou*, 1994]. The system evolved by underthrusting of Iberia under Europe leading to the development of the Pyrenees [*ECORS Pyrenees team*, 1988; *Roure et al.*, 1989; *Muñoz*, 1992] and by thrusting of Apulia over Europe leading to the development of the Alps [*Nicolas et al.*, 1996]. According to *Vergés et al.* [1995] and to *Meigs et al.* [1996], the Pyrenean system main phase of building took place from the lower Eocene up to the upper Oligocene (i.e. from 55 to 25 Ma). On the other hand, thrusting in the Alps took place since the Cretaceous up to the Miocene. However, it is established that the Ligurian Tethys was already closed at the beginning of Eocene times [*Nicolas et al.*, 1996].

In the "Dorsale Calcaire" of the internal Tell-Rif (belonging to the ALKAPECA terranes and forming the northern margin of the southern Tethys), the first compressive tectonic event occurred only at the end of the Eocene [*Raoult*, 1968; *Bouillin*, 1977; *Wildi*, 1983]. Traditionally, this tectonic event has been interpreted as resulting from the "collision" of the ALKAPECA terranes with the African margin [*Bouillin*, 1986; *Aïté and Gélard*, 1997]. On the contrary, *Vergés and Sàbat* [1999]

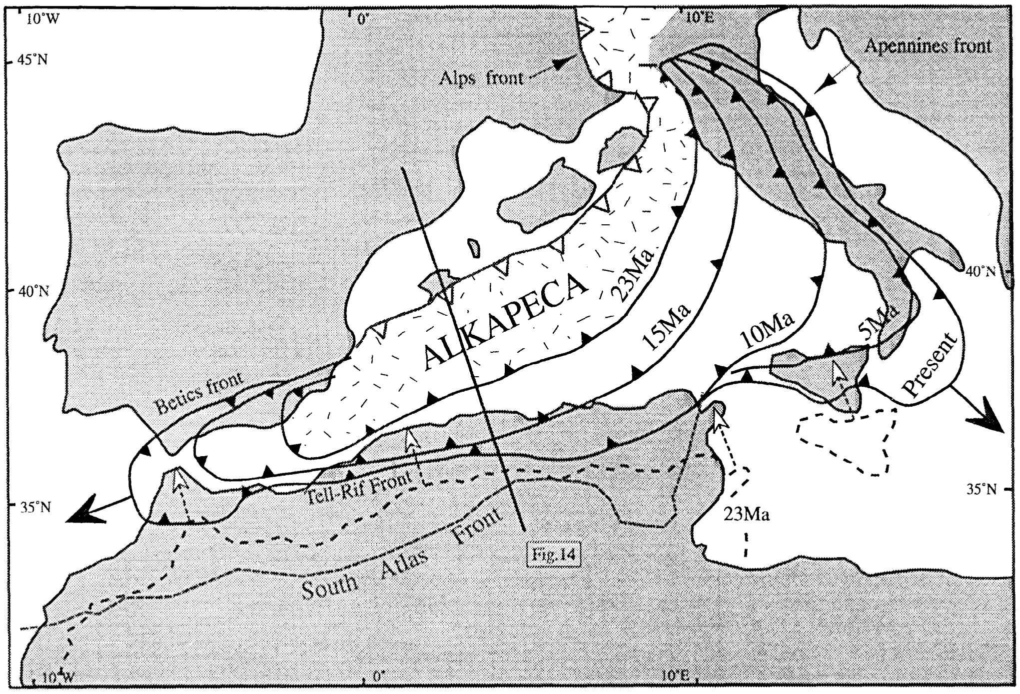

Figure 13. Schematic map illustrating the successive geometry of the Tell-Rif-Apennine subduction starting in the western Mediterranean during Oligocene (strongly modified for the western region from *Doglioni et al.*[1999]). See discussion in the text. ALKAPECA: Alboran, Kabylies, Peloritan, and Calabria.

consider that the accretion of ALKAPECA to Africa is much later. This complex problem regards two related questions: the significance of the metamorphism observed in the basement complexes of the internal Tell-Rif (as well as in the Betics) and the interpretation of the Algerian and Alboran basins.

Following the "classical" interpretation (summarized by *Comas et al.* [1999]), the high-pressure/low-temperature mineral assemblages occurring in these complexes resulted from Cretaceous and Paleocene subduction and then direct collision between Africa and Europe. In the same way the later low-pressure/high-temperature metamorphism is interpreted as resulting from subsequent exhumation. In contrast, *Doglioni et al.* [1997, 1999] emphasize the great differences between the Alps-Betics and the Maghrebides (i.e., Tell-Rif)-Apennines orogenic systems. According to these authors, we assume that if a collision zone existed in the western Mediterranean region, it must be sought in the prolongation of Alpine Corsica in relation with the closing of the Alpine Tethys (i.e., "collision" between Apulia and Europe). The history recorded in the metamorphic nappe complexes of the internal Betics-Rif [e.g., *De Jong*, 1991; *Platt and Vissers*, 1989; *Vissers et al.*, 1995; *Comas et al.*, 1999] and of the Kabylies [*Saadallah and Caby*, 1994] gives evidence for such an early collisional belt. However, the subsequent translations, leading to the breaking up and collapse of the metamorphic cores, have obscured its actual geometry.

The choice on kinematic scenario governs the interpretation of the Algerian and Alboran basins as postcollisional or as back arc basins. The age of opening of the Algerian basin is not well constrained. However, it is established that it postdates, at least partially, the opening of the Valencia Trough where middle Miocene strata overstep most of the major normal faults [*Vergés and Sàbat*, 1999]. So, if one accepts that accretion of ALKAPECA terranes to Africa was achieved before the Algerian basin opened, then a divergence of Africa with respect to stable Europe would be recorded. Such a scenario disregards plate kinematics, which indicates constant convergence [*Le Pichon et al.*, 1988; *Dewey et al.*, 1989; *Dercourt et al.*, 1993; *Ricou*, 1994].

As late Eocene compressional structures are unknown in the Tell-Rif external zones [*Leblanc*, 1979; *Vila*, 1980; *Wildi*, 1983; *Frizon de Lamotte*, 1987; *Favre*, 1995], we will consider that the late Eocene compression observed in ALKAPECA terranes and in the Atlas and Sahara domains is related to the initiation (or at least to an acceleration) of the subduction of the Maghrebian Tethys underneath the Balearic margin and that the ALKAPECA domain was, at that time, attached to it. Moreover, the initiation of rifting and calc-alkaline subduction-related volcanism began in the Valencia Trough during upper Oligocene [*Sàbat, et al.*, 1997]. This is in agreement with a model by *Ziegler et al.* [1998] showing that inception of a new subduction zone generates compressional deformations within the conjugate lower plate.

Thus the onset of Maghrebian Tethys subduction, as well as the late Eocene Atlas event of North Africa, seems linked to a first blockage within the ALKAPECA domain which was no longer able to absorb the Africa-Europe convergence. Taking into account the geometry of the Variscan chain [*Piqué et al.*, 1998a], we will assume that the Atlas late Eocene event is due to the reactivation of an old Variscan fault zone already used during the Triassic-Liassic rifting. The NE-SW folds are parallel to the inherited trends but also normal to the convergence vector

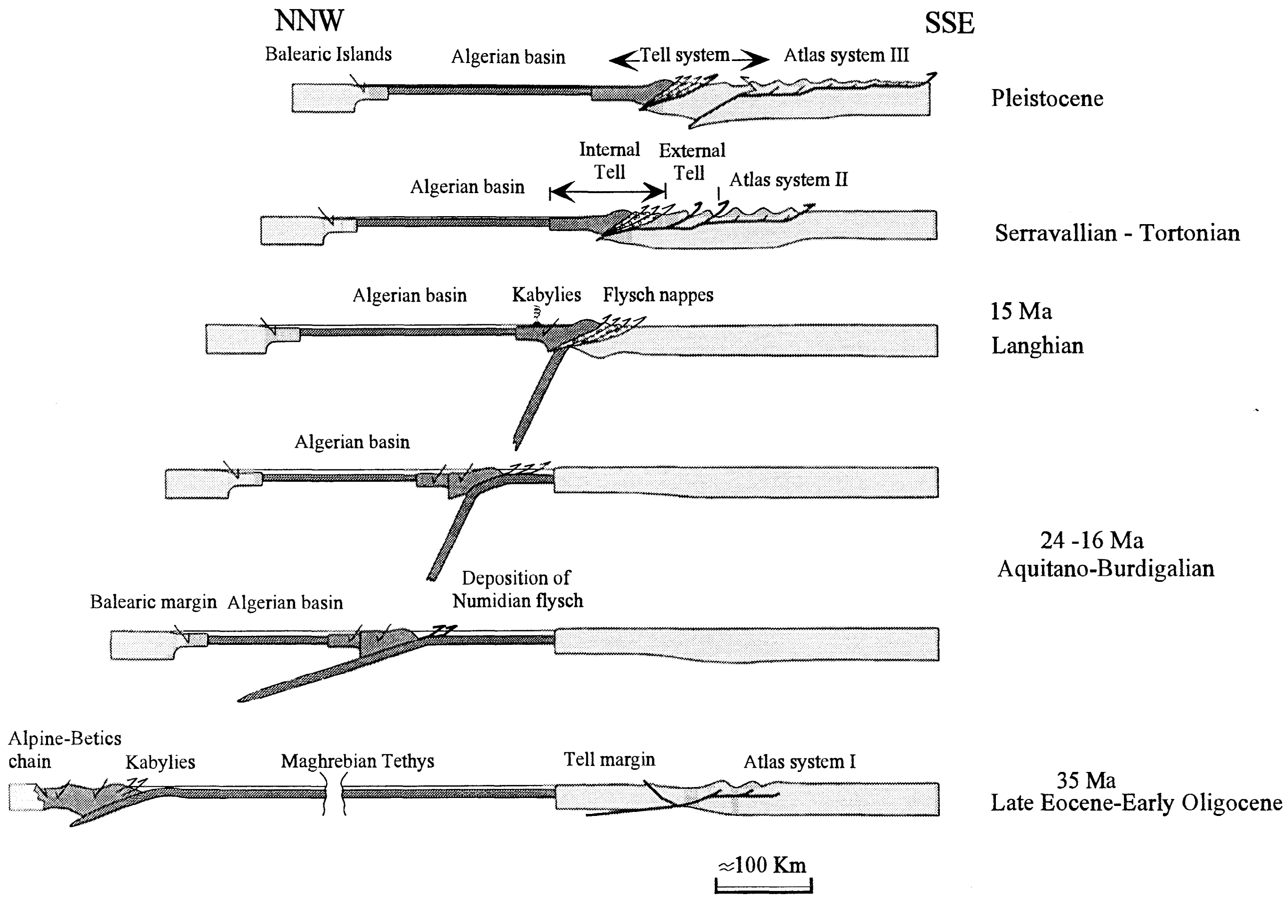

Figure 14. Kinematic model illustrating the proposed scenario along a transect from the Balearic margin to the Sahara platform (modified and completed from *Vergés and Sàbat* [1999]). The transect is located in Figure 13. The model comprises six steps from late Eocene to Pleistocene. See explanations in the text.

of Africa with respect to stable Europe if we account for the change in the convergence direction occurring at ~35 Ma [see *Le Pichon et al.*, 1988].

During the upper Oligocene-lower Miocene, Africa-Europe convergence is mainly absorbed along the Iberian-Balearic margin (including the ALKAPECA domain). The subduction-related back arc extension is marked within the ALKAPECA domain and expressed by normal faults located offshore as well as onshore [*Comas et al.*, 1992; *Watts et al.*, 1993; *Tricart et al.*, 1994; *Saadallah and Caby*, 1994; *Aïté and Gélard*, 1997]. Southward, the residual Maghrebian Tethys was characterized, at that time, by the deposition of the Numidian flysch [*Vila et al.*, 1995; *Esteras et al.*, 1996]. Following *Lonergan and White* [1997], we will assume that roll-back of the subducted oceanic lithospheric slab led firstly to accretion of Kabylies to Africa (Figures 13 and 14). These authors assign this accretion to 18 Ma. However, a younger age (15 Ma) is compatible with the data if we consider that the thrust tectonics, subsequent to accretion, developed during the Serravalian in the Tell [*Wildi*, 1983] and up to the Tortonian in the Atlas (see above). Moreover, *Piqué et al.* [1998b], synthesising magmatic data, show that the older "post-collision" calc-alkaline magmatic activity of Langhian age (15 Ma) in the Kabylies.

After the accretion of the Kabylies the Tethyan slab split into two different parts: one of which rolled back to the east controlling opening of the Tyrrhenian Sea and the second of which rolled back westward controlling opening of the Alboran Sea [*Lonergan and White*, 1997] (Figure 13). This scenario is consistent with "postcollision" magmatism at ~10 Ma in the Rif [*Piqué et al.*, 1998b] and with westward and eastward thrusting observed in the external Betics-Rif during upper Miocene and in the outer Calabrian Arc up to now (Figure 13). Such a scenario is quite different from the one developed by *Doglioni et al.* [1997, 1999] suggesting that the whole North Africa orogen results from the eastward retreat of the Apenninic subduction. This process alone does not give an adequate explanation for the westward translations observed in the Betics-Rif system during Neogene times [*Frizon de Lamotte et al.*, 1991; *Vissers et al.*, 1995; *Lonergan and White*, 1997).

In our interpretation the extensional structures developed within the ALKAPECA terranes during lower to middle Miocene and in the backarc west Mediterranean basins must be considered as being driven by the retreat of the subduction slab [*Doglioni et al.*, 1999; *Vergés and Sàbat*, 1999] and not by post-collisional collapse as proposed by some authors [*Platt and Vissers*, 1989; *Tricart et al.*, 1994; *Comas et al.*, 1999].

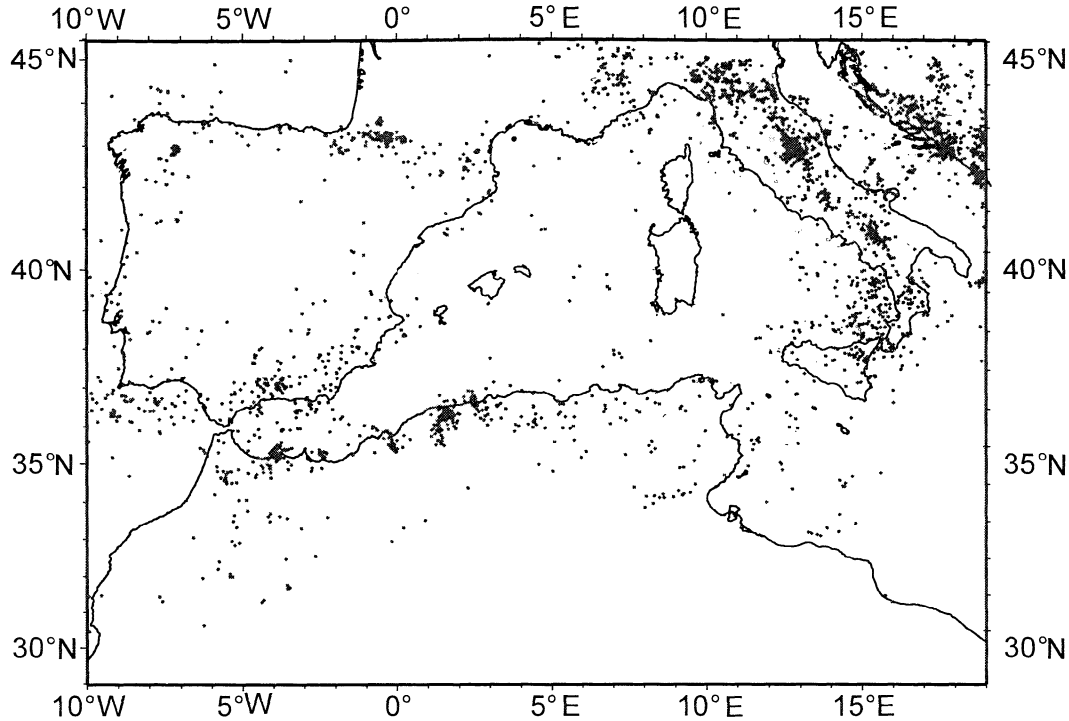

Figure 15. Map showing the repartition of the present-day seismicity within the western Mediterranean region. Seismicity is concentrated along the Tell-Rif and Apennines systems but affects diffusely the whole domain from the South Atlas Front to the Pyrenees.

In eastern Algeria and Tunisia, thrusting propagated southward since 15 Ma from the front of the Kabylies through the former Tethys margin (Tell) and then through the Atlas. Thrust propagation is recorded by successive foreland basins that have been progressively incorporated into the orogenic wedge [*Ghandriche*, 1991]. We have shown that in the Eastern Atlas, folds of various orientations have successively developed during this period owing to complex inherited patterns (Figure 11). In any case, it appears that the Eastern Atlas acted, at that time, as the foreland fold-thrust belt of the Tell system.

However, if we follow the Atlas system along strike, it is clear from maps (Figure 14) that oblique continental accretion occurring in the Rif (with a left-lateral component) during the upper Miocene cannot be responsible for southvergent thrusting in the High Atlas during the Pleistocene and lower Quaternary. Consequently, we must consider that the Tell-Rif system, which looks like an accretionary prism rather than a collision belt, was no longer active at that time. The effects of the Europe-Africa convergence were apparently transferred southward into the African continent where the convergence is responsible for the main uplift of the entire Atlas Mountains.

Therefore we consider that during the middle Miocene to lower Quaternay period, convergence of the Africa and Europe plates was absorbed in two different ways: (1) the subduction roll back of the Maghrebian Tethys lithosphere responsible for the divergent translations of Rif and Calabria and the coeval opening of the Algerian, Alboran, and Thyrrenian basins during the Miocene-Pliocene (Figure 13) and (2) the growth of the Atlas Mountains during Pleistocene and lower Quaternary.

Since the lower Quaternary, deformation was mainly concentrated within the Atlas and, particularly, along the South Atlas Front and thus presents a quite different distribution as illustrated by the present-day seismicity (Figure 15). Although the Atlas remains active, the seismicity as well as the predominant newly formed structures is located along a zone extending from the Goringe Bank and cutting through the Alboran Sea and the external Tell Mountains [*Morel and Meghraoui*, 1996]. The Goringe Bank and Alboran Ridge anticlines, as well as the El Asnam ramp-related fold, are major structures that developed along this zone. They are interpreted as "en échelon" folds underlining the present-day plate boundary. At a regional scale this fault zone appears as an out-of-sequence thrust zone cutting through various structural domains.

5. Conclusion

Two specific stages (upper Eocene and Pleistocene) are recognized in the development of the Atlas. They correspond roughly to the initiation and the cessation of the subduction of the Maghrebian Tethys and the related formation of the Tell-Rif accretionary prism. This leads us to reconsider the classical interpretation of the North African orogeny which relates it to collision processes between the Europe and Africa plates [*Durand-Delga and Fontboté*, 1980].

In fact, and compared to other mountain belts, the tectonic development of the North African chains appears as rather original. In a collision belt, classically, the main characteristic features are as follows [e.g., *Boyer and Elliott*, 1982]: (1) The deformation propagates progressively from the suture zone to the foreland fold-thrust belt following a piggy-back sequence, (2) the kinematics of the foreland fold-thrust belt can be directly related to the "en bloc" motion of the previously deformed hinterland, (3) most recent displacements take place along the basal thrust flooring the whole orogenic system.

Probably because of different mechanical coupling between the Tell-Rif wedge and its foreland [*Ziegler et al.*, 1998], the orogenic systems of North Africa present a quite different development: (1) The deformation has been initiated within the most external part of the system (i.e., within the Atlas), (2) the kinematics of the Atlas (foreland) is not in direct relation with the translations observed in the Tell-Rif (hinterland). This is particularly clear in the western regions where the translation of the Rif system is perpendicular to the shortening observed in the High Atlas, (3) the present-day activity is concentrated along an out-of-sequence thrust zone coming from the Atlantic Ocean and cutting across the previously emplaced Tell-Rif.

The accretion of ALKAPECA terranes to Africa cannot be interpreted as a large-scale "continental collision," and certainly not as a collision between Africa and Europe. The geometry of the Rif -Tell accretionary system is linked to the geometry of the subductions responsible for its development. The geometry of the subductions is the reason why the directions of translations observed in this system are not directly related to the relative convergence of Africa with respect to Europe (Figure 13). Moreover the velocity of translations within the Rif system is 5 times greater than the velocity of the convergence.

By contrast, the Atlas remains insensitive to subduction processes active in the western Mediterranean region but has faithfully recorded the main jolts and the direction of the Europe-Africa convergence. Since the lower Quaternary, the northward motion of Africa is accommodated by shortening concentrated along the Tell-Rif margin but affecting diffusedly the whole western Mediterranean region from the South Atlas Front to the Pyrenean thrust system [*Vergés and Sàbat*, 1999] (Figure 15). It is likely that the region will evolve by subduction of the Alboran-Algerian basins under Africa and then by a "true" collision between Europe and Africa.

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