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Deep-rooted "thick skinned" model for the High Atlas Mountains (Morocco). Implications for the structural inheritance of the southern Tethys passive margin

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Abstract

A re-interpretation of the deep structure of the High Atlas is presented through integration of geophysical and geological data, highlighting the architectural significance of the southern Tethys palaeomargin. Previous crustal models suggest the occurrence of a flat intra-crustal detachment at a depth of −20 km, a zone where surface thrusts merge and below which the lower High Atlas crust appears continuous. However, within this study seismic refraction data, electrical resistivity surveys and gravity modelling all appear to detect a jump in crustal thickness between the High Atlas and the northern plains. We interpret these data as penetration by thrusts within the "South-Atlasic fault" zone through the lower crust to offset the Moho in accordance with a "deep-rooted thick skinned tectonic model". The "South-Atlasic fault" zone corresponds to a series of crustal-scale structures inherited from a series earlier orogenic events (Pan-African, Hercynian and Atlasic) and thus is sufficiently weak to create separate crustal blocks in the northern part of the stable African plate. This structural configuration was achieved during Cenozoic plate collision between Africa and Europe culminating in tectonic inversion of Mesozoic extensional basins, linked to the development of Tethys and opening of the Atlantic. The inherited structures typically comprise normal faults and tilted blocks affecting the upper crust, and crustal-scale detachments affecting the lower crust.

Keywords: Atlas Mountains; Rif; Thick skinned tectonics; Jurassic rifting; South-Atlasic fault; Tethys and Atlantic margins; Structural inheritance

1. Introduction

In Morocco, three main structural domains can be distinguished (Figs. 1 and 2). From North to South these are the Rif, the Atlas Mountains and the Anti Atlas. The Rifian chain faces the active margin to the Alboran Sea (Mediterranean Sea). It shows Alpine type tectonics with nappes thrust to the south. The Anti Atlas, is a marginal updoming of the Precambrian west Africa craton which acted as stable crust during Alpine orogeny. The Atlas Mountains (High and Middle Atlas) are developed from Mesozoic rift grabens. These rifts are reversed in Cenozoic times to form the present-day Atlas range, which rises to 4165 m, one of the highest peaks in the peri-Mediterranean Alpine system (Beauchamp et al., 1999). These mountain belts are separated by two rigid and stable Palaeozoic blocks: the Moroccan Meseta and the eastern Meseta. The Atlas belt is a tectonically complex area and is a part of the Eurasia−Africa plate boundary region. The regionally deformation records the convergence and collision of the African and Eurasian continents. The High Atlas is separated from the west African craton (Anti Atlas) by a clear physiographic boundary commonly referred to as the "South-Atlasic fault" ("SAF" of Russo and Russo, 1934) or as the "South Atlas front" of Frizon de Lamotte et al. (2000). At

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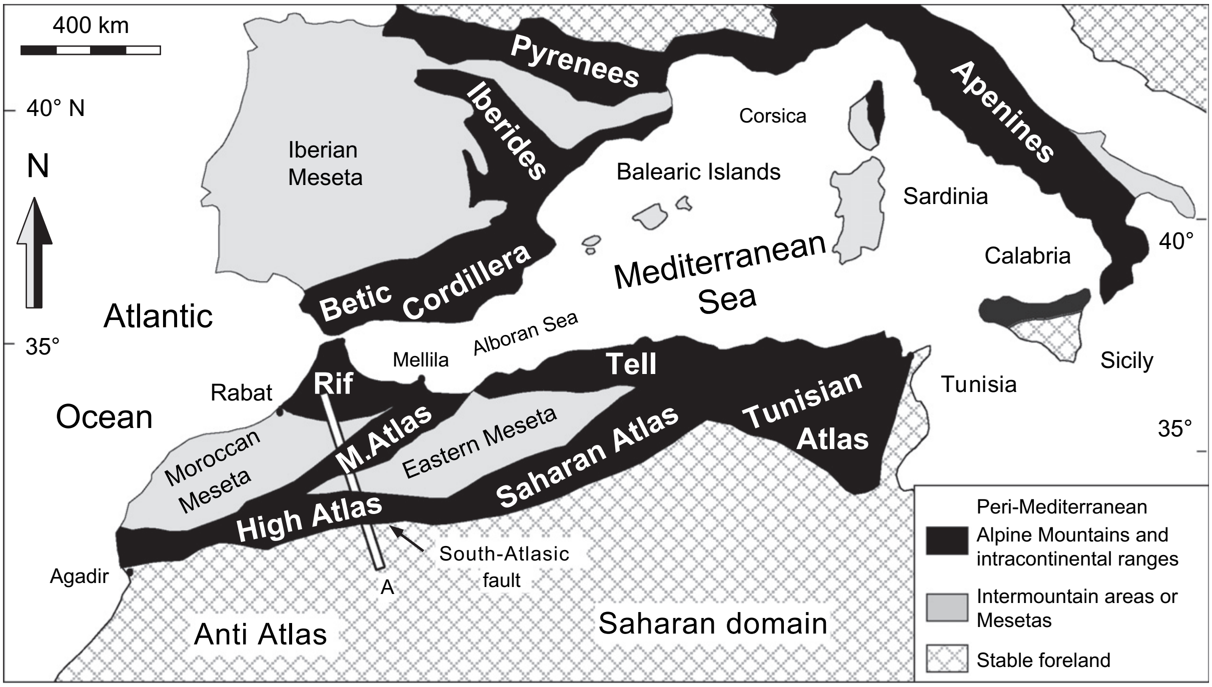


Fig. 1. Sketch map of the western Mediterranean region showing the Atlas range and the peri-Mediterranean alpine belt (after Flinch, 1996). (A) Cross sections of Figs. 6, 9A–D and 11.

present, we have no subsurface data on the prolongation of this "SAF" farther north and a number of fundamental questions remain concerning the tectonic evolution of the Atlas chain. These are

How do faults in the main part of the mountain belt continue to depth?
Are the main tectonic boundaries within the chain inherited from earlier structures that served to delimit older palaeogeographic domains?
Do low angle thrust systems found at outcrop pass steeply at depth?

The aim of this paper is to address these issues of structural geometry and tectonic evolution. It concludes with a discussion of how this new structural understanding sheds light on the geometry and timing of rifting and passive margin formation on the southern side of Tethys during the Mesozoic.

The term inversion is used here, in the sense of Coward (1994), to describe regions that have experienced a reversal in uplift or subsidence. The Atlas Mountains nicely fit this definition; after an extended period of rifting and subsidence during the Mesozoic, the Atlas rift is the locus of tectonic shortening, inversion, uplift and erosion during the later stages of the Alpine orogeny. Tectonic inversion occurs when pre-existing extensional faults or lateral discontinuities in the crust affect the architecture of subsequent compressional structures. Two models have been proposed for the folding and the uprising of the Atlas sedimentary pile. The first model (Laville and Piqué, 1991, 1992) proposes that in the Atlas axis, the folding developed in response to a transcurrent regime along the inherited N70°E basement faults. The second model favours a classical inversion of the Atlasic trough (Beauchamp et al., 1999), with a diapiric origin for the anticlines, developed in the N70°E trending Atlasic rift (Crevallo et al., 1987). The inversion of the former normal faults would result from the transition from a purely extensional to a purely compressional regime (Giese and Jacobshagen, 1992). An intermediate model, proposed recently by Piqué et al. (2002) favours transpressional deformation following transtensive opening, that resulted from a slight variation in the regional stress field.

2. Stratigraphy and tectonic evolution of the Atlasic domain (High and Middle Atlas)

The Moroccan Atlasic domain (Figs. 1 and 2) exhibits two main Atlas orogenic belts to the east: the High Atlas, trending ENE–WSW or E–W, and the Middle Atlas aligned approximately NE–SW. These inverted basins maintain their precontractional structural relationships, allowing the recognition of the synrift and postrift sequences and their initial geometry. The synthetic lithostratigraphy of the High Atlas Mountains and main regional tectonic events are summarized in Fig. 3.

2.1. Prerift Phase (Palaeozoic and Precambrian basement)

The Anti Atlas south of the High Atlas Mountains exposes Precambrian through Palaeozoic age rocks of the west African craton. The Late Proterozoic rocks were affected by the Pan-African orogeny and by a deformation in the Hercynian

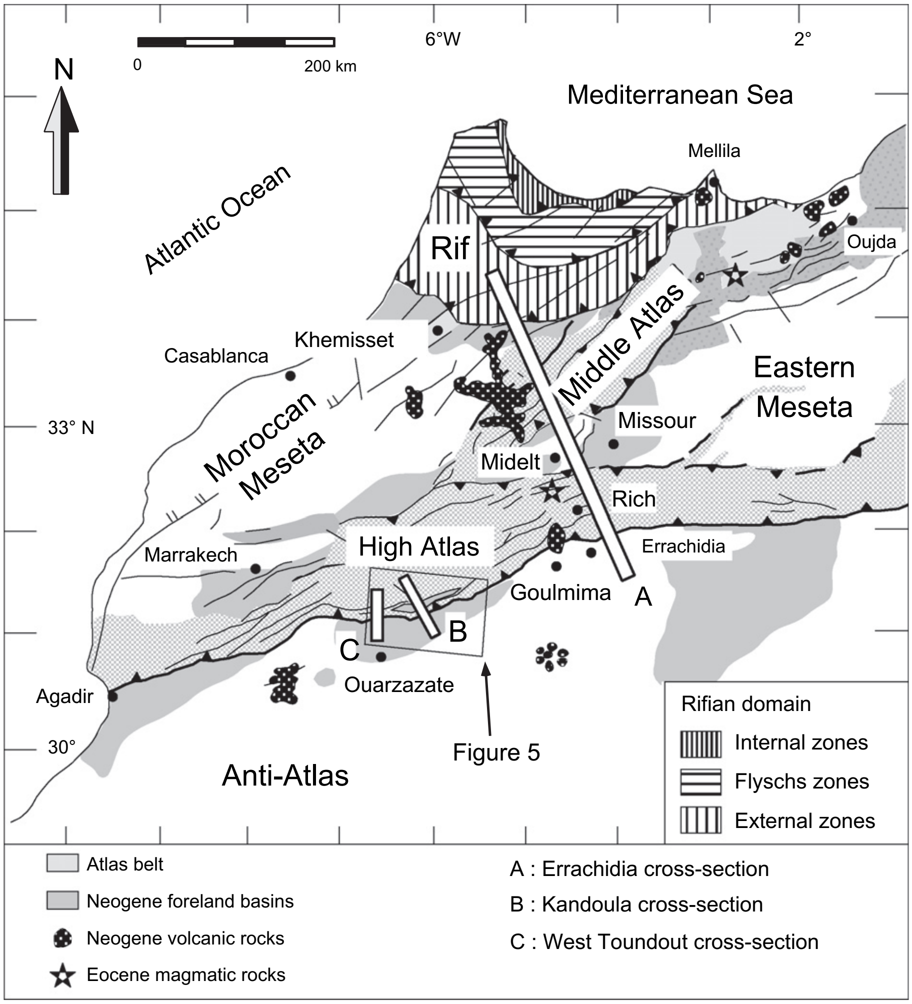


Fig. 2. Simplified structural map of Morocco indicating Neogene and Quaternary volcanic rocks, Eocene magmatic rocks and foreland basins (compiled from Harmand and Cantagrel, 1984; Piqué et al., 2000). (A) Errachidia NW–SE cross-section of Figs. 6, 9A–D and 11; (B) Kandoula NW–SE cross-section of Figs. 5, 7A and 8; (C) West Toundout N–S cross-section.

(Fig. 3). In the Late Palaeozoic, dislocation of the Pangean plate by post-convergent extension resulted in the development of extensive episodes recorded in Late Triassic–Early Jurassic synrift basins initiated by reactivation of older Hercynian–Alleghanian thrusts (Laville et al., 2004).

2.2. Synrift Phase (Triassic and Jurassic)

During most of the Mesozoic, the High Atlas domain experienced extension and rifting. First, during the Triassic, this is recorded by red beds and tholeiitic basalts. Later, during the Jurassic, it is marked by deposition of marine carbonates and shales capped by continental red beds (Michard, 1976). In Mesozoic times, the geodynamic evolution of Morocco and of the 'Atlas trough' in particular was closely related to that of the Atlantic rifting to the west and of the formation of the Tethys to the north. Either of two conflicting major geodynamic models can be used to explain the extensional structuring of the Atlasic basins in the Jurassic: Transfer fault basin model (Laville and Piqué, 1992) and Extensional basin model (''Atlasic rift''; Jacobshagen et al., 1988; Giese and Jacobshagen, 1992).

2.3. Postrift Phase (Cretaceous–present)

At the end of the Jurassic, and during the Cretaceous, large parts of North Africa were flooded by shallow seas that transgressed from the Atlantic and the Tethys Oceans. Postrift sedimentary rocks of the Atlas range are composed of transgressive shallow marine clastic, carbonate and continental sedimentary rocks, varying in age from Infra-Cenomanian to Eocene. In Upper Eocene times, the geotectonic evolution of the High and Middle Atlas changed totally. The previous rift grabens (or half-grabens) underwent compressional/transpressional deformations and, subsequently, a strong uplift happened, which is documented by the frequent appearance of Atlas pebbles within the foreland basins along the southern (El Harfi et al., 2001) and the northern rims of the High Atlas.

2.4. Timing of Atlas deformation and history of uplift

The history of uplift of the High and the Middle Atlas is clearly reflected by the sedimentary filling of the adjacent Cenozoic foreland basins, which largely contain molassic and

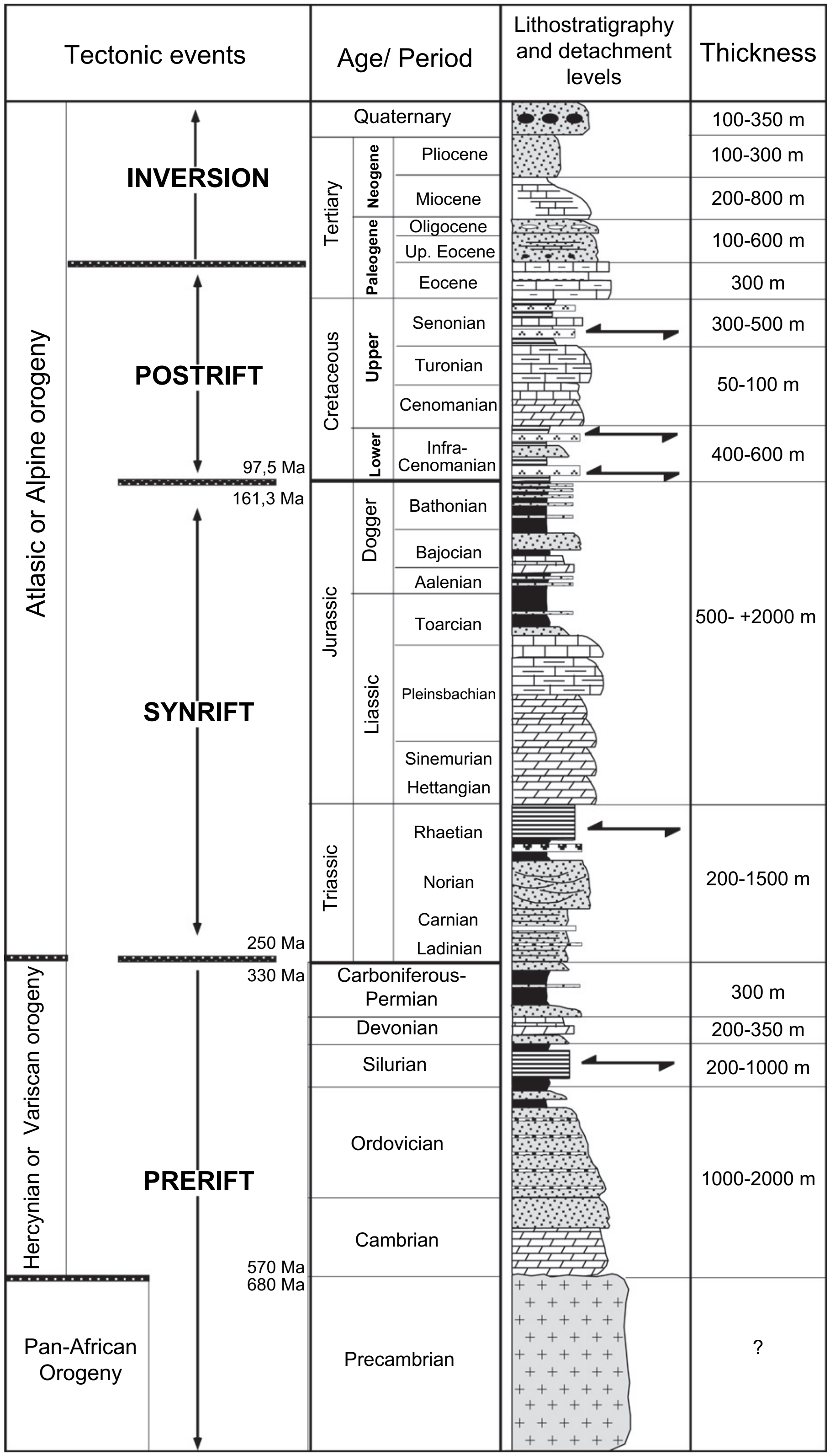


Fig. 3. Synthetic stratigraphic column of the central and eastern High Atlas Mountains, showing the main tectonic events and detachment levels.

lacustrine deposits (Charrière, 1990; El Harfi, 1994; El Harfi et al., 2001; Fig. 4). Thrusting and inversion during the Cenozoic formed the current topographic relief of the High and the Middle Atlas Mountains.

Because of the absence of Tertiary deposits in the Errachidia area (cross-section A of Fig. 2), an accurate timing of deformation cannot be established, and it is impossible to discuss the effects of the tectonic events observed in the eastern High Atlas

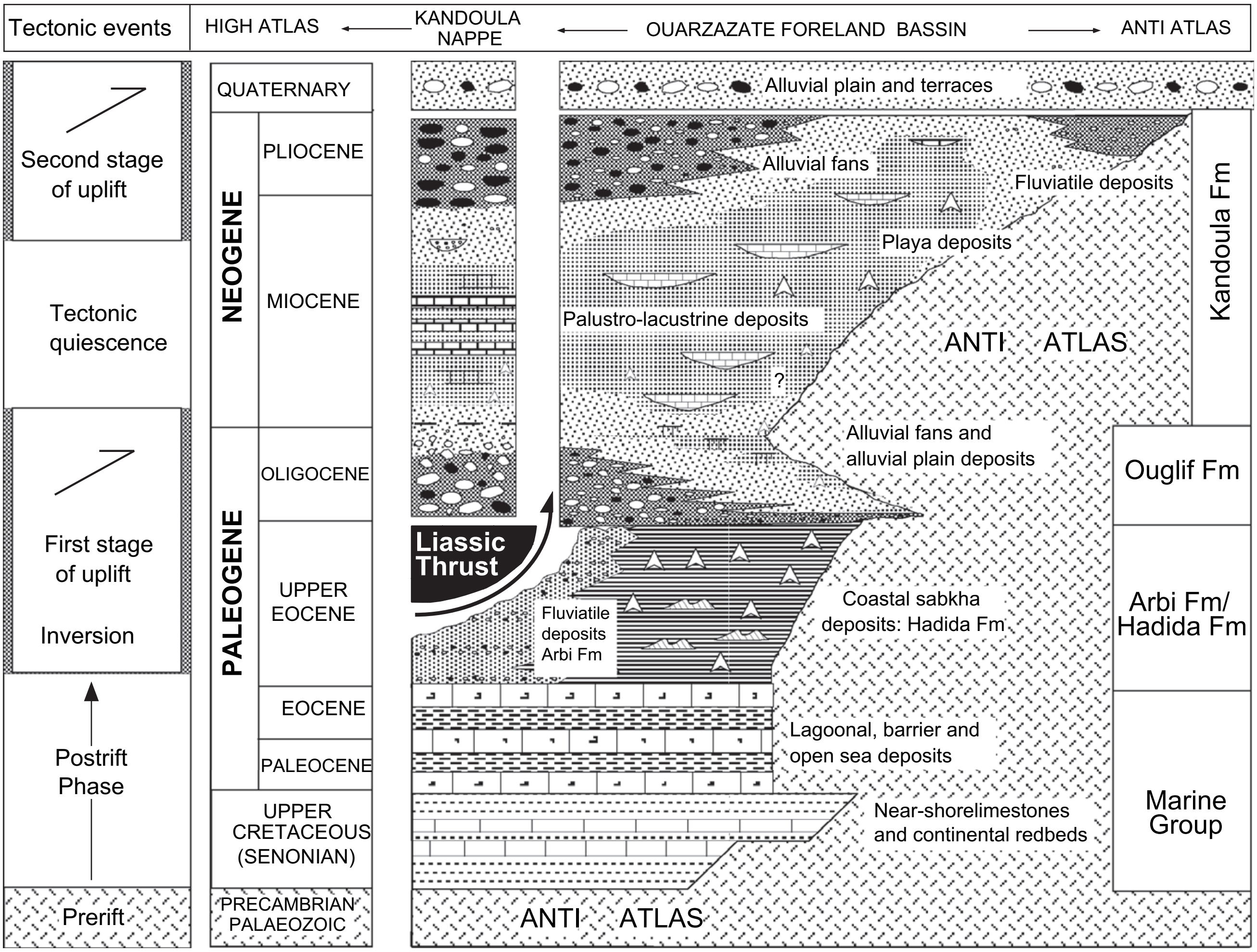


Fig. 4. Generalized chronostratigraphic diagram of the Ouarzazate foreland basin and the Kandoula thrust sheet-top basin. The diagram shows the vertical and lateral changes in the different lithostratigraphic units (El Harfi et al., 2001).

(Saint Bezar et al., 1998). However, the tectono-stratigraphic analysis of the Ouarzazate foreland basin shows the existence of two distinct stages in the development of the Atlas system corresponding to two episodes of rapid inversion and uplift of the axial zone of the High Atlas (El Harfi et al., 1996, 2001; Figs. 4 and 5). Furthermore, it shows that after a first step of activity during Upper Eocene and Oligocene and a phase of relative quiescence characterized by widespread lake deposits at the end of the Miocene, the main deformation took place during the Pliocene and the Lower Quaternary.

2.5. Structural geometry and kinematics analysis of the South Atlas front

The geometry and kinematics of the South Atlas front will be examined from east to the west in three key areas, located on Fig. 2.

2.5.1. Errachidia area

In the Errachidia area (Fig. 6), considered as representative of the eastern High Atlas, the superimposition of the South Atlas front on the southern limit of the Atlas Mesozoic basin is demonstrated by Teixell et al. (2003). The two frontal thrust sheets of the High Atlas contain a dominantly carbonatic Jurassic succession that is about 1000 m thick (Lachkar, 2000). The southern most unit (T1) shows a frontal overturned fold that was interpreted as a fault-propagation fold detached at the Triassic, and overrides a much reduced and undeformed Mesozoic succession characteristic of the Saharan platform. Whereas the second thrust (T2) is associated to a persistent shift of the regional elevation of Jurassic beds and must already involve the basement. Within the foreland, the thin Mesozoic series form an horizontal tableland overlying Palaeozoic and Precambrian basement.

Westward, the main feature of the Goulmima area (Fig. 2), studied by Saint Bezar et al. (1998), is that the southernmost thrust of the High Atlas remains blind below frontal tip line folds. Based on 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.

2.5.2. West of Toundout area

The region situated west and north-west of Toundout (cross-section C of Fig. 5) is characterized by the highest

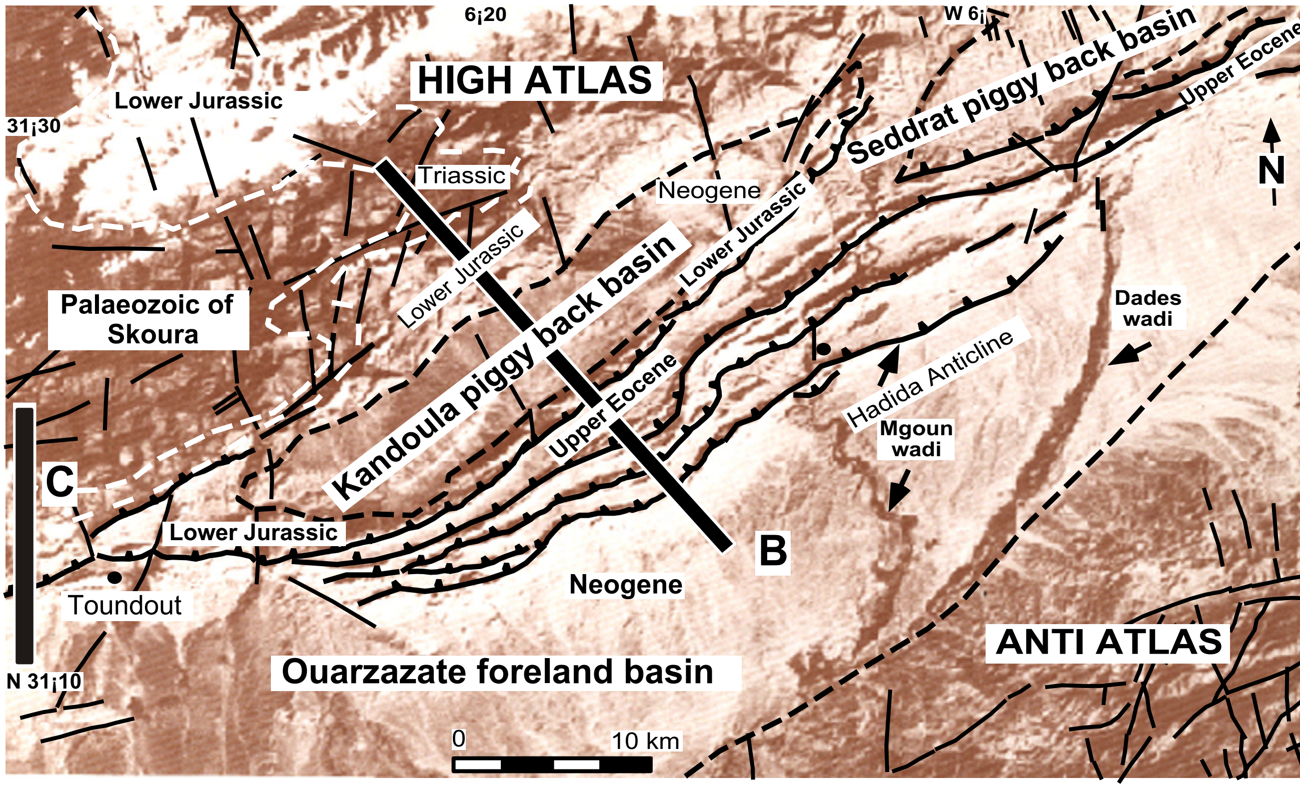


Fig. 5. Structural interpretation of the southern margin of the central High Atlas Mountains, Ouarzazate foreland basin, Kandoula and Seddrat thrust sheet-top basins and Anti Atlas (LANDSAT -1- Scene E 1551 - 10251, MSS Canal 7. 1/500 000; from El Harfi, 1994, 2001; see Fig. 2 for section location). The darker rocks in the southeast corner of the image are Precambrian and Palaeozoic basement of Anti Atlas (Jbel Saghro). The darker rocks in the northwest corner of the image are Palaeozoic rocks of Skoura culmination and Triassic outcropping in the axis of the large ramp anticline of Jbel Mgoun (up to 4000 m) that has been uplifted by the reactivation of synrift faults. The Palaeozoic rocks in the core of this anticline have been intensely deformed by what is interpreted to be a series of fault duplexes (Beauchamp et al., 1999). (B) Kandoula NW–SE cross-section of Figs. 7A and 8; (C) West Toundout N–S cross-section.

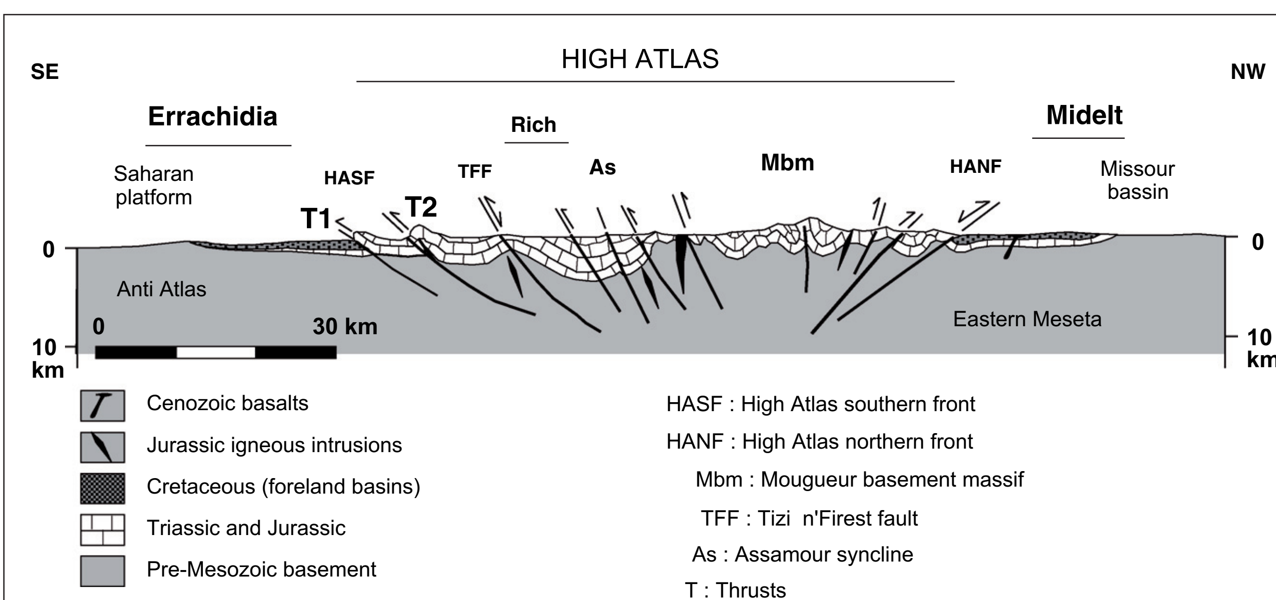


Fig. 6. Detailed structural and geological SE–NW cross-section Midelt–Rich–Errachidia through the eastern High Atlas of Morocco (see Fig. 2 for section location; modified from Brede and Heinitz, 1986; Teixell et al., 2003).

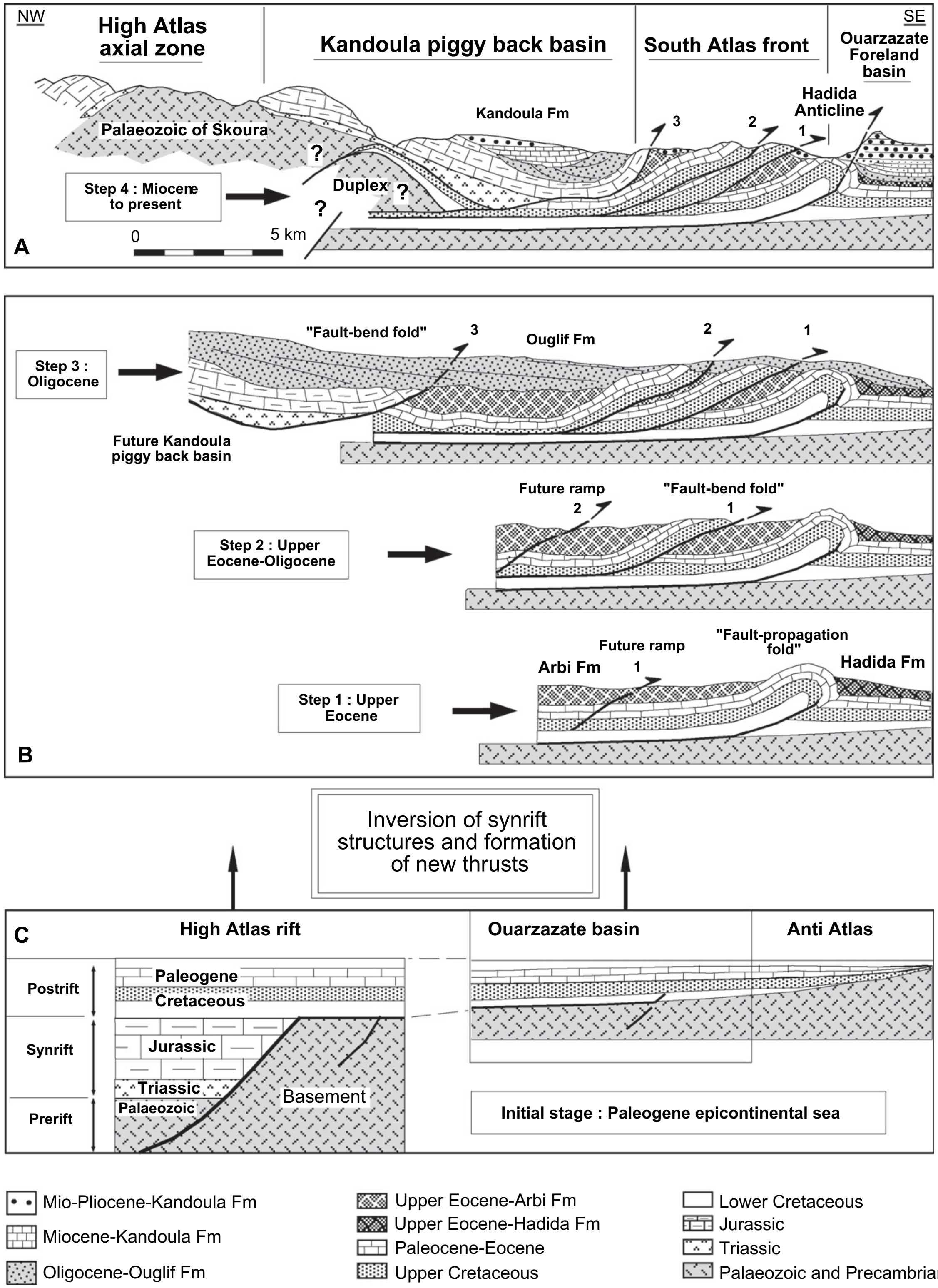


Fig. 7. (A) NW–SE cross-section through the southern part of the central High Atlas and the northern part of the Ouarzazate foreland basin (from El Harfi, 2001; additional source of data: ONAREP seismic profiles, Errarhaoui, 1998; Beauchamp et al., 1999; Frizon de Lamotte et al., 2000; see Figs. 2 and 5 for section location). (B) Schematic kinematic evolution of the cross-section A based on recent tectono-stratigraphic analysis of El Harfi (2001). (C) Simplified restored version of the cross-section A to the pre-compressional state.

topographic relief. It also shows a deeper erosional level and the basement is exposed in the large culmination of Skoura. Reduced Jurassic red beds and Liassic carbonates are locally preserved in the Highest mountain peaks. The High Atlas overrides the Ouarzazate Tertiary basin toward the south. In the northern margin of this basin, there is a complex system of thin skinned forethrusts and backthrusts that involve Cretaceous and Cenozoic rocks (Errarhaoui, 1998). The structural style of the substratum, inferred from observations made in Skoura area, is dominated by steep faults inherited from the rifting episode and more or less subsequently inverted (Beauchamp et al., 1999; El Harfi, 2001).

2.5.3. Kandoula and Seddrat areas

Using subsurface data, Beauchamp et al. (1999) and El Harfi (2001) show a frontal tectonic repetition of the Upper Cretaceous–Eocene sequence along the active margin of the Ouarzazate basin (Figs. 5 and 7A). This duplication of the stratigraphic pile is overlain by the ‘‘Kandoula thrust sheet-top basin’’ and, eastward, by the ‘‘Seddrat thrust sheet-top basin’’ (El Harfi, 2001). These later basins are consisting of Triassic red sandstone and Liassic carbonate supporting Oligocene and Neogene continental deposits (alluvial fans and palustro-lacustrine deposits; Fig. 4).

The Hadida anticline (Figs. 5 and 7A) is the southernmost unit which separates the Ouarzazate foreland basin from the South Atlas front. The initiation and the formation of the Hadida Anticline (fault-propagation fold) is assumed to be early in the Upper Eocene (Fig. 7B). This is based on the differentiation, during the Upper Eocene times, of two adjacent sub-basins: the Arbi in the north and Hadida to the south. They are filled distinctly, and respectively, with clastic and lacustrine deposits (El Harfi et al., 2001).

It is proposed here that the ‘‘Liassic Kandoula and Seddrat basins’’ developed firstly as fault-bend fold with an upper flat located within Upper Cretaceous series (c.f. Errarhaoui, 1998; Frizon de Lamotte et al., 2000). The slip transmitted progressively southward, accommodated by the duplication of the Upper Cretaceous–Eocene sequences. However, the blockage and bypass of the initial ramp during the propagation of the decollement (assumed in this study, to be in the Hadida Anticline) can explain the out-of-sequence thrust fault observed in the superimposed stratigraphic pile (Fig. 7B, C). It is proposed here that the complex thrust sequence leading to the present-day geometry of the ‘‘South Atlas front’’ is due to difficulties in thrust propagation toward the Ouarzazate basin.

Recently, the structural model of Beauchamp et al. (1996, 1999) relates Moroccan Atlas deformation to basement fault

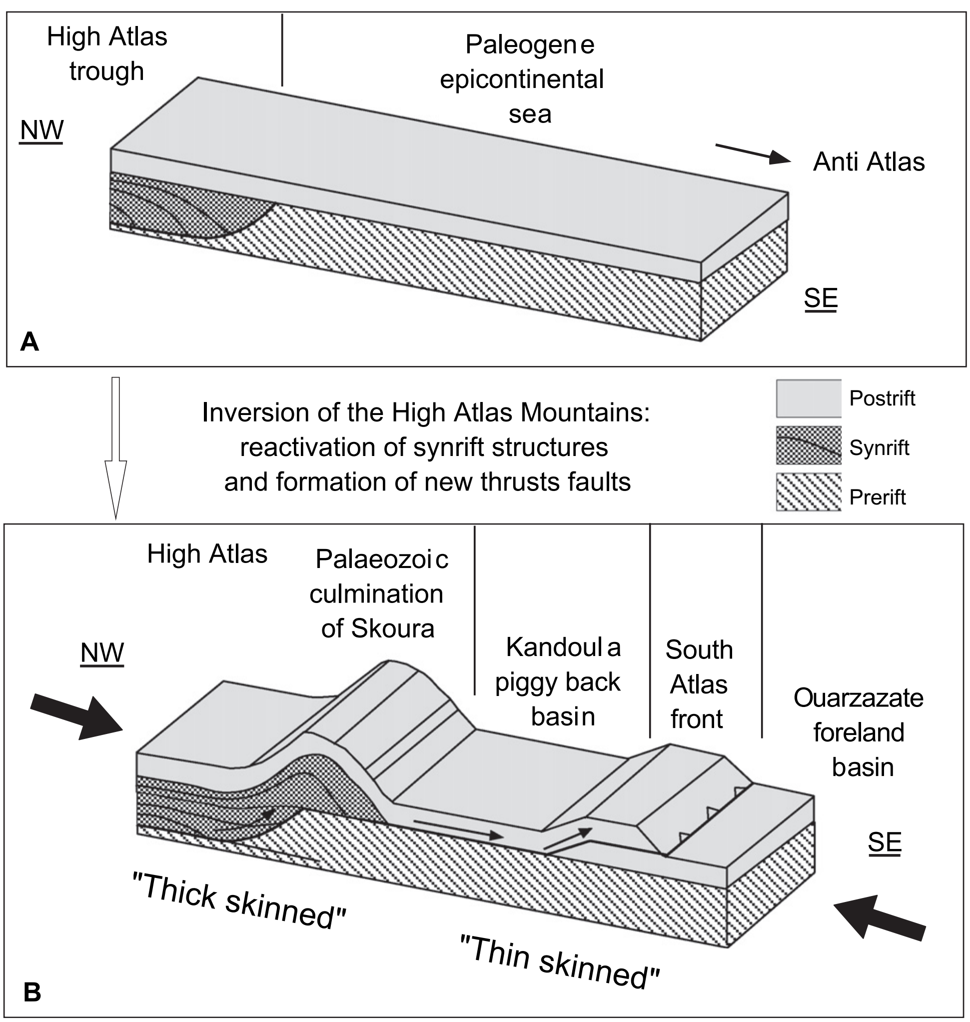


Fig. 8. Inversion and tectonic interpretation of thin skinned style of the southern High Atlas Mountains (Ouarzazate area; modified and completed from Beauchamp et al., 1999) (A) Postrift phase of a synrift half graben. (B) Inversion phase: compression normal to the half graben reactivates the synrift fault, moving synrift rocks up the hanging wall until a new thrust fault forms between the postrift and prerift rocks. Synrift strata are transported over the newly formed ramp, creating a fault-bend fold.

reactivating, and emphasises the thin and thick skinned nature of compressive deformation and the major role played by detachment levels located within Triassic and Cretaceous evaporites. The frontal thrusts branched northward from a deep decollement which, near the South Atlas front, cuts down into the basement (Frizon de Lamotte et al., 2000).

2.6. Mode of control exercised by basement faults on the development of thick and thin skinned thrusts

The comparison between the three South-Atlasic sections of the High Atlas (A, B and C in Fig. 2) leads to the proposition of an initial geometry of the Errachidia (Fig. 6) and west of Toundout segments (Fig. 5C) close to that of a half-graben. This implies that the major Triassic and Liassic normal faults were situated along the present-day Atlas boundary/Mesozoic Saharan platform and Atlas boundary/Mesozoic–Cenozoic Ouarzazate basin, respectively.

The proposed basement faults suffer two successive movements: a normal slip during the rifting stage and a reverse slip (or strike-slip) during the inversion. This is a thick skinned model. However, the geometry of Kandoula (and Seddrat) segment (Figs. 5B and 7) is very different. Errarhaoui (1998) shows seismic profiles crossing the South Atlas front (Kandoula and Ouarzazate basins) where a basement fault of Triassic age is clearly imaged by subsurface data. Inversion of the central High Atlas basin does not lead only to the reactivation of this deep and crustal fault, but to the development of a new, flat and shallow faults branched on the Triassic and Cretaceous decollements. This results in the development of thin skinned structures (Fig. 7A–C). It is clear that the border thrusts do not exactly coincide all over their extent with the palaeogeographic limits. This is the case, for instance, in Ouarzazate area where the ‘‘South-Atlasic fault’’ is commonly made up of flat-lying thrusts developed late in the history of the Upper Eocene–Neogene shortening without relationships with the basin boundaries (Fig. 8). In conclusion, the architecture and the tectonic styles of the Moroccan South Atlas front change laterally from thick to thin skinned models.

Based on the study of seismic and boreholes, the northern border of the High Atlas was described by Benammi et al. (2001). They concluded that the basement rocks and the sedimentary cover were deformed together in a thick skin style of deformation. The deformation style in the Middle Atlas and in the North Atlas front reveals that basement is involved in the compressional deformation, in a dominantly thick skinned tectonic style with a stacking of crustal units (Beauchamp et al., 1996). Convergence continues until the present times, as indicated by field evidence and active seismicity (Sébrier et al., 2006).

The thick skin style is a common trend along the Moroccan Atlas, where thin skinned thrusting is limited to the southern border of the High Atlas, and especially in the Kandoula and Seddrat areas. Such a tectonic scenario, supporting the thin skinned model, as in Ouarzazate foreland basin, is recorded in eastern Algeria and in Tunisia (Outtani et al., 1995; Bracène et al., 1998).

3. Stratigraphy and tectonic evolution of the Rifian domain

3.1. Presentation

The Rifian domain, fringing the Mediterranean Sea, represents the western part of the Rifo-Tellian chain (Figs. 1 and 2). This folded chain extends over the entire length of the Maghreb and its history is integrated into that of the western Mediterranean (Bouillin, 1986). A preliminary subdivision comprises, from west to east, the Rifian (north Morocco) and Tellian (northern Algeria and Tunisia) segments. Schematically, the Rifian structural domain can be divided into three major units, from north to south (Piqué, 2001; Fig. 2):

The ‘‘internal zones’’, sometimes called ALKAPECA domain (for ALboran, KAbylies, PEloritan and CAlabria), are only fragments of a stack of nappe complexes, including crystalline units and their sedimentary cover and peridotite flakes known in the Betics (Spain) and in Algeria. These zones belong to an exotic continental unit known as the Alboran block.

The ‘‘flysch zones’’ are the sedimentary sequences composed of turbidites and radiolarites deposited on a deep marine environment. Intercalated with these deposits are basic and ultrabasic magmatic rocks, which Bouillin (1986) has compared with an ophiolitic suite. More recently, Durand-Delga et al. (2000) confirmed the presence of ophiolitic Jurassic basic rocks associated with Rifian flyschs and concluded that the zone was at least partly oceanic.

The ‘‘external zones’’, make up an extremely complex domain including the intra, meso- and pre-Rifian units and the Rifian nappes.

3.2. Rifian tectonic history

In all the zones of the chain, the Alpine structures are undoubtedly complex because of the great number of allochthonous units, as well as the different times and mechanisms of their stacking. The sequence of thrusting is complex and involves the development of out-of-sequence thrusting at the boundaries between the main zones (Frizon de Lamotte et al., 2000). The Rifian system is considered to be an Alpine-type orogen resulting from opening then closing of an ocean followed by a continent–continent collision. The major stages of the Rifian 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).

4. Previous crustal model interpretations of the High Atlas Mountains

The first crustal model was proposed by Warme (1988). On a transverse section of the High Atlas (Midelt–Errachidia transect), at the Middle–Upper Liassic boundary, he shows

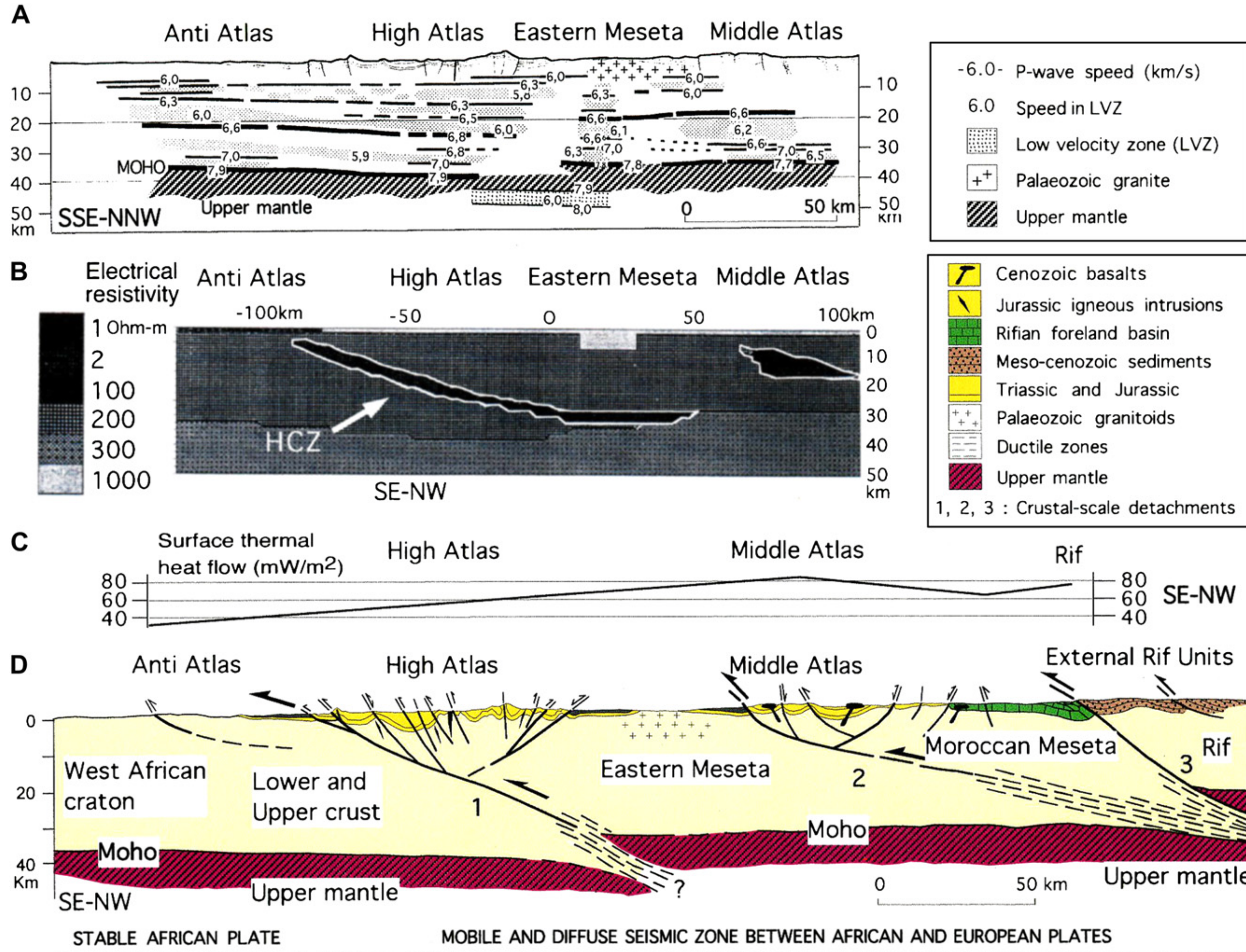


Fig. 9. (A) SSE–NNW crustal section of the High and Middle Atlas Mountains (after Wigger et al., 1992; see Figs. 1 and 2 for section location). Maximum crustal thickness (42 km) is found under the northern border of the High Atlas. Note the presence of the subcrustal velocity inversion at 45–50 km depth between the lower crust and the upper mantle above the High/Middle Atlas. (B) SE–NW electrical resistivity model across the High Atlas (after Schwarz et al., 1992; see Figs. 1 and 2 for section location). Electrical resistivity of the middle and lower crust beneath the Anti Atlas was determined to about 200 ohm m, the same as for the lower crust of the High and Middle Atlas. The model has a highly resistive (1000 ohm m) uppermost mantle. This Model (S100) shows a rather steeply dipping high conductivity zone (HCZ), stretching from the southern border of the High Atlas almost down to Moho depth (36 km) beneath the eastern Meseta. (C) SE–NW surface thermal heat flow across the High and the Middle Atlas (after Rimi, 1999). It seems to indicate a heat flow data which, although scarce in the High Atlas, yield values of 54 mW/m^2, and much higher in the Middle Atlas (85 mW/m^2). (D) SE–NW conceptual crustal section through the major intracratonic ranges of the Atlas system of Morocco: High Atlas, Middle Atlas and Rif (modified and adapted from Giese and Jacobshagen, 1992; Fig. 1 for section location). The comprehensive crustal section of the High Atlas is based on available geological data and a re-interpretation of geophysical data (El Harfi, 2001). The transect is supplemented by data collected by Bernini et al. (2000) for the Middle Atlas, Seber et al. (1996) and Michard et al. (2002) for the Rif and MIDSEA (Mantle Investigation of the Deep Suture between Eurasia and Africa) Project (Marone et al., 2003).

an arrangement in tilted blocks, separated by listric faults which are rooted at depth on a detachment fault slightly dipping to the north. The High-Atlasic trough would thus result from a pure extension, with a principal stress directed NW–SE. Jacobshagen et al. (1988) and Giese and Jacobshagen (1992) proposed the existence of a flat, intra-crustal detachment below the eastern High Atlas at some −20 km, without continuing to greater depth. Beauchamp et al. (1999) also suggested that Atlas thrust penetrated only to the middle crust, leaving the lower crust unfaulted. However, both Makris et al. (1985) and Wigger et al. (1992) modelled an asymmetric lower crustal structure beneath the High Atlas. They proposed that the lower crust gradually increases in thickness northwards beneath the southern boundary of the chain, but that the thickness changes abruptly beneath the northern boundary. These authors did not provide an explanation for this finding.

Based on a detailed geological and structural transect across the central High Atlas and the southern foreland basins and a synthesis of geophysical data mainly the ones acquired by German teams (Schwarz et al., 1992; Wigger et al., 1992), El Harfi (2001) proposed a new interpretation of the deep structure of the eastern High Atlas (Midelt–Errachidia transect). In this new model, the jump in crustal thickness (depth of Moho discontinuity) between the High Atlas and the northern plains is interpreted by the author as due to a major, north dipping thrust fault penetrating the lower crust and offsetting the Moho discontinuity. This model implies large-scale overthrusting events that involved the entire crust in a ramp fashion under pure compression or transpressive deformation (El Harfi, 2001; El Harfi et al., 2004). This result is confirmed, more recently by TRANSMED-Atlas-transect I (Ouarzazate transect; Frizon de Lamotte et al., 2004; Zeyen et al., 2005) and by gravity modelling results (Errachidia transect; Ayarza et al., 2005).

5. Compilation of geophysical data along a traverse crossing the High and the Middle Atlas Mountains of Morocco

5.1. Seismic refraction studies

A compilation of the seismic refraction profiles along a NNW–SSE section (Wigger et al., 1992; Fig. 9A) demonstrates that there exists a lateral variation in the distribution of the Low Seismic Velocity Zones (LVZs), and the maximum (7.9–8 km/s) is beneath the northern part of the High Atlas, in the area of maximal crustal thickness. The LVZs in the upper crust under the High Atlas dip slightly to the north, but a continuous transition to the NNW is not proven by the data. Moreover, the observed seismic data cannot prove how far the LVZ's sequences continue to the north beneath the Middle Atlas. Another major observation is the presence of the subcrustal velocity inversion and the LVZ at 45–50 km depth between the lower crust and the upper mantle above the High/Middle Atlas (Fig. 9A). This is supported by the relative low Pn velocity of 7.7–7.9 km/s. Additional information has been recently obtained by receiver function studies of Sandvol et al. (1998) and Van der Meijde et al. (2003). The results of Van der Meijde et al. (2003) show that the Mohorovicic (Moho) discontinuity under Midelt (in the plains north of the High Atlas) is located at 39 km. Sandvol et al. (1998) found, in fact two velocity jumps, at 36 and 39 km; the shallowest one was interpreted by them as the crust–mantle boundary, whereas the deepest one remained uninterpreted.

5.2. Moho depth and crustal thickness

Seismic refraction studies provided the first estimates of crustal thickness across the High Atlas. Tadili et al. (1986) modelled refraction/wide angle data from an experiment carried out in 1975, concluding that the crustal thickness varies from 25 km along the Atlantic coast of Morocco to 40 km near the central High Atlas. The recent result of the international MIDSEA (Mantle Investigation of the Deep Suture between Eurasia and Africa) project is the elaboration of a new map for the Moho discontinuity in the Eurasia–Africa plate boundary region (Marone et al., 2003). The reliable results, obtained for the northern African coasts, show that the western part of the African continent is characterized by a rapid change from a relatively deep Moho (down to 42 km) below the Atlas Mountains range to the thin crust (<20 km) of the southwestern Mediterranean Sea. According to detailed seismic refraction studies of Wigger et al. (1992), the crustal thickness beneath the High Atlas is about 35–40 km, being about 30–35 km in the peripheral plains (Fig. 9A).

5.3. Gravity data and isostatic state

Gravity and magnetic data displayed together with digital topographic data were provided by Makris et al. (1985), Tadili et al. (1986) and Beauchamp et al. (1999). A remarkable result of these investigations is the asymmetry of the High Atlas Mountains and the fairly homogeneous crustal thickness, without large crustal roots in spite of the high topography. As summits may exceed 4000 m in the High Atlas and 3000 m in the Middle Atlas, several authors have proposed that the Atlas Mountains are in an uncompensated isostatic state (Gomez et al., 2000). Recently, seismic wide angle and receiver function results, obtained by Ayarza et al. (2005), have been used as constraints to build a gravity-based crustal model of the central High Atlas of Morocco. Ayarza et al. (2005) concluded that, gravity modelling suggests moderate crustal thickening and a general state of Airy isostatic undercompensation.

5.4. Magnetotelluric investigations

The electrical resistivity structure of the crust and upper mantle of the Atlas Mountains system was studied by Schwarz et al. (1992) using magnetotelluric and geomagnetic deep soundings. The final electrical model obtained by the authors shows a crustal low resistivity layer with total conductance (thickness–resistivity ratio) of about 2000 Siemens which stretches from the southern border of the High Atlas towards the Middle Atlas (Fig. 9B). In this model, the highly conductive zone (HCZ) dips steeply northwards plunging deep into the upper mantle beneath the eastern Meseta. This model suggests that there is a deep or crustal-scale fault that may extend down to the upper mantle.

6. New interpretation and discussion

6.1. Proposed model

Previous crustal models of the High Atlas suppose the existence of a flat, mid-crustal detachment at some −20 km where all the surface thrusts merged and below which the lower crust was continuous (Giese and Jacobshagen, 1992). However, several questions are left unanswered by these models and a number of features require further explanation:

The occurrence of a ramp in the upper mantle (a jump in crustal thickness; Fig. 9A, B) beneath the northern border of the High Atlas. This ramp is shown up both by electrical resistivity modelling (Schwarz et al., 1992), seismic refraction data (Wigger et al., 1992) and gravity modelling (Makris et al., 1985; Ayarza et al., 2005);

The origin of the subcrustal velocity inversion (LVZ), observed at 45–50 km depth, between the lower crust and the upper mantle above the High/Middle Atlas (Fig. 9A);
The origin of the two receiver-function velocity jumps of Sandvol et al. (1998) and Van der Meijde et al. (2003). Their results show that the Moho discontinuity under Midelt is located at 36 and 39 km;
The occurrence of a significant gravimetric anomaly (values of less than −150 mgal) for the southern slopes of the central and eastern High Atlas (Ayarza et al., 2005);
The continuity of the LVZ layer at a depth of 10–20 km interpreted by Giese and Jacobshagen (1992) as horizontal detachment. The observed seismic refraction data (Fig. 9A) do not indicate how far the LVZ layer in the upper crust under the High Atlas continues to the north beneath the Middle Atlas. Indeed, no continuous transition to the NNW is proven by the data.

These issues can be addressed using the electrical model (HCZ, 2D S100; Fig. 9B) as an aid for the interpretation of the deep structure of the High Atlas. There is a clear correlation between the high conductive structure (HCZ), the jump in crustal thickness, revealed by seismic refraction data, and the southernmost boundary zone of the High Atlas chain (''South-Atlasic fault'' or ''SAF'').

The relation between the high electrical conductivity (HCZ) and the rapid change in crustal thickness (from 40 to 30 km; Fig. 9A) may be interpreted by thick skinned tectonics. It is suggested here that the jump in crustal thickness reflects a crustal-scale thrust fault (''SAF'') that penetrates the lower crust and offsets the Moho (''deep-rooted thick skinned model'' of El Harfi, 2001; El Harfi et al., 2004; Fig. 9D). In this solution, the conductive layer (HCZ), which can be interpreted as a zone of intense deformation with high fluid pressure, reaches Moho depths beneath the eastern Meseta.

The new interpretation developed here supports models of large-scale overthrusting that has involved the whole crust under pure compression or transpressive deformation. This interpretation implies that the lower crust of the NW African plate has been overthrust (subduction?) by the Moroccan and eastern Mesetas microplates, with crustal imbrication or a Moho duplication beneath the High Atlas. The geometry of the large crustal faults and their extension to depth points to the occurrence beneath the northern border of the High Atlas of a deep lithospheric detachment which roots in the lower crust and the upper mantle (Fig. 9D).

Finally, the occurrence of the subcrustal velocity inversion (LVZ), observed at 45–50 km depth, between the lower crust and the upper mantle, could be seen as a result of the beginning of a partial continental subduction (crustal imbrication or a Moho duplication) of the NW African plate under the Moroccan microplate (El Harfi, 2001; Chorowicz et al., 2001). In the same case of the High Atlas Mountains, the two receiver-function velocity jumps at 36 and 39 km of Sandvol et al. (1998), have been interpreted as a Moho duplication by Ayarza et al. (2005).

In addition, Tertiary and Quaternary alkaline volcanism (Harmand and Cantagrel, 1984; Ibhi, 2000) has been identified in the Middle Atlas, the Moroccan Meseta, the eastern Meseta, the northern part of the High Atlas, and in the Anti Atlas (Fig. 2). The tectonic context in which this volcanism appears is not fully understood yet and was explained by Harmand and Moukadiri (1986) as the result of upper mantle deformation which permitted partial melting in a compressive regime.

This interpretation is also supported by Hatzfeld and Frogneux (1981) and Medina and Cherkaoui (1991), who describe the depth range of 30–55 km beneath the High/Middle Atlas contact area as a seismic gap continued by an active seismic zone down beyond 60 km. Although, the authors cannot exclude the possibility that deep seismicity in this region represents the remains of subducting crust.

Recently, Ellouz et al. (2003) expect widespread diffuse crustal heating throughout Morocco, with a local increase in the large Atlas systems and within the Atlantic rifted margin. They relate to active Tertiary volcanism and magmatism the present-day high heat flow values >80 mW/m^2 (Fig. 9C) registered in the Middle Atlas.

6.2. *Conceptual crustal section through the Atlas system of Morocco: an assessment*

Some fundamental characteristics can be seen or deduced from the transect in Fig. 9D.

The continental crust becomes progressively thinner from the SE (35–40 km beneath the Anti Atlas) to the NW (20–25 km beneath the Rif). The lithospheric detachments of the High Atlas, the Middle Atlas and the Rif dip to the north and NW. The imbrication of the deep crustal root does not follow the general direction of the High Atlas, but is oriented obliquely, that is NE–SW. This oblique trend is actually parallel to many internal faults within the High Atlas, and to the general direction of the Middle Atlas chain (Ayarza et al., 2005). The Rifian domain contains ophiolitic Jurassic basic rocks associated with flyschs and indicates that the zone was at least partly oceanic (Bouillin, 1986; Durand-Delga et al., 2000).

The northward thinning of the crust and the direction of dip of the various crustale-scale detachments cannot be fully ascribed to the Atlasic orogen and clearly underline the influence of the inherited structure of Mesozoic rifting. This interpretation is supported by the palaeogeographic organisation of the different units of the transect controlled by the Jurassic primary architecture of the southern Tethys palaeomargin. The different structural units correspond from SE to NW (Fig. 9D) to

- The West African craton of the Anti Atlas and the Saharan domain;
- The High Atlas compartment (first Mesozoic rift basin);
- The eastern Meseta, (stable and rigid Palaeozoic block);
- The Middle Atlas compartment (second Mesozoic rift basin);
- The Moroccan Meseta (stable and rigid Palaeozoic block);
- The small Khemisset Jurassic basins (not crossed by the transect; Fig. 2). They occur in northwestern part of Morocco, and they represent the third Mesozoic rifts system;

- The thrust and nappes units in the Rifian domain, which contain ophiolitic Jurassic basic rocks and considered as remnant of the oceanic realm.

In palaeogeographic reconstructions (Dercourt et al., 2000; Durand-Delga et al., 2000), it is thought that from the mid-Jurassic onwards, the Middle Atlantic and the Ligurian Ocean basins (Tethys) were connected by a narrow E–W ‘‘oceanised zone’’ (Fig. 10). This was situated in the southern part of the western Mediterranean on the boundary or within the ALkA-PEKA continental block which separated Europe and Africa at that time (Bouillin, 1986; Durand-Delga et al., 2000).

7. Structural inheritance in the Atlas Mountains from the Tethyan palaeomargin

7.1. The NW corner of Africa between Tethys and the central Atlantic

The Moroccan tectonic fabric, before the beginning of Africa/Eurasia convergence, constitutes an atypical segment of the Tethys margin system (Fig. 10). Its southwesternmost part is directly linked to two separate tectonic events (Olivet, 1996). In the west, the central Atlantic rifting episode started during the Late Triassic–Early Jurassic. In the north, the Tethys margin formed from late Triassic times. The correlation in time and space between the formation of these two margins is not completely clear, since the two phases partly coexist and interfere. The present-day ‘‘external zones’’ (e.g. 3; Fig. 2) of the Rif are considered to be the former southern margin of the Tethys Ocean. The ‘‘flyschs’’ domain is a remnant of the deposits of the oceanic realm. The covers of the present-day ‘‘internal zones’’ are remnants of the northern margin of the Tethys (Durand-Delga et al., 2000).

During the Middle to Late Jurassic, the southern limit of ‘‘European and north American continents’’ was situated to the north of the future ‘‘flyschs domain’’, which is itself the site of the emplacement of basic and ultrabasic magmas and the uplift of ophiolitic series. Studies of Jurassic rocks in the ‘‘European continent’’ lay emphasis on the pure opening (normal component) of the starved basins established to the south of the south-European margin, where the substrate is oceanic (Durand-Delga and Fontboté, 1980). Unfortunately, the collapse of the south-European margin and the Rifo-Tellian

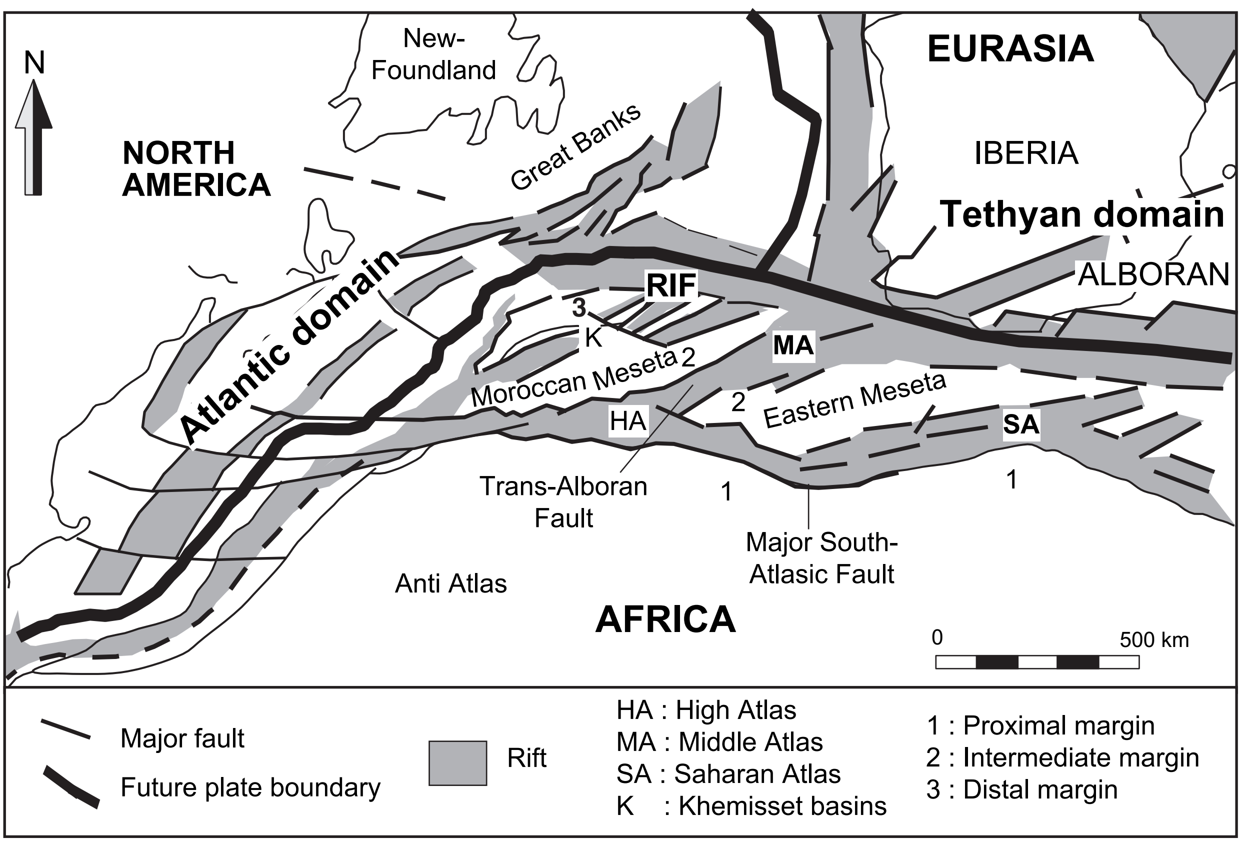


Fig. 10. Sketch map of Mesozoic rifting along the northern African margin and adjacent regions (modified from Favre, 1995; Dercourt et al., 2000). At this time, the Atlas domain of Morocco constituted the southern margin of the Tethys. Its polyphase development recorded the geodynamic evolution of the African plate and the correlative opening of the Tethys and Atlantic Oceans. At the Jurassic time, Morocco corresponded to a triple point, at the intersection of ‘‘central Atlantic rift’’ (between Africa and North America) and the western oceanised extremity of the Tethys (between Africa and Eurasia). Later, in the Cenozoic, the eastern passive margin of the central Atlantic Ocean remained undeformed, while the southern margin of the Tethys in Morocco was affected by the Alpine orogeny. Since the ‘oceanised’ opening boundary fault witness (distal margin: 3), separating Africa from Eurasia, has been intensively reworked and integrated in the allochthonous units of the Rifian belt, while the Atlasic domain (proximal margin: 1) remained better preserved.

compression obscures the initial form and geometry of these basins, and especially those of the NW African margin. Farther south, however, the contemporary basins of the Atlas are less distorted and their pre-orogenic structure can be resolved.

To interpret the palaeogeography of the NW part of the African margin, account must be taken of both the N–S extension related to the Tethyan opening, and the NW–SE oriented extensional processes associated with the Atlantic. Both Tethyan and Atlantic extensional mechanisms co-existed over a 40–50 my period, from the Carnian to the Early Dogger (Ellouz et al., 2003). This interference between two different directions of extension can explain the complex organisation of the inferred stress field and the primary architecture created in the NW corner of Africa, especially, the obliquity of the most Jurassic rifts (grabens or hemi-grabens) to both the Tethyan and the Atlantic margins.

Jurassic extensional structures were concentrated in deformation zones that were several hundred kilometres long but comparatively narrow (Fig. 10). This process has produced a series of NE–SW half-grabens in NW Africa, which were mostly the intracontinental Atlasic, Khemisset and Rifian basins (rifts; Fig. 10). On a regional scale, the entire continental passive margin of NW Africa appears symmetrical both to the northwestern Eurasian and to the northern American continental margins (Newfoundland–Iberian block). The southern border of this great graben was controlled by the major ''South-Atlasic boundary fault'' (''SAF''). Given this geodynamic context, we propose a model for reconstructing the primary architecture of the southern Tethys passive margin, during Jurassic extension (Fig. 11).

7.2. Synthetic restored cross-section

A precise restored version of the transect (High Atlas – Rif; Fig. 11) to the pre-compressional state is not possible because of the strong deformations which affect the Rif expressed by recumbent, complex and regional overthrust folds and nappes (HP metamorphism). However, the comparison of the deformed and restored sections of the High and Middle Atlas (Beauchamp et al., 1996, 1999; Arboleya et al., 2004) yields an Alpine shortening of some 36 km for this transect of the African plate, distributed in 31 km for the High Atlas and 5 km for the Middle Atlas. The Middle Atlas value is comparable to that previously obtained by Gomez et al. (1998). As demonstrated by the balanced cross-section of Beauchamp et al. (1999), overthrusting at the margins of the central High Atlas nearly triples the magnitude of horizontal shortening across the 105 km wide mountain belt from 10% to 25%. Overthrusting was an important shortening

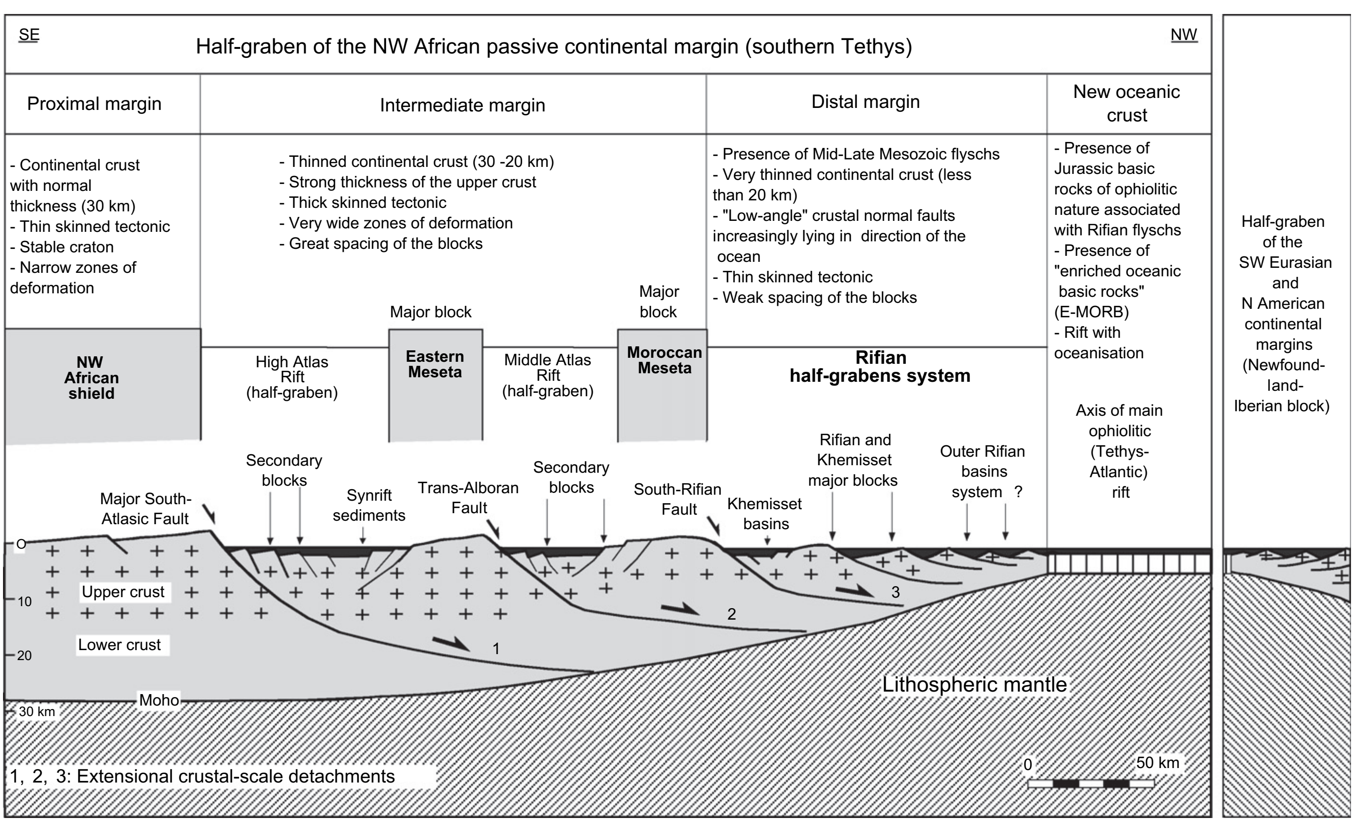


Fig. 11. Model of the extensional southern Tethys passive margin during Mesozoic rifting. A synthetic reconstruction of the restored version of the transect to the pre-compressional state is presented only for the High and Middle Atlas Mountains (datum: base of the Cretaceous; modified from Giese and Jacobshagen, 1992; Zouine, 1993; Errarhaoui, 1998; Beauchamp et al., 1999). The geometry of the Rifian structures is difficult to restore to the pre-compressional state. The High and Middle Atlas rifts are bounded by major normal faults which connect up at depth with extensional crustal-scale detachments dipping gently northwestwards towards the ocean (axis of the main Tethys–Atlantic rift; see Figs. 1 and 2 for section location). The geometry of tilted blocks (synsedimentary tectonics; horizontal dimensions…), fault walls and sedimentary thicknesses in this reconstruction is based on several workers in the High and Middle Atlas.

mechanism for the High Atlas range, accommodated on steep faults. This thrust system is interpreted as penetrating into the deep crust and offset the Moho discontinuity.

A degree of uncertainty arises from the obliquity of some of the individual structures to the section line. However, out-of-plane movements in the High Atlas were considered to be minor (around 4%) by Zouine (1993) and Arboleya et al. (2004). In the case of the Middle Atlas, Khemisset basins and the Rif, obliquity of structures to the regional compression direction is larger, so the restoration results are only approximate. The precursors of the High Atlas, the Middle Atlas and the graben system of the Rif appear as rather symmetrical extensional basin, flanked by marginal zones with much reduced sedimentary thickness.

8. Primary architectural model of the southern Tethys passive margin

At the time of the Atlasic rifting, the thinning of the NW African continental crust initiated the formation of tilted blocks bounded by normal faults and transfer faults (Fig. 11). Two types of tilted block can be distinguished in the reconstruction of the Atlasic passive margin: major blocks and secondary blocks (El Harfi, 2001; Chevalier, 2002).

8.1. Major blocks (wavelength 50 km and more)

The main rigid blocks namely the Moroccan Meseta and the eastern Meseta are specified by wavelengths 50 km and more. The characteristics of the major blocks from SE to NW show that these units are half-grabens tilted to the SE (Fig. 11). This structuring individualises High and Middle Atlas troughs in the place of the eastern Meseta and Moroccan Meseta, respectively. The major blocks are bounded by major border faults dipping NW towards the rift axis, that is, towards the future Tethys–Atlantic Ocean. The ‘‘South-Atlasic fault’’ and the ‘‘Trans-Alboran fault’’ are two of these major normal palaeofaults (Jacobshagen, 1992).

The outer Rifian and Khemisset major blocks are located in the oceanic ‘‘distal part’’ of the NW African palaeomargin, probably, in the meeting point between the Atlantic and the Tethys Oceans (Favre et al., 1991; Fig. 11). Durand-Delga et al. (2000) thought that the ‘oceanised’ opening, connecting the Mid-Atlantic and the Ligurian oceanic domain (Tethys), may have been up to 100 km wide. Because of a higher crustal thinning rate towards the axis of the main rift, the Rifian blocks are bounded by ‘‘low-angle’’ normal faults. Later and during the Cenozoic collision between Europe and Africa, these structures will be reactivated to become large thrusts (South-Rifian faults) and will be involved in the complex internal structure of the Rifian nappes (Fig. 12).

8.2. Secondary blocks (4–30 km wavelength)

These are the most characteristic Mesozoic structures of the High and Middle Atlas troughs. They produced a contrasting palaeogeography with alternating shallows, carbonate and clastic platforms, and deep subsiding basins. The synsedimentary faults bounding the tilted blocks dip to the NW and to the north but also to the SE and to the south (Fig. 11). In the Mesozoic, the entire Atlas domain was characterized by two main sets of directions controlling syntectonic sedimentation. N35°E–N45°E dominated in the Middle Atlas, whereas N70°E was preponderant in the High Atlas (Brede et al., 1992).

The timing of movement along these normal palaeofaults and block tilting has been identified in the field from many markers of extensional synsedimentary structuring (progressive unconformities, divergent onlap features, sedimentary thicknesses in fault walls, condensed strata, turbidites, olistholiths, slumps etc.…). Proust (1962) reported Triassic and Jurassic activity of normal faults along the northern and southern borders of the High Atlas. The presence of a normal component is indicated by many observations showing Early–Middle Jurassic collapses to the north along the southern border of the High Atlas trough and to the south along the northern border (Warme, 1988). Subsequently, several workers have found evidence for synsedimentary tectonics and tilted blocks in the High Atlas (Hadri, 1993; Lachkar, 2000 and among others) and in the Middle Atlas (Beauchamp et al., 1996; Gomez et al., 1998).

The model of crustal deformation of the southern Tethys passive margin set out in this paper (Fig. 11) is characterized by the presence of tilted blocks affecting the upper crust and crustal-scale detachments in the lower crust dipping towards the ocean. Such structures are characteristic of present-day divergent margins and are reported in the north of the Bay of Biscay (Montadert, 1984), the Gulf of Lion (Benedicto et al., 1996; Vially and Tremolières, 1996), the west-Iberian margin off the coast of Galicia in Spain (Thomas et al., 1996) and on the northern Tethys fossil margin analysed in the western and central Alps (Froitzheim and Eberli, 1990; Chevalier, 2002).

It can be noticed also that the main Tethys–Atlantic rift resulted in oceanisation, whereas the secondary rifts of the High, the Middle Atlas and the system of half-grabens of the Khemisset and outer Rif (Favre et al., 1991) aborted. A similar pattern is found on the margins of the Alpine Ocean and on the present-day margin of Galicia (Thomas et al., 1996) where only one of two or three rifts evolved to oceanisation while the others fail.

9. Conclusions

It can be concluded that the results of surface geology and geophysical investigation along a traverse which crosses the High Atlas (Midelt–Rich–Errachidia) suggest the concept of deep rooted overthrusting that has comprised the whole crust (‘‘deep-rooted thick skinned model’’ of El Harfi, 2001). This model is confirmed, more recently in the Ouarzazate area by TRANSMED-Atlas-transect I (Frizon de Lamotte et al., 2004). The dip of the fossil plate that could explain the abrupt displacement of the Moho would be north and northwestwards. This agrees with the deduced subduction

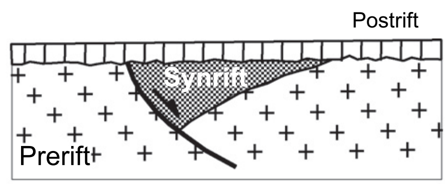
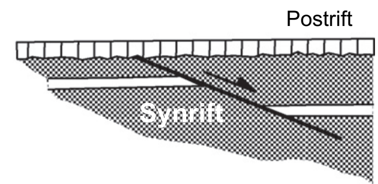
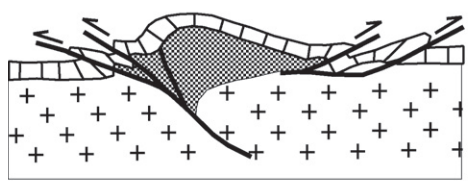
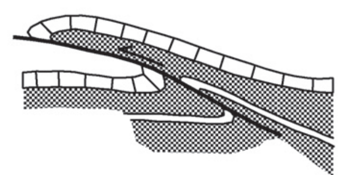
SE Half-graben of the NW African passive continental margin (southern Tethys) NW		
Margin	**Proximal and intermediate**	**Distal**
Rift basins	**High and Middle Atlas**	**Rif**
Mesozoic extensional half-grabens		
Cenozoic inversion of extensional half-grabens		
Geometry of normal faults	- Inversion of listric or planar normal faults	- Inversion of "low-angle" normal fault
Shortening and structures	- Weaker general shortening - Overall steep dip of the outcropping structures	- Structural vergence to the South - Relatively thin units detached from their substratum
Deformation	- Abnormal contacts with a developed cover, especially on the edges of the chain - Thick with minor thin skinned tectonics	- Stacking of a great number of allochthonous and complex nappes and units - Thin skinned tectonics with complex thrusts
Folds	- Folds with steeply-dipping axial planes	- Synschistosity folds, with flat or gently dipping axial planes
Metamorphism	- Absence of metamorphism	Presence of HP-type metamorphism

Fig. 12. Faults associated with the Mesozoic half-graben system of NW Africa occur as listric or planar and ‘‘low-angle’’ normal faults. Thin (Ouarzazate area) and thick skinned tectonics are frequently formed by the reactivation of listric or planar faults in the proximal and intermediate margins. Large thrusts, recumbent nappes and allochthonous units of the Rifian domain are the result of reactivated ‘‘low-angle’’ normal faults in the distal margin, during the Cenozoic collision between Europe and Africa.

directions beneath the Pyrenees and the Alps (Roure and Choukroune, 1998; Zeyen et al., 2005).

Positive inversion mechanisms that have been detailed by many workers (Cooper et al., 1989; Coward, 1994) may be applied to the High Atlas Mountains. Mainly these include the reactivation as reverse faults of old graben and half-graben extensional structures. The major northwest dipping lithospheric fault that we interpret to offset the Moho discontinuity (‘‘thick skinned’’ thrust) corresponds to the inverse reactivation of old extensional faults and Mesozoic crustal-scale detachments.

The NE–SW trending structure characteristic of many of the internal thrust faults of the High Atlas and of the Middle Atlas is inherited from normal faults created during the Mesozoic. This rifting phase is inferred to have been perpendicular to this trend

(Piqué et al., 1987). The deformable ''South-Atlasic fault'' zone reflects the influence of crustal structures inherited from pre-existing orogenies (Pan-African, Hercynian, Atlasic); as a result, this zone is sufficiently weak to create separate major crustal blocks in the northwestern part of the stable African plate.

As is the case for other margins where Tethys and Atlantic rifting coincided (North Atlantic/Norwegian and North Sea), in NW Africa rifting processes extended over a long time and were spatially widely distributed. The rifting history of this area, during the Triassic–Jurassic, is linked to the evolution of the Tethys and Atlantic Oceans, from rifting to oceanic accretion and not only to a pure Atlantic driving of the African plate. The rifting phases occurred over about 40–50 my on the southern Tethys margin itself, to a maximum of 70 my along the Atlantic segment (Ellouz et al., 2003) resulting in regional complex stretching of the margin.

The southern Tethys passive margin was dissected by major tilted blocks that include the Moroccan and eastern Mesetas bounded by major extensional boundary faults. Those faults determined the position of the High, the Middle Atlas and the Rifian troughs (rifts). The surface faults continued to depth as crustal-scale detachments dipping gently towards the Tethys–Atlantic Ocean.

The Jurassic structural zoning of the southern Tethys margin included from SE to NW a proximal, intermediate and distal parts with respective increases in the rate of crustal stretching toward the Tethys–Atlantic Ocean. This primary architecture controlled the different rates (magnitudes) of Cenozoic shortening recorded in the Atlas Mountains, in the Khemisset basin and in the Rifian domain (Fig. 12). Consequently, the Atlas system of Morocco and the Rifian chain correspond to an inverted array of Mesozoic basins, inherited from the southwestern Tethyan continental margin.

As a part of north Africa, the Moroccan margin registered three successive geodynamic episodes relative to the movement of the African plate with regard to the adjacent plates. First, its separation, during the Mesozoic, from the North America plate during the Late Triassic–Early Jurassic, was contemporaneous with its dissociation from the European plate during late Triassic times (Olivet, 1996). Second, the NW African craton was subjected to extension associated with crustal thinning, which controlled the development of a great half-graben in the position of the future Atlas range of north Africa. This geodynamic context is responsible for the opening and individualisation of extensive (transtensive?) sedimentary basins in the Rifian, the Khemisset system and the Atlasic rifts, thereby isolating the more stable and major blocks of Moroccan and eastern Mesetas. This event is marked by the emplacement of mafic magmas in the Atlas troughs (proximal part) and of ophiolitic suites in the Rifian troughs (distal and oceanic part). This great half-graben of NW Africa is limited, to the NW, by the oceanised crust of the Tethys–Atlantic Ocean, and to the S and SE, from the African shield, by the ''SAF''. It is therefore difficult to quantify the individual contributions of the Tethys and Atlantic oceanisation, and consequently to relate the extensive history of the NW corner of Africa only to a pure Atlantic eastward drift of Africa plate. Third and finally, the convergence of North Africa with the European plate, from the Upper Eocene, generated progressive horizontal shortening of the mountain belt. Unlike the 'Atlantic and Tethys rifts' that evolved into oceans, ''the Atlas, the Khemisset and the Rifian rifts'' failed and aborted. In these sites, however, the strong Cenozoic shortening has obscured the Mesozoic sedimentary and structural pattern in the Rifian domain (distal part) and contributed towards the hiding and the masking of Khemisset basins system. In the Atlasic domain (proximal part), these early structures remain better preserved. It can be concluded that better resolution of belt's crustal structure will give better restoration of the old basin geometry and improve our understanding of their tectonic evolution.

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