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# Tectonic shortening and topography in the central High Atlas (Morocco)

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[1] Three cross sections of the Moroccan High Atlas illustrate the structural geometry and relationship between tectonic shortening and topography in this Cenozoic intracontinental mountain range. The structure is dominated by thick-skinned thrusting and folding, essentially by inversion of Mesozoic extensional faults and by buckling of both pre-Mesozoic basement and its sedimentary cover. Detached, thin-skinned thrusting is limited and apparently related to basement underthrusting, which did not always create structural relief. Despite the high topography, tectonic shortening is moderate, with faults and folds being spaced and separated by broad tabular areas. Section restoration indicates that shortening decreases along strike from east to west in the High Atlas, while topographic elevation generally increases. This inverse correlation suggests that crustal thickening does not fully explain the observed topography and suggests a mantle contribution to uplift. This is supported by geophysical indications of a thin lithosphere and by alkaline volcanism in the vicinity. Mantle-related uplift, which occurs in a broad region, may also explain the scarce foreland basin record adjacent to the High Atlas. The relief of the Atlas Mountains is interpreted as a combination of crustal isostatic and dynamic topography. *INDEX TERMS:* 8005 Structural Geology: Folds and folding; 8010 Structural Geology: Fractures and faults; 8110 Tectonophysics: Continental tectonics—general (0905); 9305 Information Related to Geographic Region: Africa; *KEYWORDS:* intraplate tectonics, inversion, folding, topography, High Atlas, Morocco. **Citation:** Teixell, A., M.-L. Arboleya, M. Julivert, and M. Charroud, Tectonic shortening and topography in the central High Atlas (Morocco), *Tectonics*, *22*(5), 1051, doi:10.1029/2002TC001460, 2003.


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## 1. Introduction

[2] The Atlas Mountains of North Africa are considered as type examples of intracontinental chains [*Mattauer et al.*, 1977; *Rodgers*, 1987; *Ziegler et al.*, 1995]. They formed during Cenozoic times in the interior of the African plate ahead of the Rif-Tell plate boundary orogenic belt (Figure 1a). Like other intracontinental chains, the Atlas experienced moderate crustal shortening and exhumation. However, they have a very high topography, with peaks of over 4000 m.

[3] Many earlier studies have dealt with aspects of the evolution of the Atlas system from rift or transcurrent sedimentary troughs during the Mesozoic, to compressional belts of tectonic inversion in Cenozoic to recent times [*Choubert and Faure-Muret*, 1962; *Mattauer et al.*, 1977; *Schaer*, 1987; *Laville*, 1988; *Jacobshagen et al.*, 1988; *Laville and Piqué*, 1992; *Beauchamp et al.*, 1996, 1999; *Frizon de Lamotte et al.*, 2000; *Gomez et al.*, 2000] (among others). However, some important tectonic questions are still unsolved, including the compressional structural style, whether it is thin-skinned or thick-skinned, the mode of basement involvement, the total orogenic shortening, the origin of the elevated topography, and the reasons for the scarce development of peripheral foreland basins. Our discussion of these points is based on three new cross sections through the High Atlas of Morocco combined with geological and geophysical considerations from neighboring areas (the Moroccan Meseta, the western end of the Algerian High Plateau and the Anti-Atlas, Figure 1a). The modes of compressional deformation in the interior of the continent and the contribution of crust and mantle to topography are addressed more generally in the discussion.

## 2. Geological Cross Sections

[4] The Alpine foreland in front of the Rif-Tell orogen consists of several intracontinental mountain belts and plateaux. The High Atlas forms the most prominent mountain chain (Figure 1a), extending for 2000 km in a roughly west-east direction from Morocco into Algeria and Tunisia, where the range is known as the Saharan Atlas and the Tunisian Atlas respectively. This belt, together with its NE-SW branch in Morocco known as the Middle Atlas, is

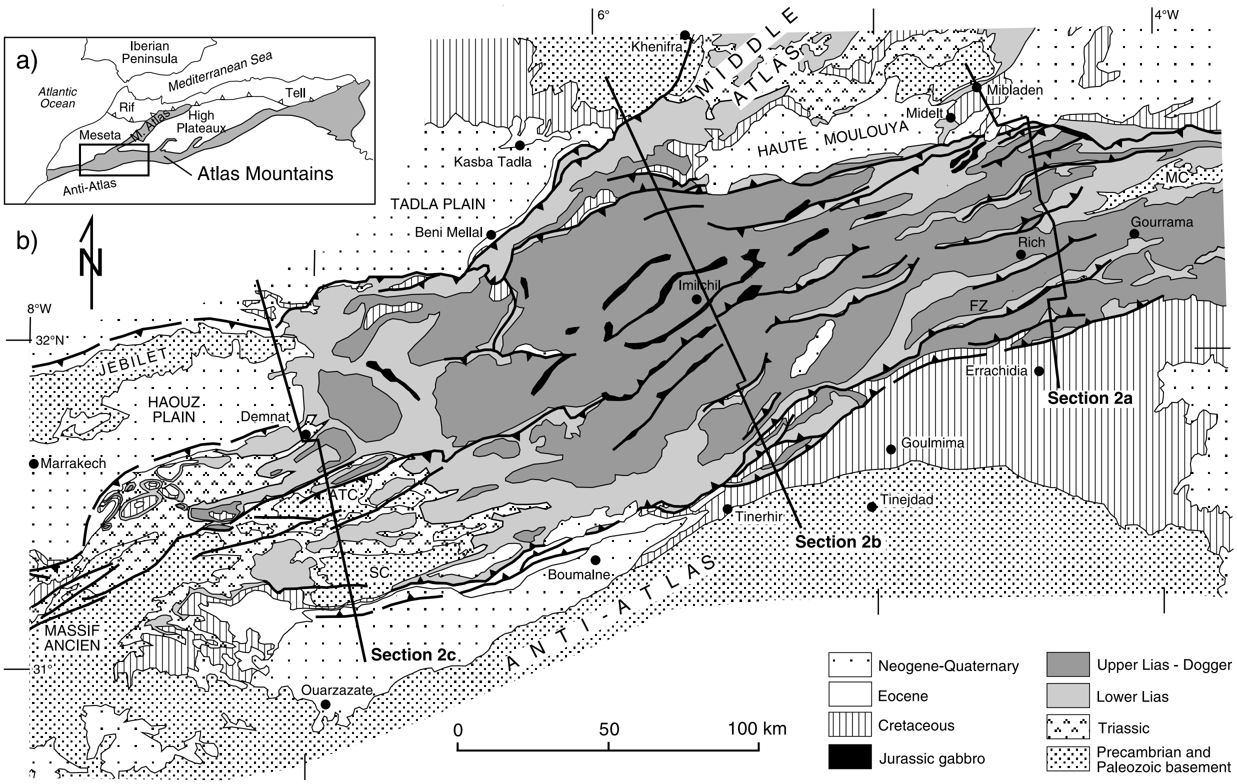


**Figure 1.** (a) Location sketch map of the Atlas Mountains in the North African foreland. (b) Geological map of the central High Atlas, indicating the section lines of Figure 2. ATC, Ait Tamlil basement culmination; SC, Skoura basement culmination; MC, Mougueur basement culmination; FZ, Foum Zabel thrust.

composed at the surface of folded and thrust Paleozoic and Mesozoic rocks, mostly Jurassic in age. Flanking the High and Middle Atlas, the Moroccan Meseta, the Moulouya Plain and the Anti-Atlas are wide uplifts with much less Alpine deformation, but reaching altitudes of over 2000 m. In this section we will focus on the geology of the central High Atlas, a ca. 100 km wide belt concentrating most of the Alpine deformation (Figure 1b).

[5] We describe here three transects of the High Atlas that illustrate the structural variation along the range (Figure 2). The cross sections are oriented NNW-SSE, perpendicular to the general trend of the belt, and parallel to the direction of regional convergence in the African plate during the Cenozoic [*Mattauer et al.*, 1977; *Gomez et al.*, 2000]. They are based on detailed field data acquisition along a stripe of geological mapping some km wide following the lines of section, and were originally constructed at a scale of 1:50,000. The sections use a simplified stratigraphy, based on the lithostratigraphic units of larger rank, which have a tectonosedimentary significance. Surface data were extrapolated to the pre-Mesozoic basement using the parallel fold methods and taking into account stratigraphic variations when detected. There is a lack of seismic or well information for most of the range. Subsurface data are restricted to bordering basins [e.g., *Jabour and Nakayama*, 1988; *Beauchamp et al.*, 1996, 1999]; the structure of these marginal basins in our transects is constrained by the high structural elevation of the basement, which either crops out or lies at shallow depth (Figures 1b and 2).

## 2.1. Regional Stratigraphy and Sedimentary Evolution

[6] The location of the High Atlas coincides with a Triassic-Jurassic trough [*Choubert and Faure-Muret*, 1962; *Piqué et al.*, 2000]. The chain is essentially made up of Jurassic rocks in outcrop (Figure 1b). Pre-Mesozoic basement, formed by rocks affected by the Hercynian and/or Panafrican orogenies, crops out in some culminations (Mougueur, Ait Tamlil, Skoura, Figure 1b) and in a large massif in the High Atlas of Marrakesh called "Massif Ancien" [*Jenny*, 1985; *Froitzheim et al.*, 1988; *El Kochri and Chorowicz*, 1988].

[7] During most of the Mesozoic, the High Atlas domain experienced extension and rifting, first during the Triassic, as recorded by red beds and tholeiitic basalts, and later during the Jurassic, with the deposition of marine carbonates and shales capped by continental red beds. Though largely hidden below the Jurassic, the Triassic is thickest around the Massif Ancien, the Skoura culmination, and the southern Middle Atlas (up to 1000 m). The reactivation of previous Hercynian faults has been considered responsible for the location of the Triassic basins, under a NW-SE

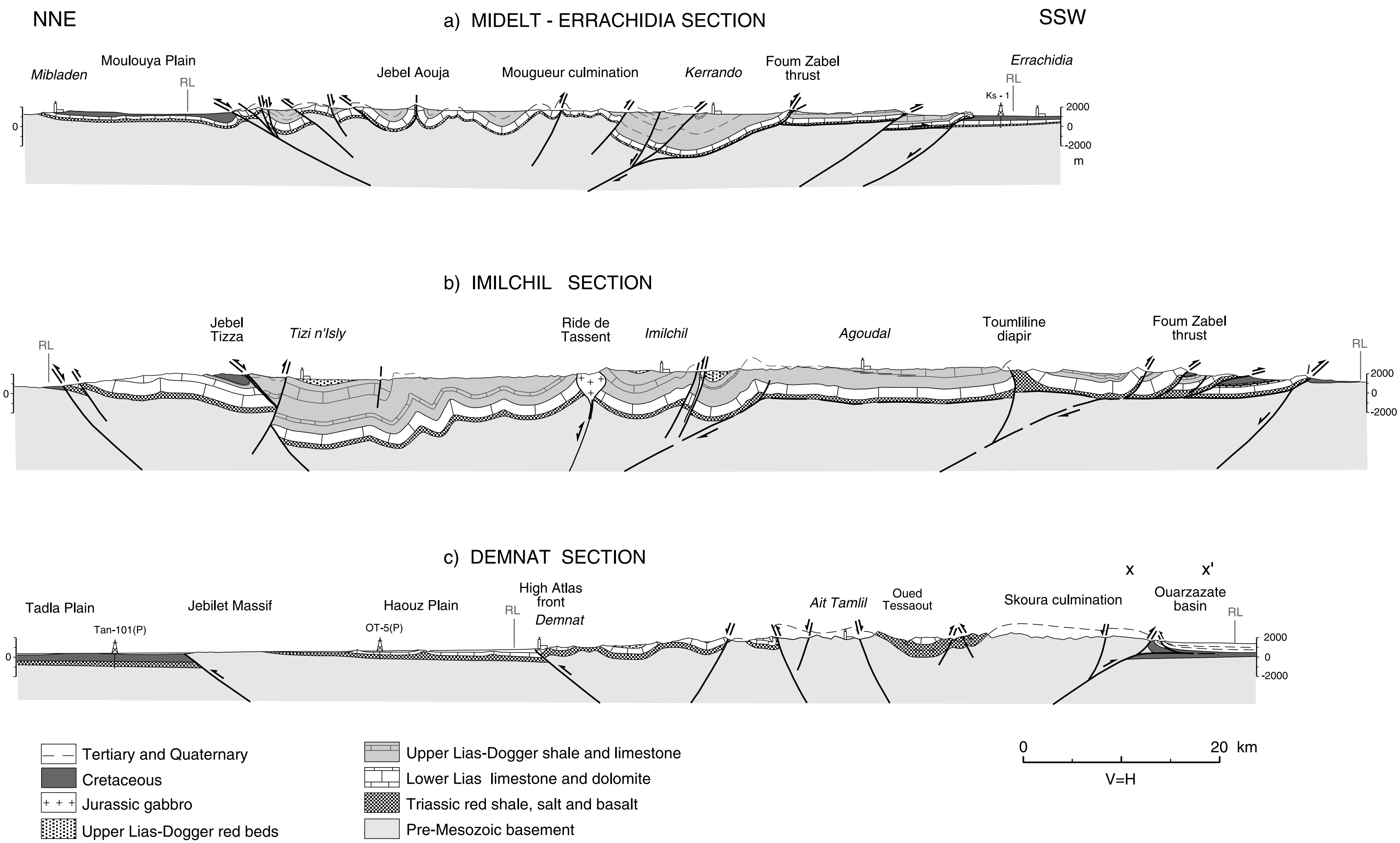


**Figure 2.** Serial geological cross sections through the High Atlas of Morocco (location in Figure 1b): (a) Midelt-Errachidia section, (b) Imilchil section, and (c) Demnat section. Segment x–x′ in 2c is adapted from *Errarhaoui* [1997]. RL, reference lines for restoration (see Figure 6). See color version of this figure at back of this issue.

tensional field [*Laville and Petit*, 1984; *Beauchamp*, 1988; *Piqué et al.*, 2000].

[8] Sedimentation of transgressive lower Liassic platform carbonates, a continuous key level consisting of a few hundred meters of limestone and dolomite, sealed the Triassic rift basins [*Laville et al.*, 1995; *Piqué et al.*, 2000]. During the late Liassic, the platform was drowned and disrupted [*Brechbühler et al.*, 1988; *Warme*, 1988; *Igmoullan*, 1993; *Poisson et al.*, 1998; *Souhel et al.*, 2000]. Consequently, subsiding basins started to differentiate, essentially coinciding with the trace of the present-day High and Middle Atlas. The Jurassic basins can be grouped into two main provinces located on either side of an emerged Massif Ancien [*Choubert and Faure-Muret*, 1962; *Du Dresnay*, 1971; *Warme*, 1988; *Stets*, 1992]. To the west, the basin was open to the Atlantic and related to its passive margin, and to the east there were several epicontinental troughs connected to the Tethys. From Toarcian to Bajocian times up to 5000 m of open marine shales (marls), calciturbidites and reefal limestones accumulated in the central High Atlas, while to the west, in the area adjacent to the Massif Ancien, emersion and terrestrial sedimentation dominated. Red beds indicating generalized filling and regression became widespread in the Bathonian [*Choubert and Faure-Muret*, 1962; *Jenny et al.*, 1981].

[9] During this period, the Atlas troughs were characterized by a complex fault pattern, developed under a NW-SE tensional field. Two models have been proposed to explain the overall basin structure: (1) a basin system dominated by movement along ENE-WSW strike-slip faults, with the generation of a complex set of compressional and tensional stepovers and pull-apart basins [*Mattauer et al.*, 1977; *Laville*, 1988; *Laville and Piqué*, 1992], and (2) a transtensional (oblique) rift, with internal NE-SW oriented extensional faults, oblique to the basin axis [*Warme*, 1988; *El Kochri and Chorowicz*, 1996]. Fanning of beds and low-angle unconformities are common in Jurassic sediments [*Jenny et al.*, 1981; *Laville*, 1988]. Crustal thinning was accompanied by alkaline gabbroic intrusions, dated as mid to late Jurassic [*Hailwood and Mitchell*, 1971].

[10] Jurassic rocks are usually absent beyond the southern and northern borders of the Atlas ranges. The Cretaceous is composed of a basal red bed unit followed by a platform limestone level dated as Cenomanian-Turonian. Although largely eroded in the interior of the High Atlas, these sediments form a tabular, expansive body that we believe covered the entire Atlas domain and overlapped the basement of its margins (Saharan platform, Anti-Atlas and Meseta), representing a postrift setting. This view is consistent with the geophysical imaging of the equivalent Jurassic basins in the American Atlantic margin, which are overlain by a marked postrift unconformity close to the Jurassic/Cretaceous boundary [i.e., *Hutchinson and Klitgord*, 1988]. The uppermost Cretaceous again consists of terrigenous red beds, indicative of the beginning of the Alpine compression to some authors [*Laville et al.*, 1977; *Froitzheim et al.*, 1988; *Amrhar*, 1995].

[11] Upper Paleogene to lower Quaternary continental deposits are contemporary with the main episode of compressional deformation and uplift, and are found essentially in the plains bordering the High Atlas [*Monbaron*, 1982; *Görler et al.*, 1988; *Fraissinet et al.*, 1988; *Morel et al.*, 1993; *El Harfi et al.*, 1996]. They reach a maximum thickness of 1200 m in the Ouarzazate syncline (Figure 1b). Younger Quaternary terrace gravels are still affected by compressional deformation, as shown in the northern borders of the High Atlas of Marrakesh and Midelt by *Dutour and Ferrandini* [1985] and *Morel et al.* [1993, 2000]. Despite these occurrences, large areas of the plains bordering the High Atlas lack Cenozoic deposits, being the Mesozoic or even the Paleozoic rocks exposed at the surface. (Figure 1b). The limited foreland basin record contrasts with the high topography of the Atlas. Interestingly, small conglomerate and lacustrine limestone inliers attributed to the Neogene are preserved in the interior of the High Atlas [*Zouine*, 1993], occasionally tens of km away from the borders (i.e., Rocher La Cathédrale, Jebel Tizza, Tasraft, etc.). These may reach thicknesses of over 500 m, and suggest that part of the High Atlas may once have been buried by synorogenic sediments but was later reexhumed.

### 2.2. Structural Geometry and Kinematics

[12] The rocks exposed in the High Atlas were deformed in upper crustal conditions, with very weak or no metamorphism. Compressional deformation is heterogeneously distributed: narrow deformation bands constituted by anticlines or thrust faults are frequently separated by broad synclines or tabular plateaux [*Schaer*, 1987]. Variations in Mesozoic stratigraphy and thickness across many thrust faults attest to their origin as synsedimentary extensional faults, which later experienced tectonic inversion. The trend of the main individual thrusts and folds is NE-SW, slightly oblique to the general trend of the range (Figure 1b). Consistent with the original dip of the Mesozoic basin-bordering faults, the present High Atlas is a doubly-verging chain, although an uplifted central axis or internal zone cannot be defined. In fact, due to the moderate degree of inversion, in much of the interior of the range, the basement is at lower structural elevation than in the peripheral forelands (Figure 2). A somewhat homogeneous erosion level in the central High Atlas results in Jurassic rocks dominating in outcrop (Figure 1b). West of Demnat, due to a thinner cover, the erosion has reached the level of the Paleozoic, that crops out extensively (''Massif Ancien''). An inspection of the geological map in this area reveals that basement is involved in the compressional deformation (Figure 1b), in a dominantly thick-skinned tectonic style characterizing the Moroccan Atlas [*Frizon de Lamotte et al.*, 2000]. The cross sections described below constitute the first series of detailed transects which cross the entire High Atlas and illustrate the structural variation along strike.

#### 2.2.1. Midelt-Errachidia Section

[13] This easternmost section shows many of the common characteristics of the transects (Figure 2a). The southern thrust front overrides a much reduced Mesozoic succession characteristic of the Saharan platform. South of the cross section, in the Tafilalet valley, Cretaceous sedimentary rocks directly overlie the Precambrian or Paleozoic

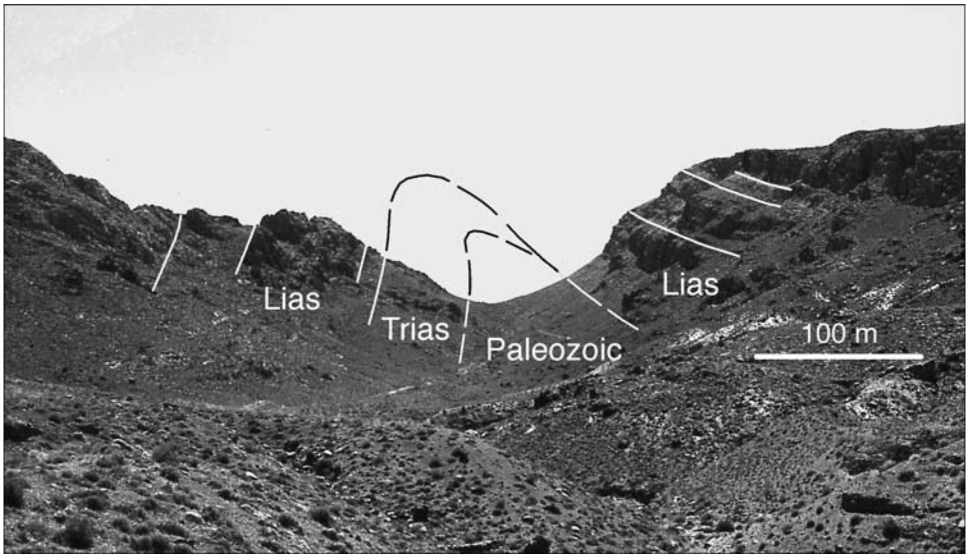


**Figure 3.** Field photograph of the western end of the Mougueur basement culmination, showing a tight anticline defined by Mesozoic beds, with Paleozoic slates occupying the core. The photograph is looking west; north is to the right.

of the craton, although at Errachidia, the Ks-1 well drilled a thin Triassic and Jurassic carbonatic to evaporitic series [*Jossen and Filali-Moutei*, 1992] (Figure 2a). The two frontal thrust sheets of the High Atlas contain a dominantly carbonatic Jurassic succession that is about 1000 m thick, overlain by mid to late Jurassic red beds. The southernmost unit shows a frontal overturned fold that was interpreted as a fault propagation fold detached along the Triassic by *Saint Bezar et al.* [1998], whereas the second thrust is associated to a persistent shift of the regional elevation of Jurassic beds and must already involve the basement. Farther north, the Foum Zabel thrust carries a much thicker Jurassic succession (at least 4000 m of shales with minor limestones), which describes the kilometric-scale Kerrando syncline (Figure 2a). The stratigraphic variation and the presence of basement slivers along the thrust indicate that it derives from the inversion of a basement-involved extensional fault. Triassic rocks along the sole of the fault define a hanging wall flat. Beds in the footwall are truncated in a high-angle ramp, but the relationships between hanging wall and footwall (where Jurassic beds are at much higher regional elevation) suggest that the Kerrando syncline cannot be attributed to this footwall ramp, but must have been inherited from the Mesozoic extension instead. A gentle unconformity within Jurassic beds in the northern part of the syncline supports this view.

[14] North of the Kerrando syncline there is a marked increase in structural relief, and lower Liassic limestones define a large anticlinorium that represents the western continuation of the basement culmination of Mougueur (Figures 1b and 2a). As seen in the culmination, the individual folds, although tight, involve the basement (Figure 3). In some cases, competence contrast between basement and cover may result in disharmony at the Triassic level and large-scale hinge collapse of the lower Liassic carbonates (e.g., Jebel Aouja, Figure 2a). Finally, on the northern border of the range, north-vergent thrust faults coexist with Jurassic normal faults that were not reactivated (Figure 2a). The High Atlas is separated from the Moulouya Plain (western end of the Algerian High Plateau) by a frontal thrust that overrides the overturned limb of a syncline in Cretaceous beds. This corresponds to footwall ramp geometry, and thus limits the degree of superposition and slip along the frontal thrust. The reduced and tabular Mesozoic succession of the Moulouya Plain, with a few hundreds of meters of Triassic and Jurassic rocks, can be traced from the Paleozoic outcrop of the Mibladen mining district.

#### 2.2.2. Imilchil Section

[15] In this transect, the southern frontal thrust also brings Jurassic rocks over the Cretaceous cover of the Saharan platform. The southern part of the High Atlas in this transect is made up of a 20 km wide thrust imbricate fan that is detached from the basement at the Triassic level. This detachment is evident in outcrop (expressed by hanging wall flats) and in cross-section construction, since the top of the basement keeps the same structural elevation below the successive thrust sheets (Figure 4). However, there are stratigraphic variations across each of the imbricate thrusts, and the frontal of them carries a small Paleozoic shortcut [*Hadri*, 1997; *Poisson et al.*, 1998]. This indicates reactivation of Mesozoic extensional faults, but their lower parts were probably left beneath the thin-skinned imbricate system.

[16] North of the frontal imbricates, there is a broad tabular region at a mean topographic altitude of over 2000 m (Figure 2b). This tabular area is only disrupted by a Triassic salt diapir at Toumliline, which originated during the Jurassic extension and was later slightly deformed during the Alpine compression. There is a surprising absence of a basement culmination that could act as a root zone of the frontal, detached imbricates described above

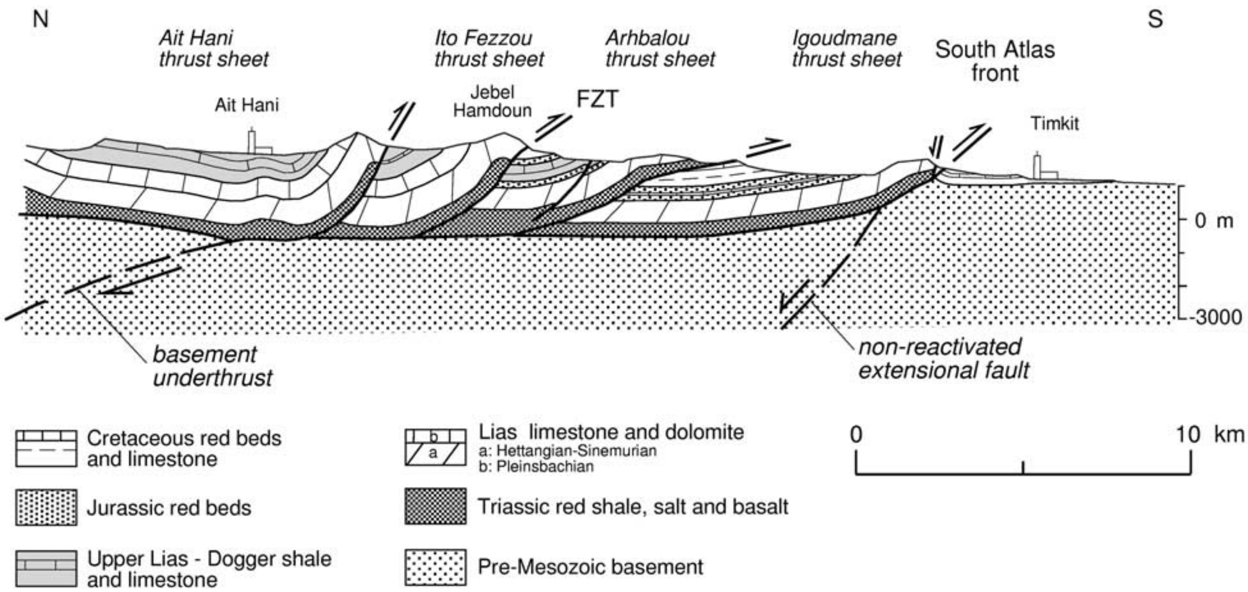


**Figure 4.** Cross section of a thin-skinned thrust system at the southern margin of the High Atlas north of Tinejdad. This section is a blowup of Figure 2b (see text for explanation). FZT, Foum Zabel thrust.

(Figure 2b). One possible explanation is that basement was shortened at the rear of the cover imbricate fan by underthrusting (the footwall moving down), without creating much structural relief.

[17] The central part of the section, close to Imilchil, is characterized by more regular, open folding (Figure 2b). Variations in structural elevation suggest that basement must be involved. Late Jurassic red beds are preserved in synclines, and show large-scale bed fanning and thickness variations indicative of growth folding. Comparable features in the same formation were described to the west of the transect by *Jenny et al.* [1981], and in older, Dogger marine sediments by *Laville* [1988]. This author relates them to synsedimentary shortening within a strike-slip system, whereas *Jenny et al.* [1981] interpret them as formed within an extensional regime, an interpretation we tentatively retain. Jurassic elongate gabbro plutons and dykes are occasionally found in cores of the intervening anticlines or faults (Figures 1b and 2b). These were intruded along Jurassic extensional openings [*Laville*, 1988], their dominant trend being NE-SW. Although some doming in host rocks may have been produced during intrusion, in our view the tight folding of sedimentary layers around the magmatic bodies was essentially produced during the later compressional episode. Bedding-perpendicular dykes in fold limbs irrespective of bedding dip supports this view.

[18] This folded area is bordered to the north by a north-vergent thrust, associated to a marked stratigraphic change [*Fadile*, 1987] (Figure 2b). Jurassic sediments pass from over 6500 m in the hanging wall to some 2000 m in the footwall, where they are overlain by Cretaceous and Neogene rocks (Figure 5). Although unconformable on overturned Cretaceous beds of the Tizza footwall syncline, Neogene sediments are also cut by the thrust fault, indicating their syntectonic nature. This footwall unit is in continuity with the Middle Atlas, and to the NNW is bordered by the frontal thrust that overrides the Tadla Plain, where, in a

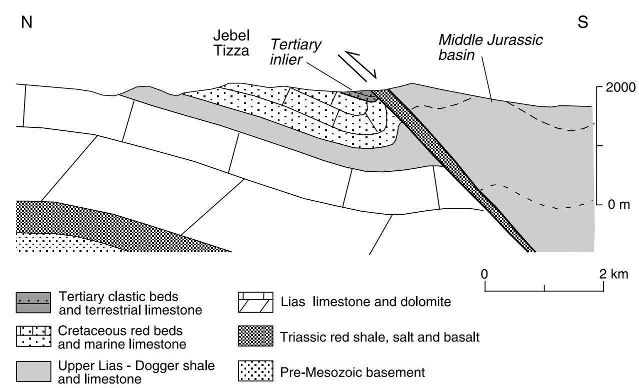


**Figure 5.** Detailed section of the Tizi n'Isly-Jebel Tizza area (northern part of section 2b), illustrating a long-lived fault history characteristic of High Atlas. The thrust fault is of Cenozoic age because it involves an outlier of lacustrine limestone, conglomerate, and shale which, by correlation with similar outcrops described by *Zouine* [1993], can be attributed to the Neogene. However, these rocks are unconformable on the Cretaceous of the overturned limb of the footwall syncline, indicating a previous episode of thrusting, and in addition, the thrust derives from the inversion of a major Jurassic extensional fault, as indicated by thickness variations of the upper Lias-Dogger sediments.

symmetrical disposition with respect to the southern margin, the Cretaceous lies directly over the Paleozoic basement.

#### 2.2.3. Demnat Section

[19] This westernmost transect has the highest topographic relief, and also shows a deeper erosional level, the Pre-Mesozoic basement being exposed in two large culminations (Skoura and Ait Tamlil) (Figures 1 and 2c). Basement is covered by a reduced Mesozoic succession preserved in some synclines, where the Jurassic red beds (here starting in the Toarcian) directly overlie the lower Liassic carbonates [*Le Marrec*, 1985]. The structure of the High Atlas at this transect is characterized by open folds, cut by high-angle thrust and normal faults (Figure 2c). From the Mesozoic beds preserved in the highest mountain peaks near the section line, it can be deduced that the top of the basement culminations had a tabular structure, with very little internal shortening. The High Atlas overrides the Ouarzazate Tertiary basin toward the south. In the northern margin of this basin, the Jurassic is absent and there is a complex system of thin-skinned forethrusts and back thrusts which involve Cretaceous and Cenozoic rocks [*Laville et al.*, 1977; *Fraissinet et al.*, 1988; *Errarhaoui*, 1997; *Beauchamp et al.*, 1999; *Frizon de Lamotte et al.*, 2000]. North of Demnat, the relief of the High Atlas experiences a marked recess, passing to the Haouz erosional plain. The plain is covered by a thin Neogene to Quaternary veneer (Figure 2c), which, although unconformable and largely undeformed, is thrust by the High Atlas [*Dutour and Ferrandini*, 1985; *Petit et al.*, 1985]. Paleozoic rocks appear again at the surface in the Jebilet Massif (Figure 1b), which according to *Chellaï and Perriaux* [1996] and *Morel et al.* [2000] is bordered to the north by a buried thrust fault. Boreholes located north of the Jebilet drilled the top of the basement at some 1000 m below sea level, under a Triassic, Cretaceous and Neogene cover in continuity with the Tadla Plain [*Jabour and Nakayama*, 1988]. This cover shows a gentle dip to the south, and east of the transect it is clearly thrust by the High Atlas mountain front at Beni Mellal (Figure 1b), of which the Jebilet is the western continuation. The Jebilet is a chain of low hills, and from a geomorphic point of view the High Atlas is considered to begin at the latitude of the town of Demnat.

#### 2.2.4. Comparison of the Sections

[20] The comparison of the three cross sections shows several common features and some differences. Along the three transects, the inverted Triassic and Jurassic basins of the High Atlas are thrust outward over Cretaceous postrift rocks which overlie basement or very reduced Triassic-Jurassic remains. Within the chain, there is a predominantly thick-skinned tectonic style, and most thrust faults and buckle folds, not necessarily fault-related, involve basement. Folds are markedly noncylindrical, occasionally with strong axial plunges, making correlations of individual structures between transects difficult. Displacement along individual thrust faults is moderate, as faults show common ramp-over-ramp relationships and die out laterally within a few kilometers. This does not favor the low-angle thrust interpretations for the northern margin of the High Atlas proposed by *Lowell* [1995] and *Beauchamp et al.* [1999]. Detached, thin-skinned thrusting is limited, restricted to a narrow belt in the southern margin of the chain (Figures 2b and 2c), and not always related to basement culminations hinterlandward. In general, section construction results in a less shortened basement compared to the cover; additional shortening mechanisms such as ductile flattening must have occurred in the former. Owing to this and the occurrence of buckling, we believe that there are no simple angular relationships between folds and basement faults. The eastern, Midelt-Errachidia cross section shows relatively closely spaced folds and thrusts, which to the west become more spaced and separated by broad, undeformed units. However, the most conspicuous variation between the cross sections concerns the erosional level. Exhumation appears greater in the westernmost section, where basement is frequently brought to the surface. This occurs without a correlative increase of apparent compressional deformation (actually it becomes milder), and is largely because the Mesozoic succession was originally thinner in this segment of the chain. The relationship between these features and the amount of tectonic shortening is further discussed in the following sections.

### 2.3. Section Restoration and Shortening Estimates

[21] The cross sections of the High Atlas have been restored to the precompressional state, to gain insight into the two-dimensional configuration of the Mesozoic basin and to evaluate the magnitude of the Alpine contraction (Figure 6). Restoration is based on bed lengths, essentially from competent lower Liassic strata, and assumes a basin system dominated by inclined extensional faults with moderate dip. Evidence of low-angle normal faulting was not found. The reference horizontal is placed at the top of the Jurassic rift sequence, since the overlying Cretaceous has homogeneous facies and thickness wherever present. The geometry of Triassic sediments is poorly constrained in the eastern transects, and they are represented as forming a tabular body with thickness based on the outcrops at the Mougueur culmination and the Ks-1 well at Errachidia. Some uncertainty concerning bed lengths and translations arises from the obliquity of internal structures with respect to the general High Atlas trend, which may have induced out of the plane movements. However, *Zouine* [1993] evaluated these potential movements to be minor (of the order of 4%, which was within the margin of error of his measurements), making restorations admissible. Jogs in the section lines (Figure 1b) are located in areas with tabular or very simple structure, and hence they have no significant effect on the calculations.

[22] Thickness variations of Jurassic rocks define a rather symmetrical extensional basin in profile, with central, rapidly subsiding zones (essentially during late Liassic to Dogger times), flanked by marginal zones of reduced thickness and, occasionally, rift shoulders without Jurassic between the basement and the postrift, expansive Cretaceous (Figure 6). Maximum basin width and depth is inferred in the Imilchil section (Figure 6b). The thickest accumulation preserved in the Midelt-Errachidia transect is

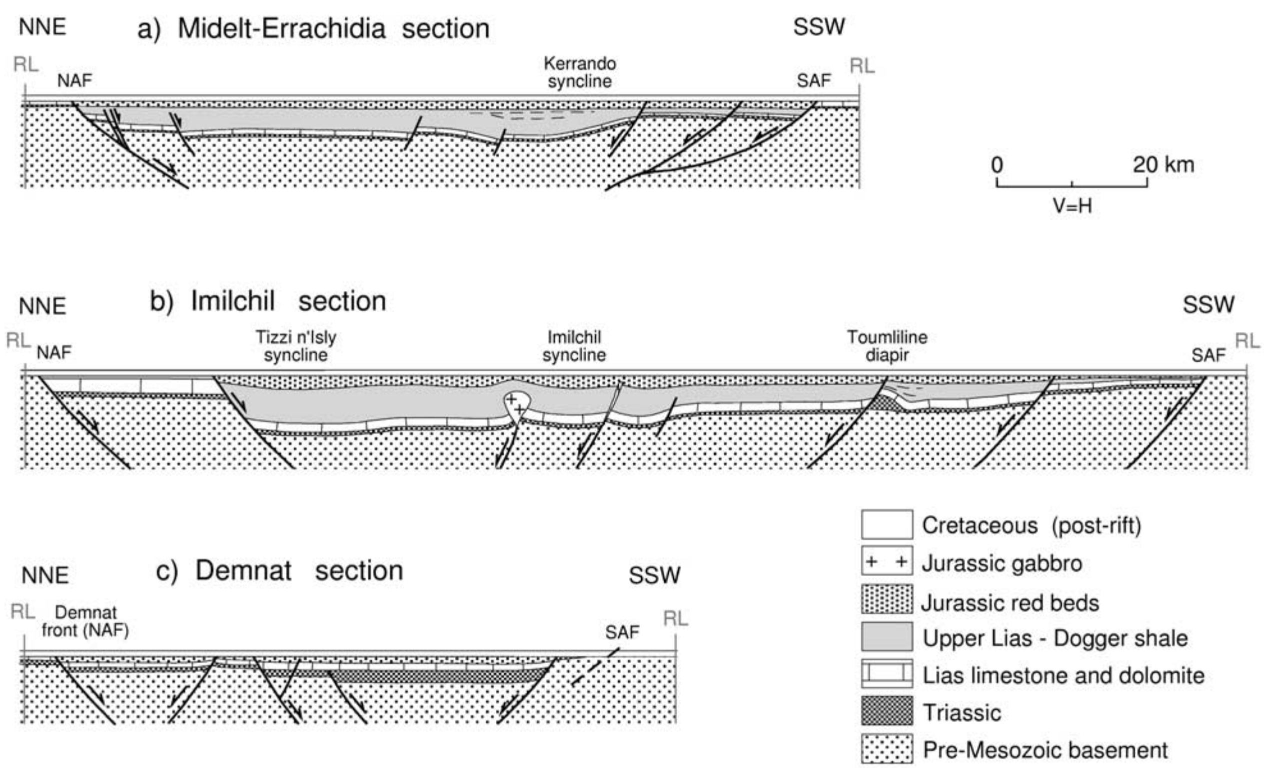


**Figure 6.** Restored versions of the cross sections of Figure 2 to the precompressional state: (a) Midelt-Errachidia section, (b) Imilchil section, and (c) Demnat section. Figure 6c corresponds to the High Atlas segment of Figure 2c, i.e., to the area located south of the mountain front at the town of Demnat (the Haouz Plain and Jebilet Massif are not included). The Jurassic red beds were drawn based on their geometry and thickness in neighboring outcrops because of their discontinuous preservation. NAF, northern High Atlas front; SAF, southern High Atlas front; RL, reference lines for restoration.

in the Kerrando syncline (Figure 6a), which is interpreted as a crustal-scale extensional ramp basin caused by inflections of the normal faults at depth [e.g., *Guimerà et al.*, 1995]. On the other hand, the Demnat transect has a much thinner Jurassic sequence; red beds directly overlie the lower Liassic carbonates, and, in nearby outcrops where the succession is more complete, these red beds have a maximum thicknesses of close to 1000 m or less [*Jenny et al.*, 1981]. As mentioned earlier, these beds have fanning geometries, where thickness increases away from the major faults (Figures 6b and 6c), suggesting gentle extensional fault propagation or drape folding predating inversion [e.g., *Mitra*, 1993].

[23] A few studies deal with the amount of shortening in the High Atlas. *Brede et al.* [1992] gave different values of percentage shortening across the range (decreasing from 20 % at Midelt to 10% west of Beni Mellal), although these authors neither documented actual, complete sections nor restored versions to preorogenic times for comparison. *Zouine* [1993] calculated total shortening values of 10–15% (in this case decreasing from west to east) based on excess area estimations of a Neogene erosional surface. Again, no documenting cross sections were provided, and these results were claimed to be underestimations by *Gomez et al.* [2000]. *Beauchamp et al.* [1999] presented a restored transect drawn west of Beni Mellal, where they arrived at a much higher shortening value of 36 km (25%).

[24] Comparison of our present-day and restored sections yields a total shortening of 26 km (24%) for the Midelt-Errachidia section, almost 30 km at Imilchil (which represents 18% as shortening is distributed across a wider area), and some 13 km (15%) for the High Atlas of Demnat (i.e., the part of section 3c located to the south of the mountain front at the town of Demnat). The basement parts of the latter section are restored on the basis that, in remaining Jurassic outliers nearby, the beds show always a tabular disposition (e.g., Jebel Rat [*Jenny*, 1985]). The values obtained cannot be gross underestimations as (1) there are no indications of regionally significant internal deformation in Jurassic competent strata, (2) most thrust faults appear in ramp-over-ramp configurations, thus precluding the possibility of overlooking hidden footwall flats or eroded hanging wall leading edges.

[25] The scarce preservation of Jurassic red beds adds some uncertainties concerning basin thickness or the

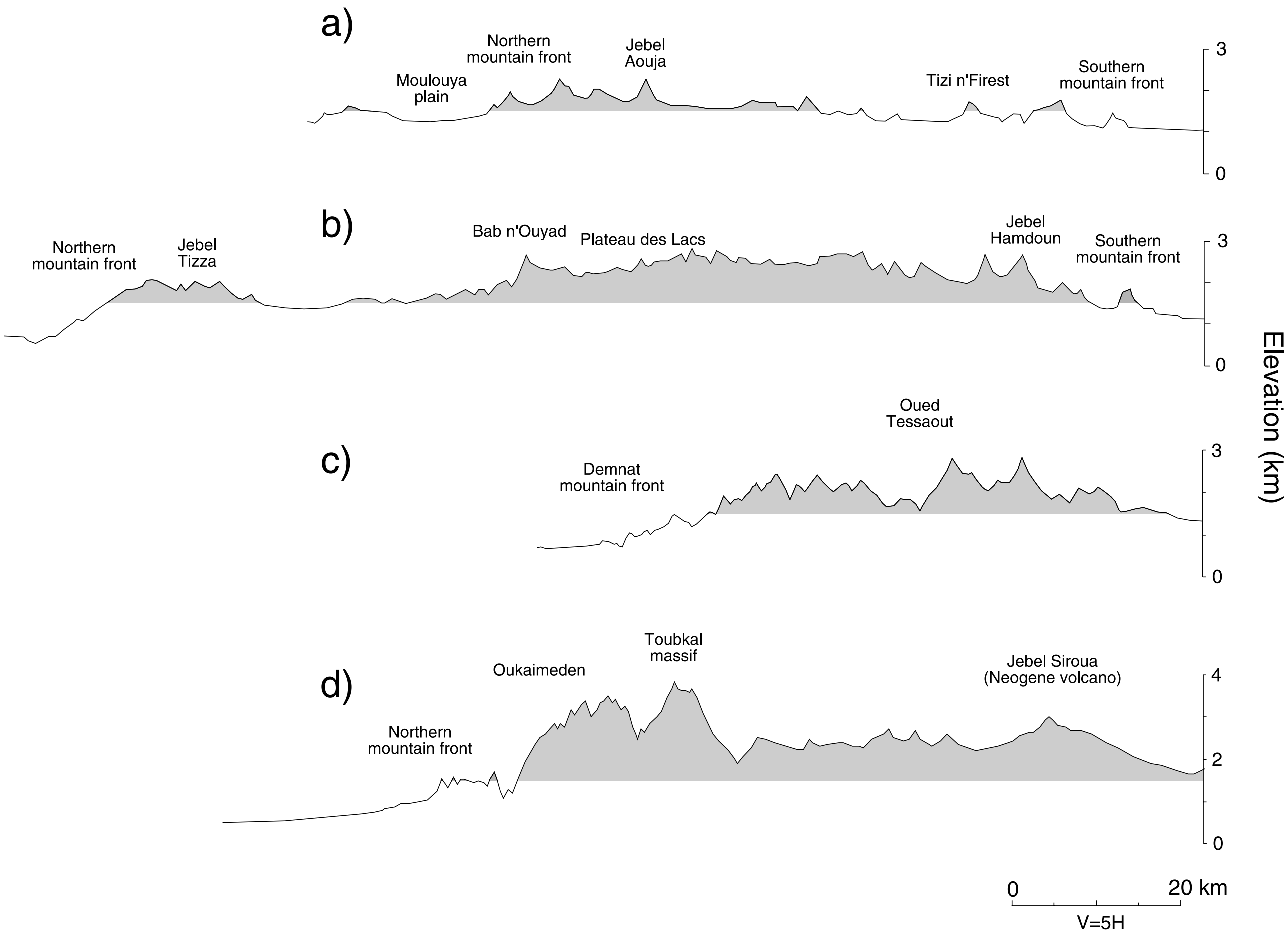


**Figure 7.** Serial topographic profiles of the High Atlas, showing the general increase of relief from east to west along the range (see Figure 8 for location). (a–c) Profiles coinciding with the Midelt-Errachidia, Imilchil, and Demnat geological cross sections, respectively, the latter including only the High Atlas part. (d) Profile running from the Massif Ancien of the High Atlas to the Siroua volcanic massif in the Anti-Atlas, in a zone where there is no clear boundary between these mountain chains. For reference, the percentage of the profile that is above 1500 m (shaded) is 59% at Midelt; it is comparable in the Imilchil and Demnat profiles (83% and 80%, respectively) and increases farther west to profile d (93% above 1500 m).

amount of extension and shortening. For this unit, we have used average thicknesses from the nearest outcrops in each section, but have calculated the error margins in the shortening calculations that would have arisen from a 50% error in the thickness assigned, finding these to be minor. If red bed thickness was 50% greater than assigned, shortening would then have been underestimated by 1.2, 1.8 and 0.7 km for the Midelt, Imilchil and Demnat sections respectively (that is, 0.8, 0.9 and 0.7%). Conversely, if the thickness was 50% smaller, shortening would have been overestimated by 0.5, 0.2 and 0.7 km (0.3, 0.2 and 0.6%) in the three sections. Similarly, we have evaluated the effect of varying the dip of the major faults within the restored sections. A systematic decrease of 10° would represent an extra shortening of 1, 2.3 and 1.2 km in the Midelt, Imilchil and Demnat sections, whereas a systematic increase of the same magnitude would represent a shortening decrease of 1.1, 1.8 and 1 km, respectively.

[26] We suggest that the higher shortening value in the *Beauchamp et al.* [1999] section, which was traced in between our Imilchil and Demnat sections, is probably an overestimation arising from poor imaging of fault cutoffs in the seismic lines of the marginal zones. These admit alternative, more conservative interpretations [e.g., *Zouine*, 1993], especially considering that some of their far-traveled thrusts die out laterally after short distances [e.g., *Choubert et al.*, 1977].

[27] Our data indicate that percentage shortening decreases from east to west along the High Atlas. This tendency is further supported by the fact that the western High Atlas (the Massif Ancien and farther west) shows less Alpine deformation structures, without major thrust faults [*Froitzheim et al.*, 1988], and in addition, in a section of the

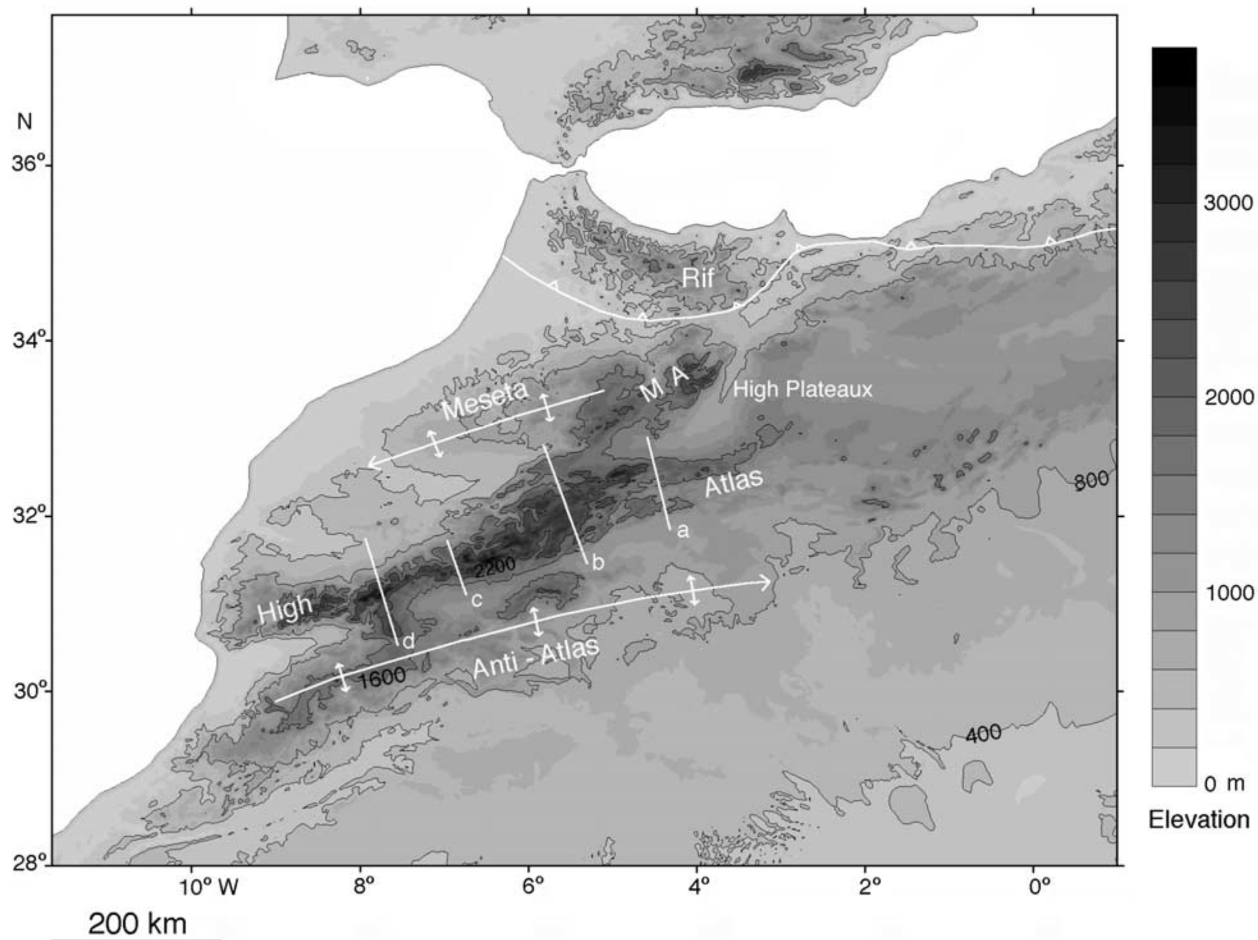


**Figure 8.** Topographic map of northwest Africa, showing the intracontinental mountains and elevated plateau in front of the Rif orogen (derived from the U.S. Geological Survey GTOPO30 database, available at http://edcdaac.usgs.gov/gtopo30/gtopo30.html). Indicated are the High Atlas profiles of Figure 7 and a tectonic interpretation of the Anti-Atlas and Meseta upwarps as large (lithospheric-scale) buckle folds. MA, Middle Atlas.

southern and most deformed part of the High Atlas of Agadir, near the Atlantic coast, *Frizon de Lamotte et al.* [2000] estimated a shortening of about 10%.

## 3. Crustal Shortening and Topography

[28] The topographic elevation of the Moroccan High Atlas increases along strike from east to west, from the Tamlelt area, close to the Algerian border, where the range is at a mean elevation of about 1000 m, to the Massif Ancien, where summits such as the Jebel Toubkal surpass 4000 m. This trend is illustrated by the topographic profiles shown in Figure 7, some of which follow the geological cross-section lines.

[29] The Midelt transect is mostly at a mean elevation of more than 1500 m above sea level, whereas much of the Imilchil section lies at an altitude close to 2500 m (Figure 7). In the Demnat area, the relief is stronger; several summits exceed 3000 m, although there are deep valleys as well, and the mean elevation is comparable to that of the Imilchil profile. The relief increases conspicuously farther west, to the high peaks of the Tizi n'Tichka and Toubkal massifs (Figures 7 and 8).

[30] This topographic trend correlates inversely with the tectonic shortening recorded along the High Atlas. As is particularly evident between the Midelt-Errachidia and Imilchil sections, while the topography increases, the percentage shortening (and presumably, crustal thickening) decreases. This indicates that not all the observed topography can be directly related to tectonic convergence and crustal thickening, and there must be an additional factor contributing to the uplift of this mountain chain.

[31] Seismic refraction studies of the Atlas system failed to detect large crustal roots [*Wigger et al.*, 1992], and based on gravity and isostatic anomaly maps, some authors have suggested that much of the High Atlas is isostatically

uncompensated [*Van den Bosch*, 1971; *Makris et al.*, 1985]. According to *Wigger et al.* [1992], the crustal thickness beneath the High Atlas of Midelt-Errachidia is about 35–38 km, being about 33–36 km in the peripheral plains [*Makris et al.*, 1985; *Wigger et al.*, 1992; *Sandvol et al.*, 1998]. Comparable values were estimated from combined gravity and seismic data for the western segments of the chain [*Makris et al.*, 1985; *Tadili et al.*, 1986]. This supports the fact that tectonic thickening cannot explain all the observed elevation, and also precludes crustal thickening by magmatic addition.

[32] The isostatic state of the High Atlas Mountains can be evaluated by calculating the changes in crustal thickness during the Jurassic rifting and the Cenozoic compression, and comparing the values to the present topographic elevation. The crustal thickness after Jurassic rifting is given by

$$t_{\mathrm{cr}} = t_{\mathrm{co}} 1/\beta + t_s, \tag{1}$$

where $t_{\mathrm{co}}$ is the original crustal thickness, $t_s$ is the basin sedimentary thickness and $\beta$ is the stretching factor, that gives the crustal thinning [*Bickle and Eriksson*, 1982]:

$$1/\beta = 1 - [t_s(\rho_s - \rho_m)/t_{\mathrm{co}}(\rho_c - \rho_m)], \tag{2}$$

where $\rho_s$, $\rho_c$ and $\rho_m$ are the average densities of sediments, crust and mantle respectively. Simplifying to one-dimensional calculations, we can take 5 km to be the average thickness in the Jurassic basin center and an original crustal thickness of 34 km, based on the stable areas outside the Atlas chains [*Tadili et al.*, 1986; *Wigger et al.*, 1992]. Mean density values of 2.5, 2.8 and 3.3 g/cm$^3$ for sediments, crust and mantle, yield a preorogenic crustal thickness ($t_{\mathrm{cr}}$) of about 31 km.

[33] Assuming plane strain, ~24% of Cenozoic average shortening at the Midelt-Errachidia transect gives a compressional crustal thickness of over 40 km, and ~18% at Imilchil gives ~38 km, values that, combined with erosion, are reasonably close to the published geophysical results. According to Airy isostatic principles, the relationship between the elevation ($h$) and the thickness of the crustal root ($t_r$) under a mountain range is given by

$$t_r = \rho_c h/\rho_m - \rho_c, \tag{3}$$

which with the above densities means some 4.5 km of root to support each km of elevation above sea level. With a mean elevation of over 1500 m, great part of the Atlas of Midelt-Errachidia is close to a compensated state or only slightly uncompensated (in terms of Airy-type mechanisms), whereas most of the Imilchil transect is over 1 km too high, an imbalance that would be even larger in the higher segment of the Demnat transect.

## 4. Origin of Topography

[34] To understand the origin of the topography of the High Atlas, it must be taken into account the fact that the current elevated area of the North African foreland exceeds the extent of the Atlas folded belts (Figure 8). The High Plateaux and part of the Moroccan Meseta are higher than 1000 m, whereas the Anti-Atlas reaches altitudes of 2000 m. All these regions are generally under a positive isostatic anomaly [*Van den Bosch*, 1971; *Makris et al.*, 1985].

[35] The Meseta and the Anti-Atlas basement massifs did not escape Alpine orogeny completely, although they were deformed very mildly. Their geometry conforms to very large (100 km scale), gentle undulations that, according to their wavelengths can be attributed to crustal or lithospheric-scale folding [e.g., *Cloetingh et al.*, 1999]. This geometry can be perceived from the surface morphology shown in Figure 8. In the case of the Anti-Atlas, the disposition of Paleozoic and Cretaceous strata indicates that large-scale buckle folding had a long history: it started during late Paleozoic times (when the complex doming of the Anti-Atlas and the synformal Tindouf basin were formed), and persisted until recent times, as indicated by the geomorphologic evidence of uplift presented by *Stablein* [1988]. This folding process represents an alternative deformation mechanism of the continental interior, in addition to the fault zone reactivation that occurred in the High Atlas, and its recent activity in the Meseta and Anti-Atlas may partly explain the relatively high topography of these areas.

[36] Other geological and geophysical data shed light on the origin of relief in the North African foreland. Analyses of seismic wave travel times by *Hoernle et al.* [1995] and *Seber et al.* [1996] revealed low velocities in the upper mantle suggesting that the lithosphere is relatively thin beneath the Atlas ranges. This is consistent with localized high values of heat flow [*Ramdani*, 1998], and with the presence of strongly subsaturated alkaline volcanism, of Tertiary and Quaternary age, which even includes carbonatite occurrences [*Agard*, 1973; *Harmand and Cantagrel*, 1984]. This volcanism, although scarce in the High Atlas, is abundant in the peripheral Meseta, Middle Atlas, Moulouya and Anti-Atlas areas. These two lines of evidence suggest that topography may at least be partially supported thermally, thus indicating a mantle source for uplift in the North African foreland of Morocco. It is worth noting that neotectonic features [*Dutour and Ferrandini*, 1985; *Harmand and Moukadiri*, 1986; *Morel et al.*, 2000] and earthquake fault plane solutions [*Coisy and Frogneux*, 1980; *Medina and Cherkaoui*, 1991] all indicate a compressional stress field in recent times, so extension can be ruled out as a cause for lithospheric thinning.

[37] In the light of these facts we suggest that, in addition to moderate crustal thickening and lithospheric folding, which can account for the difference in elevation between the mountains and peripheral high plains, the residual topography may ultimately be caused by a mantle upwelling (a mantle plume?) and thus be dynamically supported. A strong positive geoid anomaly in the Atlas system (after data from the TOPEX 2 min grids, available at http://topex.ucsd.edu), supports this view. We further suggest that mantle-related uplift, together with lithosphere folding, has

counterbalanced the flexural subsidence associated to the High Atlas thrust loads, and explains the poor development or preservation of foreland basin sedimentary deposits.

## 5. Conclusions

[38] The principal modes of compressional deformation and uplift in the interior of continents are well illustrated by the High Atlas and surrounding mountains: (1) the reactivation of older discontinuities (in this case the inversion of rift-transtensional troughs), and (2) crustal-scale or lithospheric-scale buckling of more intact portions of cratons. During High Atlas inversion, crustal shortening and thickening took place to a certain degree, and this might explain differential topography with respect to the flanking plains. Crustal thickening also explains the gap of recent volcanism in the High Atlas, which on the other hand is abundant in the peripheral areas. However, the moderate values of shortening and their poor correlation with the surface elevation suggest that crustal thickening cannot explain the full topographic pattern. The Midelt-Errachidia transect, with relatively lower mean elevation, higher percentage shortening and an uppermost crust composed of Mesozoic sedimentary rocks may be closer to isostatic equilibrium than segments located to the west. Gravity modeling is in progress to test this. The potential degree of isostatic uncompensation increases westward in the central High Atlas, as does the abundance of volcanism in the surroundings, all suggesting an increasing role of a thermally anomalous mantle in elevating topography. This, coupled with geophysical data supporting a thin lithosphere, indicates that both crust and mantle contributed to surface uplift: a combination of crustal isostatic and dynamic topography may explain the origin of the mountains of northwest Africa.

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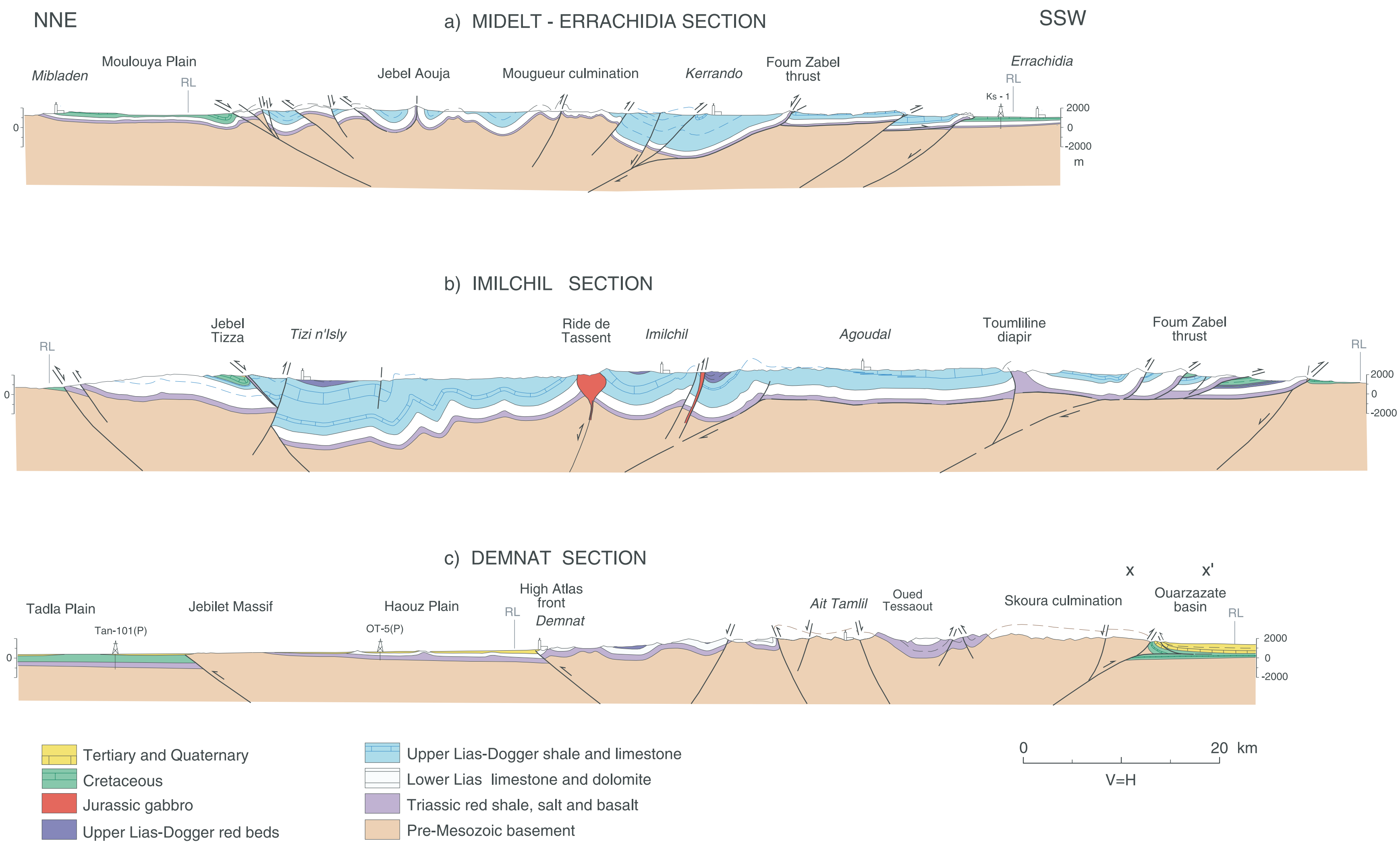


**Figure 2.** Serial geological cross sections through the High Atlas of Morocco (location in Figure 1b): (a) Midelt-Errachidia section, (b) Imilchil section, and (c) Demnat section. Segment x–x′ in 2c is adapted from *Errarhaoui* [1997]. RL, reference lines for restoration (see Figure 6).