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Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological-geophysical transect

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Abstract. The High Atlas Mountains of North Africa were formed over a major intracontinental rift system that had extended from what is now the Atlantic margin of Morocco to the Mediterranean coast of Tunisia. The Atlas rift system began in the Triassic and was active through the Jurassic. The inversion phase of the Atlas rift system began in the Early Cretaceous and extended into the present. The major uplift phase occurred between 30 and 20 Ma (Oligocene-Miocene) and corresponds to the Alpine orogenic event. The uplift and inversion of the Atlas rift system resulted in a shortening of the rift basin by a minimum of 36 km. A restoration of the deformed cross section indicates the original Atlas rift basin was approximately 113 km wide, comparable to the width of the present-day Red Sea. Synrift and postrift sedimentary rocks were uplifted by the reactivation of synrift normal faults, with further shortening along newly formed thin-skinned thrust faults. Structures formed by the reactivation of synrift faults resulted in structures with different geometries than those created by newly formed fault-bend and fault-propagation faults. Shortening across the High Atlas Mountains involved a partitioning of strain, with the greatest magnitude of shortening occurring along the margins of the High Atlas Mountains.

1. Introduction

Intracontinental rift systems are important tectonic features as they may focus strain during plate convergence along nearby plate boundaries. The style and magnitude of the rifting (pure shear or simple shear) and the orientation of the rift system relative to plate convergence contribute to the magnitude of the shortening and the style of folds and thrusts. Rift system geometry may control the vergence or bivergence of thrusting during inversion. Stresses transmitted during plate convergence may be accommodated by decoupling of the crust at different levels depending on the variation of crustal thickness and yield strengths associated with the rifts. Multiple levels of detachment may result.

The High Atlas Mountains of Morocco were formed by the reactivation of a major intracontinental rift system during convergence of the African and European plates during the Cenozoic [*Dewey et al.*, 1989]. The convergence of the North African and Eurasian plates during the Cenozoic through Present may be partially accommodated by shear along Proterozoic zones of weakness in the crust [*Milanovsky*, 1981; *Ziegler*, 1982; *Laville and Pique*, 1991; *Giese* and *Jacobshagen*, 1992]. Tectonic inversion occurs when preexisting extensional faults or lateral discontinuities in the crust affect the architecture of subsequent compressional structures. The architecture of these structures depends upon the geometry and azimuth of the preexisting structure relative to the orientation of the subsequent compressional events [*Williams et al.*, 1989; *Vially et al.*, 1994; *Beauchamp et al.*, 1996]. In certain cases listric or low angle normal faults will reactivate with a reverse sense of motion, particularly if the fault is orthogonal to the azimuth of compression. Higher angle normal faults may act as a buttress that focuses strain. The focusing of strain related to compression across the higher angle normal fault may lead to the formation of new low angle thrust faults or footwall shortcut faults [*Williams et al.*, 1989; *Beauchamp et al.*, 1996].

This paleo-Atlas rift system extended from what is now the Atlantic margin of Morocco, to the Mediterranean coast of Tunisia (~2000 km) (Figure 1). The paleo-Atlas rift system was oblique to the orientation of the Atlantic rift during the Triassic-Jurassic. The oblique orientation of the rift system resulted in structures that are representative of transtensional deformation such as pull-apart basins, extensional ramps and en echelon faulting [*Schaer and Rodgers*, 1987; *Laville and Pique*, 1992; *Vially et al.*, 1994; *Beauchamp et al.*, 1996]. The greater part of Atlas rift system

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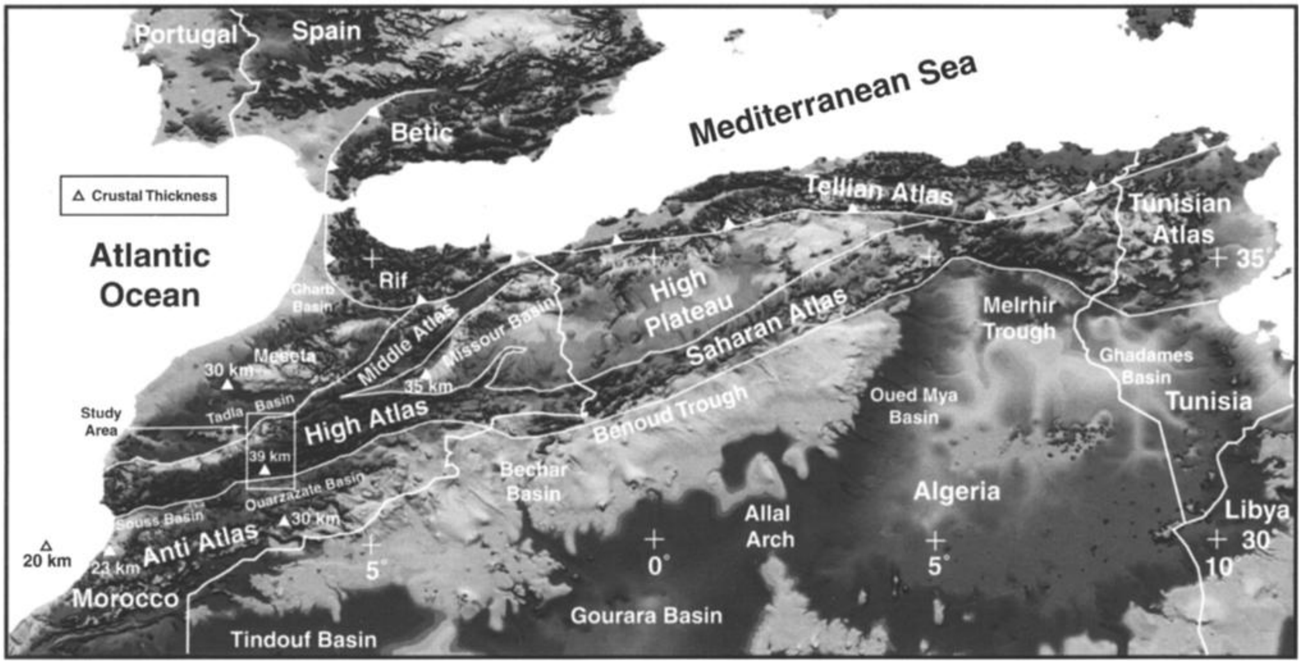

Figure 1. Map of North Africa showing the location of the region of transect A-A' (box), the High Atlas, Anti-Atlas, Tellian Atlas, and Betic Rif orogens overlain on digital topography. The white lines refer to tectonic boundaries for the original Mesozoic rift system and the extent of uplift and shortening in the Atlas mountains. Estimates of crustal thickness in Morocco are also shown.

was oriented orthogonal to the direction of plate convergence and compression during the Tertiary. Regions of the Atlas such as the Middle Atlas mountains were oriented oblique to the direction of plate convergence during the Tertiary, and may have been subjected to oblique compression [*Beauchamp et al.*, 1996; *Gomez et al.*, 1996; *Gomez et al.*, 1998].

A geophysical-geological cross section (Transect A-A', Plate 1) was constructed across the central High Atlas Mountains of Morocco to develop a better understanding of how intracontinental mountain belts are formed (Figure 2). The construction of a balanced and restorable cross section across the High Atlas Mountains enables the study of the kinematic history of the mountain belt, as well as how strain resulting from convergence between the European and African plates is partitioned. Based on the interpretation of previously unpublished seismic reflection profiles and field work, the shortening between the Ouarzazate basin (regional pin line) and the Tadla basin (local pin line) on the deformed and restored sections (A-A') is estimated to have been ~36 km (Plate 1). The amount of shortening across what is interpreted to have been the original High Atlas rift basin (113 km in width, restored section) was estimated between reference lines (A) and (B) (Plate 1). Values for shortening indicate that the Atlas rift basin was shortened by a minimum of ~32%. Previous estimates of the amount of shortening across the High Atlas had been between 10 and 20% [*Brede*, 1992].

The balanced section is constrained by data from Cenozoic age rocks over most of the transect. Shortening in older rocks (Mesozoic-Paleozoic) along the line of the section is interpretive. Shortening within the Paleozoic is represented by a series of duplexes that mechanically achieve the overall uplift necessary to balance the section, but the duplexes are not intended to suggest the true structural complexities in the Paleozoic and older rocks (shown as dashed lines, Plate 1). This interpretation of the shortening in Paleozoic and older rocks is believed to represent the most likely model for crustal thickening. The absence of high heat flow in the Atlas mountains precludes the possibility of a thermally supported uplift.

Shortening across the High Atlas Mountains resulted in a partitioning of strain, with the greatest magnitude of shortening occurring along the margins of the High Atlas. This partitioning of strain has been recognized elsewhere in the Saharan Atlas of Algeria [*Vially et al.*, 1994], and in similar age inverted mountain belts (Venezuelan Andes) of South America [*Colletta et al.*, 1997]. Two pin lines and loose lines were placed along transect A-A' to determine the magnitude of shortening across the High Atlas (Plate 1). Thrusting and inversion resulted in the concentration of shortening along the two margins of the High Atlas over a distance of 10 km (Plate 1, Transect A-A') with little shortening along the section in the interior of the High Atlas Mountains.

Many foreland fold and thrust belts have only one sense

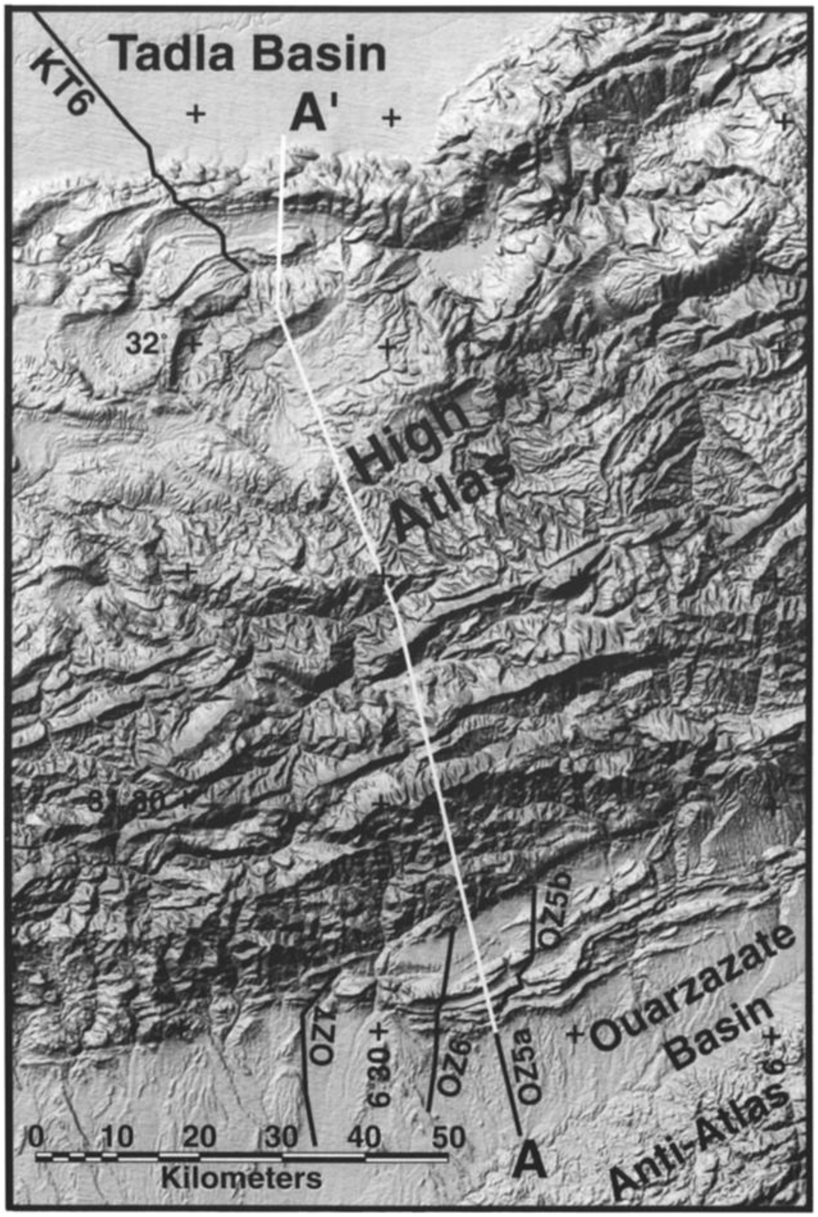

Figure 2. Map showing the location of seismic reflection profiles (KT-6 in the Tadla basin, OZ-5a and OZ-5b, OZ-6 and OZ-7) used in the study and the location of transect A-A'.

of thrusting and vergence. Intracontinental mountain belts are commonly bivergent and have a more complex history for the timing and sequence of thrusting. The bivergence of thrusts in intracontinental mountain belts may be related to the reactivation of rift fault systems that are often bivergent [e.g., *Bally*, 1984; *Fraissinet et al.*, 1988; *Hayward and Graham* 1989; *Jossen and Filali-Moutei*, 1992; *Vially et al.*, 1994; *Colletta et al.*, 1997]. A bivergent intracontinental mountain belt implies that at some point in the orogen there should be a null point on either side of which the sense of vergence is reversed. Bivergent thrusting/inversion may result in foreland basins developing on both sides of the orogen. Tectonic thickening of the upper crust along the margins of the High Atlas Mountains by thrusting and fault reactivation occurred without the development of significant foreland basins, indicating a strong lithosphere. The High Atlas mountain belt is a bivergent mountain belt that has been affected by moderate to strong inversion [*Gauthier*, 1960; *Rolley*, 1978; *Vially et al.*, 1994; *Lowell*, 1995].

2. Regional Setting

2.1. Topography and Crustal Thickness Variations

Crustal thickness variations across the High Atlas highlight one of the most remarkable aspects of this orogen. Despite topography locally in excess of 4 km elevation, the crust apparently is nowhere in excess of 40 km thick. The crustal thickness outside of the High Atlas Mountains has been estimated to be 35 km from single station earthquake seismological data near Midelt in the Missour Basin [*Sandvol et al.*, 1996]. A refraction study parallel to the axis of the High Atlas Mountains [*Makris et al.*, 1985] found that the crust beneath the Meseta (Figure 1) is 30 km thick, thinning to between 18 and 20 km along the Moroccan shelf. South of the High Atlas Mountains, crustal thickness below the Anti-Atlas is 30 km, thinning towards the Atlantic margin (24 km). Gravity data were modeled in two dimensions along two profiles across the High Atlas indicating crustal thickness between 34 and 38 km. *Makris et al.* [1985] noted that these gravity models indicate a lack of isostatic compensation relative to the topography of the High Atlas Mountains. Later refraction studies were oriented perpendicular to the High and Middle Atlas Mountains [*Wigger et al.*, 1992]. The crust beneath the High Atlas Mountains was interpreted to be 38-39 km thick, with an average crustal thickness to the north of the mountains of 35 km, with a possible low-velocity zone at depths of 10-15 km [*Wigger et al.*, 1992].

These results can be interpreted in concert with electrical resistivity studies that indicate the presence of a north dipping layer of low resistivity between 10-15 km beneath the High and Middle Atlas Mountains [*Schwartz et al.*, 1992]. This high conductivity, low-velocity layer was interpreted by *Schwartz et al.* [1992] as a crustal detachment that could also be related to the previously mentioned low-velocity zone and may be equivalent to the base of the Paleozoic as interpreted on transect A-A' (Plate 1).

2.2. Prerift Phase

The Anti-Atlas south of the High Atlas Mountains expose Precambrian through Paleozoic age rocks of the West African craton. The Late Proterozoic rocks were affected by the Pan-African orogeny that occurred between 680 and 570 Ma [*Pique et al.*, 1993]. The Paleozoic stratigraphic section exposed in the large anticlinorium of the Anti-Atlas was part of Paleozoic Gondwana. These rocks of the West African craton were then affected by deformation in the Hercynian orogeny (330-250 Ma).

Precambrian granites and volcanics outcrop in the Anti-Atlas at the southern end of the transect A-A' of this study (Figure 2). Similar Precambrian rocks have not been recognized to the north of the High Atlas Mountains. Igneous rocks exposed north of the High Atlas Mountains are granites of Hercynian age [*Michard*, 1976]. The absence of Precambrian rocks in outcrop or penetrated by wells may indicate a much thicker Paleozoic sequence north of the High Atlas Mountains. It is also possible that there is no Precambrian crust north of the Atlas Mountains.

During the Early Paleozoic to Middle Devonian, western and central Morocco was a shallow platform distinguished by the deposition of epicontinental facies. Silurian age sedimentary rocks are composed of thick black, organic-rich graptolitic shales [*Michard*, 1976] and thin sandstone and crinoidal limestones (Figure 3). These Silurian shales range from 200 to 1000 m in thickness and provide an important sedimentary sequence for the nucleation of thrusts and duplexes within the Paleozoic rocks. The Late Devonian to Early Carboniferous was characterized by transtensive sedimentary basins [*Pique et al.*, 1993] that are not uniformly distributed throughout Morocco. Permo-Carboniferous rocks may have been eroded from constraining bends (flower structures) and preserved in the releasing bends (pull-apart basins) [*Jabour and Nakayama*, 1988]. Wells drilled to the north of the High Atlas along transect A-A' (DRZ-1 and KMS-1) did not encounter rocks of Devonian to Carboniferous age (Figure 2).

2.3. Synrift Phase

The Atlas rift system began in the Triassic and was active depositionally through the Jurassic [*Michard*, 1976; *Manspeizer et al.*, 1978; *Schaer and Rodgers*, 1987; *Laville*, 1988]. The synrift sequence of the Atlas began with the deposition of continental Triassic clastic rocks composed of interbedded sandstones, shales, anhydrite, and volcanics [*Manspeizer et al.*, 1978; *Beauchamp*, 1988; *Medina*, 1988]. The basal Triassic is composed of continental red beds which are locally overlain by thoeliitic basalts (~400 m in western Atlas) [*Manspeizer et al.*, 1978]. Thicknesses of Triassic strata range between 4000 and 5000 m in the western High Atlas, with thicknesses of less than 1000 m in the central High Atlas, eastern High Atlas, Middle Atlas Mountains, and Missour basin [*Manspeizer et al.*, 1978, *Beauchamp et al.*, 1996] (Figure 1). Triassic age rocks are absent in the Ouarzazate basin and Anti-Atlas south of the High Atlas Mountains along the southern part of the cross section A-A' (Figure 2). Well data in the Tadla basin (DRZ-1 and KMS-1) indicate that Triassic rocks north of the High Atlas have thicknesses of between 350-450 m. Triassic rocks exposed along the line of the transect (A-A') at the northern and southern margins of the High Atlas Mountains are less than 500 m in thickness.

Triassic evaporites provide a weak stratigraphic interval that controlled the development of Cenozoic contractional structures. In the field, Triassic evaporitic rocks are intensely deformed and sheared. The evaporites and associated basalts characteristically have a high-amplitude response on seismic reflection data, which is a key marker for interpretation [*Beauchamp et al.*, 1996].

Synrift Jurassic sedimentary rocks were deposited within the Atlas rift system only in the present High Atlas Mountains, with the exception of the Missour basin and the High plateau [*Laville*, 1988; *Beauchamp et al.*, 1996] (Figure 2). Late Jurassic and Early Cretaceous sedimentary rocks are present offshore along the passive margin of Morocco, as extension continued unabated during the successful opening of the Atlantic [*Broughton and Trepanier*, 1993]. Jurassic

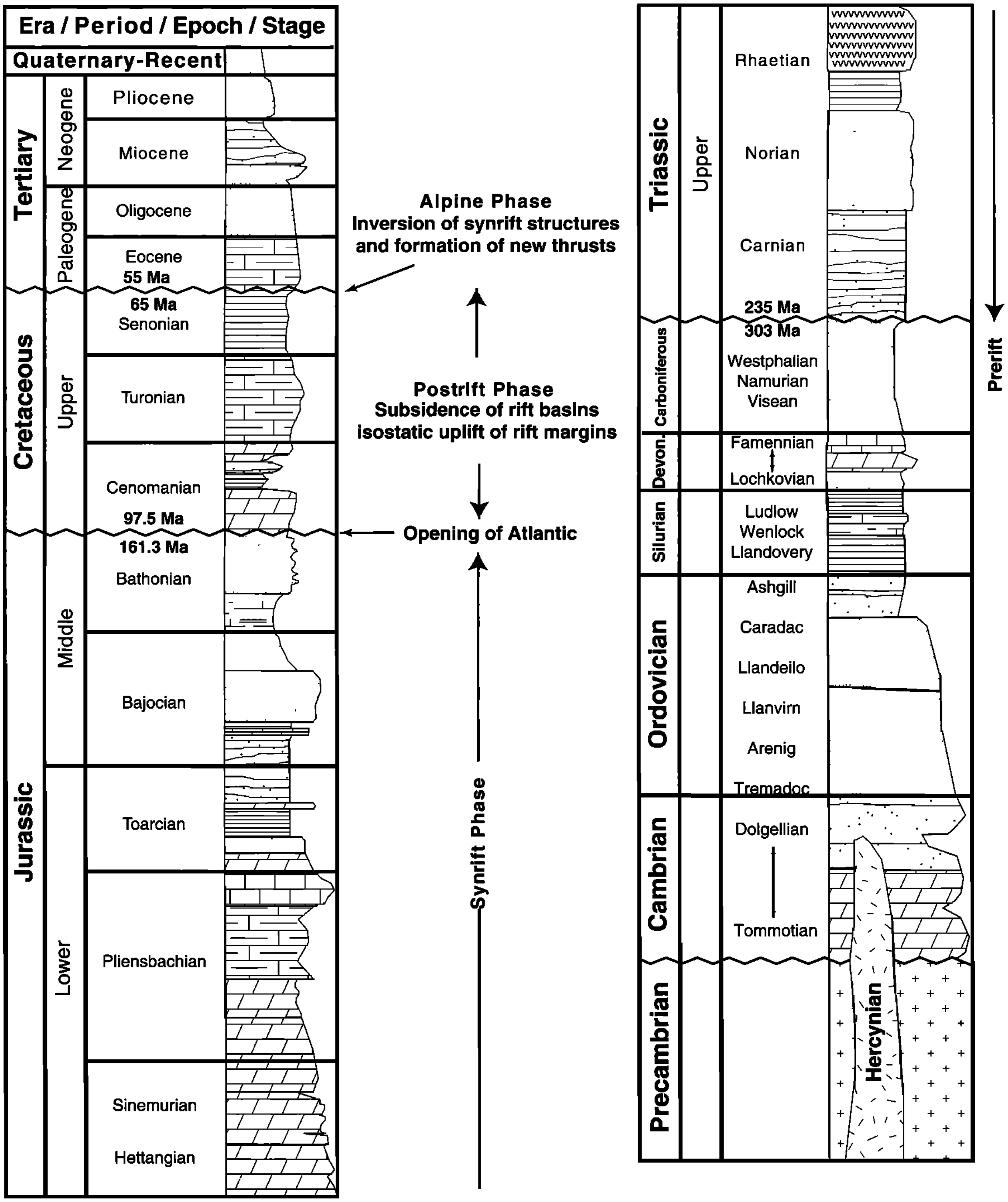

Figure 3. Generalized stratigraphic section of the central High Atlas Mountains compiled from geological maps [*Saadi et al.*, 1977, 1979, 1985; *Fetah et al.*, 1990], measured sections [*Michard*, 1976; *Rolley*, 1978; *Jabour and Nakayama*, 1988; *Ensslin*, 1992], and well data (Tadla basin DRZ-1 and KMS-1).

sedimentary rocks display complex facies distributions, varying both normal to the Atlas rift basin axis as well as along strike. Jurassic sedimentary rocks of the eastern High Atlas (Figure 1) are generally more calcareous and range from shallow water reef carbonates to deep water calcturbidites in the center of the rift basin [*Crevallo et al.*, 1987]. The central High Atlas Mountains to the west expose Middle Jurassic (Dogger) sedimentary rocks most of which are clastic. Thick sandstone and shale intervals are common [*Rolley*, 1978]. Late Jurassic sedimentary rocks are composed of thick massive limestones and dolomites (Liassic). Upper Jurassic (Oxfordian-Kimmeridgean-Tithonian)

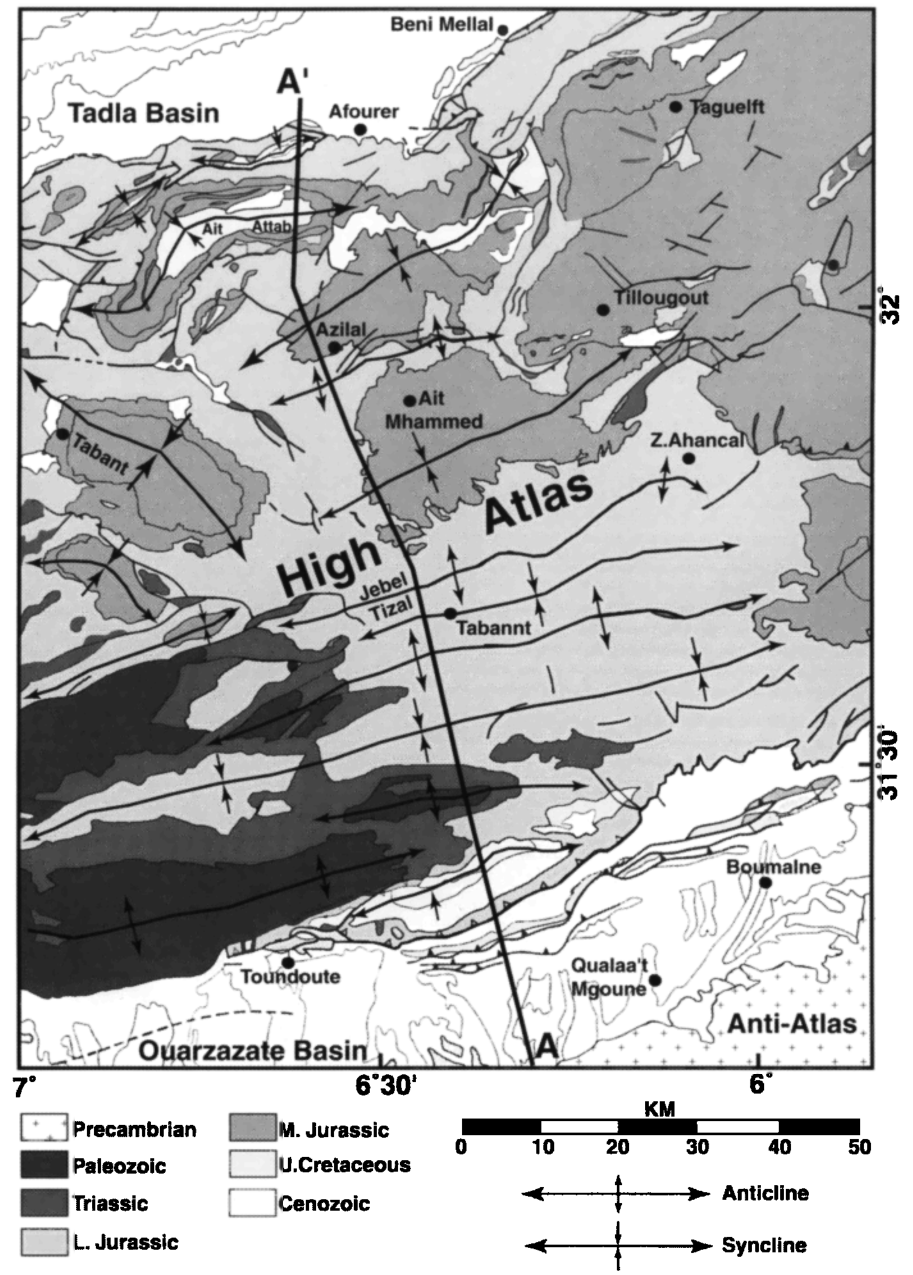

Figure 4. Generalized tectonic and geological map of the central High Atlas showing the location of transect A-A', modified after *Saadi et al.*, [1977, 1979, 1985]; *Rolley*, [1978], *Jenny* [1988] and *Fetah et al.* [1990].

through Lower Cretaceous (Berriasian-Albian) sedimentary rocks in the High Atlas Mountains and margins are not preserved, and rocks of these ages may never have been deposited (Figure 3).

The synrift Triassic and Jurassic sedimentary rocks have been interpreted by several researchers to have been deposited in a transtensional rift environment [*Crevallo et al.*, 1987; *Laville*, 1988; *Beauchamp*, 1988; *Laville and Pique*, 1991]. Extension combined with strikeslip movement has been used to explain different syntectonic sedimentary relationships such as facies changes, angular disconformities, and abrupt changes in thickness (Triassic-Late Jurassic faulting). There are also a variety of explanations for these relationships such as lateral ramps, high-relief accommodation zones, and changes in fault polarity [*Rosendahl et al.*, 1986].

2.4. Postrift Phase

Postrift sedimentary rocks of the Atlas range in age from Cenomanian to Eocene. Uplift of the Atlas Mountains began in the Cretaceous and extended into the present [*Faure-Muret and Choubert,* 1971]. Uplift of the northwestern margin of Africa during the Late Jurassic and Early Cretaceous was a regional tectonic response to the opening of the Atlantic, resulting in erosion and/or nondeposition.

Cretaceous sedimentary rocks are composed of transgressive shallow marine clastic and carbonate sedimentary rocks [*Ensslin,* 1992]. The Ouarzazate basin is the southern foreland basin of the High Atlas Mountains and attains a maximum Tertiary thickness of 1000 m [*Schmidt*, 1992]. These Tertiary rocks (Paleogene/Neogene) overlie Cretaceous rocks (Cenomanian/Turonian) composed of siltstones, fine sandstones and marls [*Schmidt*, 1992]. The Paleogene and Late Cretaceous age rocks onlap Paleozoic and Proterozoic age rocks along the southern margin of the Ouarzazate basin. Uplift in the Oligocene was related to the Alpine orogeny and was driven by the convergence of the Eurasian and African plates. This Alpine uplift was responsible for the major uplift and inversion of the High Atlas Mountains [*Brede et al.*, 1992; *Laville and Pique,* 1992]. Thrusting and inversion during the Oligocene formed the current topographic relief of the High Atlas Mountains, yet significant Tertiary foreland basins to the north and south of the High Atlas are lacking. Tertiary sediment thicknesses in the foreland regions are on the average less than 1 km. Northwest of the Ouarzazate basin, Jebel Mgoun reaches elevations of over 4000 m. The lack of development of foreland basins may indicate a high rigidity for the upper crust along the margins of the Atlas. Sediments of Tertiary age may have bypassed these basins and were deposited westward onto the Atlantic margins.

3. Structural/Geophysical Transect of the High Atlas

A geological-geophysical cross section was constructed across the central High Atlas Mountains of Morocco to study the geometry and kinematic history of the orogen. Seismic reflection data acquired by the Office National de Recherches et d'Exploitations Petrolieres (ONAREP) were reprocessed and combined with field mapping, Landsat-TM images, gravity, and magnetic data to facilitate the construction of an area-balanced cross section across the High Atlas Mountains. The location of transect A-A' (Plate 1) was chosen to parallel the direction of thrust transport (Figure 4) and to be as near as possible to the seismic reflection profiles along the northern and southern margins of the High Atlas (KT-6 in the north and OZ-5 in the south, Plate 2). Abundant examples of small-scale deformation exist along the trend of the transect and make a significant contribution to the overall shortening along the transect. Although many smaller structures were also mapped in the field along the transect, only those structures that could be shown at the scale of transect A-A' will be discussed.

Field data were collected along seismic reflection profile KT-6 using topographic maps, air photographs, Landsat-TM images, and a Global Positioning System receiver. Additional regional data were collected along dip-oriented transects across the Ait Attab syncline parallel to seismic profile KT-6 and across the High Atlas along transect A-A' (Figure 2). Geological data were also digitized from 1:100,000-scale geological maps [*Ghissassi,* 1977, *Saadi et al.*, 1979, 1985; *Fetah et al.,* 1990] and additional structural data were derived from air photographs and Landsat-TM images.

Synrift normal faults commonly can be identified by the presence of Jurassic basalts that were intruded along the fault planes, as well as by sedimentary thickening into the faults. In the field many small Mesozoic normal faults thus identified have not been later reactivated. A common characteristic of structures in the High Atlas along transect A-A' is an asymmetric style of fault-related folding. Many of the topographic ridges are directly related to structures of this nature. It is suggested that these structures were generated by fault-propagation folds that originated from underlying evaporites in the Triassic (Figure 5). These steep fold limbs are commonly faulted along strike by complex multiple fault-breakthroughs.

Gravity and magnetic data [*Makris et al.,* 1985] were used in this study in conjunction with seismic reflection data and geological data to assist in the interpretation of the first-order crustal features. Gravity data (Plate 1) displayed together with digital topographic data illustrate the asymmetry of the High Atlas Mountains (Figure 6). The regional Bouguer gravity low beneath the High Atlas and the topographically highest region of the Atlas show a correlation along the southern margin of the mountains. The gravity low extends southward into the relatively shallow Tertiary Ouarzazate basin. Given the amount of tectonic loading along the southern margin, the High Atlas Mountains appear to be uncompensated [*Makris et al.*, 1985]. The lack of a significant foreland basin [*Schmidt*, 1992] may be related to high crustal rigidity. The gravity and magnetic data correlate well to Precambrian basement rocks exposed in the Anti-Atlas south of the section (Plate 1). The Bouguer gravity data show a decrease in the asymmetrical anomaly northward along the traverse A-A' (Plate 1 and Figure 6) as does the topographic profile. Magnetic data displayed along traverse A-A' show a broad regional anomaly beneath the High Atlas Mountains of approximately 100 nT. Several smaller magnetic anomalies along the traverse are of the order of 10-15 km in wavelength and show a 20-30 nT deflection in the regional magnetic gradient (Plate 1). These magnetic anomalies may be related to gabbroic intrusives of Triassic to Lower Jurassic age.

Digital copies of seismic reflection profiles acquired for petroleum exploration by ONAREP were reprocessed for use in this study. The profiles were reprocessed from the field shot gathers for the lines north of the High Atlas Mountains (KT-6, Tadla basin), and poststack processing was carried out on the lines south of the High Atlas Mountains (OZ-5, Ouarzazate basin) (Figure 2). These seismic profiles were chosen because of their orientation with respect to the transport direction of thrusting and because they are the only seismic profiles that extend into the foothills of the High Atlas Mountains. The seismic profiles KT-6 and

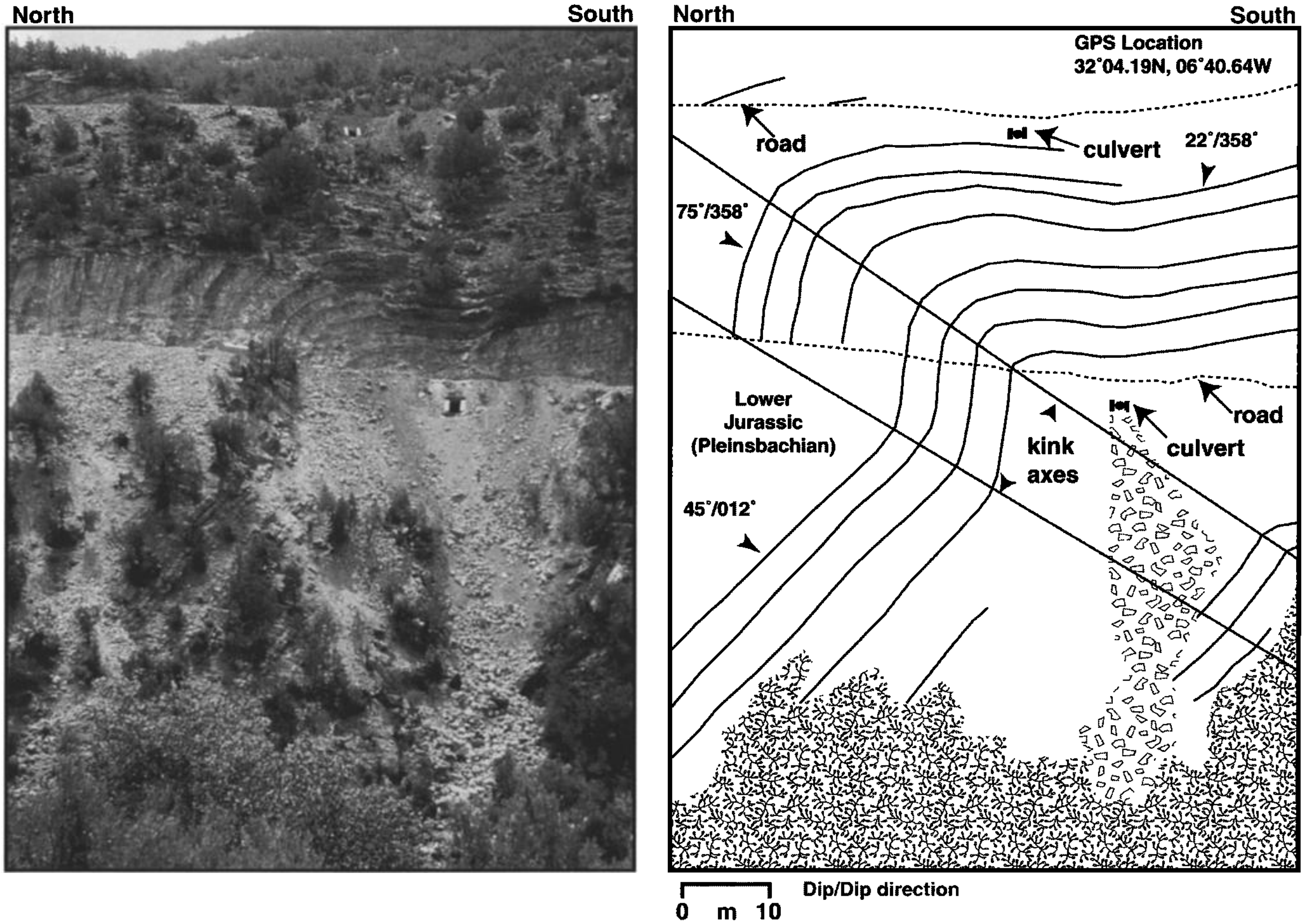

Figure 5. A photograph and corresponding geological sketch showing kink axes interpreted as a fault propagation fold at depth. These kink axes are located along the southern margin of the Ait Attab syncline and steepen the southern limb of the syncline. These kink axes are thought to result from the propagation of a newly formed fault generated in the Triassic evaporites. The location of Figure 5 is shown on Figure 7.

OZ-5 were acquired by Compagnie Génèrale de Géophysique for ONAREP using a Vibroseis source (50 m shot spacing, 96 channels, and 10-60 Hz sweep) and a split spread geometry. The profiles were migrated using a finite difference algorithm, depth converted, and displayed with no vertical exaggeration to provide an accurate display for structural interpretation and for construction of the cross section A-A' (Figures 2 and 7). Well data from the Tadla basin were used to provide stratigraphic control by creating synthetic seismograms and time-depth curves. These wells were drilled in close proximity to seismic profile KT-6 and provided stratigraphic control into Paleozoic age (Ordovician) sedimentary rocks (Figure 7).

4. Description of Structures

4.1. Northern High Atlas Mountains

Seismic profile KT-6 is oriented at about 45° to the direction of thrust transport in the northern High Atlas (Figure 7). Bedding dips seen on this line, for the most part, are apparent. Structures interpreted along profile KT-6 were corrected for true dip and then projected along strike to transect A-A' (Plate 1). The northern 20-25 km of profile KT-6 was acquired in the Tadla basin north of the High Atlas. Both wells in the Tadla basin drilled through the Tertiary, Upper Cretaceous and Triassic and into the Ordovician without encountering Lower Cretaceous or Jurassic rocks (Plate 2). Reflections correlative to the stratigraphic intervals encountered in the two wells clearly extended to within less than 1 km of the exposure of massive Lower Jurassic carbonates in the foothills of the High Atlas. No clear evidence of a décollement north of the exposed Lower Jurassic carbonates was found on profile KT-6.

The style of deformation on the north side of the High Atlas is illustrated by spectacular field relations on the northern limb of the Ait Attab syncline (Figures 7 and 8). A low-angle thrust fault dips 15° toward 210° (SSW). Massive Early Jurassic carbonates and Triassic clastic rocks in

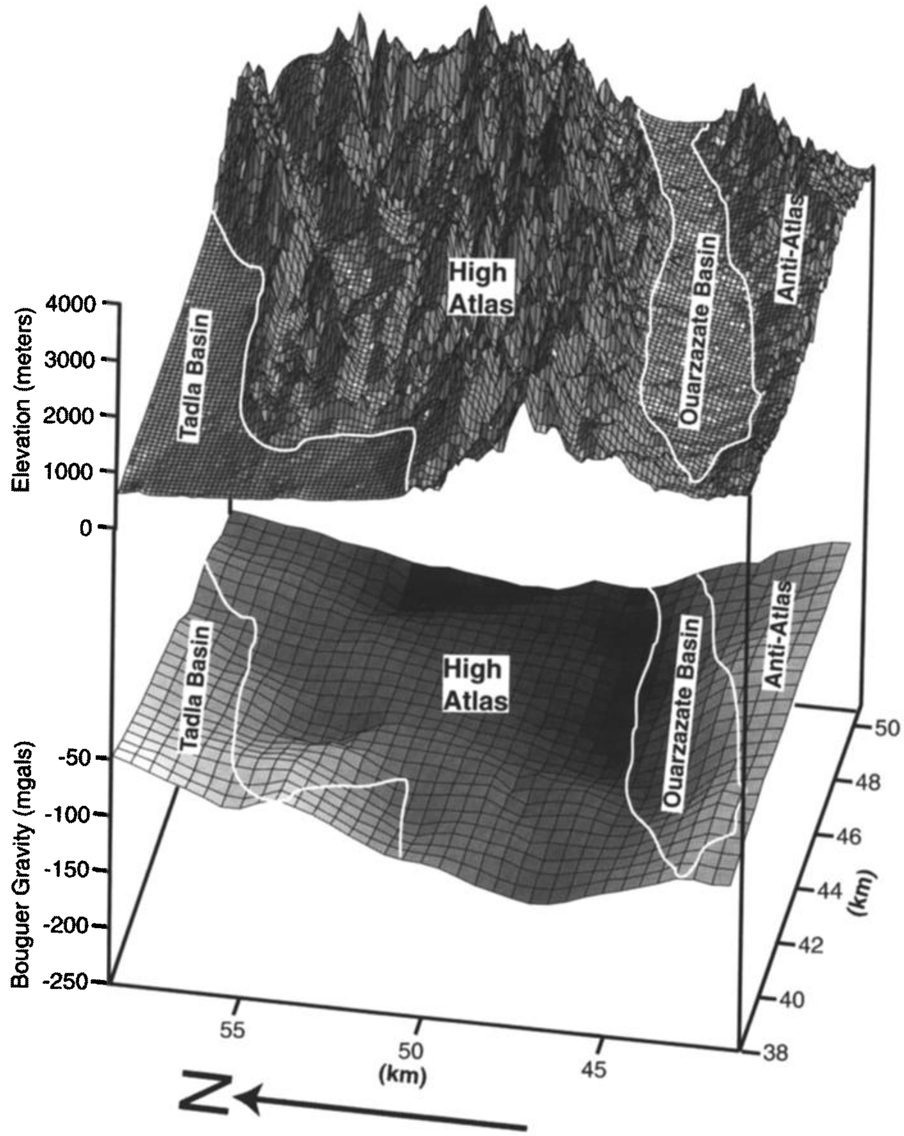

Figure 6. Perspective view of the High Atlas showing digital topography and Bouguer gravity data. The highest topography is along the southern margin of the High Atlas where gravity data indicate a regional gravity low. The lack of significant foreland basins may indicate that the High Atlas are uncompensated. The area shown is the area of study on Figure 1.

the hanging wall flat of this fault are thrust over Late Cretaceous red shales and sandstones that are Cenomanian in age (Figure 8). The thrust zone is composed of breccia several meters in thickness. This breccia is made up of angular fragments of the overlying massive dolomites and limestones that range up to 0.5 m in size. The breccia is poorly cemented except for the recent deposition of travertine. It was difficult to find kinematic indicators within the zones of fault breccia, but lineaments oriented downdip of the fault zone were common on bedding planes above the fault zone. The hanging wall of this major fault is deformed locally by fault-propagation folds that step up from the underlying thrust

The Ait Attab syncline (Figure 7) is one of the largest and most conspicuous structures along the trace of transect A-A' and is well imaged on line KT-6. Surface structural dips are in good agreement with those extracted from the seismic data (Figure 9). Thickening of the Middle and Lower Jurassic rocks across the syncline from southeast to northwest indicates that the synrift rocks were deposited in the hanging wall of an active normal fault that dipped southward into the paleo-Atlas rift basin. The folded rocks lie above high-amplitude reflectors that dip more gently than the overlying exposed rocks in the Ait Attab syncline (Figure 9). The lower reflections are interpreted as being the footwall flat of the major fault, described above, along the

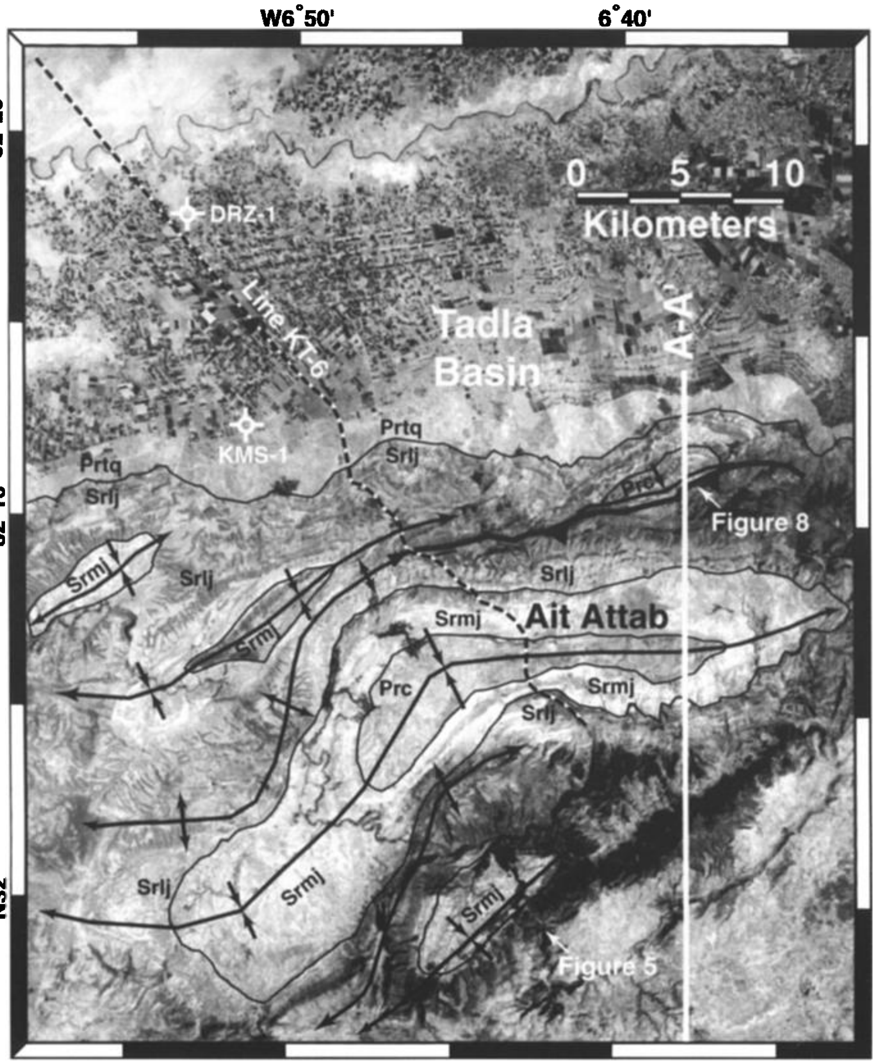

Figure 7. Simplified geological map of the Ait Attab syncline along the northern margin of the High Atlas Mountains. Shown are the locations of seismic line KT-6, wells DRZ-1 and KMS-1 and transect A-A'. Fold axes, fault traces and stratigraphic contacts are overlain on a Landsat-TM image (near infrared band 5).

northern margin of the Ait Attab syncline (Figure 7). This broad, open syncline has a wavelength of ~12 km and is interpreted to have formed between two fault-bend anticlines (Figure 9 and Plate 1). Locally, the geometry of the limbs has been complicated by small fault-propagation folds (Figure 5 and Plate 2). In three dimensions the syncline displays a classic type-2 fold interference pattern [*Ramsey and Huber*, 1987]. The first of the two fold axial surfaces strikes NE to east and the second fold axial surface strikes NW and is near vertical. It is believed these two phases of folding are the same age and were generated by one phase of deformation by the thrusting of rift sedimentary rocks northwestward over a lateral ramp.

4.2. Central High Atlas Mountains

Shortening in the central High Atlas Mountains is insufficient to explain the topography found in this part of the orogen and therefore presents an interesting problem. The small amount of shortening expressed in this region indicates that the topography of the area must be related to shortening at depth. We interpret many of the structures present in the central area of the mountain belt (Jebel Waougoulzat, Jebel Tizal, and Azilal) were formed by fault-propagation folds that ramped up from an inferred décollement in Triassic age rocks (Plate 1). The central region of the High Atlas known as Ait Mohammed is bounded by two

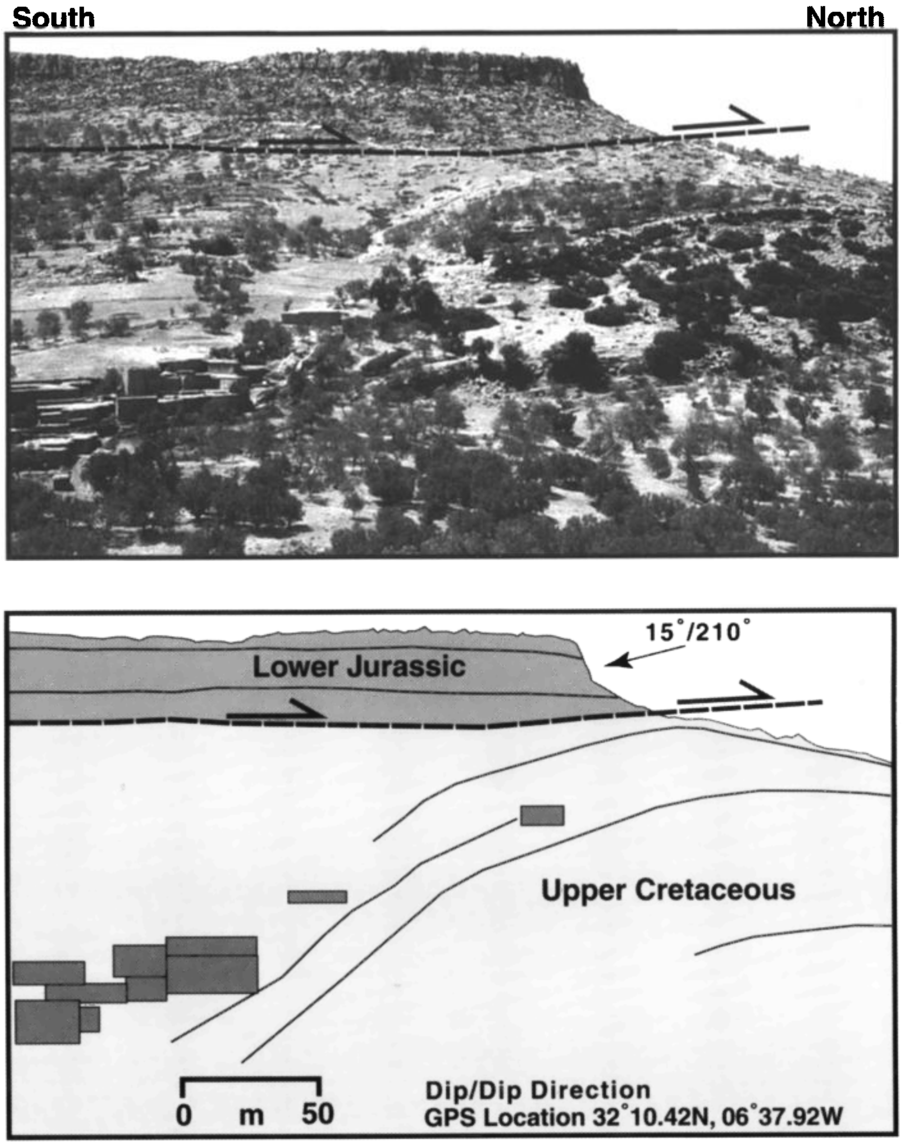

Figure 8. A photograph showing a field location along the northern margin of the Ait Attab syncline (fault is shown on Figure 7). This low-angle fault thrusts Lower Jurassic (Hettangian/Pleinsbachian) massive dolomites and limestones over Upper Cretaceous (Cenomanian/Turonian) rocks. Along strike this fault thrusts Lower Jurassic rocks in the hanging wall over Tertiary rocks in the footwall.

of these large-scale structures. The gently dipping syncline of Ait Mohammed is approximately 20 km wide. Beneath Ait Mohammed a fault duplex system was interpreted to lie in the Paleozoic part of the section so as to generate the amount of uplift present above the décollement (~12 km thickness). Without this duplex system there is insufficient sedimentary thickness to balance the area between the décollement and the surface. The duplex geometry in the Paleozoic shown in Plate 1 probably does not represent the true geometry of the deformed beds in the Paleozoic rocks, but some type of internal shortening beneath the Triassic and above the décollement is necessary to create the large relatively undeformed region of Ait Mohammed above the décollement. Other possible explanations of the uplift in the central region of the High Atlas include high-angle planar faulting, thermal uplift, or a previously existing Paleozoic or Proterozoic crustal root.

4.3. Southern High Atlas Mountains

Three seismic lines from the Ouarzazate basin were supplied by ONAREP for this study. Line OZ-5 also extends completely across the Ouarzazate basin from the exposed Precambrian of the Anti-Atlas into the foothills of the southern High Atlas Mountains (Figure 10). The southern end of line OZ-5 shows a high-amplitude reflector that dips

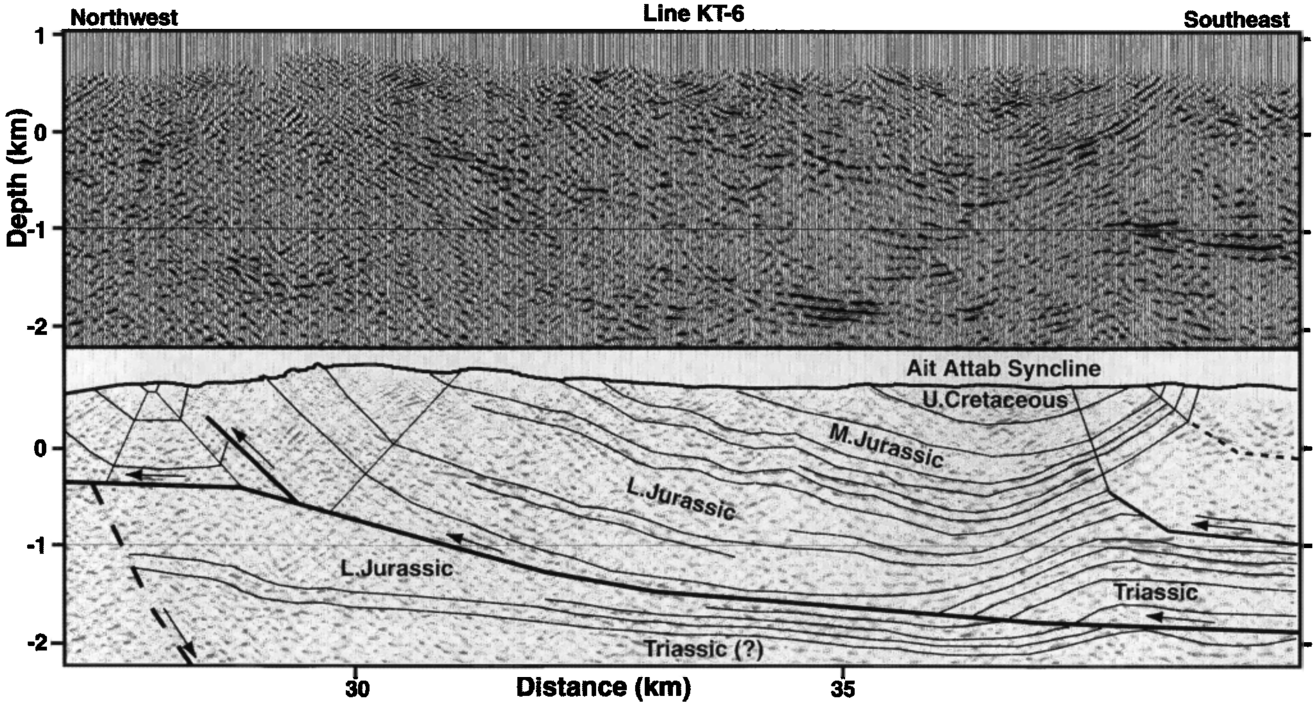

Figure 9. A detailed portion of seismic line KT-6 across the Ait Attab syncline (see Figures 2 and 7 for location). This line illustrates the folded rocks of the Ait Attab syncline in the hanging wall of a major thrust fault that verges to the northwest. The southern limb of the Ait Attab syncline is further steepened by fault propagation folds (Figure 5). Bedding above the thrust fault is not parallel to the fault along the southeast end of the line (previously deformed), which may be synrift rocks inverted along a reactivated normal fault.

15° northward (Figure 11). This reflector is interpreted as the top of basement, and above it reflectors are truncated beneath the Upper Cretaceous unconformity. Along the northern margin of the Anti-Atlas Mountains in the Ouarzazate basin there are no exposures of Triassic, Jurassic, or Lower Cretaceous rocks. The only synrift rocks in the Ouarzazate basin are Triassic and Jurassic rocks that have been thrust southward out of the rift basin during inversion in the Oligocene. The thrust faults along the southern margin of the High Atlas can be easily identified on Landsat-TM images (Figure 10). Fault-bend folds deformed the postrift rocks of the Ouarzazate basin during uplift and inversion of the Atlas rift basin. On the basis of the interpretation of line OZ-5b (Figure 12) we suggest that the base of the Upper Cretaceous is the principal décollement in the basin. The dips of the rocks and faults imaged on this line are apparent, as the seismic line is oblique to the direction of transport, necessitating correction prior to projection westward onto transect A-A'. At least two thrust faults repeat postrift rocks to the surface in the Ouarzazate basin. Because Neogene rocks unconformably overlie the hanging wall of fault A (Figure 12) but are cut by fault B, we infer an out-of-sequence progression of faulting. Further evidence for out-of-sequence thrusting is the fact that faults closer to the High Atlas Mountains are not folded by faults to the south in the Ouarzazate basin (Plate 2).

Northward along line OZ-5b the décollement at the base of the Upper Cretaceous is cut by a younger thrust that places synrift sedimentary rocks at outcrop over postrift Upper Cretaceous and Paleogene rocks (Figure 13 and Plate 2). The rocks in the footwall of this thrust were originally deposited along the platform margin of the Atlas rift (no synrift age rocks). The synrift Triassic and Lower Jurassic rocks were thrust in the hanging wall of a thrust fault from within the Atlas rift basin and transported onto the platform margin of the rift basin. It is important to be able to differentiate between a reactivated normal fault and a newly formed thrust. The northernmost of the faults along line OZ-5b is interpreted as a reactivated fault; the bedding in the hanging wall shows noticeable thickening within the Jurassic over a short distance in the hanging wall of the thrust, possibly related to synrift deposition during extensional faulting (e.g., growth fault). The rocks in the hanging wall of this reactivated fault were previously deformed during the synrift extensional phase (hanging wall rollover) of deformation and were transported up the same normal fault until a new thrust fault was formed transporting synrift rocks over the platform margin of the rift (Plates 1 and 2).

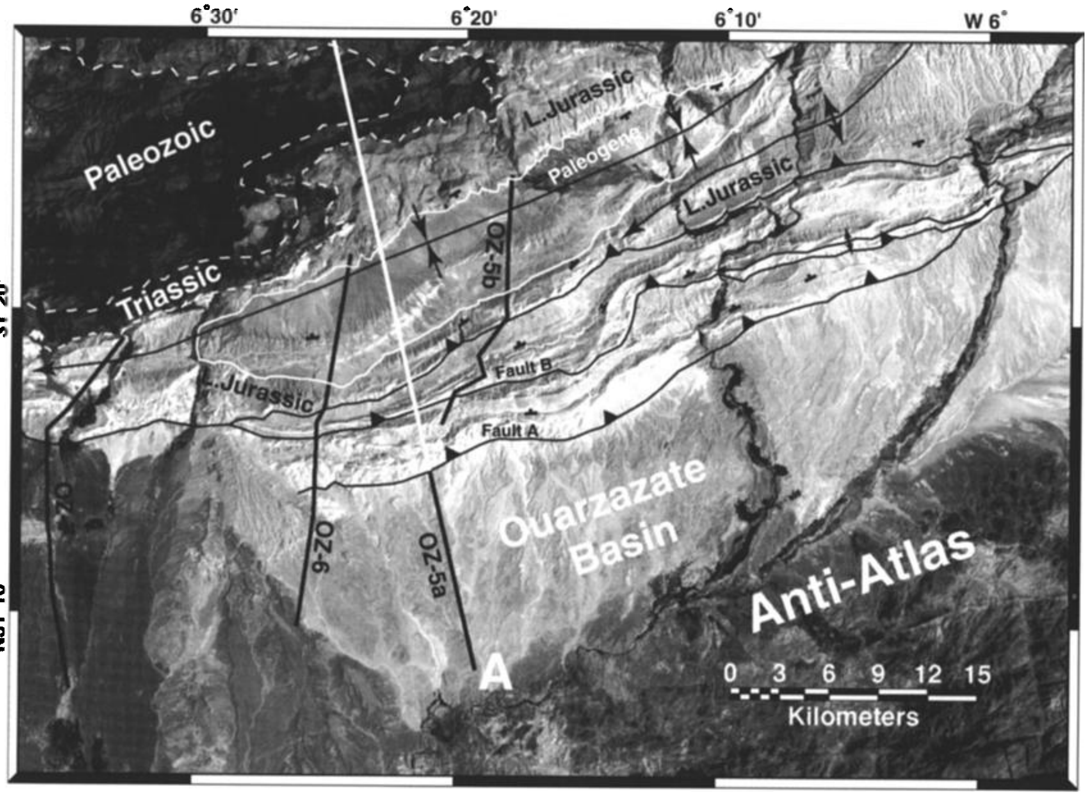

Figure 10. Tectonic map along the southern margin of the central High Atlas Mountains, Ouarzazate basin and the Anti-Atlas. Seismic reflection profiles used in the construction of transect A-A' are shown. Line OZ-5a is parallel to the transport direction of thrusting, and extends to the south near the exposure of Precambrian age rocks of the Anti-Atlas. The surface expression of thrust faults is evident on the Landsat-TM image (near infrared band 5). The darker rocks in the northwest corner of the image are Triassic and Paleozoic rocks outcropping in the axis of the large ramp anticline of Jebel Mgoun (>4000 m) that has been uplifted by the reactivation of synrift faults. The Paleozoic rocks in the core of this anticline have been intensely deformed by what is interpreted to be a series of fault duplexes (Plate 1).

5. Construction of a Geological-Geophysical Transect

This study set forth to prepare a transect across the High Atlas Mountains (A-A') that would use as much of the data available as possible and would for the first time generate a cross section across the High Atlas Mountains that is balanced and restorable. Such a transect would help to shed new light on mountain-building processes in an intraplate setting. The construction of a transect across the High Atlas Mountains was achieved by hand using several techniques that involved area balancing, line length balancing, and dip domains [*Woodward et al.*, 1989]. Kinematic methods of construction included the fault-bend/fault-propagation fold technique [*Suppe*, 1983; *Suppe and Medwedeff*, 1990] and vertical simple shear [*Gibbs*, 1983]. Vertical simple shear was used to model the reactivation of synrift normal faults, while new thrusts were assumed to have deformed by layer-parallel shear.

A number of fundamental assumptions were required in the construction of transect A-A', such as deformation by plane strain with no movement out of the plane of the section, conserved area between the deformed and restored section (area balanced), and line lengths remaining the same before and after shortening at the scale of the section [*Woodward et al.*, 1989]. Refolded folds similar to those in the region of Ait Attab (Figure 7) are common throughout the High Atlas and violate the two-dimensional assumptions inherent in cross-section balancing. However, because our transect is nearly parallel to the second fold axial surface, the effect of the refolding (resulting from extensional lateral ramps) is minimal. A depth to detachment calculation was made (approximately 12 km depth) based upon a composite thickness from well data and from measured thicknesses in the High Atlas and Anti-Atlas. This depth to detachment calculation is in agreement with a proposed detachment level based on previous geophysical studies (10-15 km)

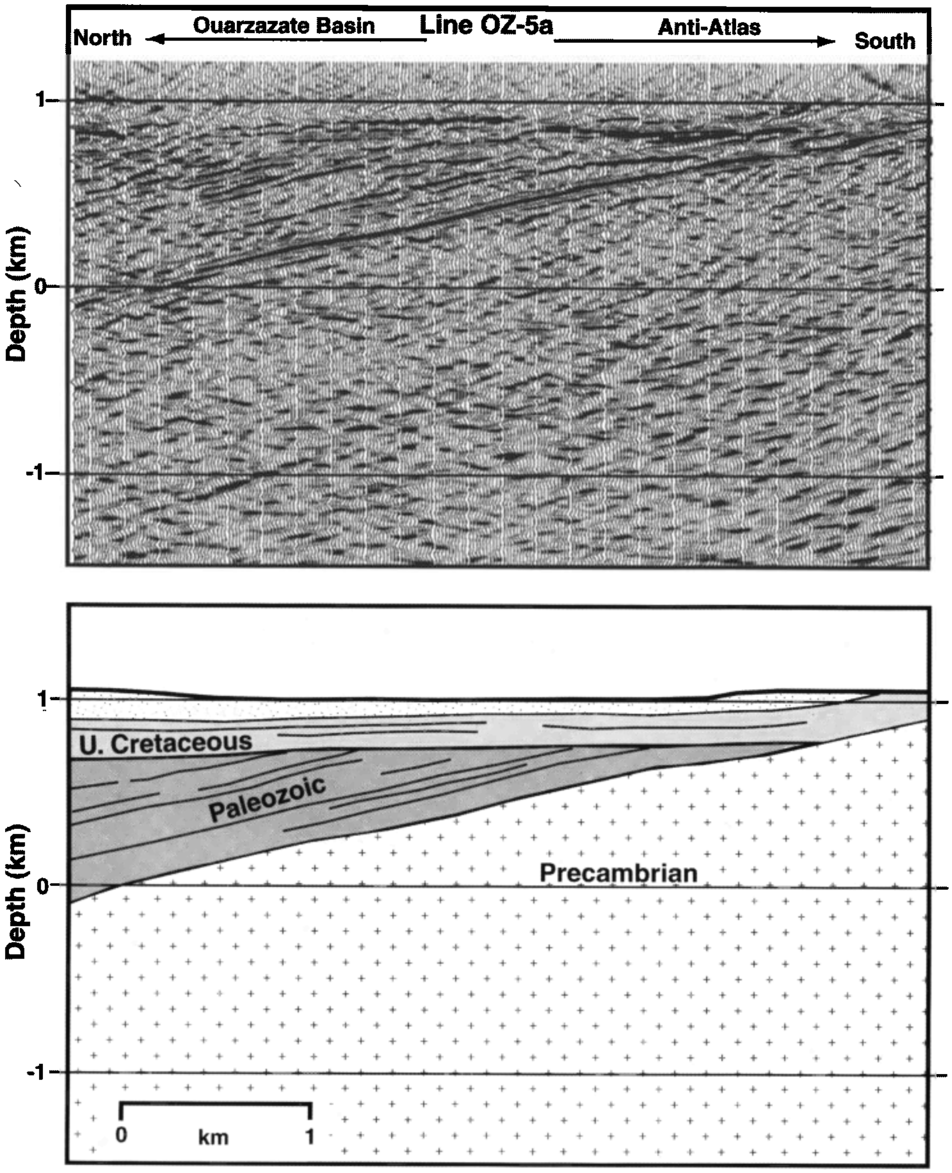

Figure 11. Seismic line OZ-5a located along the southern margin of the Ouarzazate basin and the Anti-Atlas (see Figure 10 for location). The high-amplitude reflection that dips 15° from south to north across the section is interpreted as Precambrian basement. The reflectors above the basement are interpreted as Paleozoic rocks that are truncated by the Upper Cretaceous unconformity.

[*Makris et al.*, 1985; *Wigger et al.*, 1992]. An estimate for the depth to basement (Precambrian) and dip of basement along the southern margin of the High Atlas Mountains was possible based on a seismic reflection tie to basement (Figure 11). The postrift Late Cretaceous age rocks were assumed not to have thickened significantly into the center of the Atlas rift basin during subsidence, and measured thicknesses from the margins of the High Atlas Mountains were assumed to be regionally consistent. Section A-A' was then restored to the base of the Late Cretaceous (Plate 1). The construction of the transect also assumed there was no preexisting Paleozoic or Proterozoic basin or crustal root present beneath the High Atlas. There is no indication of abnormally high heat flow beneath the High Atlas that might suggest recent thermal uplift of the Atlas Mountains. Bouguer gravity data modeled by *Makris et al.* [1985] suggest that there are no abnormal density anomalies beneath the High Atlas that in turn might suggest possible delamination based upon gravitational instability of the crustal root [*Seber et al.*, 1996].

Estimates of shortening attained from this study based upon transect A-A' (Plate1) are a minimum estimate of the

A Balanced Geological-GeophysicalTransect
of the High Atlas Mountains, Morocco

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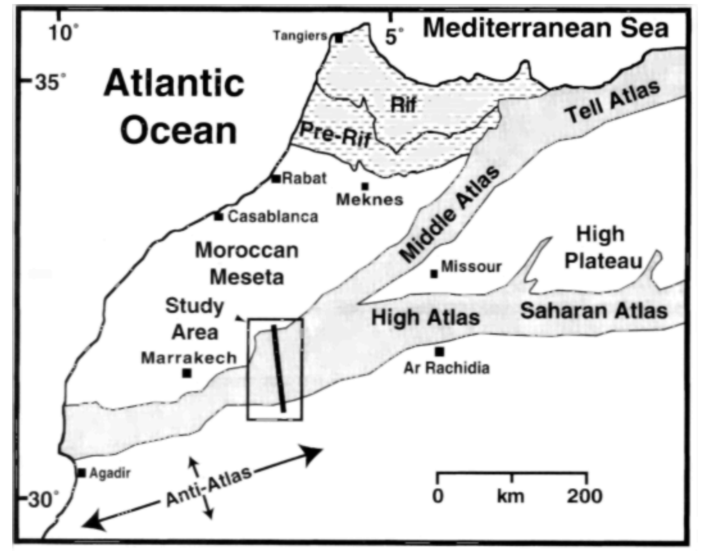

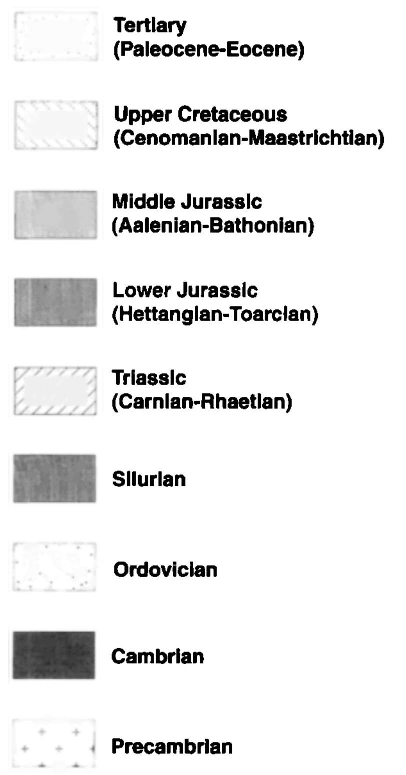

*No vertical Exaggeration

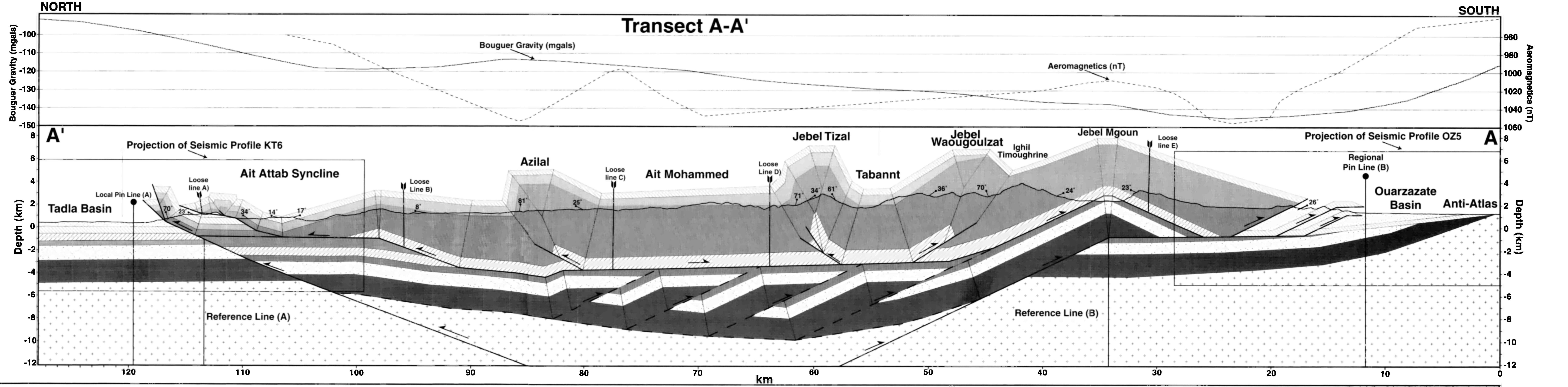

Plate 1. *Transect A-A' across the High Atlas Mountains, Morocco. Transect A-A' is an area and line length balanced cross section restored to the base of the postrift Early Cretaceous time. Local (A) and regional (B) pin lines indicate shortening of 36 km from the restored to the deformed section. Reference lines (A and B) represent the original boundaries of the Atlas rift and show a minimum shortening of 36 km. Loose lines (A-E) illustrate the partitioning of strain across the High Atlas Mountains. Loose line (A) has been transported over reference line (A) along the northern end of transect A-A'. Loose line (E) was transported southward over reference line (B). By comparison, loose lines (C) and (D) show almost no shortening inferring that the greatest magnitude of shortening is near the margins of the High Atlas Mountains This information shows that shortening across the Atlas Mountains is being achieved at shallow crustal levels along the margins and at middle to lower crustal levels in the interior of the mountain belt*

Plate 2. *Seismic reflection profiles KT-6 and OZ-5 and interpretations. Seismic reflection data were reprocessed and combined with geological data to provide subsurface control for at least 40 km of transect A-A'.*

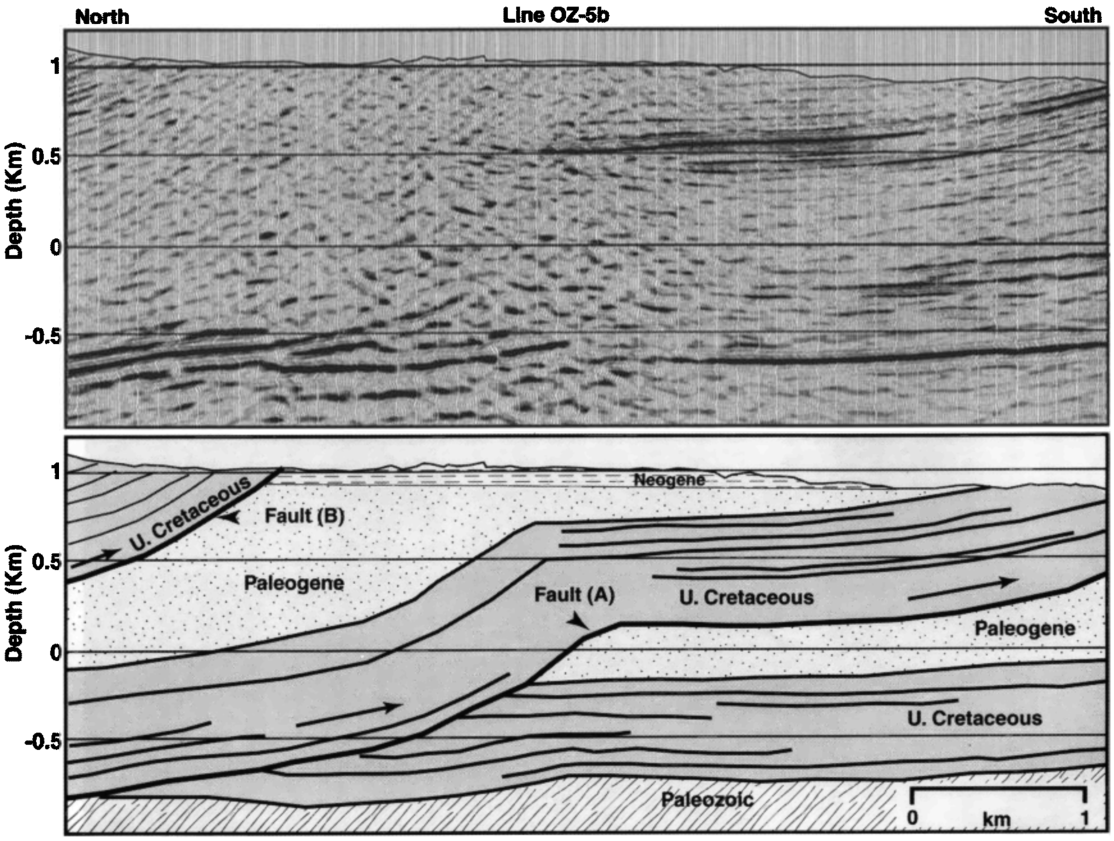

Figure 12. Seismic line OZ-5b located along the southern margin of the High Atlas Mountains (see Figure 10 for location). This seismic line was shot across several exposed thrust faults that repeat the Upper Cretaceous in the hanging walls of thrusts faults developed during the Oligocene. Fault A is believed to have occurred first with fault B occurring out of sequence, as fault B has not been folded by fault A. An out-of-sequence progression of faulting can be inferred from Neogene rocks that unconformably overlie the hanging wall of fault A and are cut by fault B.

amount of shortening across the High Atlas. A regional pin line was placed outside the orogen in the Tadla basin to the north, and a regional pin line was placed in the Ouarzazate basin south of the High Atlas Mountains. On the basis of a depth of décollement calculation of 12 km a template was created for the footwall of the original Atlas rift system (Plate 1). The depth of basement in the footwall of the template (deformed and restored sections) in (Plate 1), was derived from seismic reflection data (Figure 11). Whereas the original rift basin was composed of many normal faults with various displacements, vergence, and magnitudes, only the main rift-bounding faults are considered for reactivation along transect A-A'. Synrift sediments deformed in the hanging wall of the rift basin were transported up the southern rift bounding fault (Figure 14 and Plate 1). Thrusts with a fault-bend-fold geometry are north verging (Figures 8 and 9) along the northern margin of the High Atlas, ramping upward from evaporites in the Triassic (Plate 1).

6. Discussion

6.1. Historical Evolution

The Atlas mountain belt is the largest Phanerozoic mountain belt in Africa and is similar in size and extent to the Appalachians, Urals, and Zagros mountains. Convergent mountain belts such as the Andes and Himalayas have been studied extensively in relation to their plate tectonic origin, yet intracontinental mountain belts such as the Atlas have received comparatively less attention. Convergence of continental plates or continental and oceanic plates result in orogens near the plate margins. The convergence of the Iberian and African plates during the Alpine orogeny [*Ziegler*, 1992] transferred stresses 200-300 km through the crust to an intraplate region where strain was accommodated along crustal weaknesses formed by rifting in the Mesozoic and also in earlier Paleozoic orogenic events.

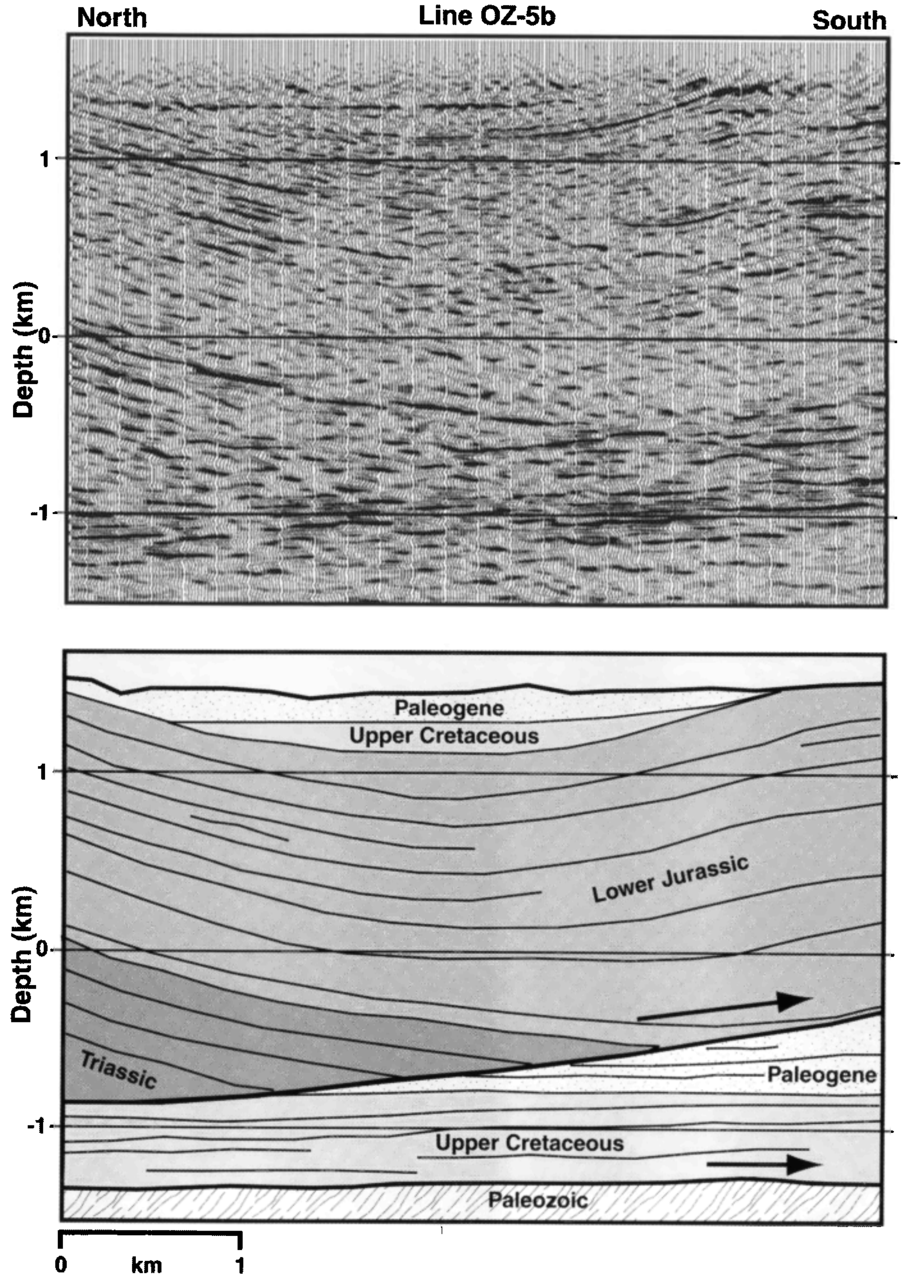

Figure 13. Seismic line OZ-5b located to the north of Figure 12 shows what is interpreted as synrift sedimentary rocks transported in the hanging wall of a new thrust (see Figure 10 for location). These synrift strata are thrust out of sequence over an earlier thrust formed between the Upper Cretaceous and Paleozoic.

The evolution of the Atlas Mountains and the resulting structural styles were predetermined by conditions established in earlier tectonic events. Triassic to Jurassic age basaltic intrusives found in the Central High Atlas Mountains yield ages of 210-182 Ma and are of a mid-ocean ridge basalt-like composition, indicating a subcontinental to suboceanic composition, with crustal contamination [*Fiechtner et al.*, 1992]. The presence of volcanic rocks intercalated within the rift strata of the Atlas Mountains indicates that crustal thinning was accomplished by a pure shear or a similar rifting mechanism (Figure 15a).

The postrift phase of the Atlas rift basin coincided with the opening of the central Atlantic Ocean (Figure 15b). During the Late Jurassic and Early Cretaceous the north-

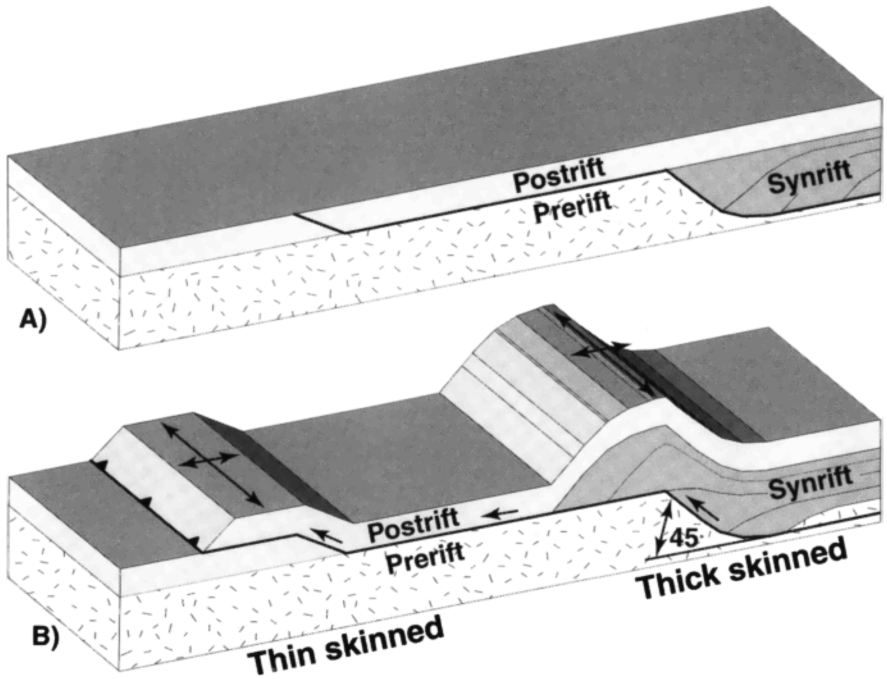

Figure 14. The tectonic evolution of the High Atlas Mountains involved newly formed (a) fault-bend folds and (b) fault-propagation folds combined with reactivated synrift faults and new thrust faults. The geometry and amount of displacement of the original normal fault control the geometry and structural style of the resulting ramp anticline. Reactivated rift faults will transport previously deformed rocks upward until a new thrust or footwall shortcut fault is formed. Synrift rocks transported in the hanging wall of these thrusts will not be parallel to the fault plane because of previous deformation. The newly formed thrusts and footwall shortcut faults are mechanically more efficient. The direction of transport is to the south (Ouarzazate basin).

western margin of the African plate was uplifted because of thermal expansion and rifting in the Atlantic. Subsidence along the Atlantic margin was amplified during the Late Jurassic to Early Cretaceous with continued rifting and spreading. Though sedimentary rocks of the Late Jurassic and Early Cretaceous are preserved offshore along the Moroccan passive margin, few sedimentary rocks of these ages (161-97 Ma) are preserved in the Atlas Mountains and margins of onshore Morocco (Figure 3).

6.2. Crustal Distribution of Shortening

Structural inversion occurs when rift faults reverse their sense of motion during subsequent episodes of compressional tectonics. Features originally generated by extension, such as half grabens, are uplifted to form positive anticlinal structures (Figures 14a and 14b). Structural relationships commonly found in rift basins such as low-and high-relief accommodation zones, pull-apart basins, extensional transfer zones and changes in fault polarity commonly produce complex three-dimensional structural geometries.

The results from transect A-A' indicate a shortening of 36 km between the local and regional pin lines (140-105 km) based on a restoration of the deformed section (Plate 1 and Figure 15). The greatest shortening (10-14 km) is confined to a narrow region (~10-15 km) along the margins of the High Atlas mountains (Figure 15c). The central High Atlas Mountains display far less shortening in the shallow crust (synrift/postrift strata) with ~12 km of shortening across ~60 km of transect A-A'. These conclusions indicate a partitioning of strain across the High Atlas Mountains that may involve the transfer of shortening from the margins at shallow depths to deeper in the middle to lower crust in the central region of the orogen. Topography in the High Atlas is lowest where there is the greatest amount of shortening (margins), while the greatest topographic relief is supported by apparent shortening in the middle-lower crust. The transfer of shortening and the accommodation of strain to the middle-lower crust in the central High Atlas may be manifested by the distribution of seismicity. Seismicity beneath the High Atlas Mountains occurs to considerable depths in the crust (< 30 km) [*Tadili and Ramdani*, 1983].

Transect A-A' was constructed using only regional structural features (e.g., Ait Attab syncline and Jebel Mgoun), and fault reactivation was only assumed along major rift-bounding faults. It has been determined that smaller faults and fault-related folds may account for an ad-

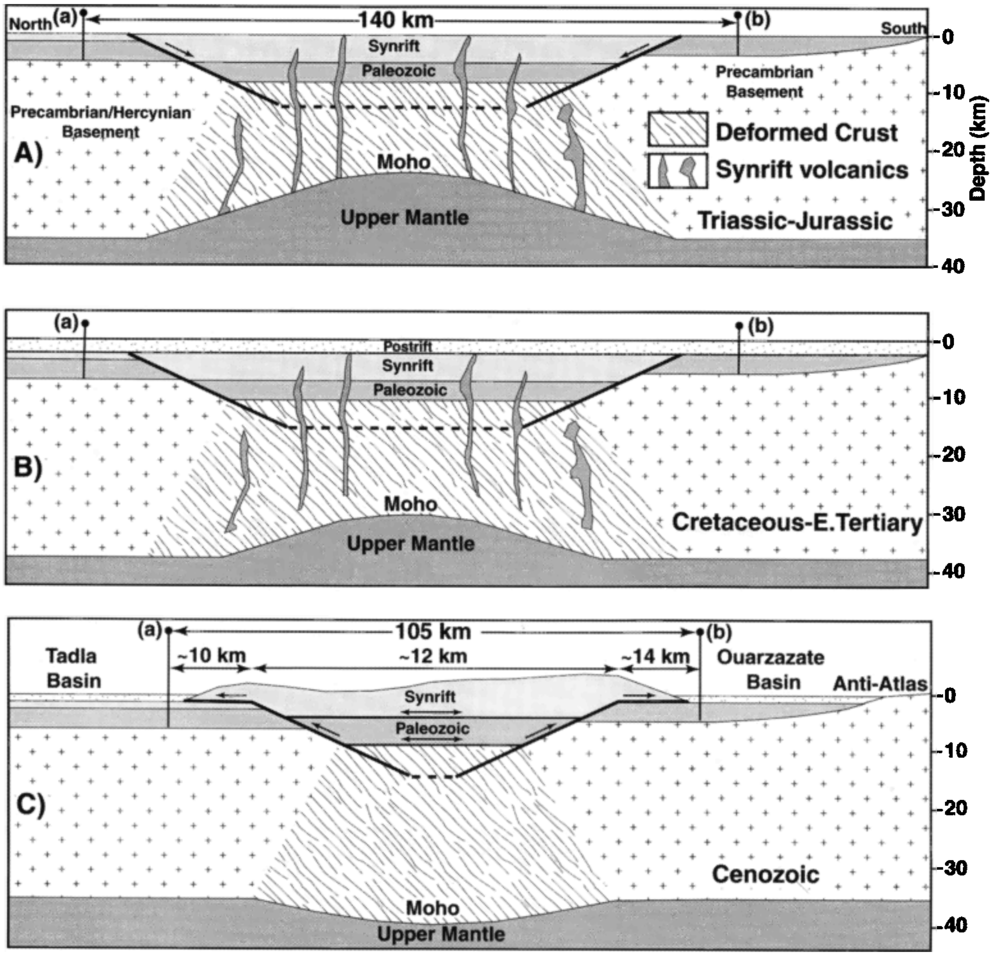

Figure 15. Schematic cross sections showing the tectonic history of (a) the Atlas synrift phase, (b) the postrift phase, and (c) the final uplift and inversion of the Atlas rift system to form the present-day Atlas Mountains. Two regional pin lines are shown which are equivalent to those in Plate 1 (line a) and (line b). The distance between these pin lines after restoration is approximately 140 km (Figure 15a), which yields the original width of the Atlas rift system. Subsidence during the Late Jurassic to Tertiary is also shown (Figure 15b). Convergence between the African and Iberian plates in the Tertiary (Miocene-Oligocene) resulted in the inversion of the Atlas rift basin by bivergent thrusting along the rift margins at shallow crustal levels over short distances (~10-14 km). Significantly less shortening in the interior of the mountain belt indicates shortening is being achieved at middle to lower crustal levels (Figure 15c).

ditional 25-40% of deformation in extensional terraines and that these same estimates may apply to shortening in compressional terraines [*Marrett and Allmendinger*, 1992]. Therefore estimates for the amount of shortening across the High Atlas based upon area balancing (Transect A-A') may be a minimum estimate, and smaller faults may yield a greater amount of shortening.

While thin-skinned deformation along newly formed thrusts occurred in the postrift sequence of the Moroccan Atlas, additional deformation by the reactivation of synrift normal faults (thick skinned) has resulted in a "hybrid" thick-and thin-skinned style of deformation (Figure 14). Faulting along the southern margin of the High Atlas Mountains and northern Ouarzazate basin (Figures 10 and 12) involves thin-skinned faults similar to those reported from the Algerian Saharan Atlas Mountains [*Outtani et al.*, 1995]. Thick-skinned deformation inverts synrift strata along pre-existing faults, often with continued transport of strata along newly formed thin-skinned faults toward the exterior of the basin (Figures 14 and 15c). The lower-angle thrusts allow for larger displacements and greater shortening than the reactivated synrift faults. Shortening is concentrated where the synrift strata are extruded over the margins of the rift and onto the adjacent platform. If reactivation of a listric synrift fault follows the upward steepening fault, a high-angle fault breakthrough may occur to accommodate shortening [*McClay and Buchanan*, 1992]. It is possible that larger normal faults at the margins of the rift have been, in part, bypassed by footwall shortcut faults [*Coward et al.*, 1991]. These newly formed thrusts may transport the upper segment of the original normal fault along the hanging wall ramp and flat without reactivation of the upper segment of the synrift normal fault [*Beauchamp et al.*, 1996].

7. Conclusions

The inversion phase of the Atlas rift system began in the Cretaceous and extended into the present. Regional Late Jurassic and Early Cretaceous uplift was in response to the opening of the Atlantic, resulting in erosion and/or nondeposition. The major uplift phase in the Atlas Mountains occurred between the Oligocene and Miocene (30-20 Ma). Synrift and postrift sedimentary rocks were shortened by the reactivation of synrift normal faults, with further shortening along newly formed thin-skinned thrust faults. These synrift rocks exhibit significant changes in thickness over a short distance, and these deformed rocks display different geometries from newly formed fault-bend and fault-propagation style folds. Transect A-A' across the High Atlas Mountains suggests shortening across the orogen (36 km) was achieved by thrusting along detachments at several levels in the upper crust (e.g., Upper Cretaceous, Triassic, and Silurian). Thrusting across the Atlas rift basin resulted in a partitioning of strain, with the greatest magnitude of shortening occurring along the rift margins and very little shortening occurring in the interior of the mountain belt. Thrusts in the High Atlas Mountains dip to the south along the northern margin (Tadla basin) and dip to the north along the southern margin (Ouarzazate basin). This bivergence may have been influenced by the original rift configuration. The minimum shortening in the central part of the topography of the High Atlas Mountains suggests significant middle-lower crust shortening.

The objective of this study was to develop a better understanding of how intracontinental mountain belts evolve. The results of this research clearly demonstrate the remarkable difference between the evolution and present structural architecture of this intracontinental mountain belt in comparison to those mountain belts associated with plate boundary processes.

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References

Bally, A.W., Tectogenèse et sismique réflexion, *Bull. Soc. Gèol. Fr.*, *7*, 279-285, 1984.

Beauchamp, J., Triassic sedimentation and rifting in the High Atlas (Morocco), in *Triassic-Jurassic Rifting*, edited by W. Manspeizer, pp. 477-497, Elsevier, New York, 1988.

Beauchamp, W., M. Barazangi, A. Demnati and M. El Alji, Intracontinental rifting and inversion: the Missour basin and Atlas Mountains of Morocco, *AAPG Bull.*, *80*, 1459-1482, 1996.

Brede, R., Structural aspects of the Middle and the High Atlas (Morocco), phenomena and causalities, *Geol. Rundsch.*, *81*, 171-184, 1992.

Brede R., M. Hauptmann, and H. Herbig, Plate tectonics and intracratonic mountain ranges in Morocco-The Mesozoic-Cenozoic development of the Central High Atlas and the Middle Atlas, *Geol. Rundsch.*, *81*, 127-141, 1992.

Broughton, P., and A. Trepanier, Hydrocarbon generation in the Essaouira basin of Western Morocco, *AAPG Bull.*, *77*, 991-1015, 1993.

Colletta B., F. Roure, B. De Toni, D. Loureiro, and H. Passalacqua, Tectonic inheritance, crustal architecture, and contrasting structural styles in the Venezuelan Andes, *Tectonics*, *16*, 777-794, 1997.

Coward, M.P., R. Gillcrist and B. Trudgill, Extensional structures and their tectonic inversion in the Western Alps, in *The Geometry of Normal Faults*, edited by A. M. Roberts, G. Yielding and B. Freeman, *Geol. Soc. Spec. Publ.*, *56*, 93-112, 1991.

Crevallo, P.D., J. E. Warme, M. Septfontaine, and R. Burke, Evolution of Jurassic carbonate platforms in an active transtensional rift, *AAPG Bull*, *71*, 543-544, 1987.

Dewey, J. F., M. I. Helman, E. Turco, D.H.W. Hutton, and S.D. Knott, Kinematics of the western Mediterranean, in, *Alpine Tectonics*, edited by M.P. Coward, D. Dietrich, and R.G. Park, *Geol. Soc. Spec. Publ.*, *45*, 265-283, 1989.

Ensslin, R., Cretaceous synsedimentary tectonics in the Atlas system of Central Morocco, *Geol. Rundsch.*, *81*, 91-104, 1992.

Faure-Muret, A., and G. Choubert, Avec la collaboration de J. Kornprobst pour la zone interne, Le Maroc: Domaine rifain et atlasique, in *Tectonique de l'Afrique*, edited by, G. Choubert, UNESCO, Paris, 1971.

Fetah, M., M. Bensaid, and M. Dahmani, Carte Géologique du Maroc-Zawyat Ahancal, p.355, map scale 1:100,000, Minist. de l'Energie et des Mines, Royaume du Maroc, Rabat, Morocco, 1990.

Fiechtner, L., H. Friedrichsen, and K. Hammerschmidt, Geochemistry and geochronology of Early Mesozoic thoeleiites from central Morocco, *Geol. Rundsch.*, *81*, 45-62, 1992.

Fraissinet, C., E.M. Zouine, J.L. Morel, A. Poisson, J. Andrieux, and A. Faure-Muret, Structural evolution of the southern and northern Central High Atlas in Paleogene and Mio-Pliocene times, in *The Atlas System of Morocco-Studies on its geodynamic Evolution, Lect. Notes Earth Sci.*, vol.15, pp.273-291, edited by V. Jacobshagen, Springer-Verlag, New York, 1988.

Gauthier, H., Contribution à l'étude géologique des formations post-liasiques des bassins du Dadés et du Haut Todra, *Notes Mém. Serv. Géol., Morocco*, 119, 212 pp., 1960.

Ghissassi, A., M. Saadi, T. Skalli, E. A. Hilali, and A. Boudda, Carte Géologique du Maroc, Afourar, p. 247, map scale 1:100,000, Minist. de l'Energie et des Mines, Rabat, Morocco, 1977.

Gibbs, A.D., Balanced cross-section construction from seismic sections in areas of extensional tectonics, *J. of Struct. Geol.*, *5*, 153-160, 1983.

Giese, P., and V. Jacobshagen, Inversion tectonics of intracontinental ranges: High and Middle Atlas, Morocco, *Geol. Rundsch.*, *81*, 249-259, 1992.

Gomez, F., M. Barazangi and M. Bensaid, Active tectonism in the Intracontinental Middle Atlas Mountains of Morocco: Synchronous crustal shortening and extension, *J. Geol. Soc., London*, *53*, 389-402, 1996.

Gomez, F., R. W. Allmendinger, M. Barazangi, A. Er-Raji, and M. Dahmani, Crustal shortening and vertical strain partitioning in the Middle Atlas Mountains of Morocco, *Tectonics*, *17*, 520-533, 1998.

Hayward, A.B. and R.H. Graham, Some geometrical characteristics of inversion, in Inversion Tectonics, edited by M. A. Cooper and G. D. Williams , *Geol. Soc. Spec. Publ.*,*44*, 17-39, 1989.

Jabour, H. and K. Nakayama, basin modeling of Tadla basin, Morocco, for hydrocarbon potential, *AAPG Bull*, *72*, 1059-1073, 1988.

Jenny, J., Carte Géologique du Maroc au 1/100,000 Feuille Azillal (High Atlas central), *Notes Mém. Serv. Géol., Morocco*, *339*, 104 pp., 1988.

Jossen, J.-A., and J. Filali-Moutei, A new look at the structural geology of the southern side of the central and eastern High Atlas Mountains, *Geol. Rundsch.*, *81*, 143-156, 1992.

Laville, E., A multiple releasing and restraining stepover model for the Jurassic strike-slip basin of the Central High Atlas (Morocco). *in Triassic-Jurassic Rifting, Continental Breakup and the Origin of the Atlantic Ocean*, edited by W. Manspeizer, Elsevier, New York, pp. 499-523, 1988.

Laville, E., and A. Pique, La distension crustale atlantique et atlasique au Maroc au début du Mésozoique: le rejeu des structures hercyniennes, *Bull. Soc. Géol. Fr., 162*, 1161-1171, 1991.

Laville, E., and A. Pique, Jurassic penetrative deformation and Cenozoic uplift in the Central High Atlas (Morocco): A tectonic model-Structural and orogenic inversions, *Geol. Rundsch., 81*, 157-170, 1992.

Lowell, J.D., Mechanics of basin inversion from worldwide examples, eds., in *Basin Inversion*, edited by J. G. and P. G. Buchanan, *Geol. Soc. Spec. Publ., 88*, 39-57, 1995.

Makris, J., A. Demnati, and J. Klubmann, Deep seismic soundings in Morocco and a crust and upper mantle model deduced from seismic and gravity data, *Ann. Geophys., Ser. A, 3*, 369-380, 1985.

Manspeizer, W., J. Puffer, and H. Cousminer, Separation of Morocco and eastern North America: A Triassic-Liassic stratigraphic record, *Geol. Soc. Am. Bull., 89*, 901-920, 1978.

Marrett, R., and R.W. Allmendinger, Amount of extension on "small faults": An example from the Viking graben, *Geology, 20*, 47-50, 1992.

McClay, K.R., and P. G. Buchanan, Thrust faults in inverted extensional basins, in *Thrust Tectonics*, edited by K. R. McClay, pp. 93-104, Chapman and Hall, New York, 1992.

Medina, F., Tilted-blocks pattern, paleostress orientation, and amount of extension related to Triassic early rifting of the central Atlantic in the Amzri area (Argana basin, Morocco), *Tectonophysics, 148*, 229-233, 1988.

Michard, A., Eléments de géologie marocaine, *Notes Mém Serv. Géol, 252*, 408 pp., 1976.

Milanovsky, E., Aulacogens of ancient platforms: Problems of their origin and tectonic development, *Tectonophysics, 73*, 213-248, 1981.

Outtani, F., B. Addoum, E. Mercier, D. Frizon de Lamotte, and J. Andrieux, Geometry and kinematics of the South Atlas Front, Algeria and Tunisia, *Tectonophysics, 249*, 233-248, 1995.

Pique, A., G. Bossiére, J.-P. Bouillin, A. Chalouan, and C. Hoepffner, Southern margin of the Variscan belt: The north-western Gondwana mobile zone (eastern Morocco and northern Algeria), *Geol. Rundsch., 82*, 432-439, 1993.

Ramsay, J.G., and M.I. Huber, The techniques of modern structural geology, in *Folds and Fractures*, vol. 2, 494 pp., Academic, San Diego, Calif., 1987.

Rolley, J.-P., Carte géologique du Maroc au 1/100,000, Feuille Afourer (High Atlas central), *Notes Mém Serv. Géol., Morocco, 247*, 103 pp., 1978.

Rosendahl, B.R., D.J. Reynolds, P.M. Lorber, C.F. Burgess, J. McGill, D. Scott, J.J. Lambiase, and S.J. Derksen, Structural expressions of rifting: Lessons from Lake Tanganyika, Africa, *Geol Soc. Spec. Publ, 25*, 25-43, 1986.

Saadi, M., E. A. Hilali, and M. Bensaid, Carte géologique du Maroc-Jebel Saghro-Dadés, map, p.161, scale 1:200,000, Minist. de l'Energie et des Mines, Rabat, Morocco, 1977.

Saadi, M., M. A. Hilali, and A. Demnati, Carte magnétique du Maroc, Interprétation du leve aéromgnétique, Bechar and Marrakech, map p.320-c, scale 1:1,000,000, Minist. de l'Energie et des Mines, Rabat, Morocco, 1979.

Saadi, M., M. Bensaid, and M. Dahmani, Carte Géologique du Maroc-Azilal, map, p. 339, scale 1:100,000, Minist. de l'Energie et des Mines, Rabat, Morocco, 1985.

Sandvol, E., D. Seber, and M. Barazangi, Single-station receiver function inversions in the Middle East and North Africa: A grid search approach, *Eos Trans., AGU, 77*, (46), 477, 1996.

Schaer, J.-P., and J. Rodgers, Evolution and structure of the High Atlas of Morocco. *in The Anatomy of Mountain Ranges*, edited by J.-P. Schaer, and J. Rodgers, pp. 107-127, Princeton Univ. Press, Princeton, N.J., 1987.

Schmidt, K.-H., The tectonic history of the pre-Saharan depression (Morocco)-A geomorphological interpretation, *Geol Rundsch., 81*, 211-219, 1992.

Schwartz, G., H.G. Mehl, F. Ramdani and V. Rath, Electrical resistivity structure of the eastern Moroccan Atlas system and its tectonic implications, *Geol. Rundsch., 81*, 221-235, 1992.

Seber, D., M. Barazangi, A. Ibenbrahim, and A. Demnati, Geophysical evidence for lithospheric delamination beneath the Alboran Sea and Rif-Betic mountains, *Nature, 379*, 785-790, 1996.

Suppe, J., Geometry and kinematics of fault-bend folding, *Am. J. of Sci., 283*, 684-721, 1983.

Suppe, J., and D. A. Medwedeff, Geometry and kinematics of fault-propagation folding, *Eclogae Geol. Helv., 83*(3), 409-454, 1990.

Tadili, B., and M. Ramdani, Computer file of Moroccan earthquakes, *Bull. Seismol. Soc. of Am., 73*, 653-654, 1983

Vially, R., J. Letouzey, F. Bénard, N. Haddadi, G. Desforges, H. Askri, and A. Boudjema, Basin inversion along the North African Margin, The Saharan Atlas (Algeria), in *Peri-Tethyan Platforms*, edited by F. Roure, pp. 79-118, Éd. Technip, Paris, 1994.

Wigger, P., G. Asch, P. Giese, W.-D. Heinsohn, S.O. El Alami, and F. Ramdani, Crustal structure along a traverse across the Middle and High Atlas Mountains derived from seismic refraction studies, *Geol Runsch., 81*, 237-248, 1992

Williams, G.D., C.M. Powell, and M.A. Cooper, Geometry and kinematics of inversion tectonics, in *Inversion tectonics*, edited by M.A. Cooper and G. D. Williams, *Geol. Soc. Spec. Publ., 44*, 3-15, 1989.

Woodward, N.B., S. Boyer, and J. Suppe, Balanced Geological Cross-Sections: An Essential Technique in Geological Research and Exploration, in, *Short Course in Geology Ser.*, vol. 6, 132 pp., AGU, Washington, D. C., 1989.

Ziegler, P.A., *Geological Atlas of Western and Central Europe*, vol. 2, 130 pp., 40 plates, Shell Int. Pet., The Hague, 1982.

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