TECTONIC EVOLUTION OF THE ATLAS MOUNTAINS
NORTH AFRICA

A Dissertation

Presented to the Faculty of the Graduate School

of Cornell University

in Partial Fulfillment of the Requirements for the Degree of

Doctor of Philosophy

by

Weldon Harold Beauchamp

January 1998

TECTONIC EVOLUTION OF THE ATLAS MOUNTAINS, NORTH AFRICA

Weldon Harold Beauchamp, Ph.D.

Cornell University 1998

The Atlas Mountains of North Africa are one of the largest intracontinental mountain belts in the world. Despite the size of this orogen (~2000 km), the basic kinematic and tectonic evolution of the Atlas Mountains has previously not been well understood. These mountains formed hundreds of kilometers from active plate margins. The formation of the Atlas Mountains was greatly influenced by a previous Mesozoic intracontinental rift system. This rift system spanned half of the African continent and was larger in breadth than the Red Sea.

This study set out to synthesize existing data and studies of the Atlas Mountains and integrate these data with new geological, geophysical and remote sensing data. The construction of a tectonic map was undertaken to define the tectonic units and terraines of North Africa. The delineation of these regions allow for the study of how they have interacted during the kinematic evolution of the Atlas system.

Geological field work was undertaken to study the kinematics of inversion tectonics and to construct a balanced geological-geophysical transect. The transect suggests shortening across the orogen (36 km) was achieved by thrusting along detachments at several levels in the upper crust. Syn-rift and post-rift sedimentary rocks were uplifted by the reactivation of syn-rift normal faults and newly formed thin-skinned thrust faults. A restoration of

the deformed cross section indicates the original Atlas rift basin was approximately 113 kilometers wide.

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 Mountains. The partitioning of strain may involve the transfer of shortening from the margins at shallow depths, to the mid-lower crust in the central region of the orogen. Thrusting in the High Atlas Mountains is bi-vergent, with thrusts dipping to the south along the northern margin, and northward dipping faults to the south.

The presence of pre-existing structural geometries such as accommodation zones, fault ramps, fault relays and *en echelon* faulting formed by rift processes will have an effect on subsequent compressional stress fields generated by plate convergence and other tectonic processes. Superposed folding which is disharmonic may in fact be a unique characteristic to inverted rift systems that result in intracontinental mountain belts.

BIOGRAPHICAL SKETCH

Weldon Beauchamp was born in Duncan, Oklahoma on March 2, 1959. He had an early interest in rocks and spent a lot of time walking around in fields picking up rocks and arrowheads and trying to figure out what they were. This went on for years until he got sidetracked and took up playing the violincello in the third grade. It was suspected that he took up the cello to torture his parents by playing the Bach suites over and over again. Scholarship offers almost sealed his fate as a musician, until his teacher informed him that while he was a good cellist, he would always make a very modest living as a musician. This information changed his career path, and he enrolled in his first geology class at the University of Oklahoma (no one explained that geologists were about as prosperous as musicians). Weldon yearned for a change of scenery so he transferred to New England College in Henniker, N.H.. Weldon earned his BA in geology at New England College while playing cello with the New Hampshire Philharmonic, and pursuing a new passion in climbing and mountaineering. It was then back to Oklahoma to earn his MS in Geology at Oklahoma State University with a Scottish advisor who wore a kilt (Nowell Donovan). Weldon got his first job as a geologist with Sun Exploration and Production Company where he worked as a geologist for ten years and travelled the world. Along the way, Weldon finally did something right and married Rennie DePenaloza. Weldon, Rennie and daughter Margot then moved to Ithaca, N.Y.. While in Ithaca Weldon was outnumbered three-to-one with the birth of Shelby, his second daughter. With the support of his family and advisor, he earned his doctorate studying one the largest mountain belts in the world.

Dedicated to my family.

ACKNOWLEDGMENTS

This study would not have been possible without the assistance of ONAREP (Office National de Recherches et d'Exploitations Petrolieres), and the Geological Survey of Morocco. None of the research in this manuscript would have been possible without my friends in Morocco, especially Ahmed Demnati and Mohamed El Alji. The author greatly appreciates input from my colleagues and friends at Cornell: Muawia Barazangi, Rick Allmendinger, William Philpot, Alex Calvert, Francisco Gomez, Tomas Zapata, Graham Brew, Marisa Vallve, Dogan Seber, Dave Steer, Eric Sandvol, Ben Brooks, Mary Parke, Bob Litak, William Philpot, Red Bradley and many others in the preparation of this manuscript. The assistance of Christine Orgren, Marisa Valve, Meredith Hansen and Carrie Brindisi were invaluable in the compilation of the Tectonic Map of the Atlas Mountains, North Africa (Chapter One). Special thanks to Steve Gallow, Matthew Kleinman , Joe Weaver and the rest of the staff who make the department run. Acknowledgment is made to the donors of the Petroleum Research Fund, administered by the American Chemical Society (ACS-PRF #29505-AC2), and to a grant from AASERT(# DAAH04-94-G-0361) for support of this research. We thank Richard G. Stanley, Z. R. Beydoun, and John A. Best for their constructive reviews of Chapter Two. In addition, we would like to thank Harold A. Illich and Geomark Research Incorporated for their geochemical analysis and help in evaluating potential source rocks in Morocco. INSTOC contributions #223 and #243

TABLE OF CONTENTS

CHAPTER TWO

CHAPTER THREE

LIST OF TABLES

LIST OF FIGURES

LIST OF PLATES

CHAPTER ONE

Tectonic Map of the Atlas Mountains, North Africa

Abstract

The Atlas Mountains is one of the largest intracontinental mountain belts (~2000 km) in the world, comparable to the Rocky and Ural Mountains. The Atlas Mountains extend from the Atlantic margin of Morocco to the Mediterranean coast of Tunisia, and are the result of the uplift of a previous Mesozoic rift system. An understanding of the crustal scale structure and kinematic evolution of these mountains has not developed, in part, due to the lack of a single detailed structural/tectonic map of the orogen. Because of the location of the Atlas Mountains in the mostly arid/desert environment of Morocco, Algeria and Tunisia, an opportunity to create a detailed map of the mountain belt using Landsat-TM satellite data was possible. A Landsat-TM mosaic, based on 40 TM scenes, was created as a base for geological interpretation for the map The extent of the Atlas Mountain belt is also covered by available detailed geologic mapping in Tunisia, Algeria and Morocco. These data combined with available literature and field data were used as ground control for the mapping of structural/tectonic features based on the interpretation of the TM mosaic.

The developed tectonic map of the Atlas Mountains resulted in the recognition of major distinct tectonic regimes within the study area of North Africa. Along the northern coast of North Africa distinct tectonic units comprise the Rif/Tellian Atlas Mountains. These coastal mountain ranges are separated from the intracontinental Atlas Mountains by the High Plateau (Morocco and

Algeria) and the Meseta of Morocco. The Tellian coastal ranges are the result of plate margin convergence (Tertiary), and are comprised of allochthonous tectonic units transported southward from the present-day Mediterranean Sea. The High Plateau and Meseta are large platform regions that remain relatively undeformed in the Tertiary.

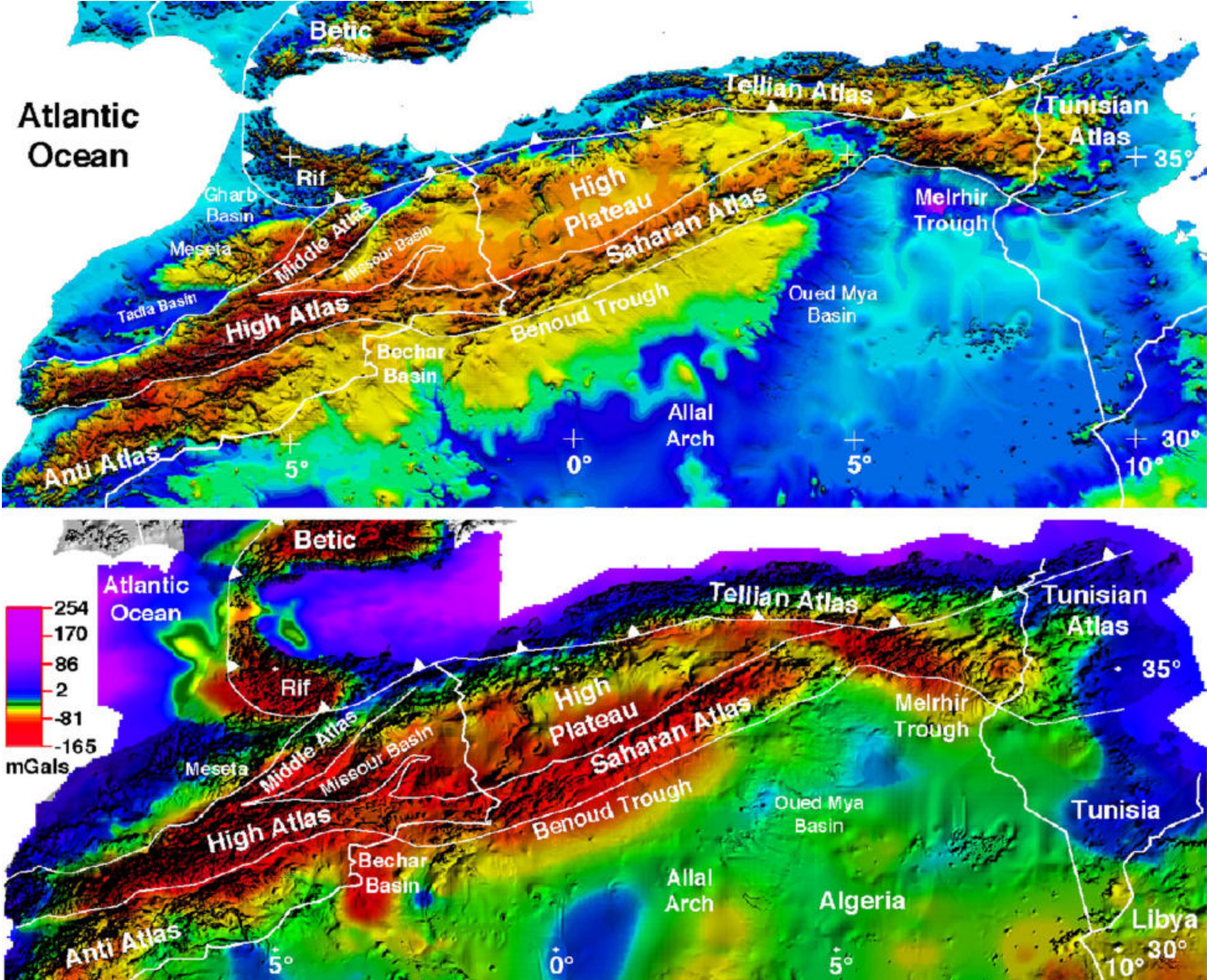
Introduction

A comprehensive structural/tectonic map of the Atlas Mountains is necessary to fully understand the evolutionary history of the mountain belt and the northern margin of Africa. While excellent detailed mapping of North Africa exists (1:50,000, 1:100,000, 1:200,000), these maps do not allow researchers to conveniently study the Atlas system in its entirety. The construction of a tectonic map was undertaken to define the tectonic units and terrains that comprise the structural architecture of North Africa. The definition of the tectonic terrains of this region of Africa allows for the study of each individual tectonic terrain and how these terrains have interacted during the evolution of the Atlas system.

Construction of a tectonic map

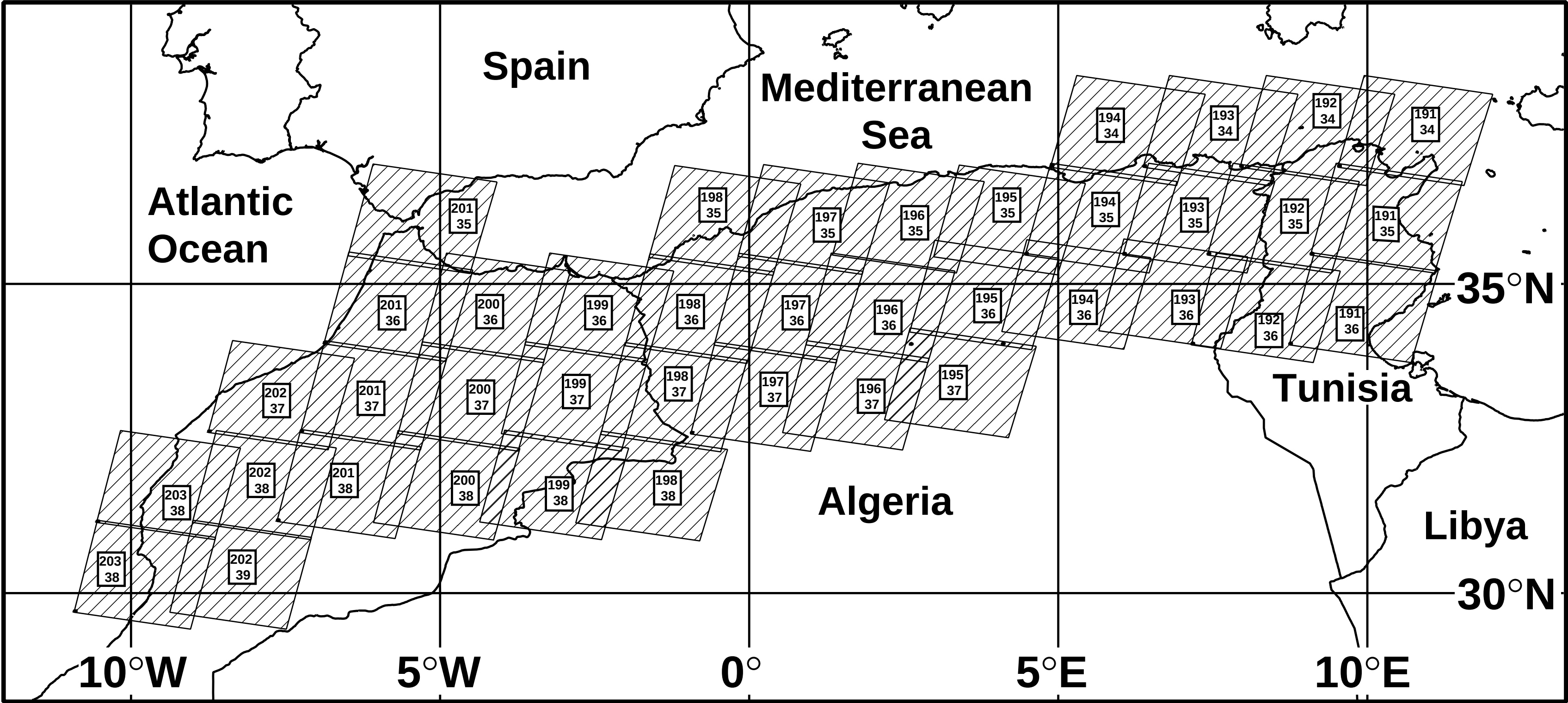
The first step in the mapping process was the selection of Landsat-TM data to cover a region of North Africa extending from Tunisia to the Atlantic coast of Morocco (Figure 1.1 and Plate 1). Images were selected for their lack of cloud cover (< 10%), high digital quality, and lack of vegetation. Forty Landsat-TM (Landsat 5, Thematic Mapper) scenes were selected to cover the Atlas Mountains (Figure 1.2 and Plate 1). The 30 meter resolution of these 40

Figure 1.1 Digital topographic map of North Africa (A), showing the tectonic terrains of the High/Saharan/Tunisian Atlas Mountains and the Rif/Tellian Mountains. Bouguer gravity data (B) show the relative location of the Atlas rift basin and associated foreland basins formed during uplift in the Tertiary Alpine orogeny.



Betic
Atlantic
Ocean
Tellian Atlas
Tunisian
Atlas
Rif
Gharb
Basin
Meseta
Middle Atlas
Missour Basin
High
Plateau
Saharan Atlas
Melrhir
Trough
Tadla Basin
High Atlas
Benoud Trough
Oued Mya
Basin
Bechar
Basin
Allal
Arch
Anti Atlas
35°
30°
5°
0°
5°
10°
Betic
Atlantic
Ocean
254
170
86
2
-81
-165
mGals
Tellian Atlas
Tunisian
Atlas
35°
Rif
High
Plateau
Saharan Atlas
Melrhir
Trough
Middle Atlas
Missour Basin
Meseta
High Atlas
Benoud Trough
Oued Mya
Basin
Tunisia
Bechar
Basin
Allal
Arch
Algeria
Libya
30°
Anti Atlas
5°
0°
5°
10°

Figure 1.2 Location map of Landsat-TM data used in the construction of a satellite mosaic of North Africa. These 40 images were processed and utilized for geological interpretation and the construction of a structural/tectonic map of the Atlas Mountains and North Africa.


Spain
Mediterranean Sea
Atlantic Ocean
Tunisia
Algeria
Libya
35°N
30°N
10°W
5°W
0°
5°E
10°E
194 34
193 34
192 34
191 34
201 35
198 35
197 35
196 35
195 35
194 35
193 35
192 35
191 35
201 36
200 36
199 36
198 36
197 36
196 36
195 36
194 36
193 36
192 36
191 36
202 37
201 37
200 37
199 37
198 37
197 37
196 37
195 37
203 38
202 38
201 38
200 38
199 38
198 38
203 38
202 39

scenes allowed mapping of geological features such as bedding, faults and the interpretation of fold axes. These images were processed and combined to form a mosaic that covered the study area (Plate 1). The final mosaic was used to trace coastlines, bedding, fold axes and faults.

Several representative scenes across the study area were selected and statistics were calculated for each scene. These statistics were used to generate scattergrams for the selection of the least correlated of the bands. Band 7 (mid-infrared, 2.08-2.35 μ) was chosen to display the red band, band 4 (near infrared, 0.76-0.90 μ) was used for display of the green band, and band 1 (blue, 0.45-0.52 μ) for the display of the blue band in a RGB color space in each scene of the mosaic. These bands resulted in a wide range of colors which yielded an excellent representation of the color, texture and pattern relationships of the geological outcrops of North Africa.

Detailed geological maps exist for much of North Africa (see Plate 2). These maps were obtained from various sources (USGS library in Reston, Virginia and the Moroccan Geological Survey) and the faults and major stratigraphic contacts were digitized as control for the interpretation of the Landsat-TM images. Earthquake focal mechanisms (M>4) displayed on the tectonic map were compiled from various published sources by Francisco Gomez (*Udias et al.,* 1976; *Tadili*, 1986; *Grimison*, 1986; *Kasser et al.,* 1987; *Swan*, 1988; *Meghraoui*, 1990; *Bobier et al.,* 1991; *Galindo-Zindivar et al.,* 1993; *Aoudia and Meghraoui*, 1995; *Buforn et al.,* 1995; *Kiratzi*, 1995; *Medina*, 1995; *Dziewonski et al.,* 1995). A GIS system (ARCINFO) was used to combine geological data digitized from detailed geological maps and field data, with data (coverages) interpreted from the Landsat-TM data.

Description of the tectonic regimes of North Africa

Rif/Tellian Atlas

The Rif and Tellian Atlas exist along the coastal regions of the Mediterranean Sea. The Rif/Tellian tectonic units of North Africa were thrust southward during the Alpine orogeny in the Tertiary (Oligocene-Miocene). The Rif Mountains are part of the "Gibraltor Arc" that surrounds the Alboran Sea (Figure 1.1 and Plate 2). The Rif Mountains (up to 1500 meters high) and the Tell Atlas of Algeria make up the southern margin of the Maghrebides orogenic belt (*Petters,* 1991). This belt extends from the Betic Cordillera of southeastern Spain, across the coastal regions of North Africa, across Sicily and into southern Italy (Figure 1.1). The Rif/Tellian Atlas are part of these allochthonous mountains characterized by numerous imbricate thrust and asymmetrical nappe structures that dip northwards and have a southward vergence. A large gravity anomaly (Figure 1.1) is associated with the Rif region, showing that it is uncompensated. *Seber et al*. (1996) proposed that this anomoly is associated with the delamination of the lithosphere.

The Rif Mountains decrease in topographic expression eastward along the Moroccan coast where the northern Middle Atlas Mountains (autochthonous) extend beneath the thrusted Rif tectonic (Plate 2). It is noteworthy that there are no significant Tertiary foreland basins associated with the thrusting of the Rif/Tellian Atlas system. While thrusting and significant shortening (e.g., crustal loading) has occurred along the Tellian coastal ranges of Algeria and Tunisia, southward of these thrusts belts the High Plateau (Figure 1.1 and Plate 2) remains at greater elevations than the coastal ranges, possibly indicating a high crustal rigidity, zero flexural rigidity (e.g., local isostacy), or uplift by some other kinematic mechanism.

The High Plateau extends from the Middle Atlas Mountains of Morocco where the Missour Basin (intermontane basin) comprises the actual western margin of the tectonic block (*Beauchamp et al.,* 1996). The Missour Basin has an average elevation of ~1500 meters which is about 500 meters lower than the average elevations of the High Atlas south of the Missour Basin and High Plateau (Figure 1.1 and Plate 2). The Tellian Atlas in eastern Algeria is the exception, where the Hodna Basin forms near the intersection of the Saharan Atlas and the Tellian Atlas. The Hodna and Cheliff basins (Plate 2) are post-tectonic basins of Late Miocene age that formed over the thrusted Early Miocene nappes (*Petters*, 1991).

North of the High Plateau along the coastal regions of Algeria is one of the most active seismic regions of North Africa. The Cheliff Basin of Algeria is the location of the largest magnitude earthquake recorded in North Africa near El Asnam (M_s=7.3) (*Meghraoui,* 1986). The shortening related to this earthquake is estimated to have been on the order of 2.2 meters. The orientation of this basin and related faulting suggests north-south to NNW-SSE compressional deformation (*Meghraoui*, 1986). Farther east, another significant earthquake occurred to the east, west of Algers in the Tipaza region at Mont Chenoua (M_s=6.1) in 1989 (*Meghraoui,* 1990). This earthquake resulted in surface ruptures of 4.0 km in length and 7 cm of vertical displacement. This earthquake was interpreted to have been the result of displacement of a thrust associated with Mont Chenoua.

As can be noted from the map accompanying this study, most of the large earthquakes in North Africa has occurred in the Rif/Tellian Atlas (Plate

2). The recognition of fault displacement associated with the hanging wall anticline of Mont Chenoua provides evidence that many of the blind thrusts related to fault related folds of the Rif/Tellian Atlas and the High/Saharan Atlas mountains have potential for future seismic activity. Structures related to neotectonic deformation were mapped in the Chellif Basin (El Asnam Fault) by (*Meghraoui*, 1986). Therefore, Landsat-TM data are an effective tool for the mapping of both active faults and older faults. TM data are an important tool for documenting active faulting and fault related folds (blind thrusts) in areas of high earthquake risk in North Africa.

High Atlas

The High Atlas Mountains of Morocco reach elevations of more than 4000 meters, and they are made up of four distinct regions. They are, from west to east, (1) the western High Atlas (Paleozoic and Hercynian age granites with a Mesozoic cover); (2) the Precambrian High Atlas with thin Mesozoic cover to the north and south; (3) the central High Atlas with thick, continental, Mesozoic syn-rift clastics; and (4) the Eastern Atlas with thick calcareous syn-rift sedimentary strata. The Western High Atlas was affected by rifting in the Atlantic margin and extends eastward with syn-rift Triassic sedimentary clastic rocks reaching 5 km in thickness (*Beauchamp*, 1988). The Eastern High Atlas is characterized by calcareous syn-rift facies, with shallow marine reefal buildups along the margins of the Atlas, grading into deep water calcturbidites in the central part of the mountain belt (*Crevallo*, 1987).

The original structural geometries of the Moroccan Atlas rift system have greatly affected the evolution of the subsequent Atlas during uplift and inversion in the Tertiary. A balanced geological-geophysical transect across the High Atlas Mountains concluded that the original rift basin had been shortened by 36 km (*Beauchamp et al.,* 1997, submitted manuscript). The restored transect indicates that the original rift basin was at least 113 km in width. Considerable shortening across this mountain belt was achieved by thin-skinned thrusting along the margins, with almost no shortening in the central region of the High Atlas in the shallow crust. It is believed that shortening and uplift of the central region of the High Atlas Mountains was achieved instead by shortening in the mid to lower crust.

Preserved in the High Atlas is the geometry of a syn-rift pull-apart basin (*Schaer and Rodgers*, 1987). This pull-apart basin was formed during rifting in the Triassic-Jurassic and has only been mildly deformed from its original extensional geometry despite significant uplift and inversion of the rift basin (Figures 1.3 and 1.4). North of this region in the Eastern High Atlas Mountains is further evidence of oblique extension during the Atlas syn-rift phase.

The Missour Basin is bounded by the NE trending Middle Atlas Mountains to the west, and on the east by a structure known as Jebel Mechkakour (Figures 1.3 and 1.5). This structure is related to a syn-rift normal fault that has been uplifted during the Tertiary, and Jurassic age strata are now exposed along its axis. The overall large scale geometry of the Missour Basin is that of a large rhombohedral shaped basin with a similar geometry as the pull-apart basin seen at Rich in the High Atlas. East of Jebel Mechkakour is the High Plateau which has been penetrated by several wells that have encountered thick sections of Triassic salt (~2000 meters)(*Beauchamp et al.,* 1996). This salt in the High Plateau has formed a salt pillow that can be seen

Figure 1.3 Location map showing the detailed regions interpreted from Landsat-TM images. These regions illustrate the structural/tectonic styles of deformation in the Atlas Mountains that were used to create the tectonic map of the orogen.

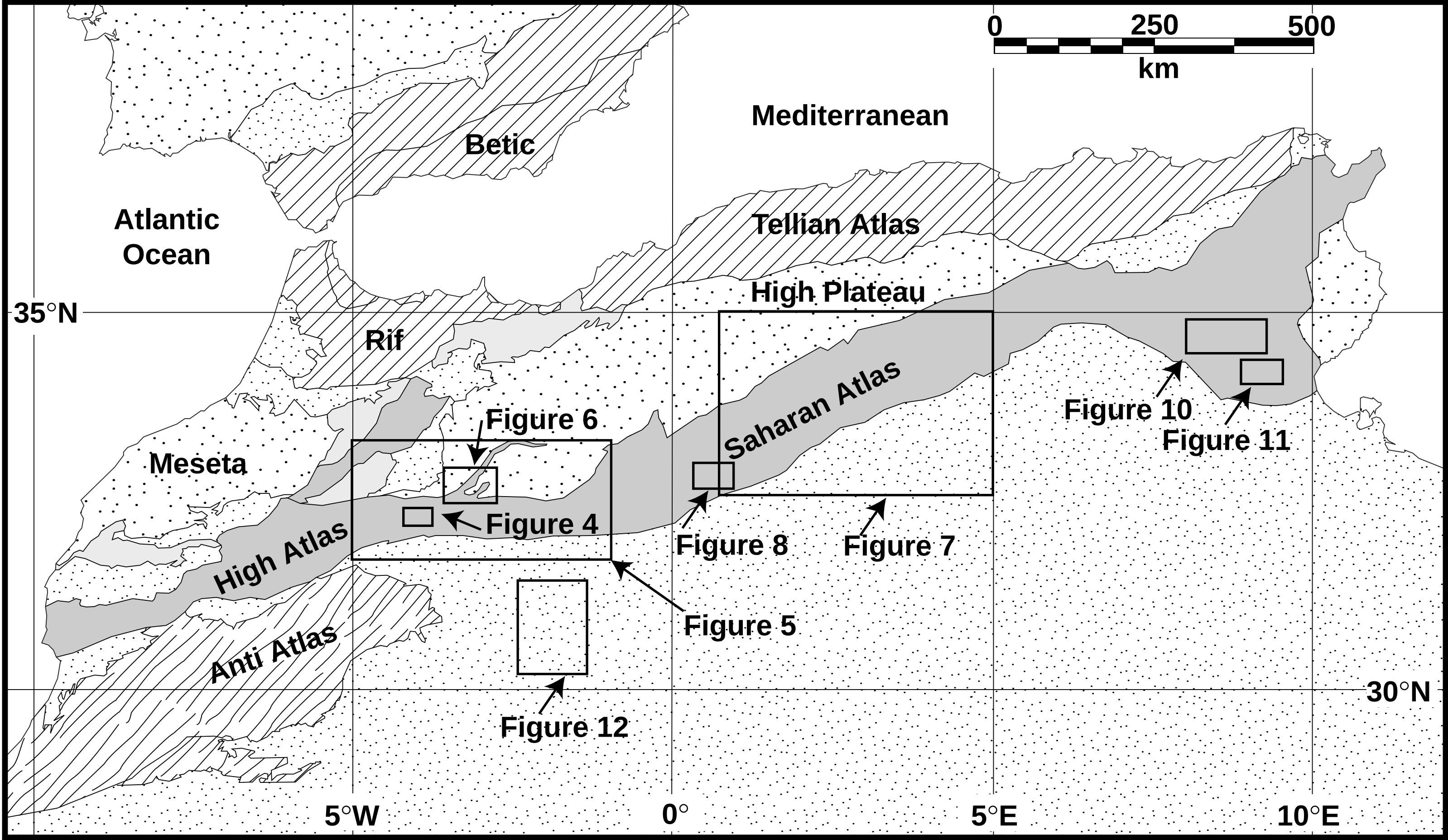

0
250
500
km
Mediterranean
Betic
Atlantic
Ocean
Tellian Atlas
High Plateau
35°N
Rif
Saharan Atlas
Figure 6
Meseta
Figure 10
Figure 11
Figure 4
High Atlas
Figure 8
Figure 7
Figure 5
Anti Atlas
30°N
Figure 12
5°W
0°
5°E
10°E

Figure 1.4 The Eastern High Atlas Mountains near the village of Rich contain a preserved pull-apart basin from the syn-rift phase of deformation in the Atlas. This basin has a rhombohedral shape and was formed by oblique extension in what today would be a NW-SE direction.

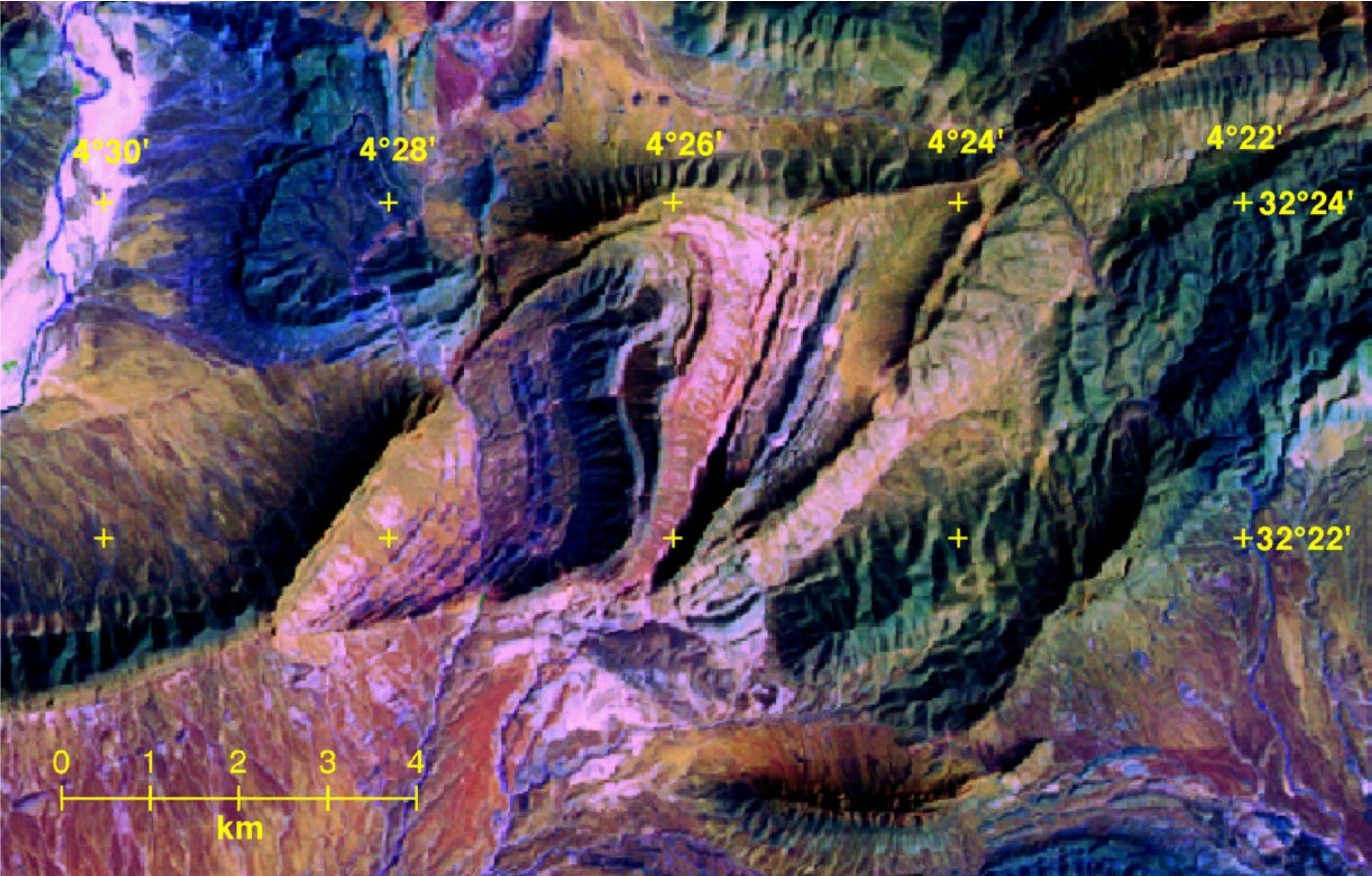

4°30'
4°28'
4°26'
4°24'
4°22'
32°24'
32°22'
0
1
2
3
4
km

Figure 1.5 The Middle Atlas Mountains and the Missour Basin form a large scale rhombohedral pull-apart basin north of the Eastern High Atlas Mountains. This geometry is repeated in the east where Jebel Mechkakour and the Saharan Atlas form another large rhombohedral basin. These large scale basins were formed by oblique extension oriented NW-SE relative to the present day Atlas.

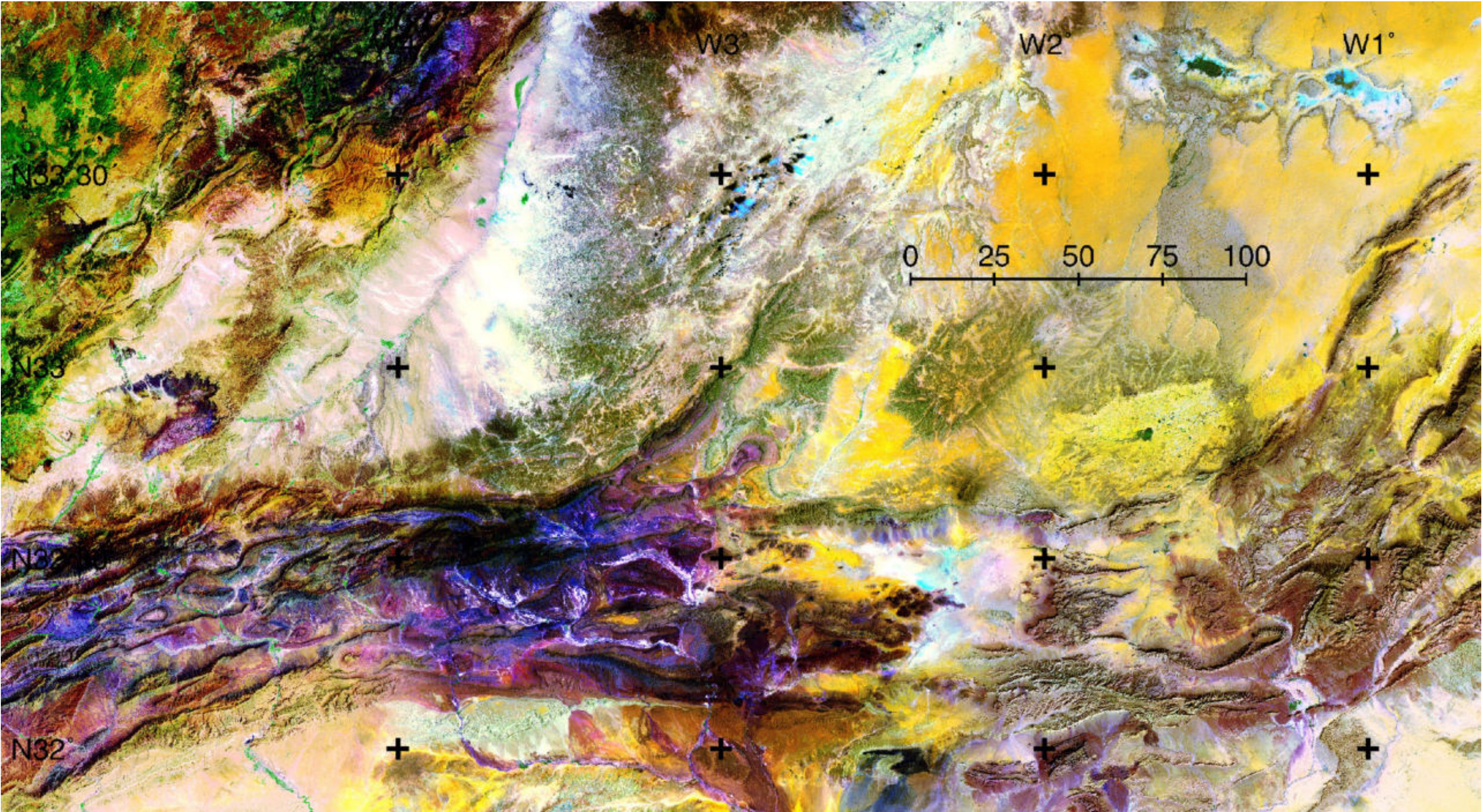
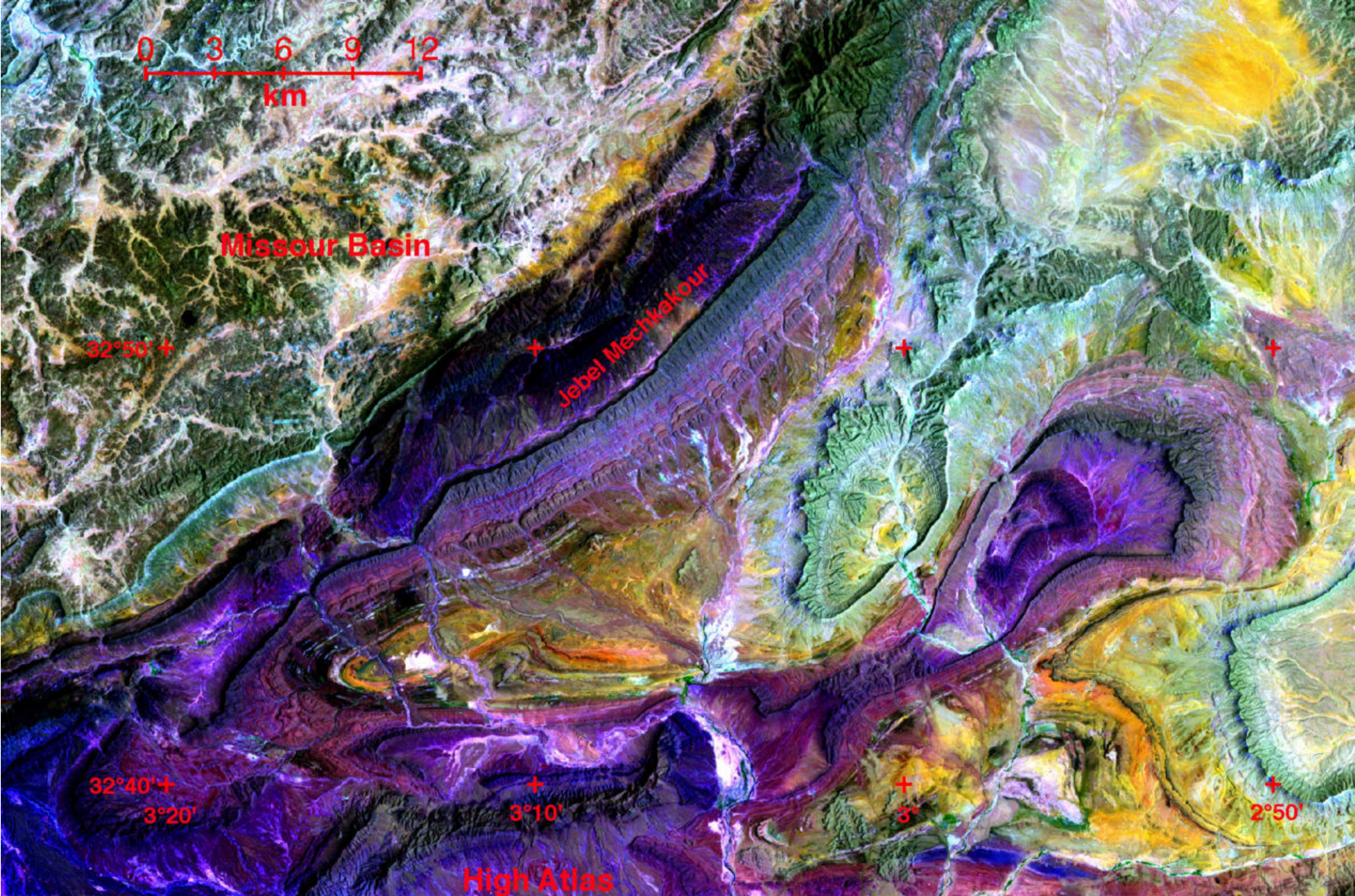

W3°
W2°
W1°
N33°30′
N33°
N32°
0
25
50
75
100

Figure 1.6 Landsat-TM image showing the structural geometry of Jebel Mechkakour. Jebel Mechkakour is an inverted syn-rift fault that was active as the margin of a rhombohedral pull-apart basin. The green colored area at the top of the image along the axis of the structure is a Jurassic age pinnacle reef. East of Mechkakour the next anticline (dark colors) is the exposure of the Jurassic above a salt pillow (known from seismic and well data).


0 3 6 9 12
km
Missour Basin
Jebel Mechkakour
32°50'
32°40'
3°20'
3°10'
3°
2°50'
High Atlas

at the surface east of Jebel Mechkakour(Figures 1.3 and 1.6). The High Plateau east of Jebel Mechkakour, like the Missour Basin, also has a large scale geometry of a rhombohedral pull-apart basin. The presence of the these pull-apart basins at different scales is strong evidence for transtensional deformation (*Crevallo*, 1987) or oblique rifting during the Mesozoic. The geometry of these features indicate that extension during the rift phase was oriented NW-SE approximately 45-60 degrees to the present Atlas Mountains.

Middle Atlas

The Middle Atlas Mountains are located 200-300 km east of the Atlantic coast of Morocco, and south of the Rif/Tellian Alpine front. Landsat-TM data show the Middle Atlas (NE-SW) bifurcating from the central region of the High Atlas (Plates 1 and 2). The Middle Atlas like the High Atlas mountains are related to the Mesozoic rifting event. The intersection of these two mountain belts may represent a triple junction that was active in the syn-rift phase, or the Middle Atlas may represent a subsidary rift basin of the Atlas rift system. Landsat-TM data show the plunge of the Middle Atlas Mountains northeastward where these mountains have less topographic expression (Plate 1). The Guercif Basin is part of the Mesozoic rift basin that has not been as strongly inverted as the southwestern part of the Middle Atlas (Plate 2).

The Middle Atlas can be divided into two provinces (*Gomez et al.,* 1996): the Tabular Middle Atlas to the northwest, and the Folded Middle Atlas in the southeast. The boundary between these two provinces is the Paleozoic-Mesozoic unconformity (*Gomez et al.,* 1996). Much of the Tabular Middle Atlas is covered by Quaternary basalts (Plate 2), and is bounded to the east by the North Middle Atlas fault. The eastern margin of the Middle Atlas is bounded by the High Moulouya and the Ait Oufella and Ksabi fault systems. The eastern margin of the Middle Atlas Mountains is bounded by the Missour Basin which is an intermontane basin that was a shelf margin/platform during the syn-rift phase of Atlas rifting (*Beauchamp et al*., 1996).

Quaternary volcanism is an important feature of the Middle Atlas Mountains. This volcanism covers about 1500 km^2 of the Tabular Middle Atlas and are alkali basalts ranging in age from 1.5 Ma to 0.5 Ma (*Gomez et al*., 1996). This younger volcanism is aligned parallel to the trend of the present-day convergence direction between North Africa and Spain. The correlation between volcanism in the Middle Atlas and active convergence in the region is not well understood.

Saharan Atlas

The Saharan Atlas, like the Moroccan Atlas, evolved from a previous Mesozoic rift system that extended across North Africa. The Moroccan High Atlas Mountains continue eastward into Algeria (Figure 1.1 and Plate 2) where the Mesozoic-Cenozoic sequence may attain a thickness of 10.7 km (*Caire*, 1974). The Saharan Atlas Mountains do not have the same topographic relief as the High Atlas Mountains of Morocco (Figure 1.1). Syn-rift age rocks of the Saharan Atlas have not been uplifted to the extent of the High Atlas, as many of the anticlinal structures of this part of the Atlas mountains only expose rocks Cretaceous and Tertiary in age. The Saharan Atlas also does not have the same level of seismic activity as the High Atlas or the Tellian Atlas. It may be that present day deformation occurs in the coastal Tellian thrust belt. The lack of present day seismic activity may also reflect upon the lack of uplift in the

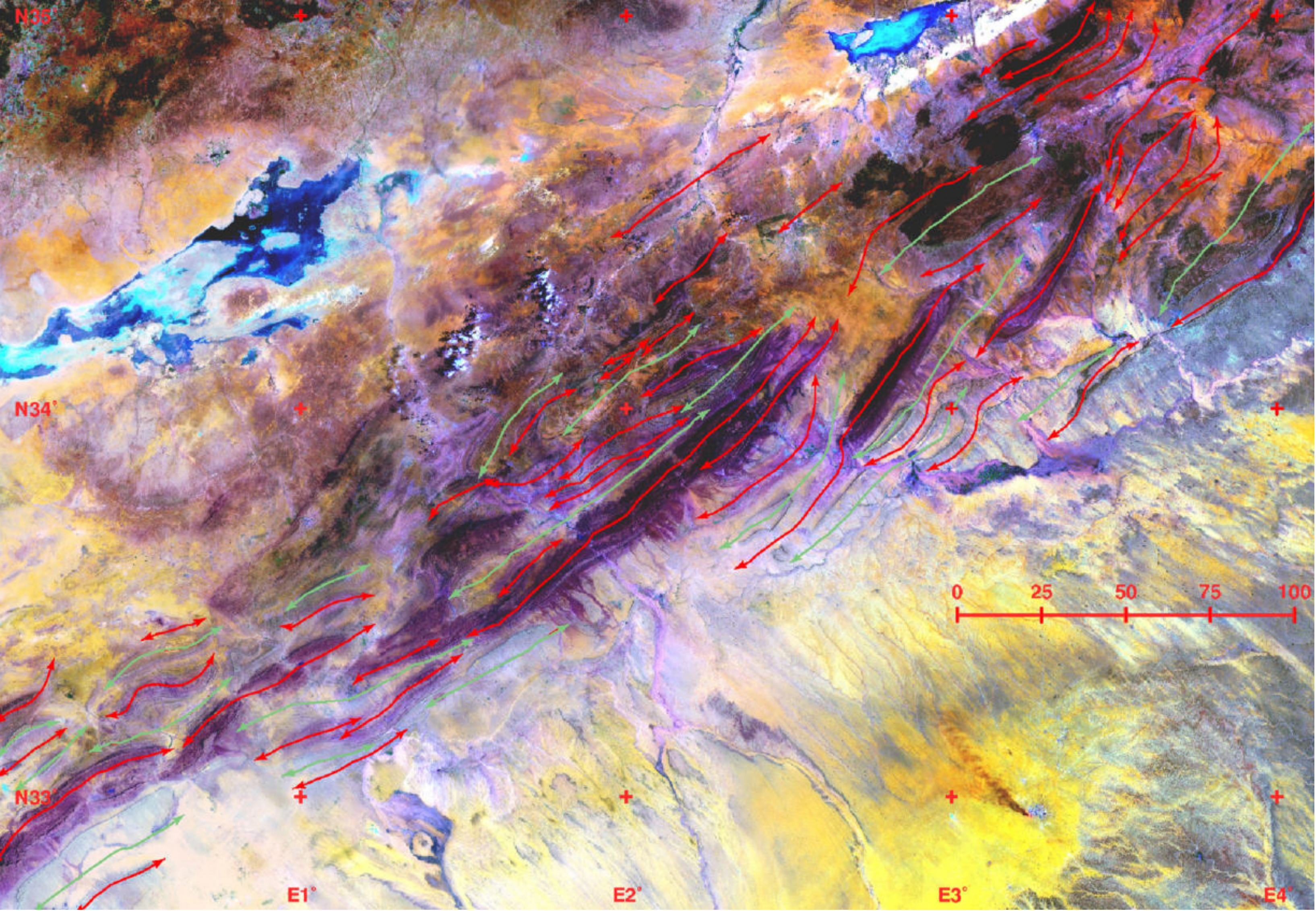
Saharan Atlas, as shortening may be accommodated by crustal deformation in the Tellian Atlas. In contrast to the High Atlas Mountains, the Saharan Atlas has a significant foreland basin developed along the southern margin of the fold belt, that may indicate less crustal rigidity than to the east in Morocco (Benoud Trough) (Plate 2 and Figure 1.1).

The Saharan Atlas Mountains are characterized by large (>50 km trend length) *en echelon* anticlines and synclines (Figure 1.7). The larger of these anticlines expose syn-rift Jurassic and occasionally Triassic age strata. Unlike the High Atlas of Morocco, the folds of the Saharan Atlas often preserve Tertiary and post-rift Cretaceous strata in the synclinal cores.

The large scale folds (~>20km trend length) are often affected by superposed folding (Figure 1.8). This superposed folding refolds the *en echelon* NE-SW trending folds about a second trend of fold axes that generally trend NW-SE. While superposed folding is usually developed by a change in the regional stress field, in the case of the Moroccan and Saharan Atlas the superposed folding is believed to be controlled by pre-existing syn-rift structural geometries. One characteristic that supports this concept is that many of the adjacent superposed folds vary in the orientation of the second phase of fold axes (Figure 1.8). This variation in the second phase of folding from adjacent structures may indicate that the original syn-rift normal faults, fault ramps, changes in fault polarity, accommodation zones and other characteristic rift structures are controlling the pattern of folding during uplift and inversion.

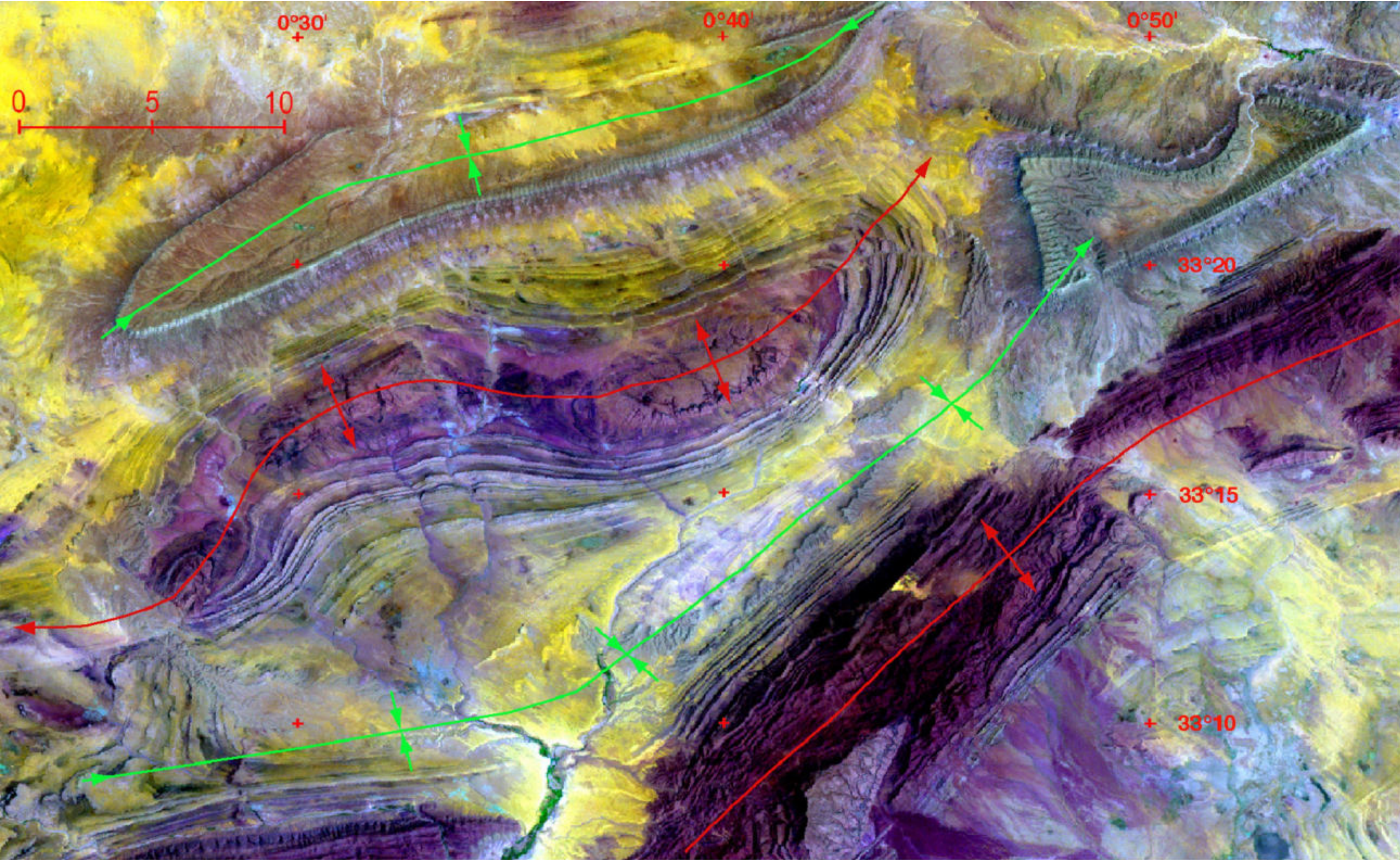
Analogue modeling of rift systems has shown that the geometry of rift structures are characteristic of the original direction of extension. Orthogonal

Figure 1.7 The Saharan Atlas Mountains of Algeria exhibit large scale *en echelon* folds. The areas that appear as bright blue to the north are large salt lakes (sabkhas). The smoke plume to the southeast of the Saharan Atlas is a natural gas flare from the Hasi Rmel gas field.



N35°
N34°
N33°
E1°
E2°
E3°
E4°
0
25
50
75
100

Figure 1.8 These large scale folds exhibit superposed folding that may have been controlled by previous syn-rift structures. The large anticline in the center of the image is trending NE-SW, and has been refolded by a second set of axes that trend NW-SE. The folds to the north and south of this anticline have been refolded by fold axes that do not affect the anticline in the center of the image.



0°30'
0°40'
0°50'
0
5
10
33°20
33°15
33°10

rift models are characterized by long, relatively straight rift border faults and short intra-rift faults that are perpendicular to the direction of extension (*McClay and White*, 1995).

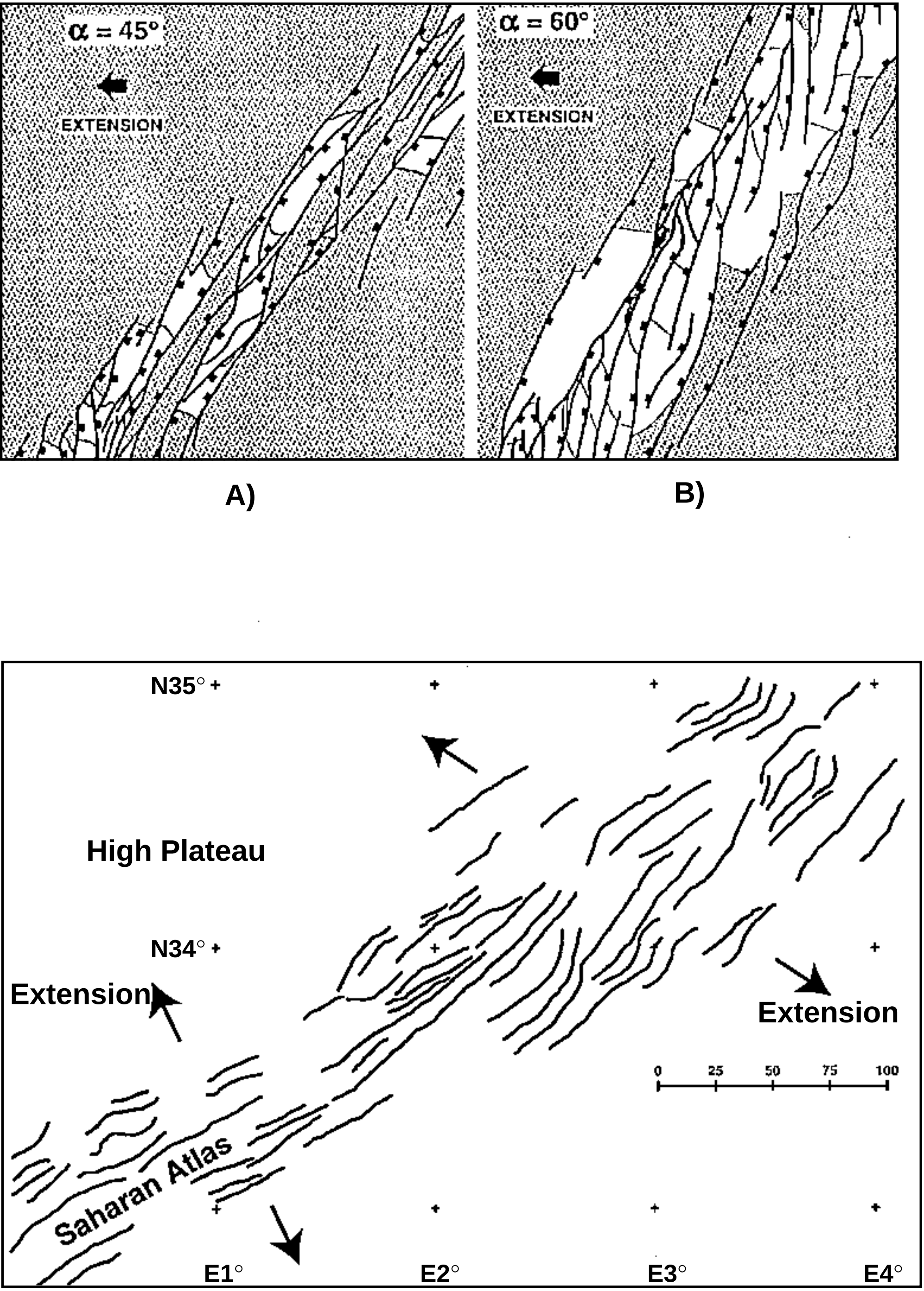
In contrast, oblique rift models result in segmented rift faults*, en echelon* fault arrays parallel to the zone of rifting and perpendicular to the direction of extension (*McClay and White*, 1995). These models compare well with faults geometries seen in younger rift systems such as the Afar and East African rift systems. The geometries shown by analogue models and in younger rift systems may have also been developed in the Atlas rift system of North Africa. The Atlas Mountains have been strongly influenced by the coincident pre-existing rift system, therefore, it is probable that upon inversion rift geometries would be inherited in the subsequent folding and compressional structural geometries.

An example of the influence of pre-existing structures on the subsequent tectonic evolution can be noted by the *en echelon* folding seen in the Saharan Atlas (Figure 1.7). If the *en echelon* folding of the Saharan Atlas and Moroccan Atlas is related to the inversion of syn-rift normal faults, then the existing *en echelon* folds may be perpendicular to what would have been the original direction of extension during rifting (Figure 1.9). Furthermore, the superposed folding that is characteristic of inversion and the partitioning of strain by pre-existing structural geometries, may have been influenced by an oblique compression relative to the original rift system.

Tunisian Atlas

The Northern margin of Tunisia is composed of nappes (Tunisian Alpine thrust

Figure 1.9 Analogue modeling of oblique extension results in rift systems that develop en echelon faulting and segmented boundaries (A and B) (after *McClay and White*, 1995). The High Atlas and Saharan Atlas both have characteristic en echelon folds that may be the result of the reactivation of syn-rift faults produced by oblique rifting.


α = 45°
EXTENSION
α = 60°
EXTENSION
A)
B)
N35°
N34°
High Plateau
Extension
Extension
Saharan Atlas
0
25
50
75
100
E1°
E2°
E3°
E4°

C)

zone) which were thrust southward during the Oligocene and early Miocene. This zone extends southward to a zone of diapirism which separates the Alpine thrusting from the Tunisian platform (Pelagian block) (*Petters*, 1991). The north-south trending fold belt of Tunisia is refereed to as the NOSA belt (North-South Axis) (*Boccaletti et al.,* 1988), and has been interpreted as related to a deep-seated left-lateral transcurrent fault. Geophysical data (gravity and seismic reflection data) and paleogeography indicate the presence of a crustal discontinuity between the Sahel plate (Pelagian platform) and the Tunisian Atlas of western Tunisia (Plate 2). The Sahel plate is a stable platform made up of a system of young (Pliocene-Quaternary) horsts and grabens (*Ben Dhia*, 1991).

The southern region of the Tunisian Atlas (Southern Tunisian Atlas or Gafsa zone) is characterized by en echelon folding near the region of Gafsa (Plate 2). This region is north of the huge salt lake of "Chott El Jerid" (Plates 1 and 2). En echelon folding in this part of the Atlas has been interpreted by *Ben Ayed* (1981) as related to a major strike-slip fault. The Gafsa fault also has associated recent seismicity (Plate 2) that has deformed by right-lateral movement. While these folds may have been influenced by recent right-lateral strike-slip movement, it is likely that these folds were originally influenced by the inversion and reactivation of preexisting rift related faults (Figure 1.10). These anticlines are asymmetric and are often cut by small displacement right-lateral strike-slip faults (Figure 1.11) (*Ferjani et al*., 1990). These structures are probably influenced by the movement of Triassic evaporites and in many cases faulting may be detached above the Triassic. North of the Gafsa region

Figure 1.10 En echelon folding in the region of Gafsa in the southern Tunisian Atlas. These folds have been proposed to have been formed by right-lateral strike-slip movement on the Gafsa fault (upper right). It is possible that these folds were formed by reactivation of earlier normal faults, and were later modified by strike-slip movement on the Gafsa fault that resulted in the second phase of folding.

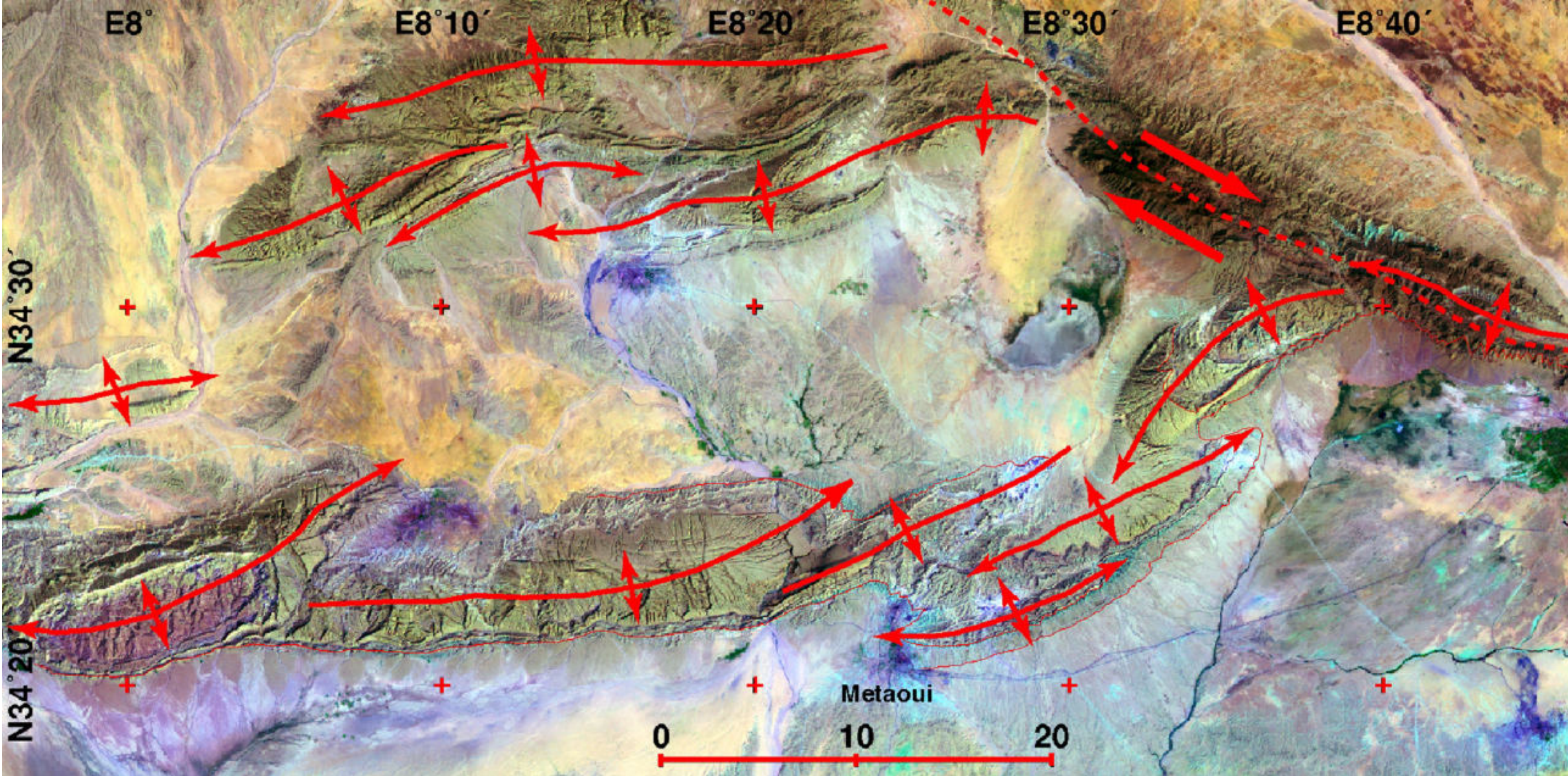
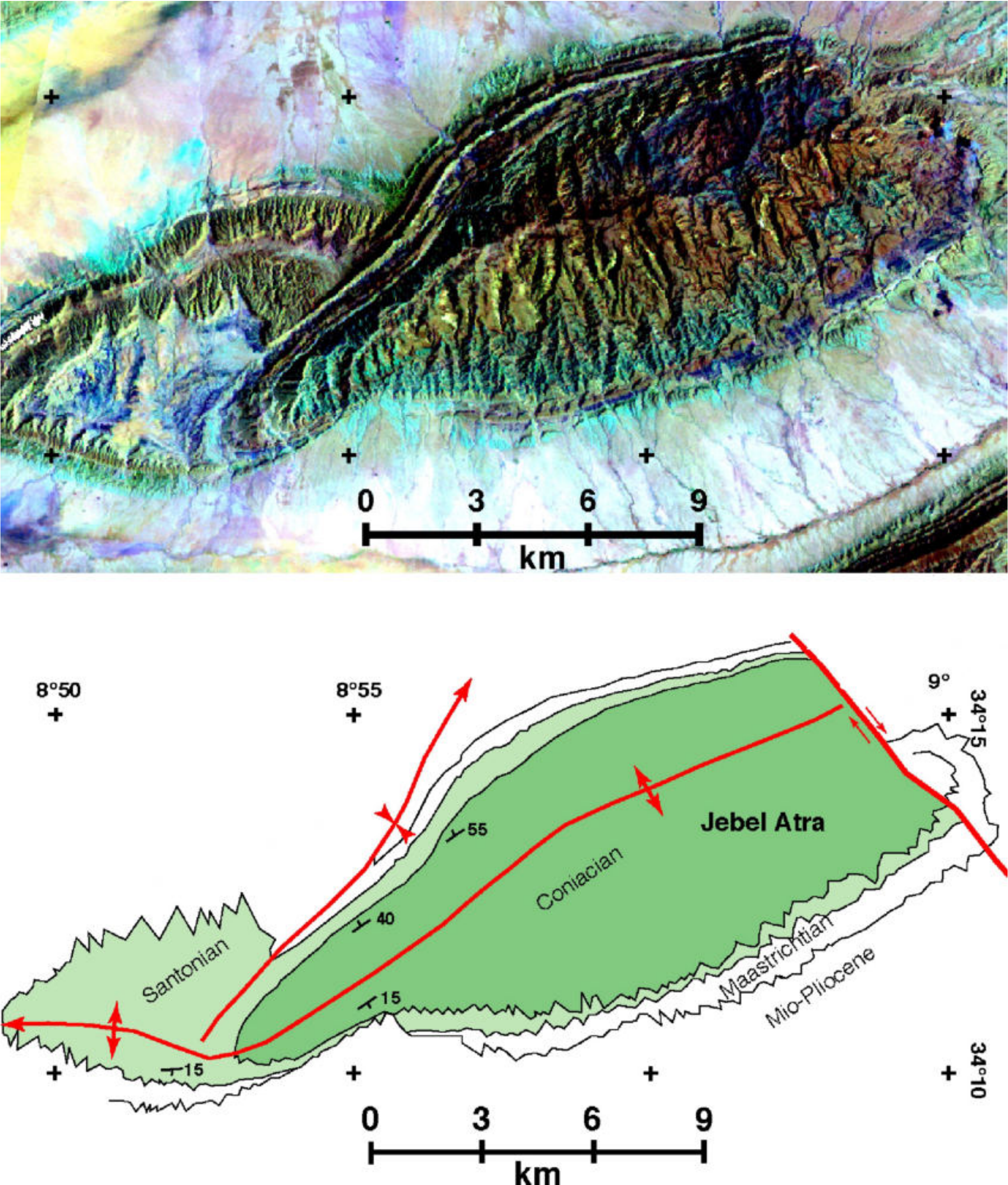

E8°
E8°10´
E8°20´
E8°30´
E8°40´
N34°30´
N34°20´
Metaoui
0
10
20

Figure 1.11 Jebel Atra in the Gafsa region north of the great salt lake of Chott El Jerif. This anticlinal structure is asymmetrical and is verging to the north. The fold has been refolded along its northern limb, and the structure has been cut by a right-lateral strike slip fault.

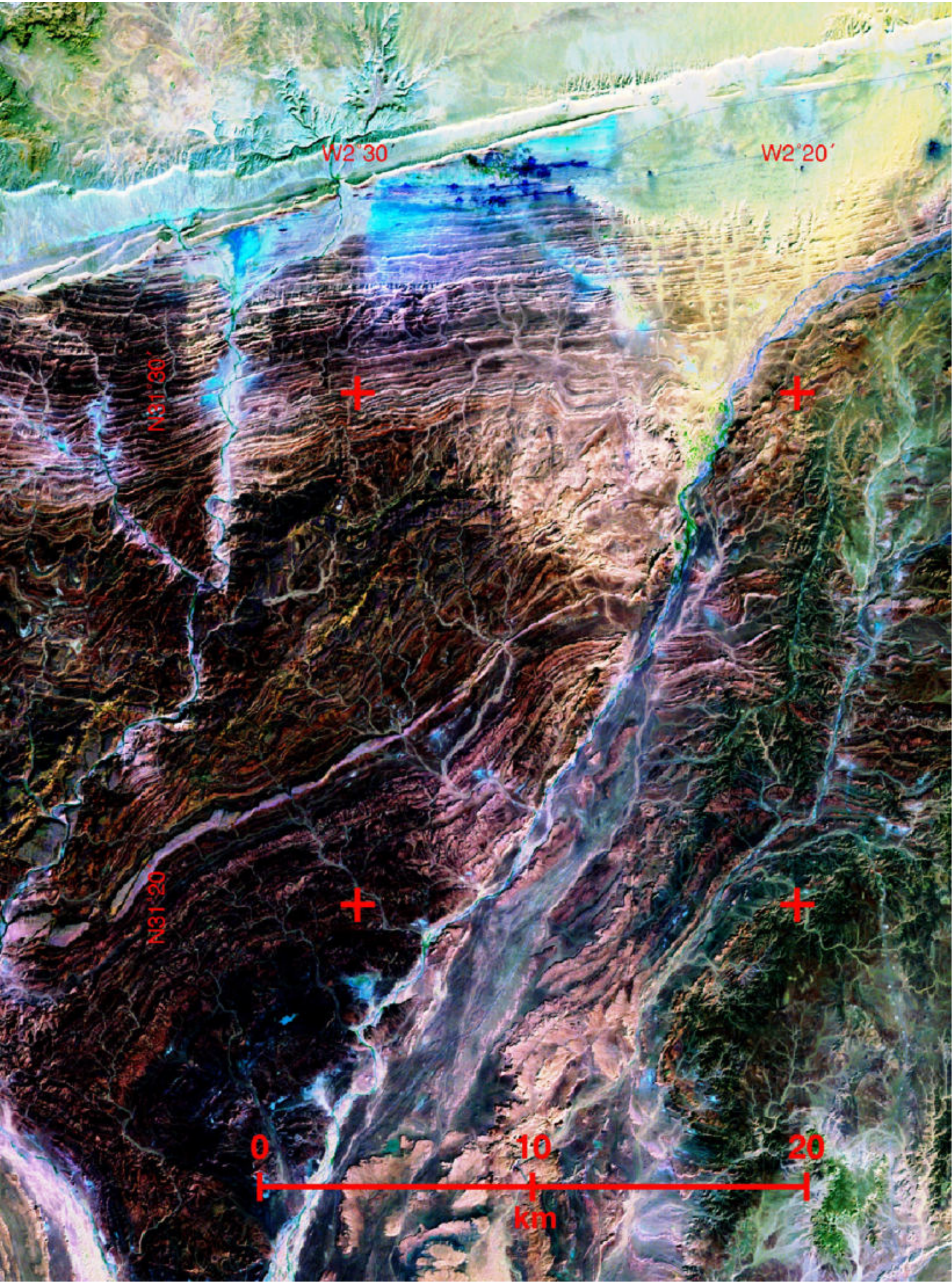

0
3
6
9
km
8°50
8°55
9°
34°15
34°10
55
40
15
15
Jebel Atra
Coniacian
Santonian
Maastrichtian
Mio-Pliocene
0
3
6
9
km

is an area referred to as Kasserine Island that is a stable region containing flat plains and plateaus separated by SW-NE trending anticlinal structures (Plate 2) (*Ferjani et al*., 1990).

Anti-Atlas

The Anti-Atlas region of Morocco is parallel to the Moroccan Atlantic coast in western Morocco, and then extends almost east-west south of Agadir and the High Atlas Mountains. These mountains are of low relief (Figure 1.1) compared to the High Atlas, and are characterized by folded Paleozoic and Precambrian rocks. The rocks of the Anti-Atlas were deformed by the Caledonian/Hercynian events, and these phases of deformation are markedly more dominant than the later Alpine deformation that uplifted the High Atlas Mountains (*Cahen et al*., 1984). The Anti-Atlas extends eastward into Algeria where it is referred to as the Ougarta belt linking the Hoggar in southern Algeria. The Anti-Atlas bounds the northern margin of the Paleozoic age Tindouf Basin. In the Anti-Atlas the Cenomanian-Turonian marine transgression was extensive and these age rocks were only slightly deformed in the later Alpine orogenic event (*Michard*, 1976) (Figure 1.12). The Cretaceous onlaps the Paleozoic along most of the northern margin of the Anti-Atlas. The Anti-Atlas Mountains of Morocco are separated from the High Atlas Mountains by the Souss and Ouarzazate basins (Plate 2). The Ouarzazate Basin was formed in the Tertiary and has generally less than 2 kilometers (*Schmidt*, 1992) of Teriary age sedimentary rocks.

Figure 1.12 The folded rocks of the Paleozoic (dark) are seen in this Landsat-TM image near the border of Morocco and Algeria. The northern area of this image shows the onlap of the Cretaceous onto the Paleozoic.


W2°30´
W2°20´
N31°30´
N31°20´
0
10
20
km

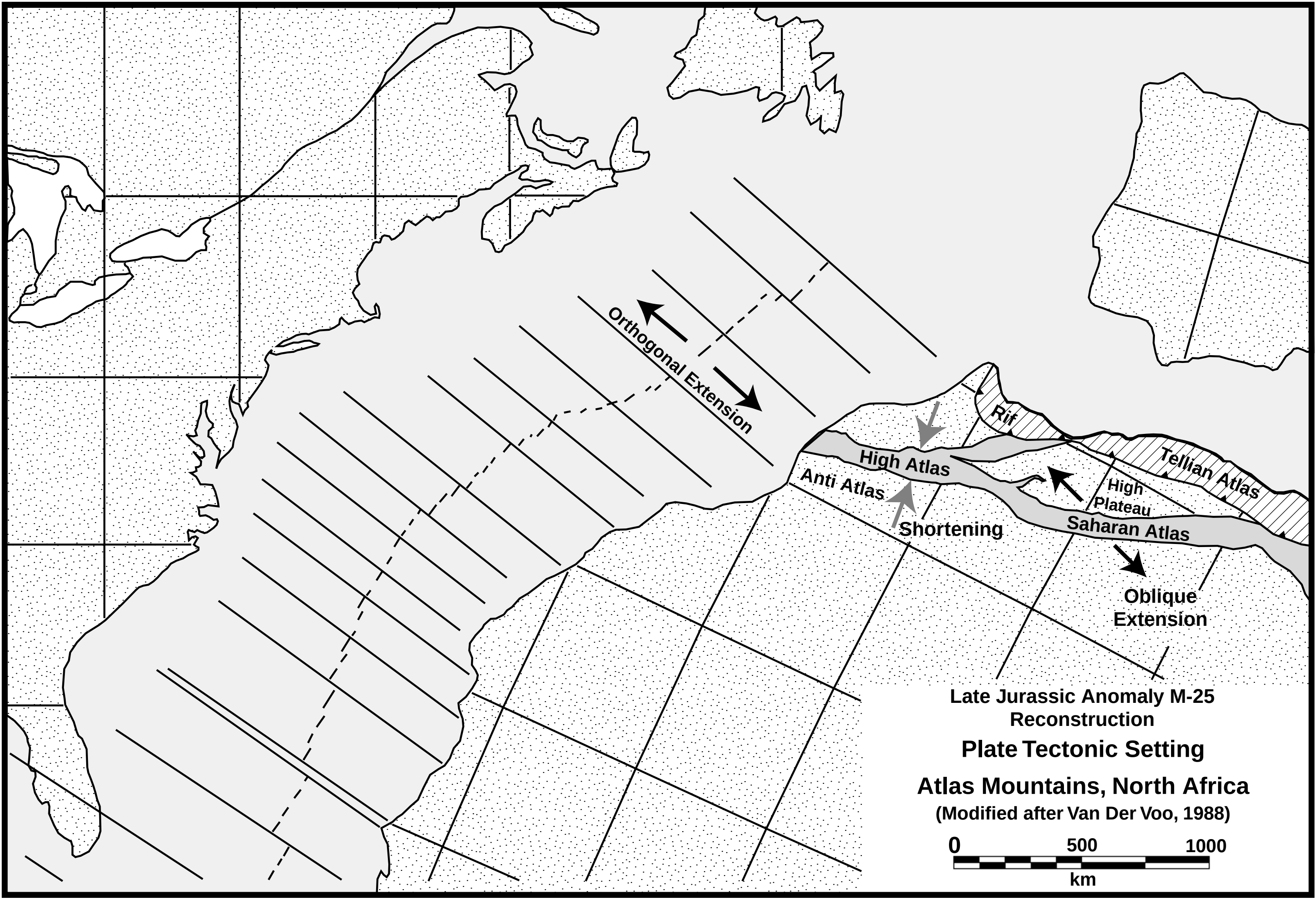
Discussion and Conclusions

The Atlas Mountains evolved from a large (>2000 km) intracontinental rift system. Evidence from geological, geophysical and remote sensing data indicates that rifting was oblique to the margins of the rift system over large regions of the Atlas rift system, with the exception, perhaps, of the Middle Atlas rift/mountains. The direction of extension was approximately NW-SE relative to the ENE-WSW trending Atlas Mountain belt (Figures 1.4, 1.5, 1.6, 1.7 and 1.9). Oblique rifting resulted in the formation of large rhombohedral pull-apart basins, fault ramps and en echelon faulting (Figures 1.5 and 1.9).

This NW-SE direction of extension is consistent with the direction of extension in the Atlantic based on paleomagnetic data (e.g., *Van Der Voo*, 1988). Extension was orthogonal to the plate boundaries of the present-day Atlantic margins (Figure 1.13). Rift basins affected by orthogonal extension are often characterized by long straight normal faults that result in large amounts of extension (crustal thinning) (*McClay and White*, 1995). Extension oblique to rift systems results in a smaller magnitude of extension across the rift system and the development of structural features such as *en echelon* faulting and pull-apart basins. It is proposed that orthogonal extension in the Atlantic rift basin during the Triassic-Jurassic period, resulted in oblique rifting in the Atlas rift system. Oblique extension relative to the Atlas rift system resulted in a slower rate of rifting than the orthogonal extension responsible for rifting in the Atlantic.

The successful opening of the Atlantic ocean was followed by uplift of the northwestern margin of the African plate. The post-rift phase of the Atlantic began in the Middle Jurassic (*Manzpeizer,* 1988) and was

Figure 1.13 Plate Tectonic setting of the Atlas Mountains, North Africa, during the Late Jurassic. Orthogonal extension in the Atlantic resulted in oblique extension relative to the Atlas rift system. Upon reactivation in the Cenozoic, the syn-rift structures of the Atlas rift system controlled the styles of deformation during inversion (Modified after *Van Der Voo*, 1988).



Orthogonal Extension
Rif
Tellian Atlas
High Atlas
Anti Atlas
High Plateau
Saharan Atlas
Shortening
Oblique Extension
Late Jurassic Anomaly M-25 Reconstruction
Plate Tectonic Setting
Atlas Mountains, North Africa
(Modified after Van Der Voo, 1988)
0
500
1000
km

accompanied by uplift and erosion of the onshore basins. This regional scale uplift resulted in the erosion or non-deposition of Upper Jurassic-Lower Cretaceous age strata over large areas of Morocco and the related Atlas rift basins (*Beauchamp et al*., 1996).

The structures formed by rifting in the Atlas rift basin greatly influenced the structural styles of deformation of the subsequently inverted Atlas Mountains of North Africa. Stresses generated by convergence along the African and European plates in the Tertiary were accommodated by the weakened crust of the Atlas rift basin hundreds of kilometers from the plate margins. *En echelon* folding and faulting in the Saharan Atlas and Tunisian Atlas (Figures 1.7 and 1.10) were likely to have been influenced by *en enchelon* faulting that resulted from oblique rifting (Figure 1.9). The structural complexity and characteristics of rift basins are inherited upon uplift and inversion. The previous structural geometries result in structural styles that are unlike those found in previously undeformed strata. Inversion of an intracontinental rift such as the Atlas may occur by layer-parallel shear in the undeformed post-rift sedimentary rocks. Vertical or inclined simple shear may approximate the deformation of the syn-rift strata that is characterized by pre-existing faults, previously defomormed rocks, and drastic changes in sedimentary thickness (*Beauchamp et al.,* 1997a).

The presence of evaporites in the Triassic age syn-rift sequence play an important role in the evolution of the Atlas Mountains, as anhydrites and saliferous rocks provide detachments between the syn-rift/post-rift sequence and pre-rift age rocks. Triassic strata thicken into the Atlas Mountains in both Morocco and Algeria (*Beauchamp*, 1988; *Assaad*, 1981). The presence of thicker evaporites in the rift basin is coincident with the thickening of the crust by thrust faults that propagate from detachments in the Triassic. The lack of deformation outside the Atlas Mountains may be related to the thinning and relative absence of Triassic rocks of evaporite facies that provide decollé ments.

Salt diapirism in the High Plateau of Morocco (Figure 1.6) and elsewhere in Algeria and Tunisia may have been influenced by the uplift of syn-rift normal faults as well as halokinesis. Salt related structures in Algeria were defined by *Assaad* (1981) as resulting from both halotectonic and halokinesis processes. Halotectonic structures exposed at the surface result from Triassic age anhydrite and saliferous shales in the hanging wall of thrusts, propagating from decollements in the Triassic during the Alpine orogeny (Oligocene-Miocene). Structures resulting from halokinesis evolved due to sediment loading and a contrast in density between Triassic age saliferous strata and younger rocks. These salt diapirs expose massive crystalline salts at the surface and are accompanied by complex fault systems (*Assaad*, 1981).

One of the most intriguing aspects of the Atlas Mountains is that the mountain belt evolved distant from margins of active plate convergence. Strain within the Atlas Mountains is partitioned vertically by levels of detachment in the upper crust within the Triassic evaporites, Cretaceous, and Paleozoic shales (Silurian) (Beauchamp et al., 1997b). The same is true across the mountain belt, as strain in many regions is concentrated along the margins of the orogen where the greatest magnitude of shortening is achieved by low-angle thrusting (*Beauchamp et al.,* 1997b).

There is a partitioning of strain across North Africa as well as within the Atlas Mountains. The Rif and Tellian Atlas Mountains make up the coastal ranges of North Africa (Figure 1.1), and are characterized by intense deformation and shortening related to the convergence of the African and European plates in the Cenozoic. These coastal tectonic regions are made up of southward verging, northward dipping thrust systems. Southward of these tectonic domains are large stable regions that are comparatively undeformed since the Mesozoic (e.g., Moroccan Meseta and High Plateau) (Figure 1.1). The Moroccan Meseta and High Plateau acted as rigid blocks that were affected only by regional uplift. These relatively stable regions were adjacent to the Atlas rift system that was weakened by the thinning and intense faulting of the upper crust in the Mesozoic. The rift systems within the North African plate were regions that would later become the templates for the focusing of strain produced from distant plate margin processes. The subsequent uplift, inversion and shortening of the Atlas rift basins resulted in the evolution of the present Atlas Mountains of North Africa.

References

Aoudia, A, and M. Meghraoui, Seismotectonics in the Tell Atlas of Algeria: the Cavaignac (Abou El Hassan) Earthquake of 25.08.1922, *Tectonophysics*, 248, 263-276, 1995.

Assaad, Fakhry A., A Further Geologic Study on the Triassic Formations of North Central Algeria With Special Emphasis on Halokinesis, *Journal of Petroleum Geology*, 4, 2, 163-176, 1981.

Beauchamp, J., Triassic sedimentation and rifting in the High Atlas (Morocco), in W. Manspeizer, ed., Triassic-Jurassic Rifting: *Elsevier*, 477-497, 1988.

Beauchamp, W., M. Barazangi, A. Demnati and M. El Alji, Inversion of synrift normal faults in the High Atlas Mountains, Morocco, *The Leading Edge*, 1171-1175, 1997a.

Beauchamp, W., Richard W. Allmendinger, Muawia Barazangi, A. Demnati and M. El Alji, Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological-geophysical transect, *Tectonics*, American Geophysical Union, (*in review*), 1997b.

Beauchamp, W., M. Barazangi, A. Demnati and M. El Alji, Intracontinental Rifting and Inversion: the Missour Basin and Atlas Mountains of Morocco, *American Association of Petroleum Geologists Bulletin*, 80/9,1459-1482, 1996

Ben Ayed, N. Interpretation structurale de la Tunisie atlasique, *Comptes Rendus des Seances de l'Academie des Sciences, Serie 2: Mecanique, Physique, Chimie, Sciences de la Terre, Sciences de l'univers*, 292,21 1445-1448, 1981.

Ben Ayed, N. and Faoud Zargouni, Carte Seismotectonic de la Tunisie, 1:100,000, Republique Tunisienne, *Ministere de 'Education de l'Enseignement superior et de l'a Recherche Scientifique Fondation Nationale de la Recherche Sceintifique*, 1990.

Ben Dhia, Hamed, Thermal Regime and Hydrodynamics in Tunisia and Algeria, *Geophysics*, 56, 7,193-1102, 1991.

Ben Ferjani, Ali, Pierre F. Burollet and Fathia Mejri, Petroleum Geology of Tunisia, *Enterprise Tunisienne dÌActivites Petrolieres*, Tunis: 1990.

Betier, M. G., L. Royer, H. Termier, R. Laffitte, Carte Geologique de Tocqueville, 1:50,000, *Le Service de la Carte Geologiquede l'Algerie*, 1953.

Bishop, William F., Geology of Tunisia and Adjacent Parts of Algeria and Libya, *American Association of Petroleum Geologists Bulletin*, 59, 3, 413-450, 1975.

Bobier, C., Viguier, C., A. Chaari and A. Chine, The post-Triassic Sedimentary Cover of Tunisia: Seismic Sequences and Structure, *Tectonophysics*,195, 371-410, 1991.

Boccaletti, M., G. Cello and L. Tortorici, Structure and tectonic significance of the north-south axis of Tunisia, *Annales Tectonicae*, II, 1,12-20, 1988.

Buforn, E., C. Sanz de Galdeano, and A. Udias, Seismotectonics of the Ibero-Maghrebian region, *Tectonophysics*, 248, 3-4, 247-261, 1995.

Cahen, L., N.J. Snelling, L. J.R. Vail, M. Bonhomme, D. Ledent, *The geochronology and evolution of Africa*, Clarendon Press, Oxford, 333-348, 1984.

Caire, A., Eastern Atlas, in A.M. Spencer (ed.), Mesozoic-cenozoic orogenic belts, *Geological Society Special Publication,* 4, 4-59, 1974.

Carte Geologique Du Maroc, 1:1,000,000, Service Geologique du Maroc, 1985.

Chefchaouni, Y.C., M. Diouri and G. Choubert, Carte Geologique du Haut-Atlas Oriental,1:200,000, *Service Geologique du Maroc*, 1963.

Chine, Ali and Khemais Ben Hassine, preparers, *Proceedings of the 4th Tunisian Petroleum Exploration Conference, Tunis*, May 1994.

Chine, Ali and Khemais Ben Hassine, Actes des Iieme Journees de Geologie Tunisienne ,Appliquee a la Recherche des Hydrocarbures, *Memoires de lÌEntreprise dÌActivites Petrolieres*, No. 3, 1989.

Choubert, G., G. Dubar and J. Hindermeyer, Carte Geologique du Haut Atlas, 1:200,000, *Service Geologique du Maroc*, 1956.

Crevallo, P.D., J. E. Warme, M Septfontaine, R. Burke, Evolution of Jurassic carbonate platforms in an active transtensional rift, *American Association of Petroleum Geologists Bulletin*, 71/5, 543-544, 1987.

Dziewonski, A.M., G. Ekstrom and M.P. Salganik, Centroid-moment tensor solutions for April-June 1994, *Physics of the Earth and Planetary Interiors*, 90, 1-2, 1-11, Amsterdam: 1995.

Ennadifi, Y, A. Boudda, and R. Dresnay, Carte Geologique du Haut Atlas DÌAnoual-BouAnane (Haut Atlas Oriental), 1:200,000, *Service Geologique du Maroc*, 1976.

Faure-Murait, A., J.L. Morel, M. Dahmani, Carte Neotectonique du Maroc, 1:100,000, Direction de la Geologie, Ministere de L'Energie et des Mines, Rabat, 1994.

Fetah, M., M. Bensaid, M. Dahmani, Carte Géologique Du Maroc-Zawyat Ahancal, 1:100,000, Royaume Du Maroc, *Ministere De L'Energie Et Des Mines*, 355, 1990.

Ghissassi, A., M. Saadi, T. Skalli, E. A. Hilali, A. Boudda, Carte Géologique Du Maroc-Afourar, 1:100,000*, Royaume Du Maroc, Ministere De L'Energie Et Des Mines*, 247, 1977.

Gomez, Franscisco , Muawia Barazangi and Mohammed Bensaid, Active tectonism in the Intracontinental Middle Atlas Mountians of Morocco:synchronous crustal shortening and extension, *Journal of the Geological Society*, London, 153, 389-402, 1996.

Galindo-Zindivar, J., F. Gonzales-Lodeiro and A. Jabaloy, Stress and palaeostress in the Betic-Rif cordilleras (Miocene to the present), *Tectonophysics*, 227, 1-4, 105-126,1993.

Grimison, N. L., The Azores-Gibraltar plate boundary; focal mechanisms, depths of earthquakes, and their tectonic implications, *Journal of Geophysical Research*, 91, 2, 2029-2047, 1986.

Jenny, J., Carte Géologique du Maroc au 1/100,000 Feuille Azillal (High Atlas central), *Notes Mémoires du Service Géologique*, Rabat, 339,104 pp., 1988.

Kasser, M., Ruegg, J.C., and J.C. Lepine, Deformations Associated with the El Asnam Earthquake of October 10, 1980, *Journal of Geodynamics*, 7, 215-219, 1987.

Kiratzi, A., Active crustal deformation from the Azores triple junction to the Middle East, *Tectonophysics*, 243, 1-2, 1-24, 1995.

Le Marrec, Alain, Carte Geologique du Maroc, 1:100,000., *Service de la Carte Geologique*, 1981.

Manspeizer, W., Triassic-Jurassic Rifting and Opening of the Atlantic: A Overview, in W. Manspeizer ed., *Triassic-Jurassic Rifting, Elsevier Press*, Part A, 41-72, 1988.

McClay, K.R. and M. J. White, Analogue modelling of orthogonal and oblique rifting, *Marine and Petroleum Geology,* 12, 137-151, 1995.

Medina, Fida, Present-day state of stress in northern Morocco from focal mechanism analysis, *Journal of Structural Geology*, 17, 7, 1035-1046, 1995.

Meghraoui, M., Seismotectonics of the Lower Cheliff Basin: Structural Background of the el Asnam (Algeria) Earthquake, *Tectonics*, 5, 6, 809-836, 1986.

Meghraoui, M., Blind Reverse Faulting System Associated With the Mont Chenoua-Tipaza Earthquake of 29 October 1989 (north-central Algeria), *Terra Nova*, 3, 84-93, 1990.

Michard, A., Eléments de géologie Marocaine, *Notes Mémoires du Service Géologique*, Rabat, 252,pp. 408, 1976.

Petters, Sunday W., Regional geology of Africa, Lecture Notes in Earth Sciences, *Springer-Verlag*,40, 532-552, 1991.

Rolley, J.-P., Carte Geologique du Maroc au 1/100,000 Feuille Afourer (High Atlas central), *Notes Mémoires du Service Géologique*, Rabat, 247, 103 pp., 1978.

Saadi, M., E. A. Hilali, M. Bensaid, Carte Géologique du Maroc-Jebel Saghro-Dadés, 1:200,000*, Royaume du Maroc, Ministere de l'Energie et des Mines*, 161, 1977.

Saadi, M., M. A. Hilali, A. Demnati, Carte Magnetique du Maroc, Interpretation du Leve Aeromagnetique, Bechar and Marrakech, 1:1,000,000, *Royaume du Maroc, Ministere de l'Energie et des Mines*, 320-C, 1979.

Saadi, M., E.M. Hilali and A. Boudda, Carte Geologique de la Chaine Rifaine, 1:500,000, *Service de la Carte Geologique*, 1980.

Saadi, M., M. Bensaid, M. Dahmani, Carte Géologique du Maroc-Azilal, 1:100,000*, Royaume du Maroc, Ministere de l'Energie et des Mines*, 339, 1985.

Schaer, Jean-Paul, and Rodgers, J., Evolution and Structure of the High Atlas of Morocco. In: Jean-Paul Schaer, and J. Rodgers, eds., The Anatomy of Mountain Ranges: *Princeton University Press*, 107-127, 1987

Schmidt, K.-H., The tectonic history of the Pre-Saharan depression (Morocco)-a geomorphological interpretation, *Geologische Rundschau*, 81/1, 211-219, 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

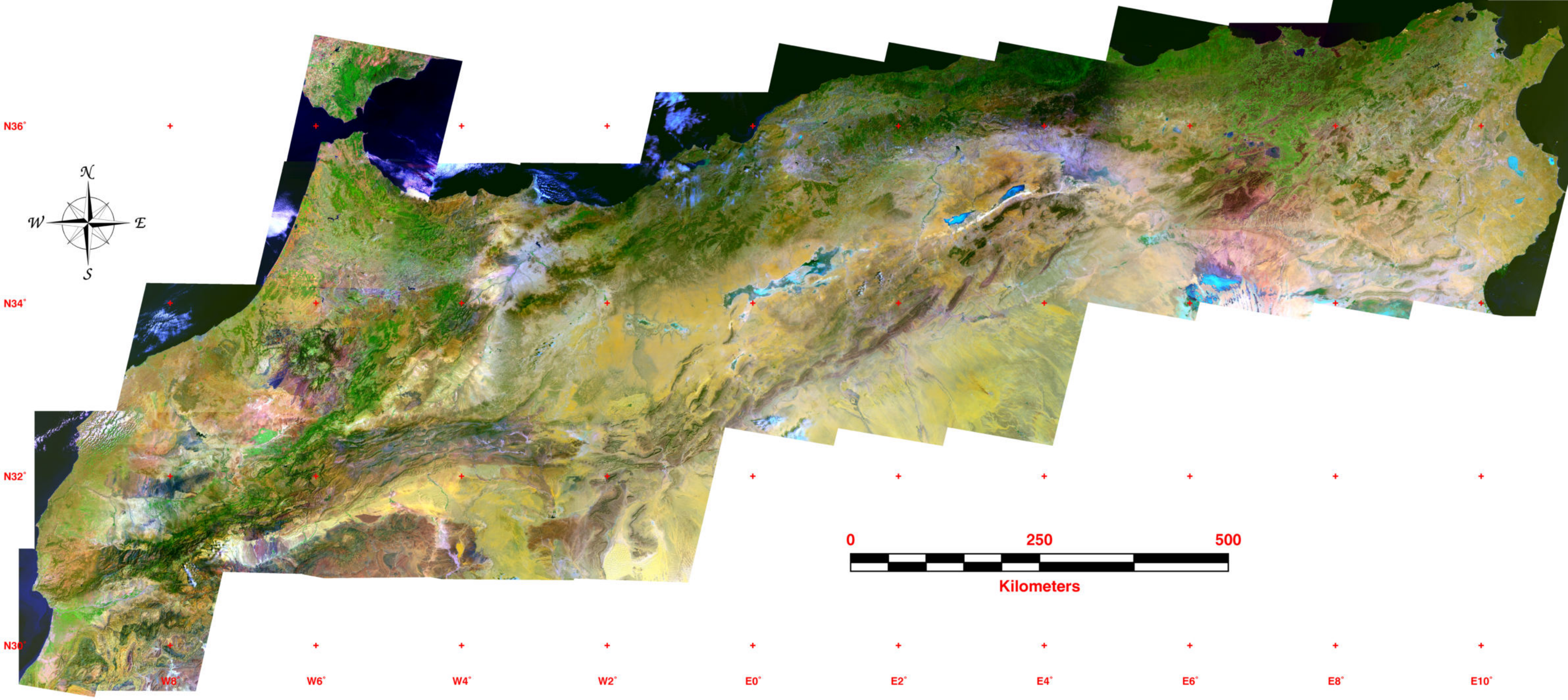
Swan, F.H., Temporal Clustering of Paleoseismic Events on the Oued Fodda Fault, Algeria, *Geology*, Vol. 16, 1092-1095, 1988.

Tadili, B.A., La sismicite du Moyen Atlas, *Gerlands Beitraege zur Geophysik*, 95, 6,469-476, 1986.

Udias, A., L. Arroyo and J. Mezcua, Seismotectonics of the Azores-Alboran region,*Tectonophysics*, 31, 3-4, 259-289, 1976.

Van Der Voo, R. Triasic-Jurassic Plate Migrations and Paleogeographic Reconstructions in the Atlantic Domain, in W. Manspeizer ed., *Triassic-Jurassic Rifting, Elsevier Press*, Part A, 29-38, 1988.

Tectonic Map of the Atlas Mountains
North Africa
Weldon Beauchamp, Cornell University-Institute for the Study of the Continents
Mediterranean Sea
Atlantic Ocean
Alboran Sea
Portugal
Spain
Betic
Morocco
Algeria
Tunisia
Libya
Strike-Slip earthquake
Thrust earthquake
Normal earthquake
Anticline
Syncline
Fault
Country Border
Neogene/Quaternary Volcanics
Cenozoic Basin
Rif Tectonic Domain
Atlas Tectonic Domain
Paleozoic
Precambrian Basement
Paleozoic Basin
Cenozoic Igneous and Metamorphic Rocks
Cheliff Basin
High Plateau
Hodna Basin
Melrhir Trough
Chott El Jerid
Sillon Des Chotts
Benoud Trough
Hassi Messaoud Ridge
Hassi Allal Arch
Ougarta Ridge
Ghadames Basin
Tindouf Basin
Bechar Basin
Anti Atlas
Ouarzazate Basin
Souss Basin
Haouz Basin
Tadla Basin
Meseta
Missour Basin
Guercif Basin
Rharb Basin
Sahel
NOSA Fold Belt
Gulf of Tunis
Cap Bon
Gulf of Hammamet
Gulf of Gabes
Sevilla
Granada
Malaga
Cadiz
Almeria
Aguilas
Cartagena
Tanger
Rabat
Casablanca
Meknes
Khenifra
Midelt
Boujad
Safi
Marrakech
Essaouira
Agadir
Tarhjicht
Oujda
Tlemcen
Sidi-Bel Abbes
Oran
Mostaganem
Alger
Blida
Bejaia
Skikda
Annaba
Constantine
Bizerte
Tunis
Mah
Sfax
Gafsa
Tozeur
El Oued
Touggourt
Laghouat
Figuig
Bechar
N
S
E
W
35
30
10
5
0
5
10


N36°
N34°
N32°
N30°
N
W
E
S
W8°
W6°
W4°
W2°
E0°
E2°
E4°
E6°
E8°
E10°
0
250
500
Kilometers

CHAPTER TWO

Intracontinental Rifting and Inversion: the Missour Basin and Atlas Mountains of Morocco[1]

Abstract

The intracontinental High and Middle Atlas mountain belts in Morocco intersect to form the southern and western margins of the Missour basin, an intermontane basin, formed as a result of the uplift and inversion of the Mesozoic Atlas paleo-rifts. These rifts were areas where the crust was greatly attenuated and more subject to deformation in response to nearby plate boundary tectonic events. Observations based on seismic reflection profiles and wells collected over the Missour basin for hydrocarbon exploration, and field mapping were utilized to understand the basin evolution, structural styles and timing of inversion of the nearby Atlas mountains. Hercynian and Mesozoic normal faults were reactivated into high-angle reverse and thrust faults in the Mesozoic during the Jurassic, Early Cretaceous (Early Alpine phase), and the Paleogene (Late Alpine phase). The reactivation of syn-rift normal faults of the paleo-Atlas rifts inverted previous half grabens into anticlinal structures, with the axis of the half graben centered below the axis of the inverted anticline. The resulting inverted fold geometries are controlled by the geometries of the extensional planar or listric faults.

---

[1] Published as: Weldon Beauchamp, Muawia Barazangi, Ahmed Demnati and Mohamed El Alji, Intracontintental Rifting and Inversion: the Missour Basin and Atlas Mountains, Morocco, American Association of Petroleum Geologists Bulletin, 80, 9, 1459-82, 1996.

The Atlas paleo-rift system is one of the largest rift systems in Africa. There has been little hydrocarbon exploration within the Atlas mountains and the margins of the paleo-Atlas rift system. Inversion of syn-rift structures can lead to both the destruction and preservation of syn-rift traps and the creation of new hydrocarbon traps. The study of the effects of inversion in the Missour basin may lead to the discovery of footwall subthrust hydrocarbon traps in the Mesozoic sedimentary sequence of the Atlas mountains.

Introduction

Mountain belts located along convergent plate boundaries, such as the Andes or the Himalayas, have been and are still the focus of intense geological and geophysical studies. In contrast, intracontinental mountain belts, including the Atlas system in Morocco, lack even an agreed upon first-order conceptual model of their deep structure and active deformation. Geological evidence suggests such intraplate belts significantly contributed to the evolution of the continental lithosphere since Precambrian time.

Rifting during the Triassic and Jurassic was widespread around the world. The Atlas rift system of North Africa, the North Sea rift, the Andean rift system of Colombia and Venezuela and the Palmyrid rift of Syria, are just a few of the intracontinental rift systems active during the Triassic and Jurassic periods. Some of these same rift systems were inverted into intracontinental mountain belts (i.e., the Atlas mountains, Palmyrid mountains and the northern Andes). These rift systems were the focus of sedimentation during the syn-rift and post-rift phases of rifting. Rift basins contain approximately 5% of the worldÌs sedimentary volume, but they also contain 10-29% of the known hydrocarbon reserve base (~275 billion barrels) (*Katz,* 1995). This high concentration of reserves is partially due to the limited migration distance allowed by the geometries of rift systems.

The uplift and inversion of hydrocarbon bearing rifts can result in the remigration and redistribution of hydrocarbons into structures generated by the reactivation of pre-existing faults formed during rifting. It is important to have a good understanding of the geometry of structures formed by the reactivation of syn-rift faults, as these structures have the potential to trap significant amounts of hydrocarbons. It is necessary to generate models for the development of individual inverted structures, and the uplift and inversion of entire rift systems such as the Atlas mountains, to better understand the potential of unexplored intracontinental rifts and mountain belts around the world. Our research in Morocco is a step in understanding and resolving the history and present architecture of such belts.

Geological Setting

The Mesozoic and Cenozoic geological evolution of Morocco can be viewed as a response to two major geological events: (1) The opening of the North Atlantic and the Western Tethys in the early Mesozoic, and (2) the Africa-Europe continental collision in mid-Cenozoic time (*Michard*, 1976; *Mattauer et al.*, 1977; *Bensaid et al.*, 1985; *Pique et al.*, 1987; *Jacobshagen et al.*, 1988b; *Dewey et al.*, 1989; *Westaway*, 1990). These two major events have shaped the present architecture of the four major geological structures of Morocco: The Rif fold-thrust mountain belt in the north, the Middle Atlas and High Atlas, and Anti-Atlas mountain belts of central Morocco (Figure 2.1). The Rif belt is fundamentally different than the Atlas system: The Rif is an asymmetric, Alpine-type, fold-thrust belt with numerous, well-mapped thrusts and complex nappe structures (*Loomis*, 1975; *Leblanc and Olivier*, 1984; *Morley*, 1987; *Doblas and Oyarzun*, 1989; *Ait Brahim and Chotin*, 1990; *Leblanc*, 1990; *Miranda et al.*, 1991), while the Atlas system is an intracontinental, largely symmetrical mountain belt.

Regional Tectonics and Rifting

The Atlas system evolved within the stable platform of North Africa. Two major events shaped the geological evolution of the system: Early Mesozoic extension and rifting, and Mesozoic through Cenozoic compressional-transpressional phases resulted in the inversion of the rift systems (Figure 2.2). The Atlas system is thus an intracontinental orogen that is "sandwiched" within the Proterozoic-Paleozoic northern African platform, and is fundamentally different from orogens located along convergent/collisional plate boundaries. Thrusts, strike-slip faults, and block uplift tectonics characterize the Cenozoic deformation of the Atlas system (e.g., *Schaer and Rogers*, 1987; *Dresnay*, 1988; *Fraissinet et al.*, 1988; *Jacobshagen et al.*, 1988a; *Medina*, 1988; *Giese and Jacobshagen*, 1992; *Jacobshagen*, 1992).

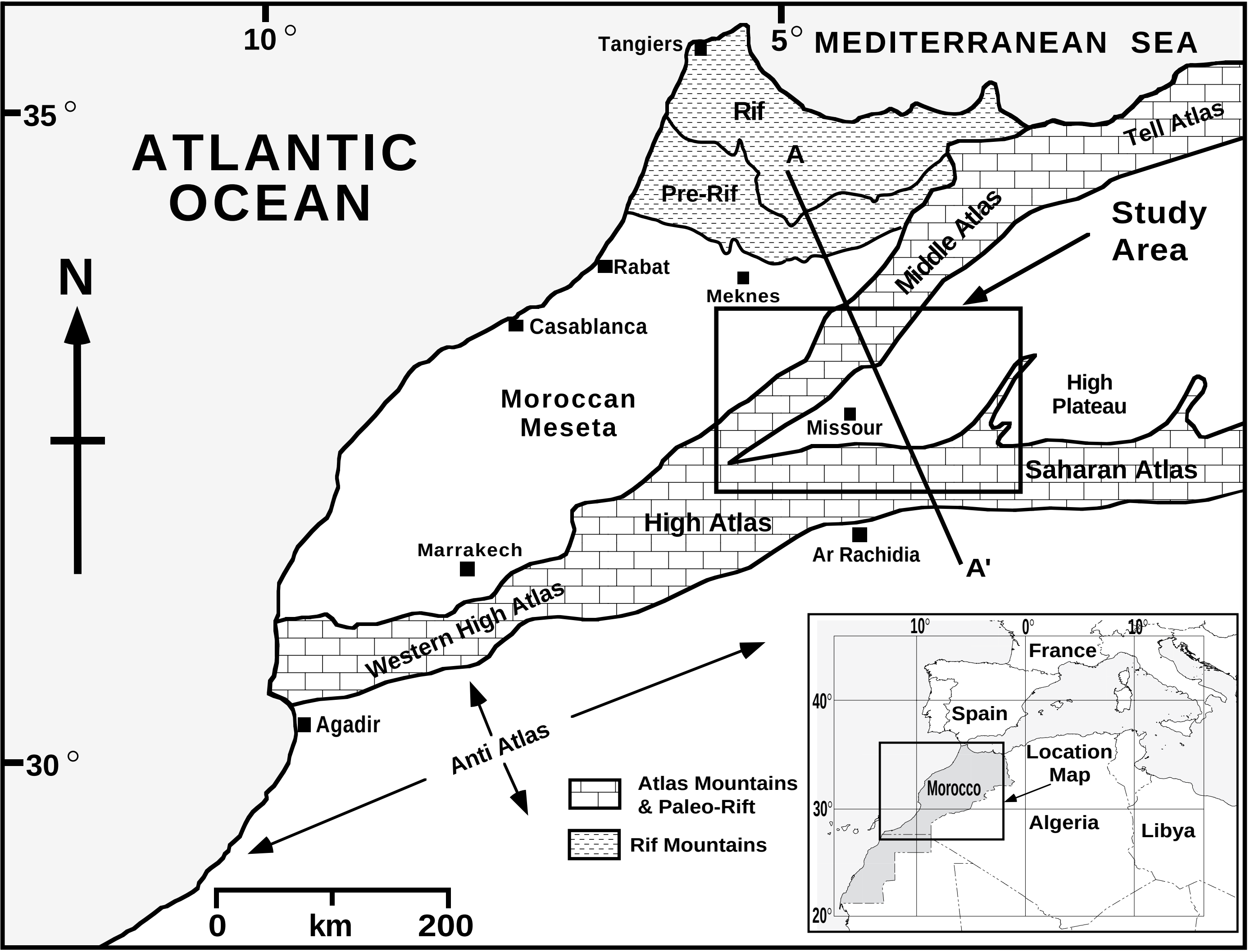
The Missour basin and the High and Middle Atlas mountain belts that form its boundaries (Figure 2.1) are examples of how large stresses can be transmitted to intraplate zones of weakness from the collision zones along the nearby plate margins. High strain rates created by the thinning of the continental lithosphere resulted in the deformation of the crust by extension and rifting in the North African plate at the end of the Permian and beginning of the Triassic (*Brede et al.*, 1992). Sedimentation rates accelerated through the Jurassic, as rifting continued with the breakup of Pangea, and the opening of the Neo-Tethys Ocean and the North Atlantic (*Ziegler*, 1982). The High Atlas developed into one of the largest of the rifts, possibly reactivated along existing weaknesses and faults formed during the Hercynian orogeny.

The High Atlas rift extends to the Atlantic margin where it forms a failed rift or aulacogen, and eastward (High Atlas/Saharan rift) across Morocco, Algeria and Tunisia (Figure 2.1). The Middle Atlas rift and mountains trend northeast, where they extend beneath the thrusted Alpine Rif allochthonous sedimentary rocks. The intersection of the Middle and High Atlas rifts/mountains may represent a failed triple junction, or a focus of thermal upwelling.

Beginning in the Late Cretaceous-Early Oligocene, dextral movement on the Newfoundland-Gibraltar fault zone increased the eastward drift of the Iberian plate (e.g., *Brede et al.*, 1992). The geometric relationship between the Iberian plate and the African plate resulted in compressional stresses that were transferred to the North African rift systems. The bulk of the intraplate stresses were absorbed by the High and Middle Atlas rift systems, resulting in shortening and subsequent inversion. The inversion of these rifts led to the reactivation of pre-existing Mesozoic and Hercynian faults into reverse and thrust faults, with an oblique-slip sense of movement. The uplift of the Middle and High Atlas rifts formed the mountains that are now the boundaries of the Missour basin (Figure 2.2).

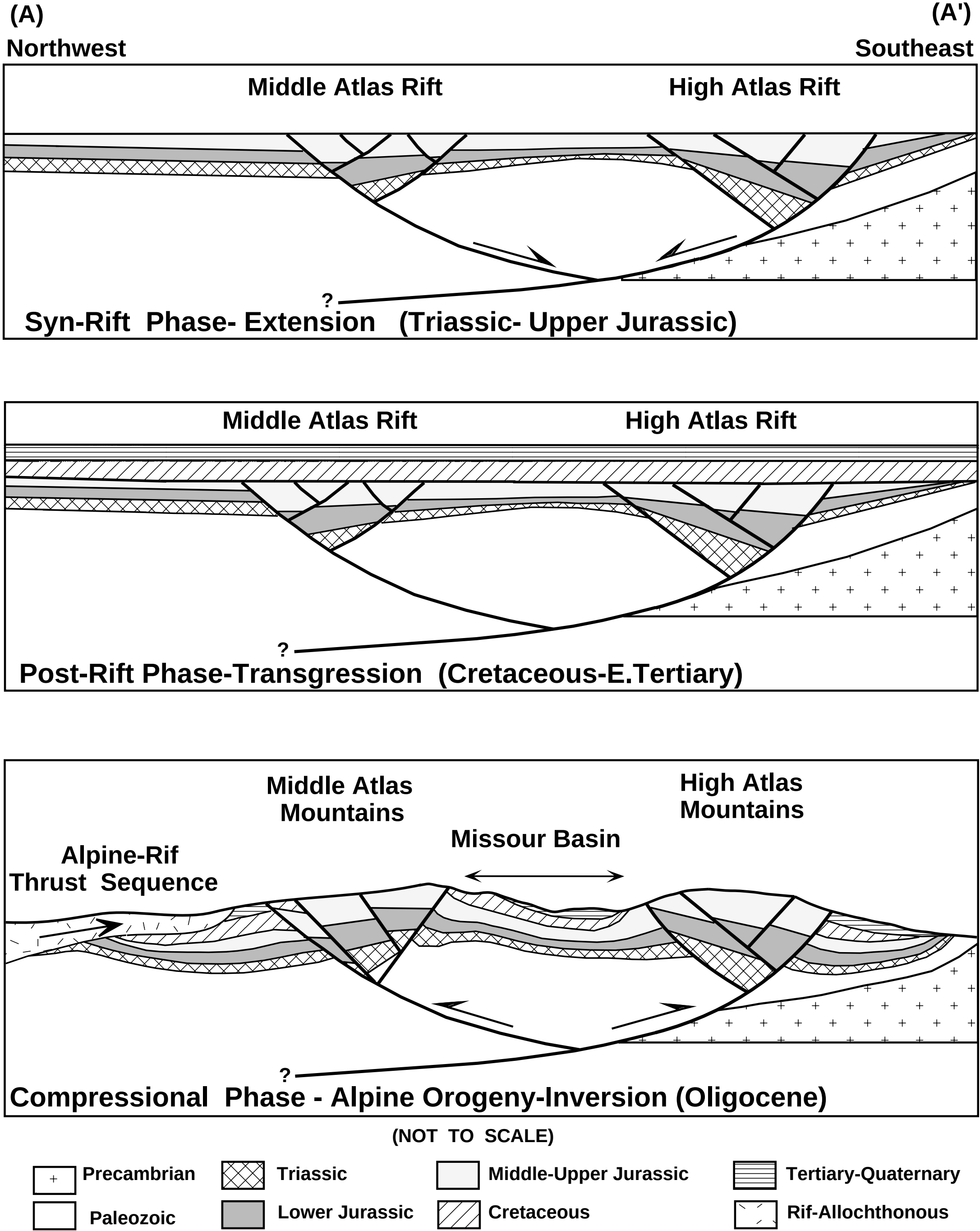
The orientation of compressional stresses relative to the orientation of rift bounding faults, resulted in transpressional deformation in both a dextral and sinistral sense (e.g., *Giese and Jacobshagen*, 1992). The sense of movement on the bounding faults of the High and Middle Atlas mountains varied depending upon the direction of plate motion between the European and African plates. Early Jurassic time was the beginning of spreading in the

Figure 2.1 Location map of the Atlas mountains and Missour basin of Morocco. The Missour basin is bounded by the Middle Atlas and High Atlas mountains.



10 °
5 °
MEDITERRANEAN SEA
35 °
30 °
ATLANTIC OCEAN
Tangiers
Rif
Pre-Rif
Tell Atlas
A
A'
Middle Atlas
Study Area
Rabat
Meknes
Casablanca
Moroccan Meseta
High Plateau
Missour
Saharan Atlas
High Atlas
Marrakech
Ar Rachidia
Western High Atlas
Agadir
Anti Atlas
N
Atlas Mountains & Paleo-Rift
Rif Mountains
0
km
200
10°
0°
10°
40°
30°
20°
France
Spain
Location Map
Morocco
Algeria
Libya

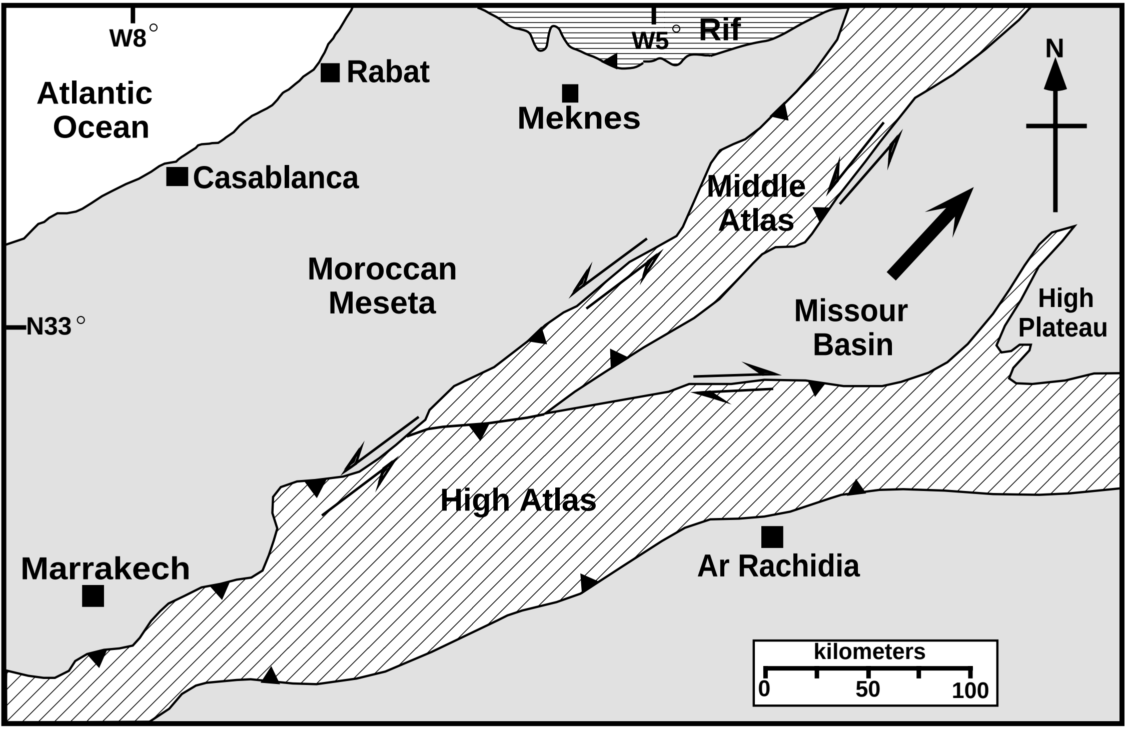
Figure 2.2 Conceptual model for the development of the Missour basin and the Atlas mountains, Morocco. See Figure 2.1 for location of A-A’.


(A)
Northwest
(A')
Southeast
Middle Atlas Rift
High Atlas Rift
?
Syn-Rift Phase- Extension (Triassic- Upper Jurassic)
Middle Atlas Rift
High Atlas Rift
?
Post-Rift Phase-Transgression (Cretaceous-E.Tertiary)
Middle Atlas Mountains
High Atlas Mountains
Missour Basin
Alpine-Rif Thrust Sequence
?
Compressional Phase - Alpine Orogeny-Inversion (Oligocene)
(NOT TO SCALE)
Precambrian
Paleozoic
Triassic
Lower Jurassic
Middle-Upper Jurassic
Cretaceous
Tertiary-Quaternary
Rif-Allochthonous

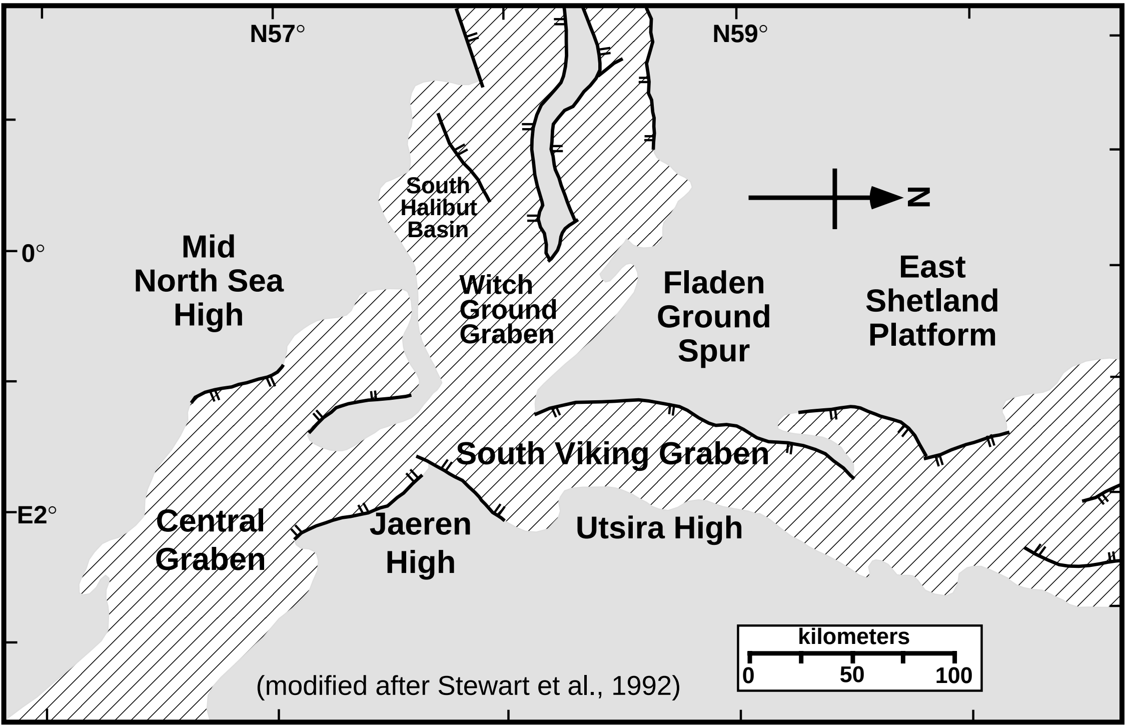
central Atlantic, while the north Atlantic was in a rifting stage. This resulted in an eastward drift of the African plate in relation to the Iberian plate that was to the north of the Newfoundland-Gibraltar transform (*Ziegler*, 1982). During the Late Cretaceous-Early Oligocene, the African plate was moving to the east with a clockwise sense of motion, resulting in transtensional deformation in the North African rift systems. Plate rotation of the Iberian plate occurred in a counterclockwise manner as the Iberian plate moved eastward along the Newfoundland-Gibraltar fault/transform. This counterclockwise rotation is evident in the opening of the Bay of Biscay along the northern margin of the Iberian plate. The African plate was moving northward with respect to the Iberian plate during the late Miocene and Pliocene (*Ziegler*, 1982), with the same counterclockwise sense of rotation, resulting in a rotation of primary stresses from approximately 180° to 120°. Faults bounding the High Atlas rift yielded a right-lateral transpressive sense of deformation during the Late Cretaceous-Early Oligocene (Figure 2.3). Folds in the Central High Atlas generally trend east-northeast, at about a 20-30° angle to the High Atlas bounding faults (*Studer and du Dresnay*, 1980), further evidence of a right-lateral phase of deformation in the High Atlas mountains. Later plate motion during the Oligocene to the Recent has been that of convergence (Betic-Rif orogen) with some wrenching, as both plates have drifted eastward at a similar rate (e.g., *Dewey et al.*, 1989).

For comparison, the High Atlas rift system is similar in size and geometry to the North Sea rift system (Figure 2.3). The actual geometry between individual rifts varies between the two rift systems. These differences may be related to the rates and direction of plate convergence between the

Figure 2.3 Comparison of the size and geometries of the Atlas and North Sea rift systems. Both rift systems were active during the Jurassic. The Missour basin (A) was a shelf margin much the same as the Fladen Ground Spur (B) of the North Sea. The Missour basin is bounded by faults that have a right-lateral component of slip to the south in the High Atlas, and by faults with a left-lateral component of slip to the west in the Middle Atlas. The result is the uplift and escape of the Missour basin to the NE.


W8°
W5°
Rif
N
Rabat
Atlantic
Ocean
Meknes
Casablanca
Middle
Atlas
Moroccan
Meseta
N33°
Missour
Basin
High
Plateau
High Atlas
Marrakech
Ar Rachidia
kilometers
0
50
100

A)


N57°
N59°
N
South
Halibut
Basin
0°
Mid
North Sea
High
Witch
Ground
Graben
Fladen
Ground
Spur
East
Shetland
Platform
South Viking Graben
E2°
Central
Graben
Jaeren
High
Utsira High
kilometers
0
50
100
(modified after Stewart et al., 1992)

B)

Iberian plate and the African plate, and the subsequent deformation. Extension began in the North Sea and the Atlas rift systems during the Triassic and Jurassic periods (Figure 2.3). The Triassic sedimentary rocks penetrated by wells in the Missour basin contain a significant amount of tholeiitic volcanics interbedded with salts (Figure 2.4). These Triassic basalts extend over most of the Missour basin, are encountered in several wells, and are clearly identifiable on seismic reflection data.

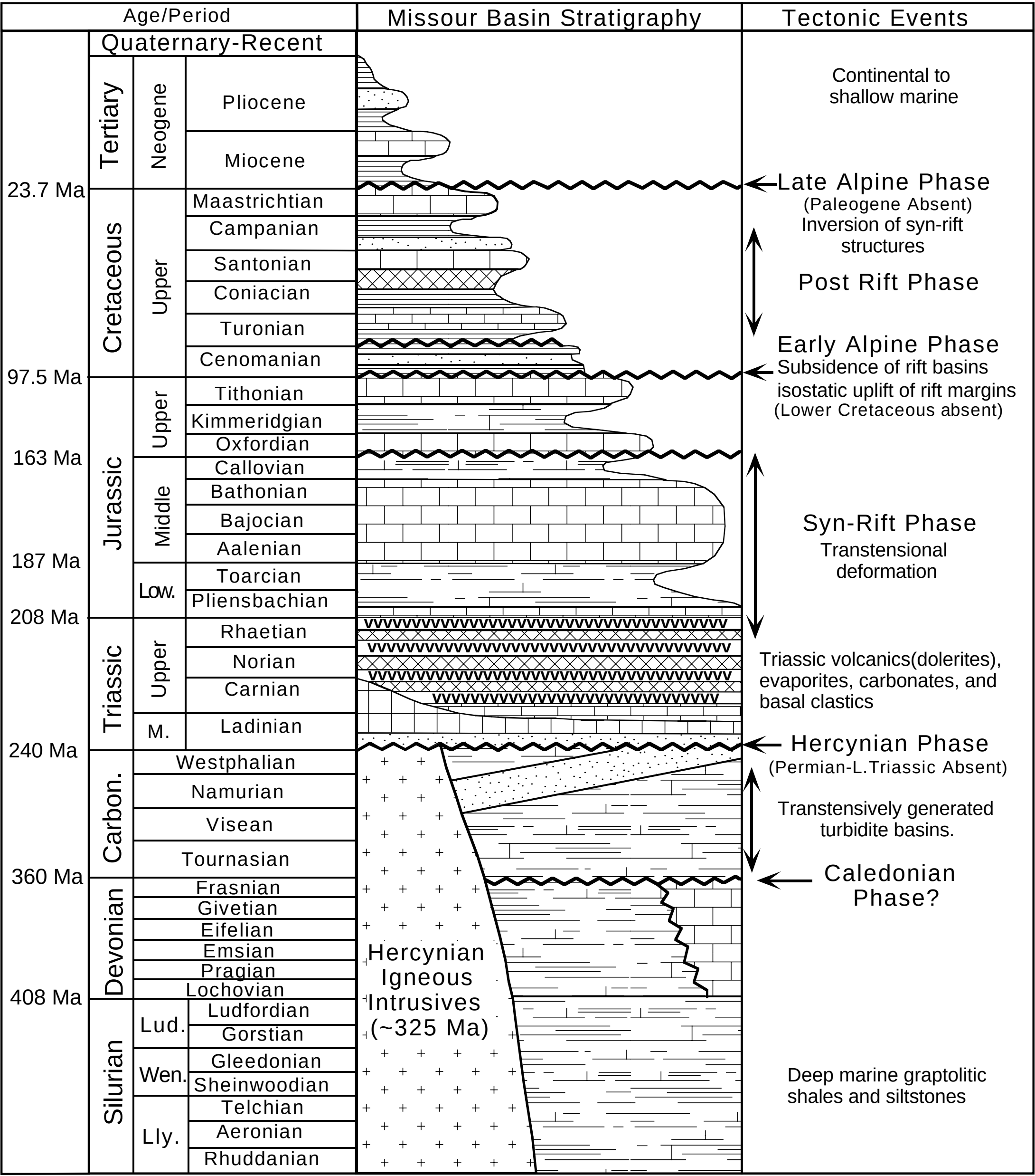
Rifting in the Atlas continued into the Middle Jurassic when subsidence continued over the rift systems during the Late Jurassic to Early Cretaceous. Lower Cretaceous sedimentary rocks are not generally preserved in the Atlas mountains, but may have been deposited in the Atlas rift systems during a post-rift subsidence phase; if so, record of these sedimentary rocks was removed by uplift and erosion. Subsidence in the Atlas rift basins was probably coupled with isostatic uplift of the adjacent platform margins (Missour basin, High Plateau and Moroccan Meseta). This isostatic uplift of the rift basin margins resulted in a thinning of the Upper Jurassic to Lower Cretaceous sedimentary rocks. The Upper Jurassic to Middle Jurassic sedimentary rocks of the Missour basin are deeply eroded along the margins of the High and Middle Atlas mountains, a result of isostatic uplift and erosion by the base Cretaceous unconformity. Well data illustrate deep truncation by the base Cretaceous unconformity into the syn-rift sedimentary rocks from the rift margins into the paleo-Atlas rift basins. This is evidence of isostatic uplift of the rift basin margins during the Late Jurassic and/or Early Cretaceous (Figure 2.4). The oldest Cretaceous sedimentary rocks encountered in the Missour basin are Cenomanian (Figure 2.4). This would suggest a subsidence phase in the Atlas rift systems that lasted 40-50 my from the Late

Jurassic to the Late Cretaceous. Subsidence slowed or ended during the Late Cretaceous, as Cenomanian to Turonian sedimentary rocks were deposited uniformly across the rift systems and the rift basin margins. Subsequent uplift in the Paleogene related to the Alpine orogeny, inverted the Atlas rift system and eroded the Lower Cretaceous sedimentary rock sequence from the present Atlas mountains. Compression and transcurrent movements generated by the relative motion of the African and Iberian plates resulted in stresses being transmitted into the African plate, the net result being the shortening and inversion of the Moroccan rift systems (e.g., *Laville and Pique*, 1992).

Regional Stratigraphy of the Missour Basin and Atlas Mountains

Five wells have been drilled in the Missour basin, three that penetrated the Hercynian unconformity and the Permo-Carboniferous clastic sedimentary rocks, composed of sandstones, shales and conglomerates. The Triassic section overlying the unconformity is recognized by a lower and upper salt series, separated by layers of basalts. The Jurassic sequence consists primarily of marine limestones, dolomites and shales. The post-rift Cretaceous-Early Tertiary sequence is made up of a shallow and marginal marine sequence of limestones, calcareous shales, dolomites and interbedded anhydritic shales (Figure 2.4). A correlation of the wells in the Missour basin clearly indicates the regional effects of inversion (Figure 2.5). The Jurassic in the Missour basin thickens dramatically towards the old Middle and High Atlas rift systems. The thick Jurassic in the center of the rift is composed mostly of shallow marine carbonates (*Studer and du Dresnay*, 1980). On what is now

Figure 2.4 Stratigraphy of the Missour basin based on well penetrations (OSD-1, RR-1, KSAB-101, KSAB-102, & TT-1, see Figures 2.5 & 2.12, and Table 2.1), seismic stratigraphy, and outcrops within the basin. Several phases of deformation are related to major unconformities in the basin.



Age/Period
Missour Basin Stratigraphy
Tectonic Events
Quaternary-Recent
Tertiary
Neogene
Pliocene
Miocene
Continental to shallow marine
23.7 Ma
Late Alpine Phase
(Paleogene Absent)
Inversion of syn-rift structures
Cretaceous
Upper
Maastrichtian
Campanian
Santonian
Coniacian
Turonian
Cenomanian
Post Rift Phase
Early Alpine Phase
Subsidence of rift basins
isostatic uplift of rift margins
(Lower Cretaceous absent)
97.5 Ma
Jurassic
Upper
Tithonian
Kimmeridgian
Oxfordian
163 Ma
Middle
Callovian
Bathonian
Bajocian
Aalenian
187 Ma
Low.
Toarcian
Pliensbachian
Syn-Rift Phase
Transtensional deformation
208 Ma
Triassic
Upper
Rhaetian
Norian
Carnian
M.
Ladinian
Triassic volcanics(dolerites), evaporites, carbonates, and basal clastics
240 Ma
Hercynian Phase
(Permian-L.Triassic Absent)
Carbon.
Westphalian
Namurian
Visean
Tournasian
Transtensively generated turbidite basins.
360 Ma
Caledonian Phase?
Devonian
Frasnian
Givetian
Eifelian
Emsian
Pragian
Lochovian
408 Ma
Hercynian Igneous Intrusives (~325 Ma)
Silurian
Lud.
Ludfordian
Gorstian
Wen.
Gleedonian
Sheinwoodian
Lly.
Telchian
Aeronian
Rhuddanian
Deep marine graptolitic shales and siltstones

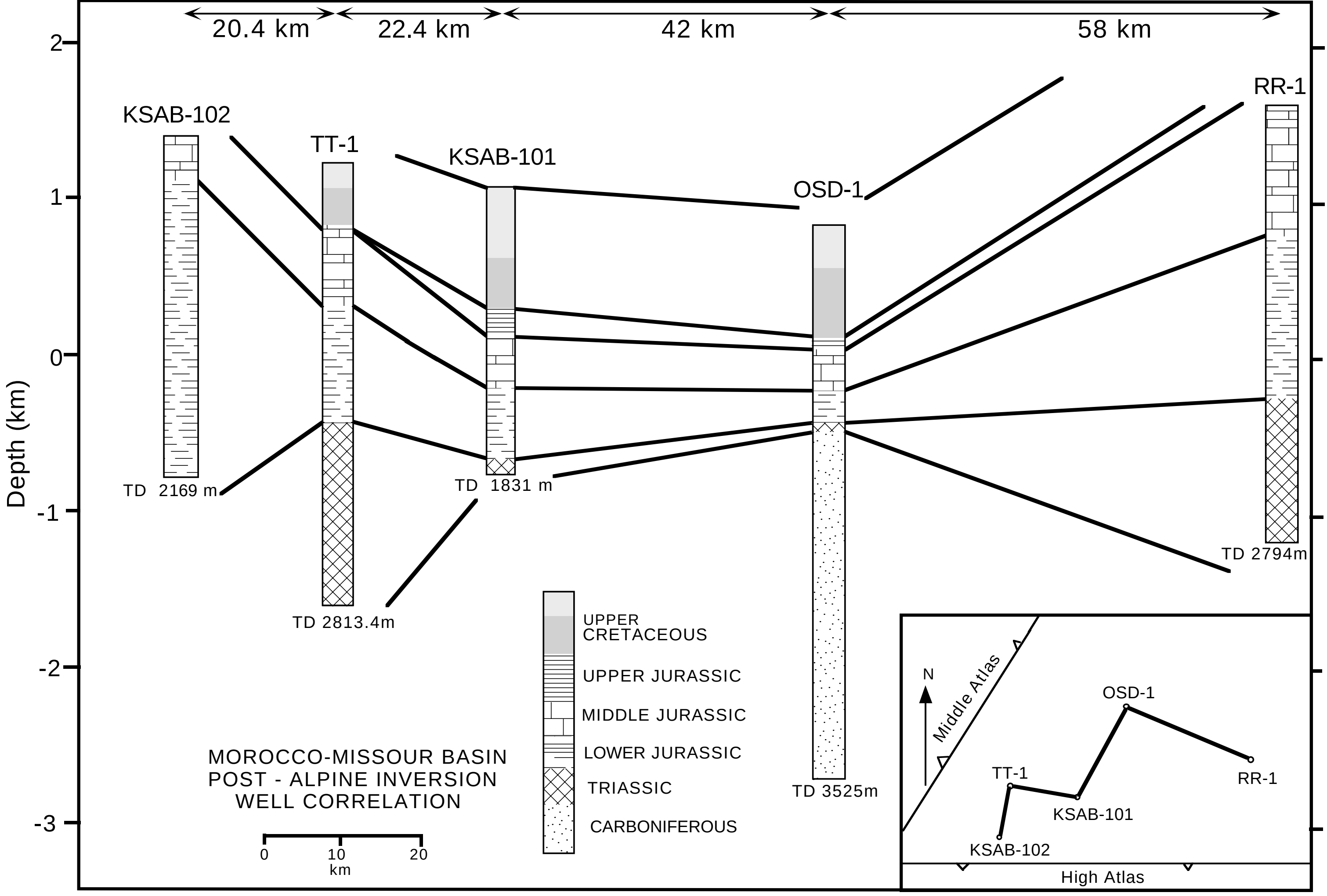
the edge of the mountain belts, Jurassic carbonates exhibit basin-margin facies such as reefs, platform and intertidal sedimentary rocks (*du Dresnay*, 1971). In the middle of the High Atlas mountains Jurassic carbonates measure up to 7 km in thickness (*Studer and du Dresnay*, 1980). Erosion in the Atlas has removed part of the Jurassic section related to inversion. Illite crystallinity in preserved sedimentary rocks (Early Jurassic) was utilized *by Brechbuhler et al.* (1988) to estimate almost 6-8 km of syn-rift and post-rift thickness in the deepest portion of the rift. The Triassic sedimentary rocks of the High Atlas attain thicknesses of 4-4.5 kilometers (*Beauchamp*, 1988). A composite thickness for the syn-rift sedimentary rocks in the High Atlas could be as thick as 10-12 kilometers, based on a measured field sections in the High Atlas .

Field Mapping and Analysis

Field work was undertaken to collect data to constrain the interpretation and modeling of subsurface data. The Missour basin and the Atlas mountains provide a means to study inversion structures in outcrop, and use the structural characteristics found to help interpret subsurface structural relationships. Multiple phases of deformation along the margins of the Missour basin indicate a complicated tectonic history that developed from the beginning of the Triassic through the early Tertiary.

The tectonic history of the Missour basin has developed in several phases of extension and compression rather than only one phase of extension in the Triassic, and one later phase of compression in the Oligocene, as has been previously believed.

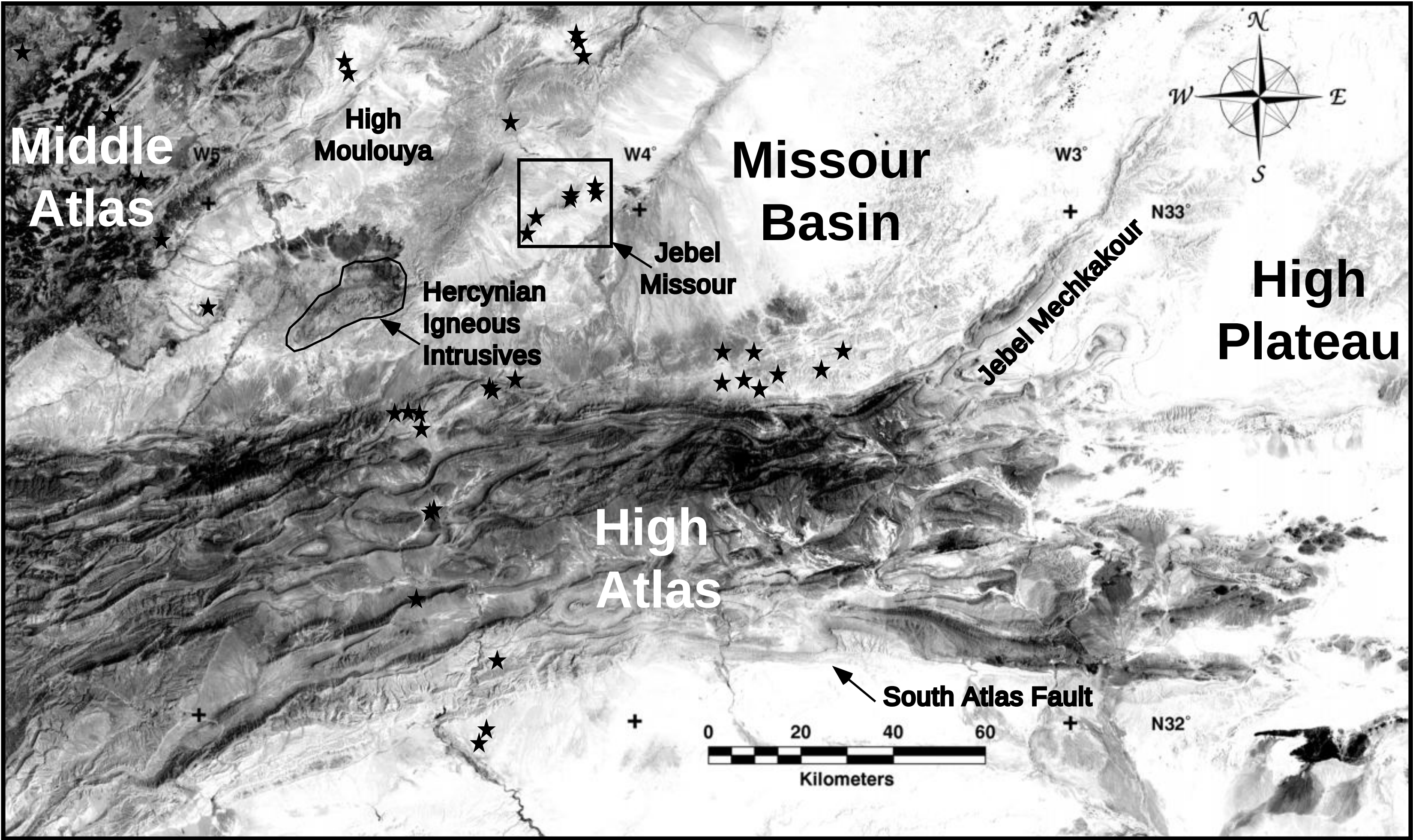
Figure 2.5 Well correlation between wells drilled in the Missour basin. Inversion can be seen by the dip of the base Cretaceous unconformity in the opposite direction of syn-rift thickening.



20.4 km
22.4 km
42 km
58 km
Depth (km)
2
1
0
-1
-2
-3
KSAB-102
TT-1
KSAB-101
OSD-1
RR-1
TD 2169 m
TD 2813.4m
TD 1831 m
TD 3525m
TD 2794m
UPPER CRETACEOUS
UPPER JURASSIC
MIDDLE JURASSIC
LOWER JURASSIC
TRIASSIC
CARBONIFEROUS
MOROCCO-MISSOUR BASIN
POST - ALPINE INVERSION
WELL CORRELATION
0
10
20
km
N
Middle Atlas
High Atlas
OSD-1
TT-1
KSAB-101
KSAB-102
RR-1

The primary goal of field work was to locate exposures of fault zones along seismic reflection profiles, and to tie the two datasets together to provide models for structural styles of inversion. Most of the exposures in the Missour basin are located along the margins of the basin (Figure 2.6). The most important exposure within the basin relative to this study is Jebel Missour (Figure 2.6). This prominent topographic ridge trends northeast and is composed of two large anticlinal structures. Cretaceous and Jurassic sedimentary rocks are exposed in the two anticlines. The most important relationship of these two folds is that they verge in the opposite direction (Figure 2.7). Air photographs used for mapping in the field illustrate the pronounced topographic expression of these anticlinal structures. The southernmost fold (F1-A) verges to the northwest, and the northernmost fold (F1-B) verges to the southeast. Both are asymmetric folds with steeply dipping to vertical limbs along one side of the fold. Strike and dip data were collected along transects across the structures and plotted to illustrate the overall geometry of the structures (Figure 2.8). The northernmost fold (F1-B) is plunging to the northeast, and the southeast limb is steeply dipping to vertical along strike of the fold (Figure 2.9). The southern anticline (F1-A) has exposures of both synthetic and antithetic faults located in the core of the fold. Both the synthetic and antithetic faults are steeply dipping high-angle faults. Fault plane lineaments indicate a reverse sense of motion on the reactivated synthetic fault (up to the northwest). These faults displace Jurassic rocks, and die out up section in the anticlinal structure. There is a distinct change in dip between the post-rift Upper Cretaceous sedimentary rocks and the syn-rift Jurassic rocks, and the examined faults do not cut the base Cretaceous

Figure 2.6 LANDSAT TM (band-5, infrared) mosaic of the Missour basin, and portions of the High Atlas and the Middle Atlas mountains. Field locations where structural data were collected are shown as stars. The eastern margin of the Missour basin is bounded by the northeast trending Jebel Mechkakour. The location of Jebel Missour is shown to the east of the High Moulouya. The structures seen in the High Atlas mountains show polyphase deformation.



Middle Atlas
High Moulouya
W5°
W4°
Missour Basin
W3°
N
W
E
S
N33°
Jebel Missour
Hercynian Igneous Intrusives
Jebel Mechkakour
High Plateau
High Atlas
South Atlas Fault
N32°
0
20
40
60
Kilometers

Figure 2.7 Air photograph of the Jebel Missour region (located on Figure 2.6). Two northeast trending anticlinal structures show two phases of deformation (F1 & F2). Fold F1-A and F1-B are verging in opposite directions. Jurassic syn-rift sedimentary rocks are exposed in the core of both anticlines separated by the base Cretaceous unconformity. The Tertiary and Quaternary sedimentary rocks are flat lying and onlap the two structures. Location of Figure 2.9 is shown near F1-B.

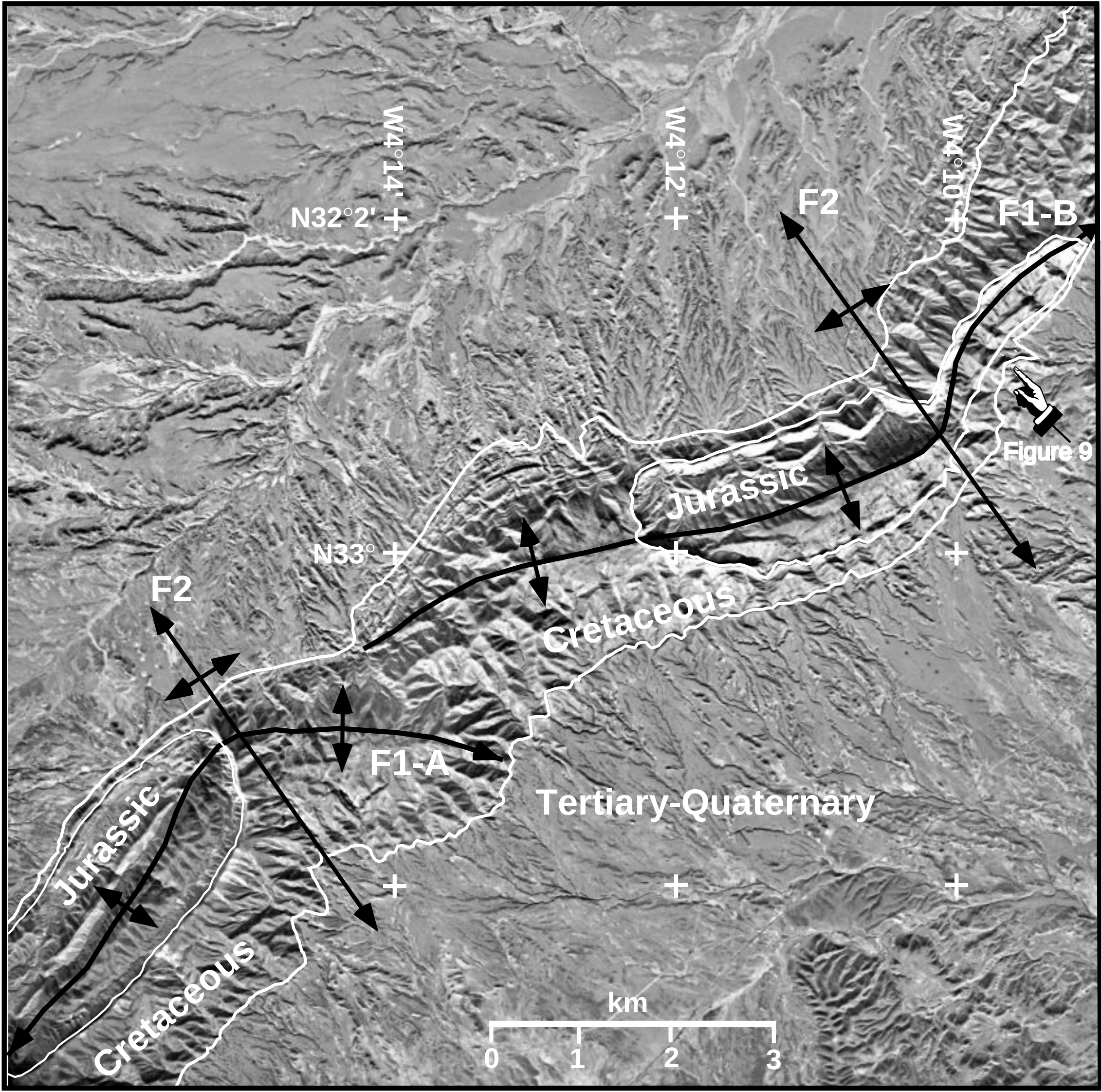
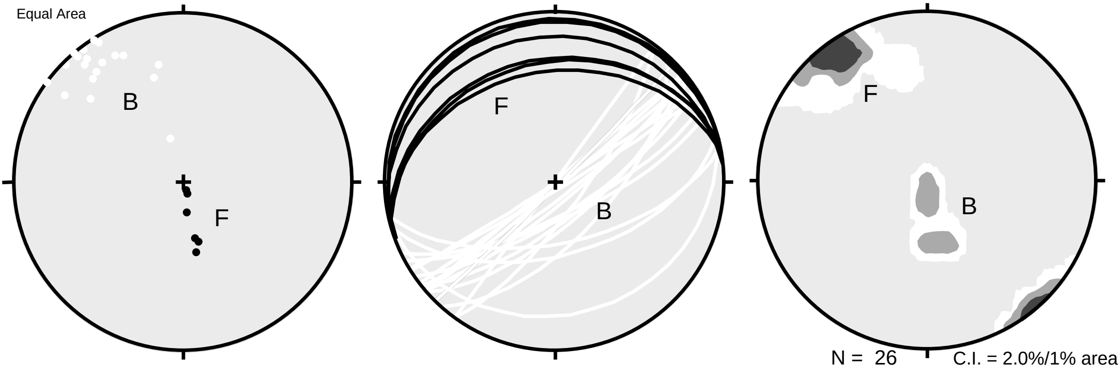
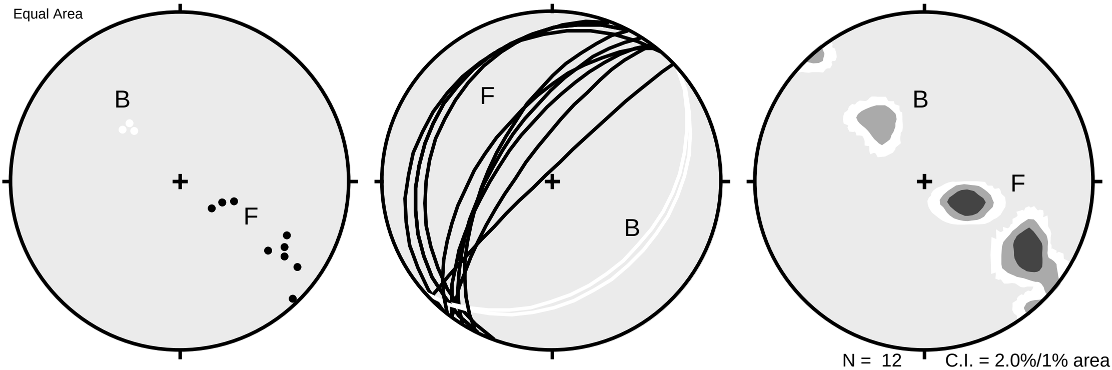

W4°14'
W4°12'
W4°10'
N32°2'
N33°
F2
F1-B
F2
Figure 9
Jurassic
Cretaceous
F1-A
Tertiary-Quaternary
Jurassic
Cretaceous
km
0
1
2
3

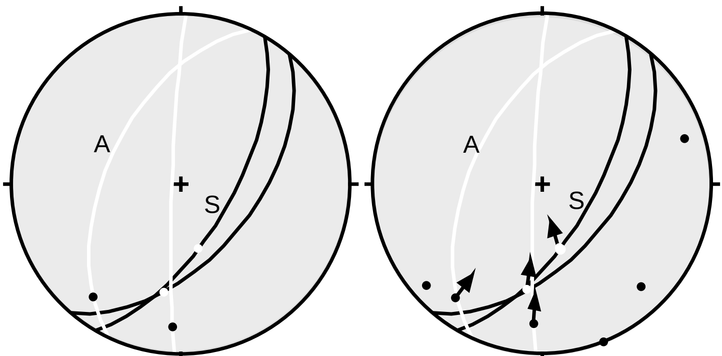
Figure 2.8 Field data collected from Jebel Missour. The northern anticline F1-B is verging to the northeast. The southern anticline(F1-A) is verging to the southwest. High angle reverse faults were measured in the core of the southern anticline (F1-A). These faults are inverted planar normal faults (synthetic and antithetic). The vergence of the folds is related to the original dip of the syn-rift normal faults.



Jebel Missour Northern Anticline - F1B
Bedding poles/planes



Jebel Missour Southern Anticline - F1A
Bedding poles/planes



Fault planes of inverted high angle
normal faults (synthetic and antithetic)

B=backlimb F=forelimb A=antithetic S=synthetic

Figure 2.9 Jebel Missour along the northeastern plunging F1-B fold. See location of photo on Figure 2.7. The fold is verging out of the picture. Steeply dipping to vertical beds can be seen along the flatirons in the right side of the photo. The fold is an asymmetrical fold formed by the inversion of a planar fault that dips to the northwest. The fault is most likely a fault propagation fold. The steeply dipping limb is complicated by several smaller folds in the foreground.

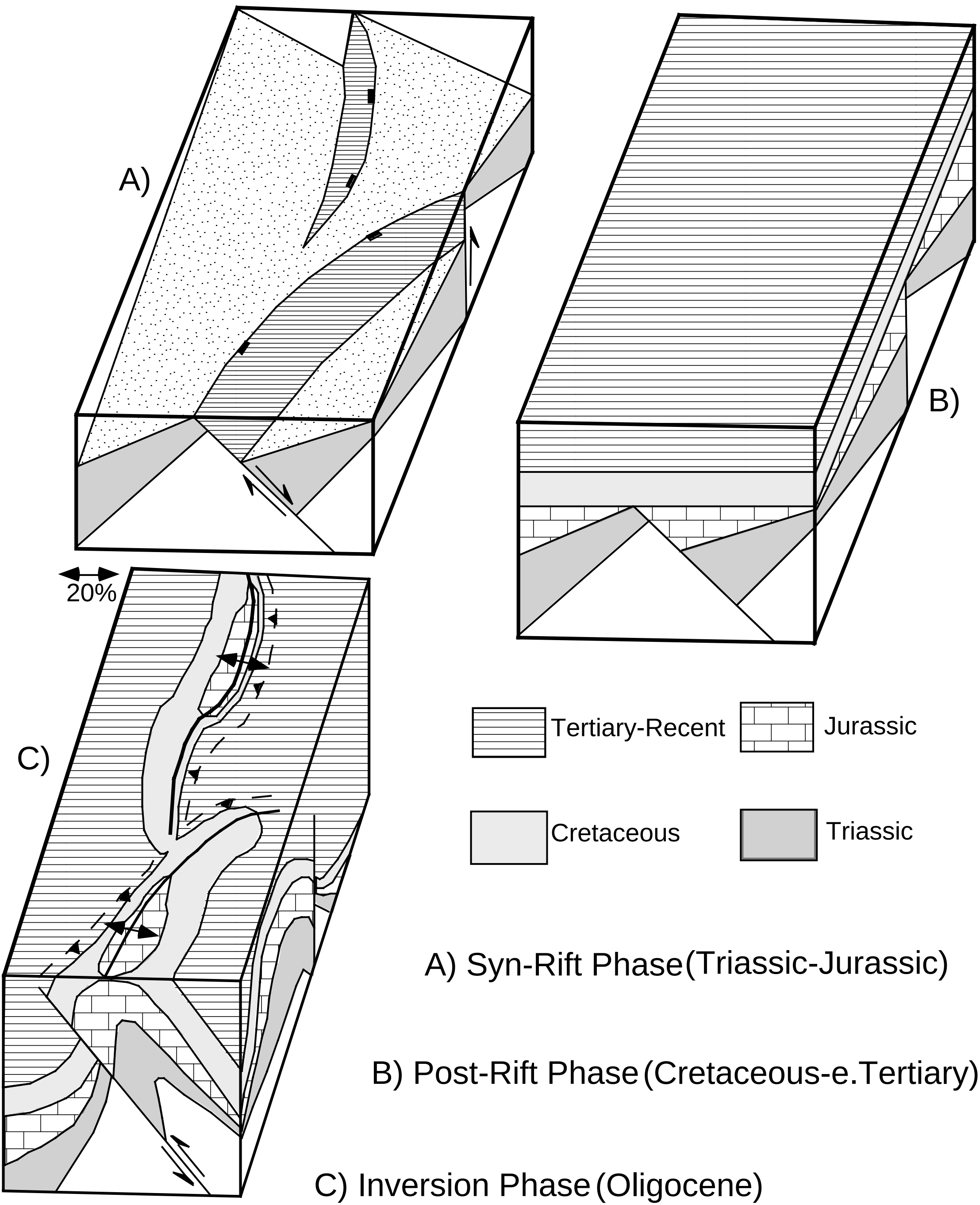

SW
NE
<30°
> 80°
0
m
50

unconformity.

The two structures (F1-A and F1-B) at Jebel Missour evolved initially from two opposing planar normal faults connected by a ramp (Figure 2.10). These two faults were active during rifting in the Triassic to Late Jurassic periods. During the post-rift phase the base Cretaceous unconformity eroded upper Jurassic syn-rift sedimentary rocks from regions of the Atlas rift system as subsidence began. Cenomanian to Turonian rocks were deposited in the Missour basin, and in the early Tertiary the two half grabens were inverted by an oblique compressional stress possibly related to the Alpine orogeny. During the late Tertiary to Recent, sedimentary rocks were deposited onlapping the existing structure formed by earlier phases of deformation. Late Tertiary to Recent sedimentary rocks are flat lying and have not been affected by any significant deformation after the Oligocene phase of uplift and compression in Morocco.

Fault lineaments and slickensides indicate a reverse sense of movement on faults measured at Jebel Missour. These lineaments overprint earlier dip-slip and oblique-slip lineaments for which a sense of movement could not be determined. Previous lineaments indicate earlier phases of deformation, and may be associated with normal and transtensive phases of deformation related to the initial phases of rifting. The Middle Atlas mountains to the west of Jebel Missour are bounded by faults that have recent movement along inverted normal faults exhibiting a left-lateral sense of shear (*Morel et al.*, 1993). These faults have been active recently, and offset Quaternary sedimentary rocks and older Neogene volcanics. Reactivated rift faults in the Middle Atlas have thrust Triassic sedimentary rocks over Pliocene and

Figure 2.10 Schematic model depicting the structural evolution of Jebel Missour. (A) Syn-rift faults result in the deposition of sedimentary rocks in an asymmetrical half graben. The two normal faults are dipping opposite one another connected by a ramp or transfer zone. (B) The post-rift phase begins the deposition of Cenomanian and Turonian age sedimentary rocks unconformably above the Jurassic. (C) Uplift and inversion reactivates opposing planar faults that both form oppositely verging fault propagation folds. The inversion of (A) has resulted in approximately 20% shortening.


A)
B)
C)
20%
Tertiary-Recent
Jurassic
Cretaceous
Triassic
A) Syn-Rift Phase (Triassic-Jurassic)
B) Post-Rift Phase (Cretaceous-e.Tertiary)
C) Inversion Phase (Oligocene)

younger sedimentary rocks. The faults related to structures at Jebel Missour have not been active since the early Tertiary. This indicates more recent convergence is being accommodated along larger fault systems in the Middle Atlas mountains.

It is difficult to ascertain the exact geometry at depth of the faults examined at Jebel Missour. Based on the asymmetrical geometry of the associated folding, the faults may represent reactivated planar faults. The inversion of a planar fault usually results in a more asymmetrical geometry than an inverted listric fault, which upon inversion forms a more open symmetrical fault-bend-fold (*Mitra*, 1993). The geometric styles generated by the inversion of listric and planar normal faults are controlled by the orientation of the maximum compressive stress. The inversion of a steeply dipping planar fault can accommodate a limited amount of shortening because of the geometry of the fault. Once shortening along a planar fault has occurred by uplift of the associated hanging wall, further stress generated by compression is accommodated by lateral strike-slip movement along the fault. It is possible to reactivate steeper faults with dips of ~40-60° only if there is a low coefficient of sliding friction (*Coward*, 1991). Because of the geometry of planar faults, it is easier to reactivate steeper faults by oblique-slip or strike-slip motion. This type of deformation may have occurred along the faults associated with Jebel Missour (Figure 2.7). The apparent second phase of folding affecting the structures is thought to be related to oblique-slip movement on the planar faults associated with the structures. The deformation of the original half grabens bounded by steeply dipping planar faults may have occurred as follows: inversion of the hanging wall half graben along a planar normal fault, folding of the post-rift Upper Cretaceous sedimentary rocks and syn-rift Jurassic rocks into an asymmetrical fault propagation fold, refolding of the asymmetrical folds (Figure 2.7, F2) by oblique-slip movement along the planar fault, and onlap of the structures by the Neogene to Recent sedimentary rocks. Jebel Missour is an example of how planar faults may respond to inversion within a particular rift system.

Topographic maps and a GPS (Global Positioning System) receiver were used to find the surface location of faults identified on seismic lines. Faults identified on seismic profiles that appear to extend to the surface were commonly covered by Recent sedimentary rocks. Many of the faults seen on seismic reflection profiles as individual faults were found to occur as zones of offset at the surface. It was possible to record dips of rocks in the hanging wall and footwall of faults in many cases, that define the overall geometry of the surface structure (Figure 2.11). Displacement in fault zones generally occurs along a system of related small faults in the zone of deformation. Slip illustrated by kinematic indicators along bedding planes in the footwall and hanging wall of faults gave a sense of shear across fault zones.

Many of the fault zones along the southern margin of the Missour basin indicate oblique-slip movement associated with the most recent phase of deformation. The sense of slip across all of the fault zones measured in the field along the southern margin of the Missour basin were either right-lateral oblique-slip, or right-lateral strike-slip movement. The age of this right-lateral deformation is thought to be associated with the early Tertiary inversion phase or older, as younger sedimentary rocks were not found to be deformed.

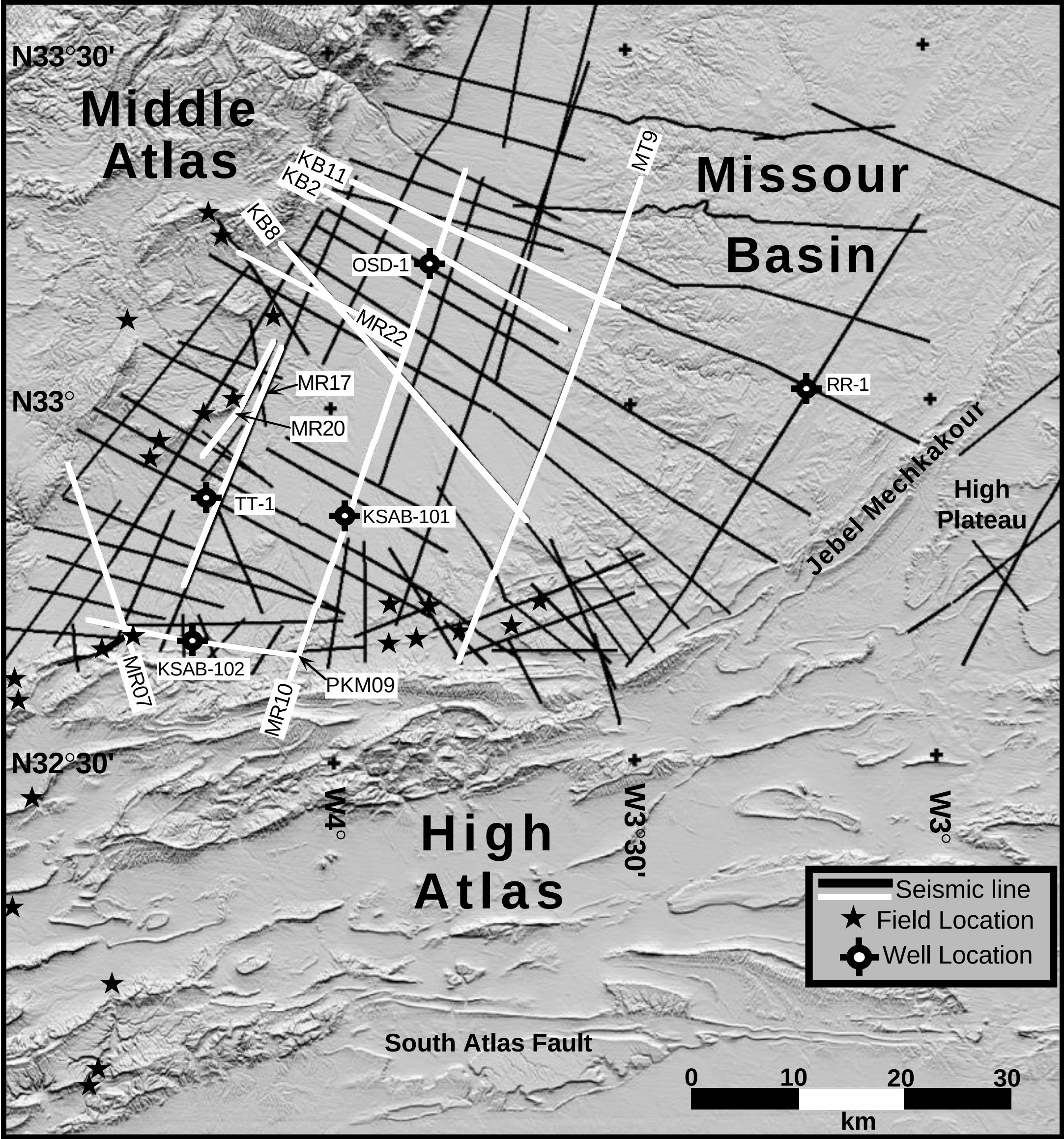
Figure 2.11 Fault propagation fold that is verging to the north near the southern margin of the Missour basin. Onlap of growth strata can be seen from north to south across the structure.


North
South
0
m
10

Lineaments and slickensides that record right-lateral oblique and strike-slip sense of movement, were found to cut across several (up to three) earlier lineations that record previous phases of movement on the same surface.

An important conclusion based on the study of the Missour basin is the concept that the entire basin may have been uplifted and inverted as a whole (Figure 2.2). Most of the deformation in the basin has occurred along the margins of the basin. Extensional features that are not near the flanks of the Atlas mountains normally do not indicate evidence of reactivation. There may have been reactivation and inversion of extensional faults in the interior of the basin, but these faults still illustrate a net extension. This observation leads to the conclusion that shortening has been accommodated by previously existing syn-rift faults within the paleo-Atlas rift and current day Atlas mountains. The density of faults observed in the field indicate a direct correlation between the degree of inversion and pre-existing syn-rift faults. This relationship may be true for the magnitude of the shortening and inversion, relative to the size, amount of throw and extent of syn-rift faults in the Missour basin and Atlas mountains. These observations may also have a correlation with the present topography in the basin (Figure 2.12). Topography is generally related to the density of faults and structure in the Missour basin as can be seen by Jebel Missour (Figure 2.7). Left-lateral oblique and strike-slip offset in the Quaternary and Neogene recorded in the Middle Atlas (*Morel et al.,* 1993), combined with right- lateral movement along the southern margins of the Missour basin, would indicate the overall relative movement of the Missour basin is to the northeast (Figure 2.3A). The concept of the Missour basin having been uplifted and inverted as a whole may be similar to the concept of escape tectonics

Figure 2.12 Map showing digital topography of the Missour basin and the adjacent High and Middle Atlas mountains, seismic reflection profiles, well data and field locations are shown. 3400 km of seismic lines were used to study the tectonic evolution of the Missour basin and Atlas mountains. The seismic lines in white are those used in this paper, both black and white lines were used in the study.


N33°30'
Middle Atlas
Missour Basin
KB11
KB2
KB8
MT9
OSD-1
MR22
MR17
MR20
N33°
RR-1
Jebel Mechkakour
High Plateau
TT-1
KSAB-101
KSAB-102
MR07
PKM09
MR10
N32°30'
W4°
W3°30'
W3°
High Atlas
Seismic line
Field Location
Well Location
South Atlas Fault
0
10
20
30
km

proposed by *Sengor et al.* (1984). The basin may have been uplifted/inverted by the culmination of several phases of deformation. The region was previously part of the shelf margin or shoulder of the Atlas rift system, and has since been uplifted and translated to the northeast.

The basin is bounded to the east by Jebel Mechkakour (Figures 2.6 & 2.12). The large-scale geometry of the basin as defined by the High Atlas to the south, the Middle Atlas/High Moulouya to the west and Jebel Mechkakour to the east, is that of a rhombohedral shape. The High Plateau to the east of the basin (Figure 2.1) also is characterized by an even more obvious rhombohedral shape. These shapes are inherent in transtensional pull-apart basins associated with rifting (*Morley*, 1995). The High Plateau basin is distinctly different from the Missour basin as there is a thick (>1000 meters) sequence of salt penetrated by wells in the High Plateau basin that is not present in the Missour basin. The anticlinal feature to the east of Jebel Mechkakour (Figure 2.6) was partially formed by the movement of Triassic salt. The distinct differences between the Missour basin and the High Plateau basin have been present since the early syn-rift phase in Morocco. The High Plateau basin was most likely an extensional pull-apart basin during the rift phase, and the Missour basin was topographically higher relative to the current High Plateau. The present Missour basin has been uplifted, inverted and translated to the northeast by a style of escape tectonics along boundaries formed during rifting.

Geological and Geophysical Analysis

3400 kilometers of seismic reflection data have been acquired in the Missour basin and were used at Cornell in this study. These surveys were acquired during a period between 1974-1986, resulting in a collection of seismic reflection data with various qualities and processing parameters. The digital post stack data of several lines were obtained from ONAREP and were utilized for further processing, migration, and depth conversion for more accurate analysis and modeling. Velocity data in the form of time/depth curves and synthetic seismograms were used to tie wells in the basin to the seismic reflection data (Table 2.1).

Table 2.1. Wells Drilled in the Missour Basin

WELLS	COORDINATE	COMPANY	YEAR	TOTAL DEPTH	FM. AT TD
TT-1	X=611.916.7 Y=252.593.2 EL=1212.6 m	S.C.P. .*	1954	2813 m	TRIASSIC
RR-1	X=704.372.2 Y=271.197.5 EL=1633 m	S.C.P.*	1965	2794 m	TRIASSIC
KSAB-101	X=632.666 Y=249.989 EL=1395.5 m	Phillips	1983	1831 m	TRIASSIC
KSAB-102	X=608.889 Y=230.609 EL=1395.5 m	Phillips	1983	2188 m	JURASSIC
OSD-1	X=648.824 Y=288.959 EL=839 m	ONAREP**	1986	3525 m	CARBONIF.

*SCP-Societe Cherifiene des Petroles

**ONAREP-Office National de Recherches et d'Exploitations Petrolieres

The combination of surface and subsurface data in the Missour basin constrain the geometry of faults recognized at the surface into the subsurface.

Seismic lines were migrated and depth converted prior to modeling, balancing or interpretation of the seismic data. Seismic lines used for modeling in this study were reprocessed by applying an FX deconvolution and coherency filters, migrating the data using a Stolt FK migration and then depth converting the line using interval velocities. This process is normally bypassed due to time constraints or the lack of digital seismic data. The migration and conversion of key seismic lines in the Missour basin proved to yield important interpretations and structural models in the basin.

Two important regional stratigraphic horizons were clearly identifiable on most seismic data in the basin. These are: (1) the base Cretaceous unconformity, and (2) the Hercynian unconformity (Figure 2.4). These horizons form the stratigraphic boundaries for the post-rift and syn-rift sedimentary sequences mapped throughout the basin. The base Cretaceous unconformity is usually identifiable on dip lines as an angular unconformity. The Hercynian unconformity is characterized by an angular unconformity between the Paleozoic and the syn-rift sedimentary sequence. The Hercynian unconformity can also be recognized as the base of the highly reflective Triassic sedimentary rocks composed of salts, anhydrites, volcanics, dolomites and clastics (Figure 2.13). Reflections from within Paleozoic strata were evident on many of the seismic sections in the basin. On several seismic sections strong reflections were present to seven seconds two-way travel time. Paleozoic structures are present on several lines in the basin. Wells drilled in the Missour basin did not penetrate deeply enough into the Paleozoic section to allow for the correlation of units within the Paleozoic. The only structures that could be defined in the Paleozoic were at the structural level of the Hercynian unconformity.

Styles of Faulting

The recognition of low-angle thrusts in the Atlas mountains has been documented on geological maps and in previous field studies. These thrusts have been interpreted previously on published cross sections as low-angle faults that detach at the top of the syn-rift Triassic salts. The results of this study indicate that these low-angle thrusts may be related to the reactivation of syn-rift listric faults that detach well below the syn-rift Triassic sedimentary rocks. Many faults in the Atlas are steep to vertical thus making it difficult to recognize the direction of dip of the fault. It is also difficult to recognize the footwall and hanging wall of the fault. While it is common for extensional faults related to Mesozoic rifting to be inverted upon compression in the Missour basin and the Atlas mountains, it is difficult in many cases to document the dip of the fault plane and the sense of movement(s) on the faults using only seismic reflection data.

This study found that it is common for large scale (>5 km, map view) faults to dip both towards the paleo-rift basins as well as out of the basins. This is a common relationship in many rift systems (*Rosendahl et al.*, 1986). When the relationship of extensional and dual fault polarity is applied to an inverted rift system, the results are that thrusts and reverse faults verge into and away from the paleo-rift basins.

It is important to recognize sedimentary relationships on seismic data that illustrate extension and active sedimentation during rifting. Otherwise, it is difficult to determine whether or not a fault is a reverse fault, thrust fault, or a reactivated syn-rift fault. Seismic line 85KB11, for example, shows a well-defined high-angle planar fault that dips to the west towards the Middle Atlas mountains (Figure 2.14). This fault is not exposed at the surface, but the highly reflective package of the Triassic sedimentary rocks above the Hercynian unconformity indicate the sense of throw and the amount of offset. The Cretaceous is not present to the west of the fault on this line 85KB11. The sense of throw on the fault is reverse, but it is difficult to determine if the fault is a reactivated normal fault or a fault that was newly formed as a reverse fault. Another consideration when interpreting reactivated faults, is the uplift of the hanging wall half graben above what has been previously referred to as the null point (*Williams et al.*,1989). A fault may have been a normal fault during rifting, with thickening of syn-rift sediments into the active extensional fault. Upon inversion the hanging wall is uplifted until the pre-rift unconformity (Hercynian) is uplifted above the pre-rift unconformity in the footwall. The null point as referred to by *Williams et al.*, would occur when there is no stratigraphic displacement of the pre-rift unconformity across a fault. This relationship can result in interpretations that assume deformation by extensional tectonics, when a significant amount of compression normal to the fault may have occurred. Identifying inversion on faults is important as inversion may affect hydrocarbon migration and trapping in either a positive or negative manner. For example, on seismic line 85KB11 the null point would occur when the base Triassic-Hercynian unconformity (Figure 2.14, A-A') is juxtaposed across the fault that is dipping to the west.

An example of a reactivated fault that has been uplifted above the null point can be seen on seismic line MR17 (Figure 2.15). This line has been migrated and depth converted to help position and restore the fault geometry. There has been an apparent uplift of the base Cretaceous unconformity to the north of the fault by at least 2 kilometers. North of the fault reflections in the Jurassic can be seen truncating beneath the base Cretaceous unconformity. The highly reflective Triassic basalts can be seen clearly above the Hercynian unconformity. The reactivation of this fault may have occurred along a previous syn-rift fault, or by reactivation of a pre-rift fault. Steeply dipping planar faults allow for a limited amount of shortening normal to the fault plane. Faults such as the fault on line MR17 reactivate by initial slip along the fault plane, inverting the hanging wall and forming a fault propagation style fold above the syn-rift fault (below the base Cretaceous unconformity). Further shortening across the fault zone was accommodated by the generation of a fault propagation fold (verging to the south). This fold was later cut by the breakthrough of the fault upwards along the forelimb of the fold. The syncline to the south of the fault on line MR17 (Figure 2.15) is the conjugate fold related to the initial fault propagation fold. If the principal compressive stress was oriented at an angle that was oblique to the fault plane, then it would be expected that oblique or strike-slip deformation might accommodate further strain across the fault (*Coward, 1991*). This style of deformation is similar to the style of deformation that was mapped at Jebel Missour (Figure 2.7).

Another more subtle criterion that indicates inversion is seen on seismic line PKM-09 (Figure 2.16). This seismic line ties well KSAB-102 drilled by Phillips in 1983 at a total depth in the Jurassic. Line PKM-09 shows a thickening of the Jurassic from east to west across the section, from 644 milliseconds to 826 milliseconds two-way travel time. This thickening occurred

Figure 2.13 Major unconformities that illustrate tectonic phases (Hercynian, base Cretaceous and Turonian?) are seen on seismic line MR-07. The Hercynian unconformity has been flattened by the inversion of the stratigraphic section. The syn-rift and post-rift sections have been rotated upwards reversing the original sense of regional dip. (Location on Figures 2.12 & Figure 2.17)

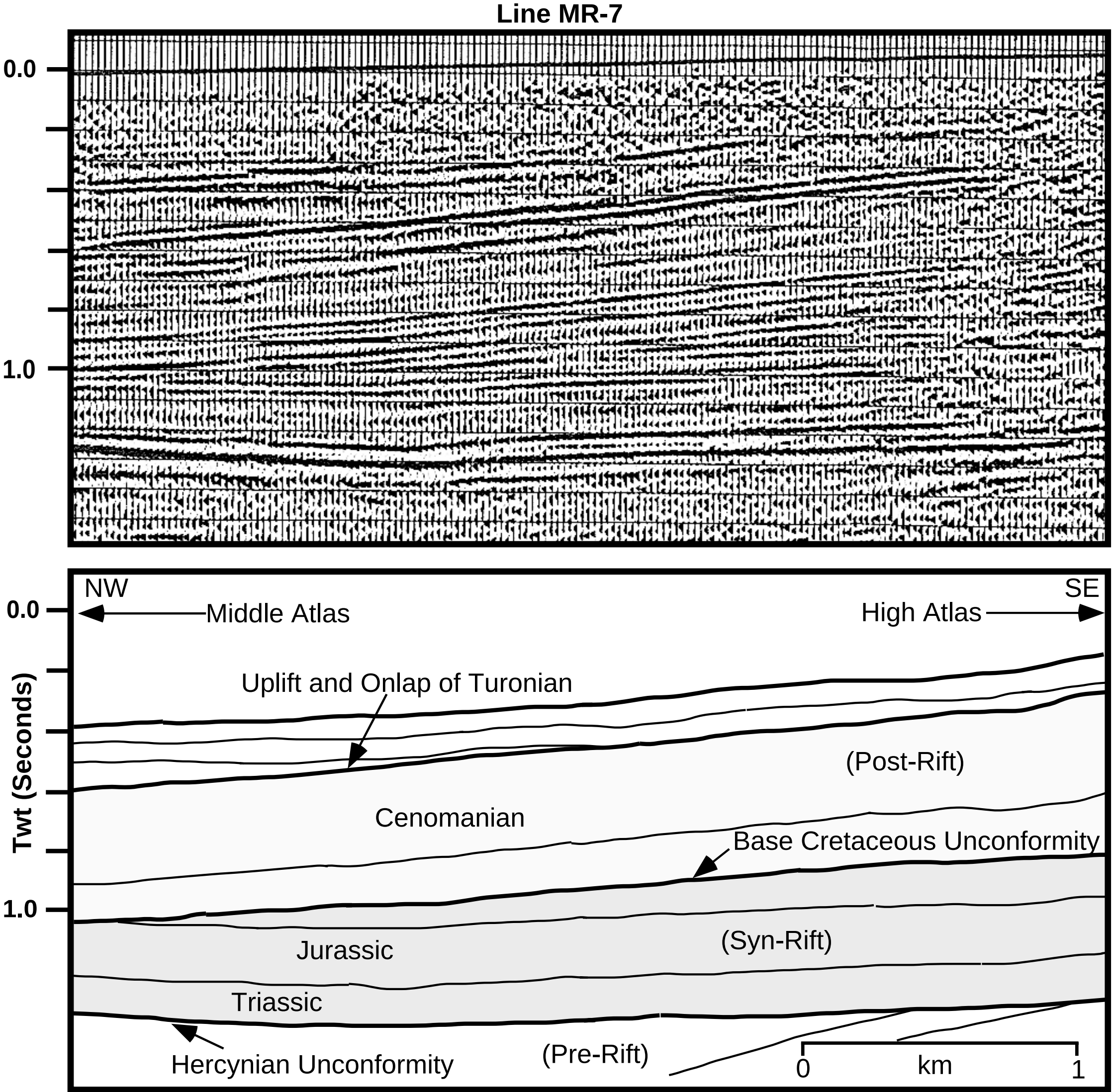

Line MR-7
0.0
1.0
Twt (Seconds)
NW
SE
Middle Atlas
High Atlas
Uplift and Onlap of Turonian
(Post-Rift)
Cenomanian
Base Cretaceous Unconformity
(Syn-Rift)
Jurassic
Triassic
Hercynian Unconformity
(Pre-Rift)
0
km
1

Figure 2.14 Seismic profile 85Kb11 located near the southeast margin of the Middle Atlas mountains (location on Figure 2.12). The null point is the point at which A-A' are adjacent across the fault. Note the uplift of the Triassic unconformity. Erosion by the base Cretaceous unconformity has removed any indication of whether the fault was a reactivated syn-rift fault, or a newly formed reverse fault.

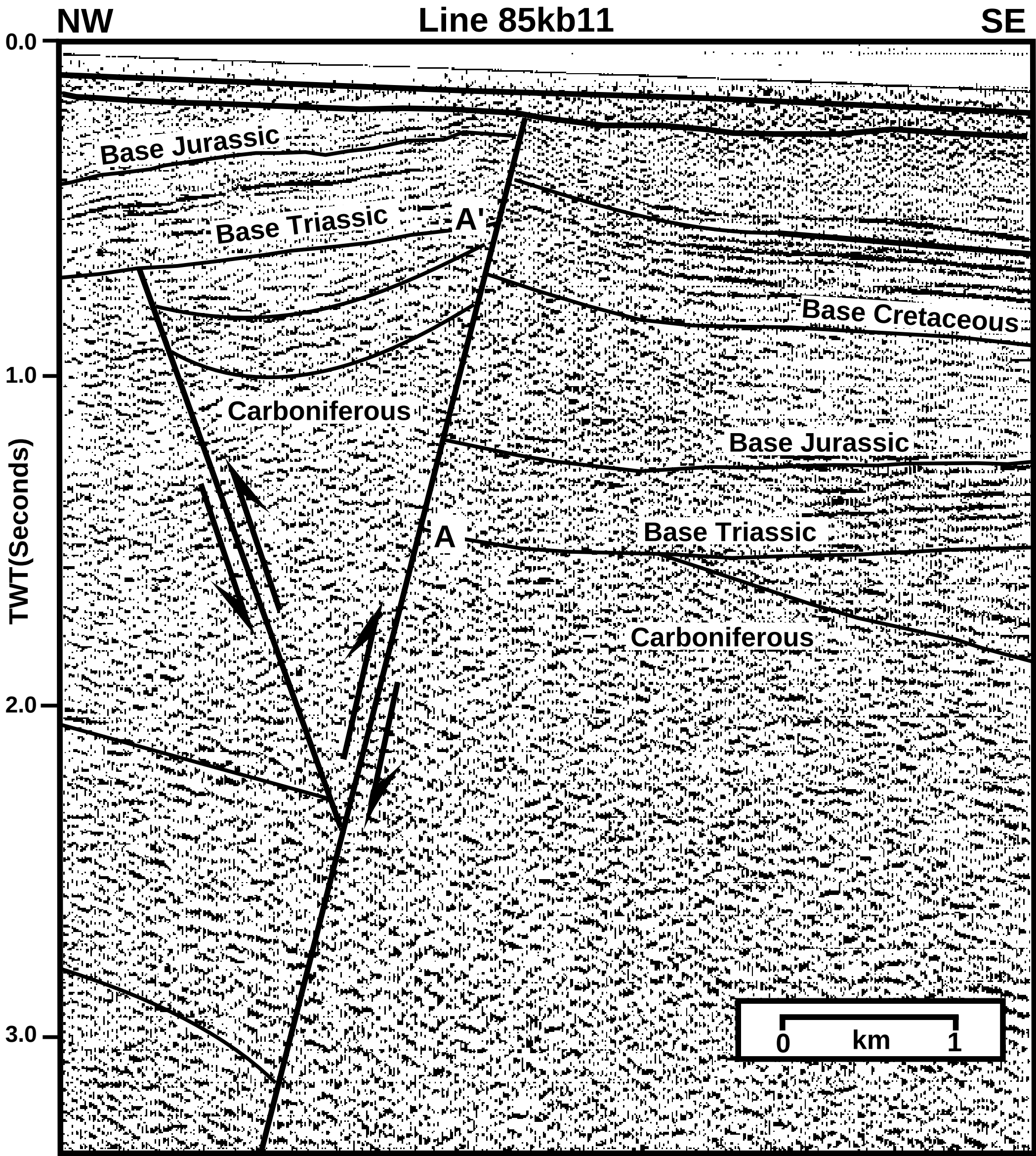
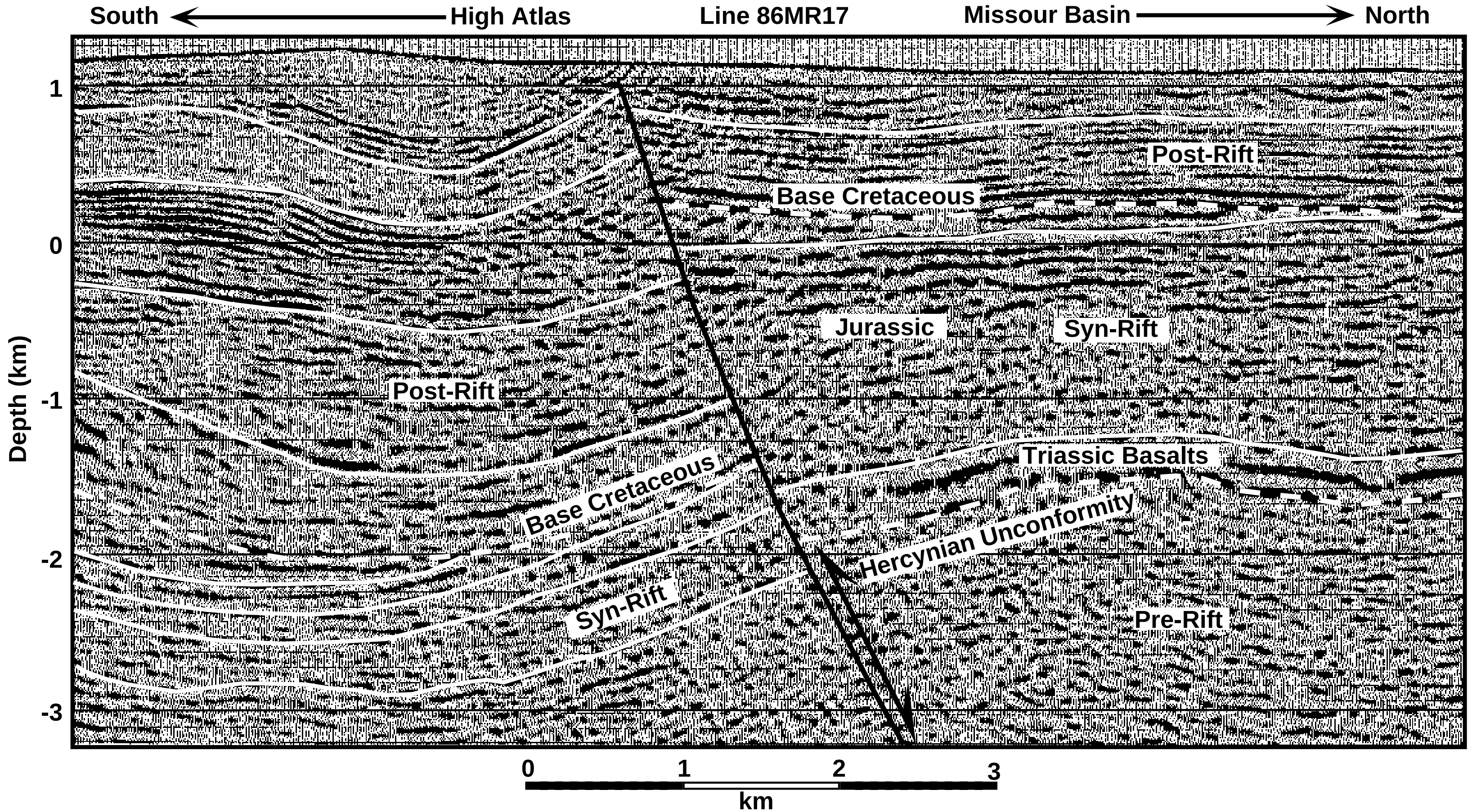

NW
Line 85kb11
SE
0.0
1.0
2.0
3.0
TWT(Seconds)
Base Jurassic
Base Triassic
A'
Base Cretaceous
Carboniferous
Base Jurassic
A
Base Triassic
Carboniferous
0
km
1

Figure 2.15 Seismic profiles MR-17 shows the uplift of the Hercynian unconformity above the null point. There has been an apparent uplift of 2 kilometers on the Hercynian unconformity. The amount of shortening normal to the fault is limited due to the geometry of the fault, and subsequent strain is accommodated by strike-slip or oblique-slip movement. (location on Figure 2.12)

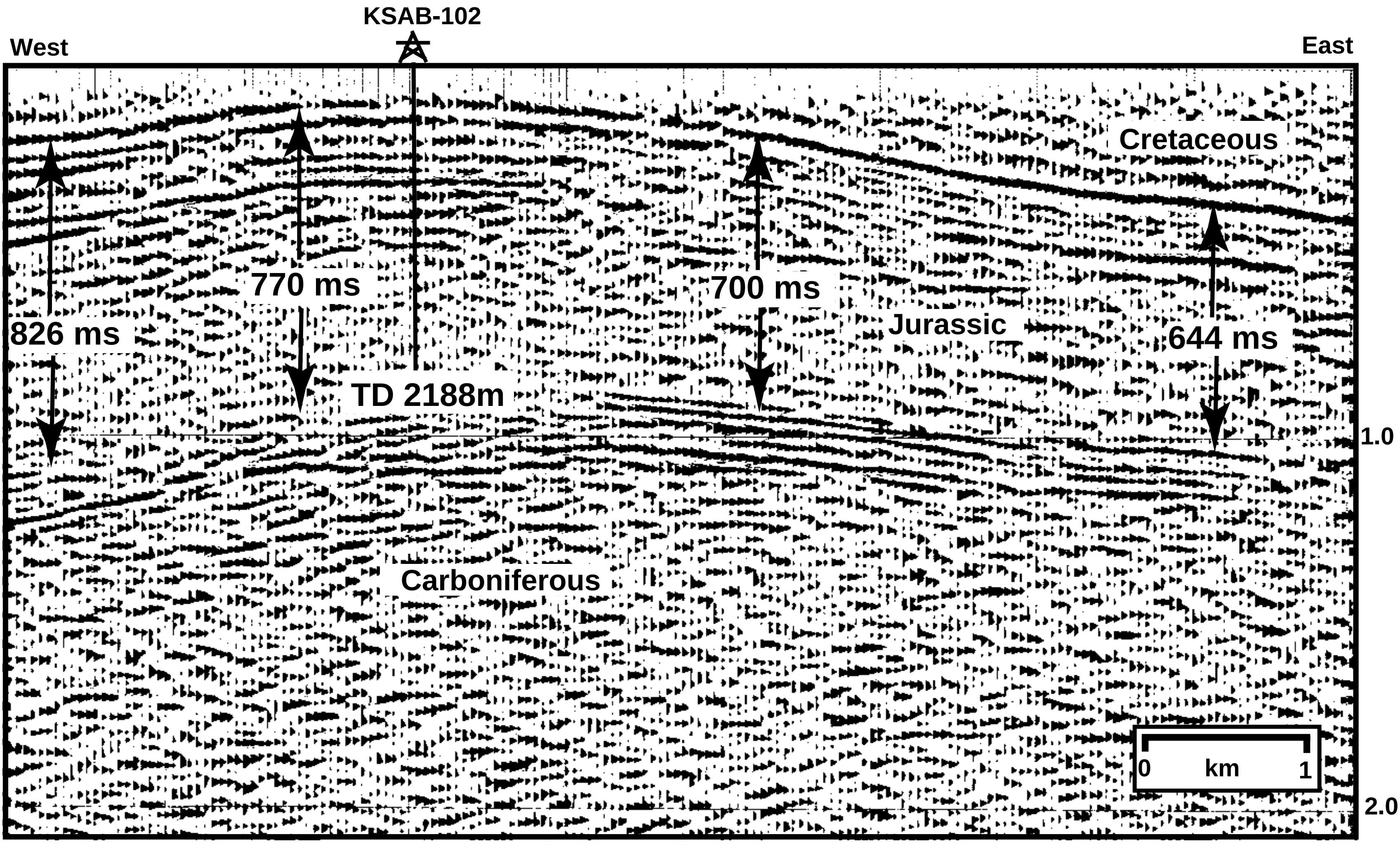

South
High Atlas
Line 86MR17
Missour Basin
North
1
0
-1
-2
-3
Depth (km)
Post-Rift
Base Cretaceous
Jurassic
Syn-Rift
Post-Rift
Triassic Basalts
Base Cretaceous
Hercynian Unconformity
Syn-Rift
Pre-Rift
0
1
2
3
km

during rifting, and the sedimentary rocks thicken towards a reactivated normal fault to the west. The inversion has resulted in an anticlinal structure (in the subsurface and surface) where there was previously a half graben.

A common feature of rifting is syn-depositional growth strata, or a progressive unconformity associated with an extensional half graben. Major unconformities such as the base Cretaceous commonly appear to have been flattened due to uplift and inversion. These unconformities originally dipped basinward as a result of post-rift subsidence (Figure 2.17). Regional seismic lines that extend across the basin were tied to wells drilled in the basin using synthetic seismograms and time-depth curves, and major stratigraphic boundaries were mapped throughout the basin. These interpreted stratigraphic boundaries were then digitized and redisplayed to produce the interpreted cross sections in Figure 2.17. One important feature as mentioned previously, is the lack of faulting in the Missour basin as a whole, with the overall deformation having occurred along the margins of the basin both during the extensional syn-rift phase, and in later inversion phases of deformation (Figure 2.17). A common inversion characteristic recognized on the seismic data in the basin is dip reversal relative to the direction of sedimentary thickening (Figure 2.13).

Evidence of a previously unrecognized phase of uplift (Turonian), or a change in sea level, can be seen on seismic line MR-7 (Figure 2.13) based on growth strata. The lack of Lower Cretaceous (Berriasian-Albian) sedimentary rocks above the base Cretaceous unconformity indicates a missing 40 my of sedimentary rocks in the Missour basin (Figure 2.13). The lack of deposition during this time might be related to isostatic uplift of the Atlas rift shoulders or

Figure 2.16 Seismic profile PKM-09. Inversion can be demonstrated by the uplift of a previous half graben. Thickening from the right to left across the section can be seen from a reflector in the syn-rift sequence to a reflector at the top of the Triassic basalts. (location on Figure 2.12)

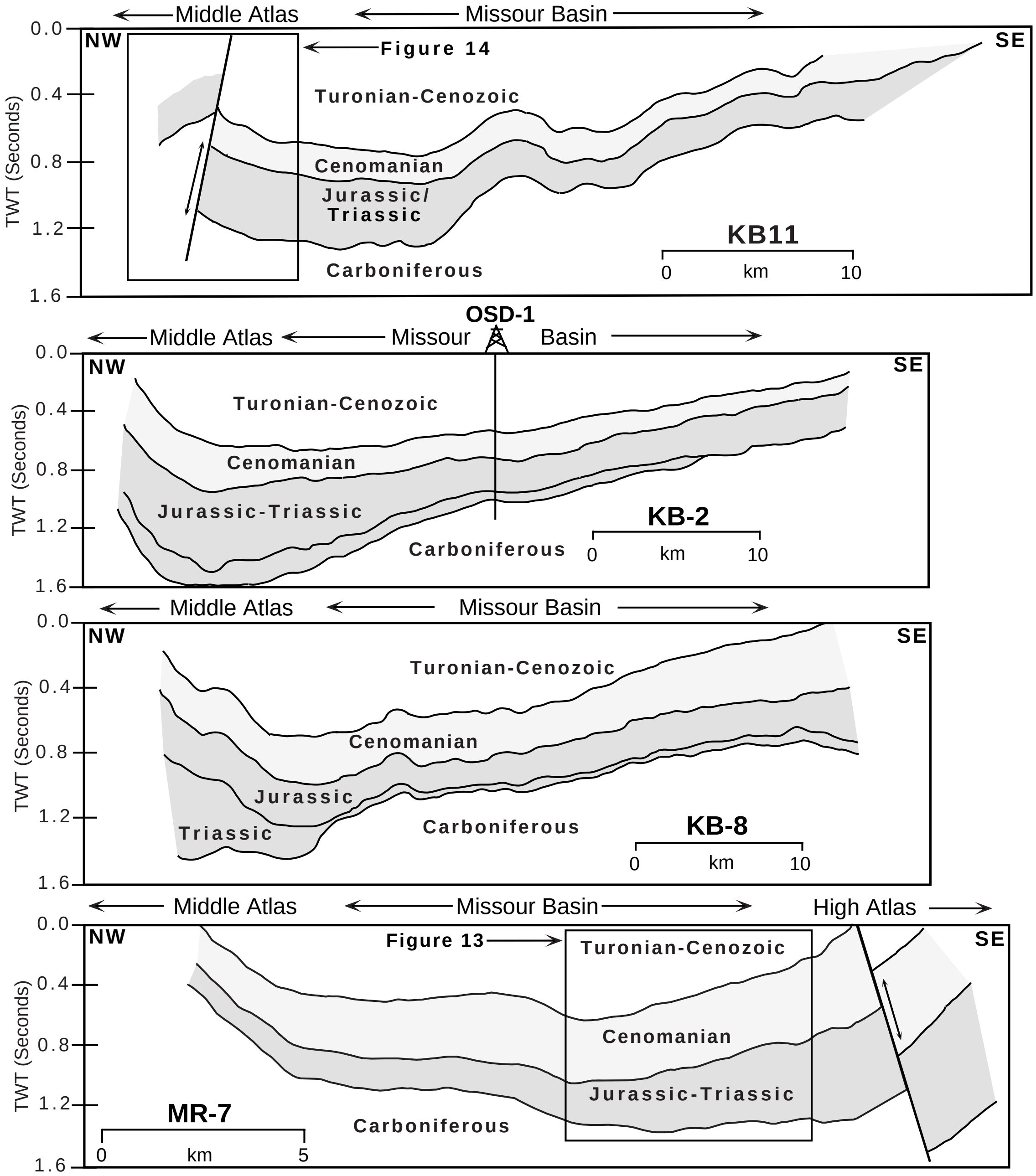

KSAB-102
West
East
Cretaceous
770 ms
700 ms
826 ms
Jurassic
644 ms
TD 2188m
1.0
Carboniferous
0
km
1
2.0

margins. The basin was positioned on such a shoulder or margin during the onset of the post-rift subsidence phase. This isostatic uplift of the rift margins (Missour Basin, High Plateau and Moroccan Meseta-Figure 2.1) during the return of the rift to thermodynamic equilibrium, would have resulted in the deposition of Lower Cretaceous sedimentary rocks into the subsiding Atlas rift basin. The Lower Cretaceous sedimentary sequence would not have necessarily been represented on the shoulders of the Atlas post-rift subsidence basin (Missour basin). Some Lower Cretaceous sedimentary rocks (Albian) are preserved in the High Atlas mountains to the south of the Missour basin. The inversion of the post-rift Atlas basin would have resulted in the uplift and erosion of the Lower Cretaceous sedimentary rocks from most of the Atlas mountains.

Fault Restorations

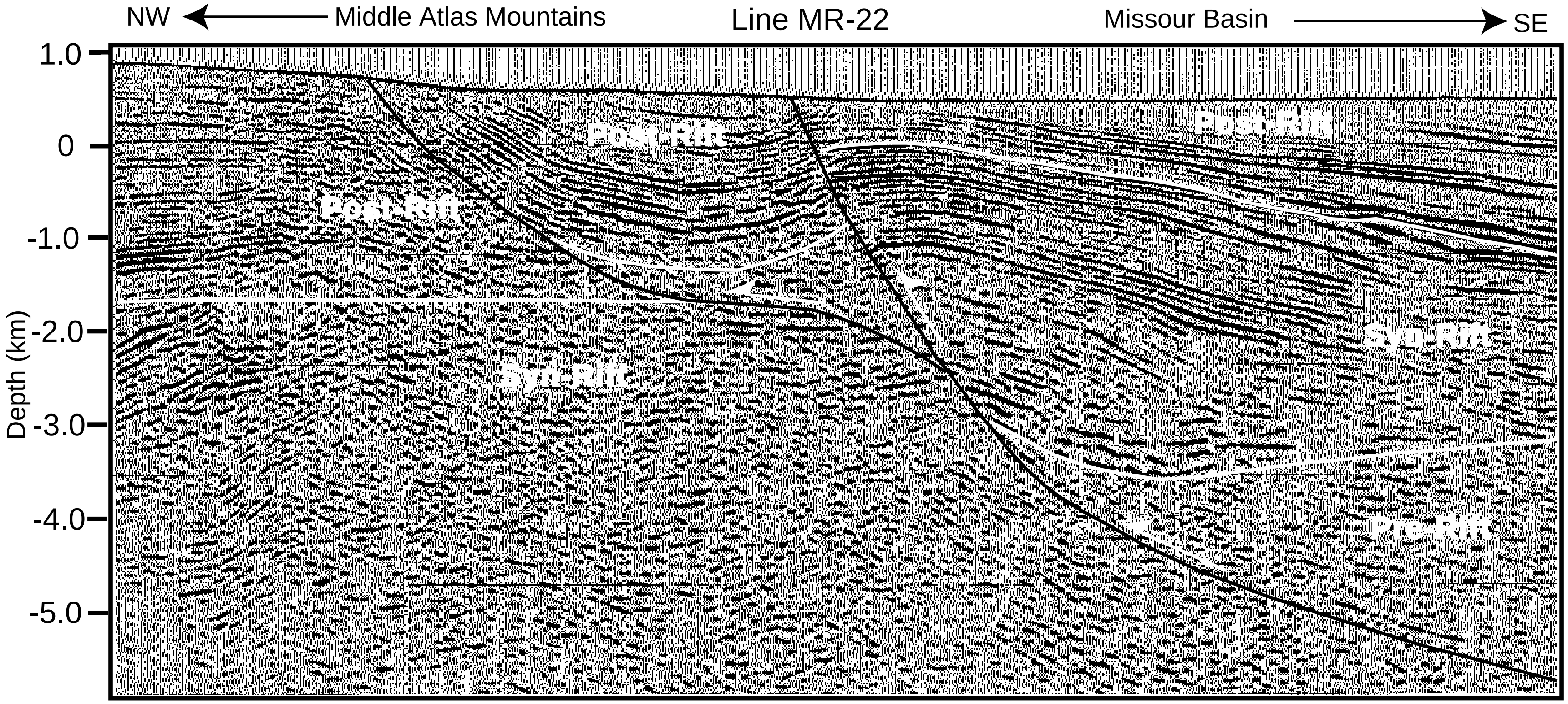
Both planar and listric faults were formed in the extensional phase of rifting of the Atlas rift systems. Listric faults interpreted in the Missour basin shallow to a detachment in the Paleozoic. One such example is seen on seismic line MR22 trends northwest along the margin of the Middle Atlas mountains (Figures 2.12 & 2.18). Thickening of the syn-rift sequence is clearly evident from the southeast to the northwest into a listric normal fault that dips to the southeast. This line was migrated, depth converted, and plotted at a one-to-one scale for modeling and balancing of the section. Line MR22 was correlated to a cross line that was then tied to well OSD-1 using a time-depth curve to the north of line MR22 (Figure 2.12). The base Cretaceous and Hercynian unconformities diverge from southeast to northwest, and the

Figure 2.17 Interpretations of regional seismic profiles across the Missour basin. Note the thickening of the syn-rift sequences (Triassic-Jurassic) out of the Missour basin into the paleo-Atlas rifts (Middle and High Atlas mountains). Regional inversion effects can be seen by the uplift of the stratigraphic section near the basin margins. The base Cretaceous unconformity has in some cases may have removed the sense of syn-rift thickening into the rift (KB-2). The location of Figure 2.13 and 2.18 are shown.



Middle Atlas
Missour Basin
NW
SE
Figure 14
Turonian-Cenozoic
Cenomanian
Jurassic/
Triassic
Carboniferous
KB11
0
km
10
TWT (Seconds)
0.0
0.4
0.8
1.2
1.6
OSD-1
Middle Atlas
Missour
Basin
NW
SE
Turonian-Cenozoic
Cenomanian
Jurassic-Triassic
Carboniferous
KB-2
0
km
10
TWT (Seconds)
Middle Atlas
Missour Basin
NW
SE
Turonian-Cenozoic
Cenomanian
Jurassic
Triassic
Carboniferous
KB-8
0
km
10
TWT (Seconds)
Middle Atlas
Missour Basin
High Atlas
NW
SE
Figure 13
Turonian-Cenozoic
Cenomanian
Jurassic-Triassic
Carboniferous
MR-7
0
km
5
TWT (Seconds)

Figure 2.18 Seismic line MR-22 was migrated and depth converted to enable the modeling of the reactivated listric normal fault seen on the seismic profile. Thickening of the syn-rift sedimentary rocks is evident from the southeast to the northwest into a listric normal fault that dips to the southeast.



NW
Middle Atlas Mountains
Line MR-22
Missour Basin
SE
Depth (km)
1.0
0
-1.0
-2.0
-3.0
-4.0
-5.0
Post-Rift
Post-Rift
Post-Rift
Syn-Rift
Syn-Rift
Pre-Rift

Hercynian unconformity has been uplifted so that it is near horizontal. This seismic line is important as it illustrates what is believed to be a common inversion structural style in the Missour basin and the Atlas mountains. Listric normal faults as illustrated on line MR22 are easier to reactivate than steeper planar faults due to a lower friction on the fault plane, particularly when the maximum compressive stress is horizontal. The listric fault on line MR22 was initially inverted by slip along the lower part of the fault, as the hanging wall half graben was rotated up the fault plane. As the dip of the fault plane steepened it became easier to generate a new fault that cut through the footwall as a lower angle thrust fault, than to continue the inversion of the hanging wall up the steeper section of the original syn-rift normal fault (Figure 2.19A). This new shortcut fault formed a ramp-flat geometry similar to that generated in a fault-bend-fold style of deformation associated with purely compressional tectonics. Continued shortening across the fault zone develops a fault-bend-fold over the ramp in the footwall. The generated fold is an open fold that is slightly asymmetrical. Further shortening across the fault zone results in the initiation of a breakthrough fault along the forelimb of the fault-bend-fold, and continued shortening is accommodated by uplift along the steeper reverse fault (Figure 2.19B). This breakthrough by reverse faulting occurs instead of further shortening along the ramp by the fault-bend-fold. Reconstruction of line MR22 to the base Cretaceous unconformity (Figure 2.19C) illustrates the geometry of the half graben prior to inversion. Stratigraphic relationships in the half graben east of the listric fault indicate there may have been a phase of uplift prior to the Cretaceous, as erosion and folding are apparent in the syn-rift sedimentary rocks. The syn-rift geometry may also represent the fold geometry associated with the syn-rift listric fault.

Using relationships from the restoration of line MR22 to a post-rift pre-inversion phase, an estimate for the amount of shortening across the fault zone can be calculated. This technique has been proven to be effective for the estimation of β as defined by *Mckenzie* (1978) for normal faults across tilted fault blocks (*Le Pichon et al.*, 1982).

Φ=angle between listric fault and pre-rift unconformity

Ψ=angle between pre-rift unconformity and a syn-rift stratigraphic horizon

$\beta = \sin\Phi/[\sin(\Phi-\Psi)$

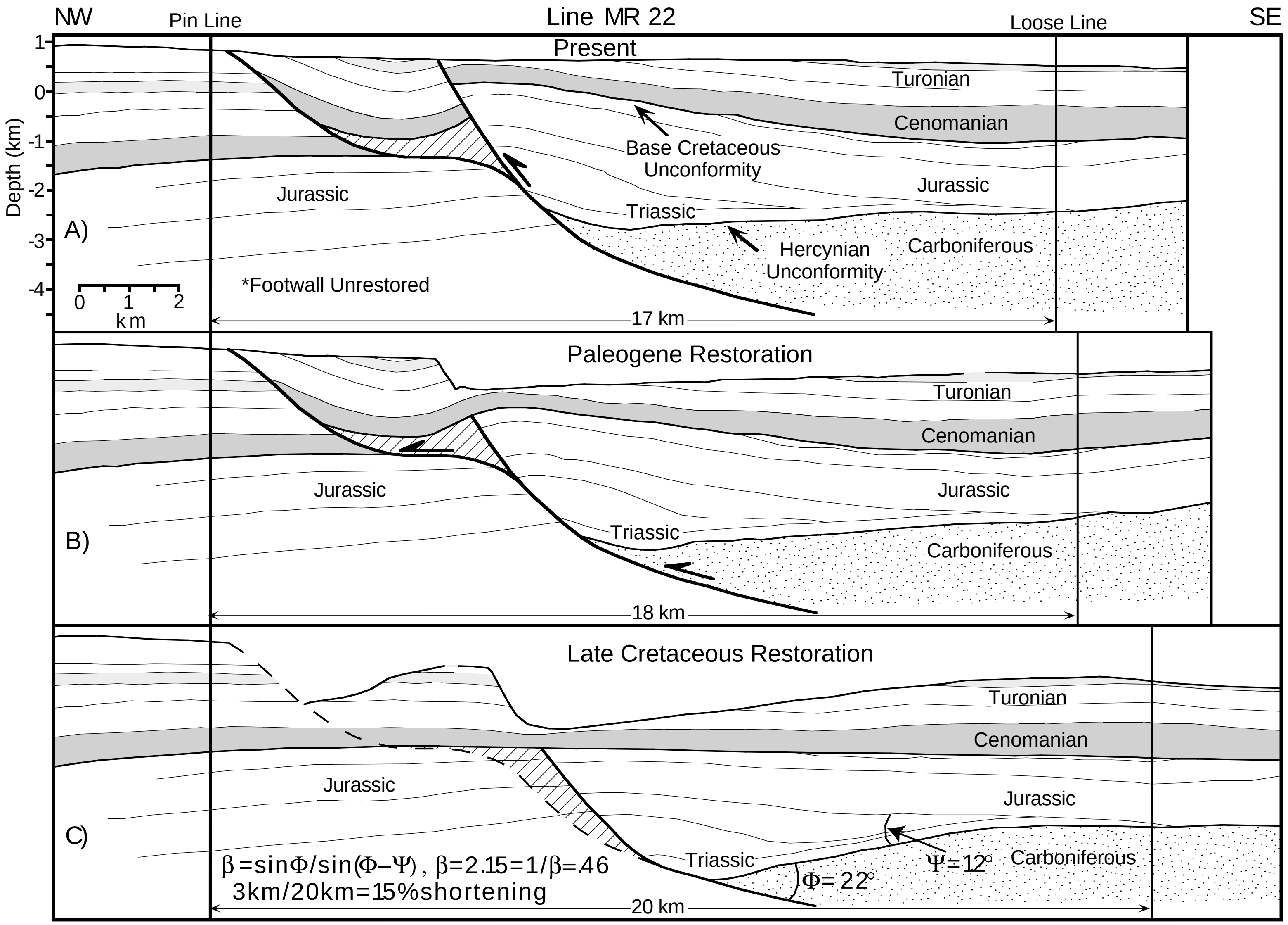
$\Phi=22°$, $\Psi=12°$, $\beta=2.15$ (stretching factor,$1/\beta = 0.46$)

The amount of stretching obtained of 2.15 for this half graben in the Missour basin is the ratio of the crust before and after extension as defined by *Mckenzie* (1978). Values of β for the North Sea (Figure 2.3) based on seismic reflection data are approximately 1.4 (*Barr*, 1987). Stretching values (β) for the Rhine graben = 1.15, the Armorican margin = 1.53, and the Afar rift system = 3.0 (*Le Pichon et al.*, 1982).

Similarly, the subsequent amount of shortening across the fault zone can be calculated using the distance between the pin line and loose line of the restoration of line MR22 (Figure 2.19). The distance between the pin line and loose line prior to restoration is 17 kilometers, and the distance before shortening is 20 kilometers. The resulting amount of shortening across line MR22 is approximately 15%.

High-angle normal faults such as interpreted on line MR17 (Figure 2.15),

Figure 2.19 Restoration of migrated and depth converted seismic reflection line MR-22. Reactivation of the syn-rift listric fault occurs until the fault steepens, and the syn-rift fault is bypassed. Shortening is then accommodated by a thrust that cuts the footwall at a lower angle (C). A fault-bend-fold forms over the new footwall ramp (B), and is later faulted along the forelimb by the reactivation of the original syn-rift normal fault (A). Shortening then occurs in the hanging wall along the high-angle reverse fault. Shortening of the original half graben is approximately 15%. The thinning factor β is approximately 0.46 that is the inverse of *Mckenzie's* (1978) stretching factor β.



NW
Pin Line
Line MR 22
Loose Line
SE
Depth (km)
1
0
-1
-2
-3
-4
Present
Turonian
Cenomanian
Base Cretaceous Unconformity
Jurassic
Triassic
Hercynian Unconformity
Carboniferous
A)
*Footwall Unrestored
0 1 2
km
17 km
Paleogene Restoration
Turonian
Cenomanian
Jurassic
Triassic
Carboniferous
B)
18 km
Late Cretaceous Restoration
Turonian
Cenomanian
Jurassic
Triassic
Carboniferous
C)
Ψ=12°
Φ= 22°
β=sinΦ/sin(Φ−Ψ), β=2.15=1/β=.46
3km/20km=15%shortening
20 km

will accommodate much less shortening with stress oriented normal to the fault plane. Steep faults in the Missour basin are accommodating strain created by horizontal compression by strike-slip or oblique-slip deformation. A significant amount of strain may be accommodated by the strike-slip and oblique-slip movement on high-angle normal faults, but the amount of strain and shortening across these faults is difficult to quantify due to movement in and out of the plane of the seismic section. The combination of both high-angle and listric normal faults in the Missour basin and Atlas mountains results in a complex tectonic history after inversion. Horizontal stress applied to the syn-rift fault systems of the Atlas have probably resulted in deformation by oblique-slip reverse movement as well as fault-bend-fold, and a thin-skinned style of deformation resulting from inversion.

Inversion Tectonics

Structural inversion related to intracontinental rifting occurs when extensional rift faults reverse their sense of motion during subsequent episodes of compressional tectonics. Features generated by extension such as half grabens are uplifted to form positive anticlinal structures.

It is important to validate that a structure is actually an inversion feature, or a newly generated compressional structure, since reactivated rift faults have prospective stratigraphic relationships. A common feature of rifting is syn-depositional growth strata, or a progressive unconformity associated with an extensional half graben. Reactivation of the growth fault by later compression normal to the fault will result in a thicker syn-rift section in the hanging wall anticline than in the footwall syncline. It is sometimes difficult to recognize inversion when uplift of the hanging wall has occurred to the point where erosion has removed any indication of the original thickening associated with the half graben.

The reactivation of normal syn-rift faults invert previous half grabens into anticlinal structures, with the axis of the half graben centered below the axis of the inversion anticline. Anticlinal structures in the Missour basin and Atlas mountains frequently represent an inverted half graben. Because of this relationship hanging wall anticlines formed by the inversion of syn-rift half grabens, will have a thicker syn-rift section in the hanging wall than in the footwall even though the hanging wall may be above the footwall. The resulting inverted fold geometry is controlled by the geometry of the extensional fault (planar or listric) and the depth of detachment (e.g., *Mitra,* 1993). Reactivated listric faults normally form inversion anticlines that exhibit fault-bend-fold geometry, and allow for greater shortening than planar faults. This style of inversion and shortening may have contributed to the creation of the high elevations (>4000 meters) in the High Atlas mountains. Inversion along listric faults can generate a compressional fault-bend-fold that is more open or symmetrical than a fault propagation fold generated by the reactivation of a planar fault (*Mitra*, 1993). Folds generated by reactivated normal faults are commonly associated with fault breakthrough along the forelimb of the anticlines, as well as footwall thrusts.

Half grabens formed during rifting are divided into a syn-rift and post-rift sequence separated by an unconformity (base Cretaceous in the Missour basin) (Figure 2.13). Sedimentary rocks associated with the post-rift sequence normally have lower dips usually associated with regional subsidence into the paleo-rift basins. The syn-rift sequence on the other hand is associated with steeper dips related to the original hanging wall fold shape and stratigraphic growth. The dips of sedimentary rocks in the inverted anticlinal structure are steeper in the core of the anticline than along the flanks of the inverted structure. Planar faults that do not develop hanging wall anticlines frequently preserve the relationships of steeper dipping syn-rift sedimentary rocks after they are inverted into fault propagation folds. The structure at Jebel Missour exhibits this relationship, where flat-lying post-inversion Neogene sedimentary rocks onlap more steeply dipping Upper Cretaceous rocks, that in turn overlie Jurassic rocks exhibiting yet even steeper dips.

Listric faults that generate hanging wall rollover folds during extension must be "unfolded" during inversion, before the hanging wall can be refolded into an inversion anticline (*Mitra,*1993). The restoration of the inverted listric fault on seismic line MR-22 (Figures 2.18 and 2.19) show that there are syn-rift reflectors indicating dips that may be related to an earlier extensional hanging wall rollover fold. It is difficult to restore an inverted listric normal fault to the exact geometry present prior to inversion, possibly due to slip out of the plane of the section.

Inverted structures are important exploration objectives, as extensional half grabens may contain source rocks as well as reservoir rocks. Hydrocarbons generated during rifting are generally trapped in the syn-rift sequence up-dip along the crests of footwall anticlines associated with extensional half grabens, such as in the North Sea. The inversion of faults associated with footwall anticlines results in the uplift of the hanging wall graben above and sometimes over the top of the original footwall anticline (Figure 2.19). The result of this inversion is the creation of a hanging wall compressional fault-bend-fold, in combination with a subthrust footwall anticline/ramp. This relationship produced by inversion tectonics creates the opportunity for stacked structural traps, both in the original footwall anticline, and in the inverted hanging wall fold. This type of potential trap could be important with significant shortening and inversion. Large amounts of horizontal shortening may place Triassic sedimentary rocks containing salts and evaporites above Jurassic source and reservoir rocks in the footwall providing an effective seal. This seal formed by the thrusting of the Triassic sequence could help to maintain traps formed during rifting in the Jurassic, and help to form a new trap that can collect hydrocarbons that are remigrated by subsequent inversion.

Hydrocarbon Potential

The Missour intermontane basin was formed by the uplift and inversion of the margins of the Atlas rift system. The stratigraphic relationships of the Missour basin are such that the distribution of syn-rift sedimentary rocks (Triassic-Jurassic) of source rock facies, as in most rift basins, would not be expected to be encountered along the margins of the rift. Most source rocks related to syn-rift phase of the Atlas rift system would have been deposited in what is now the High and Middle Atlas mountains (Atlas paleo-rift). For this reason, the most prospective areas for hydrocarbon traps in the basin are along the margins of the High Atlas and Middle Atlas mountains. These regions are important exploration fairways as hydrocarbons generated during the syn-rift and post-rift phases of the Atlas rift system would have migrated updip towards the basin margins. Hydrocarbons trapped in the original rift structures may have remigrated towards the margins of the Atlas mountains upon the inversion of the rift system. There are two potential hydrocarbon systems present in the Missour basin and Atlas mountains: (1) the Paleozoic-Triassic system sourced from the Paleozoic, sealed by Triassic salts and basalts with potential Triassic sandstone reservoirs, and (2) the post-Triassic system with potential sources and reservoirs in the Jurassic and Cretaceous. The structural models developed for the inversion of syn-rift normal faults in the Missour basin and High Atlas mountains are favorable for trapping hydrocarbons in the following scenarios:

A) The inversion of a listric normal fault may position Jurassic source rocks in the footwall of fault-bend-fold style structure. This creates the possibility of moving immature syn-rift source rocks into the oil window in the footwall of a thrust. Triassic evaporites may in some cases be thrust over source and reservoir rocks in the footwall, creating an effective top seal.

B) Inversion creates the possibility of preserving hydrocarbons trapped during rifting. As observed along the margins of the Atlas mountains, not all syn-rift faults are reactivated. Some extensional structures that trapped hydrocarbons during rifting may be preserved beneath thrust faults originating from reactivated listric faults (footwall shortcut faults).

C) Paleozoic source rocks (Carboniferous-Silurian-Devonian) are more prospective in the Missour basin and the margins/shoulder areas of the paleo-Atlas rift systems. Paleozoic source rocks in the Atlas rift would probably have been buried too deeply to have any remaining source rock potential. A lower geothermal gradient and depth of burial in the Missour basin and other rift margins may yield potential traps sourced by Paleozoic source rocks.

Hydrocarbons sourced from the Paleozoic may be trapped in Triassic sandstones and sealed by overlying and interbedded salts, anhydrites and basalts of the Triassic.

Source rock samples were collected in the field in the Middle Atlas mountains to the west of the Missour basin that indicate favorable source rock parameters (Table 2.2). Upper Pliensbachian source rocks yield total organic carbon values between 1.66-3.87%. The Tmax (Rock Eval) values of 421-437°C for these upper Pliensbachian marls (Type II) are approximately equivalent to vitrinite reflectance values of 0.5-0.6 (Ro) (*Miles,* 1989). These data indicate that the upper Pliensbachian (Dommerian) source rocks are early mature for oil generation. These source rocks may be similar to source rocks of the same age found in the Paris Basin and other parts of central and southwest Europe (*Hallam,* 1987). dditional potential may be present in Pliensbachian-Bajocian reefs developed along the margins of the basin such as are found at Jebel Mechkakour, sourced by rocks of the same age (Figures 2.6 and 2.12) (*El Alji & Ouazzaba,* 1995). Maastrichtian source rocks yield high total organic values (~18%), and have excellent potential for generating hydrocarbons when they occur in the footwall of a sub-thrust style structure. These Maastrichtian source rocks may be related to other known source rocks deposited in the Late Cretaceous during a transgressive phase, which were widespread organic-carbon rich sediments in Morocco (*Schlanger et al.,* 1987). Shales and similar deposits grade into laterally equivalent phosphorites of Late Cretaceous to early Tertiary age laid down on the southern margin of Tethys, stretching from North Africa into the Middle East (*Hallam,* 1987). The OSD-1 well drilled in the western Missour basin near the Middle Atlas mountains.

Table 2.2. Geochemical Data from the Middle Atlas Mountains and Well OSD-1

Middle Atlas Mountains

Latitude	Longitude	TOC	S1	S2	S3	Tmax	HI	OI	Age
33°25.46N	4°20.65W	3.87	0.81	2.5	0.96	437°C	323	25	Pliensb.
33°25.66N	4°20.65W	1.66	6.61	8.4	0.23	421°C	506	14	Pliensb.
33°08.53N	5°09.48W	18.12	6.21	117.4	2.22	419°C	648	12	Maastricht.

Well OSD-1

Latitude	Longitude	Depth	TOC	S1	S2	S3	Tmax	HI	OI	Age
33°11.24N	3°48.12.5W	1768m	1.92	0.09	0.27	0.23	474°C	14	11	Westphalian
33°11.24N	3°48.12.5W	2066m	1.64	0.01	0.43	0.14	452°C	26	8	Westphalian
33°11.24N	3°48.12.5W	2401m	11.44	0.52	18.7	0.62	439°C	163	5	Namurian
33°11.24N	3°48.12.5W	2556m	1.28	0.04	0.52	0.15	453°C	40	11	Namurian

TOC = weight % organic carbon   Tmax =pyrolitic yield in °C

S1,S2 = mg hydrocarbons/g rock   HI = S2*100/TOC

3 = mg carbon dioxide/ g rock   OI = S3*100/TOC

Conclusions

The integration of surface geological mapping, seismic reflection data, well data, and remote sensing has given a better understanding of the tectonic processes, timing, and structural styles resulting from the various deformational phases forming the Atlas mountains and associated basins.

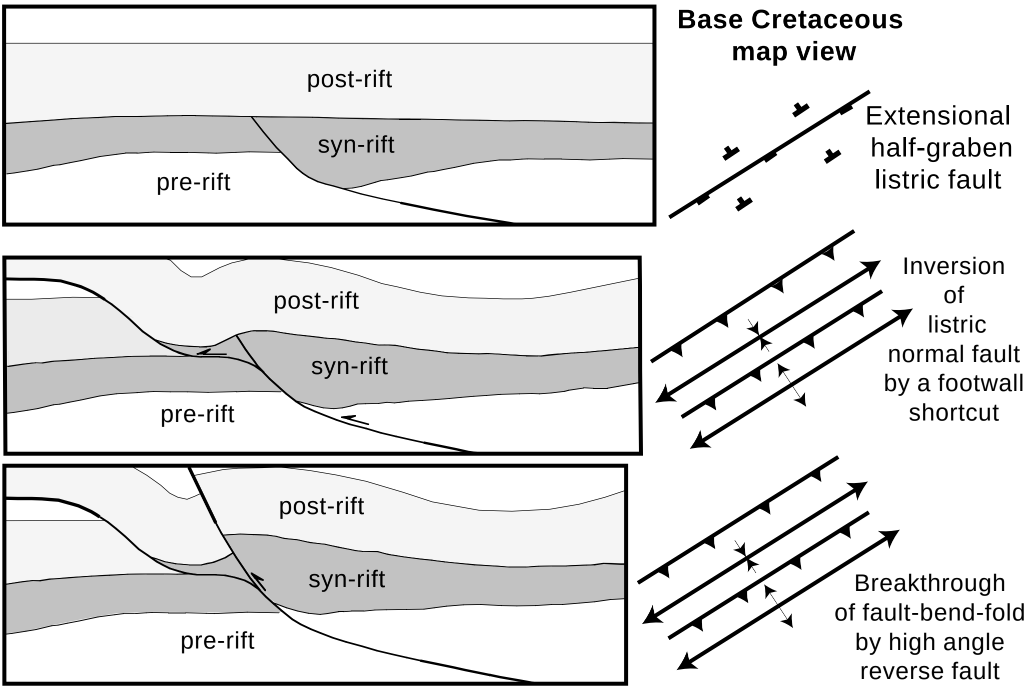
The Missour basin was a stable shelf margin separating the High and Middle Atlas rift systems. Regional shortening across the region during the Late Cretaceous and Tertiary resulted in the uplift of the entire Missour basin region contemporaneously with the uplift/inversion of the paleo-Atlas rifts. Shortening across the region has occurred mainly along the margins and the interior of the paleo-Atlas rift systems. Most of the shortening across these rift systems was along pre-existing faults formed during Mesozoic rifting or the previous Hercynian orogeny. The geometries of structures generated by inversion are controlled by the type of extensional faults formed during rifting and the orientation of the maximum compressive stresses relative to these faults (Figure 2.20). There are a variety of fault types (planar and listric), fault polarities, and fault distributions present within the Missour basin and the adjacent Atlas mountains.

Several phases of deformation resulted in the shortening and inversion of the paleo-Atlas rift system and the Missour intermontane basin; uplift related to the Hercynian orogeny, an uplift phase in the Middle Jurassic, uplift in the Early Cretaceous related to subsidence, a major uplift/inversion phase in the Early Tertiary (Paleogene), recent deformation in the Neogene-Quaternary (left-lateral oblique-slip) along the Middle Atlas-Missour basin margins, and right-lateral oblique-slip movement along the High Atlas-Missour basin margin (Figure 2.3a). The Missour basin may have been uplifted/inverted by the culmination of several phases of deformation. The region was previously part of the shelf margin or shoulder of the Atlas rift system and has been uplifted and translated to the northeast.

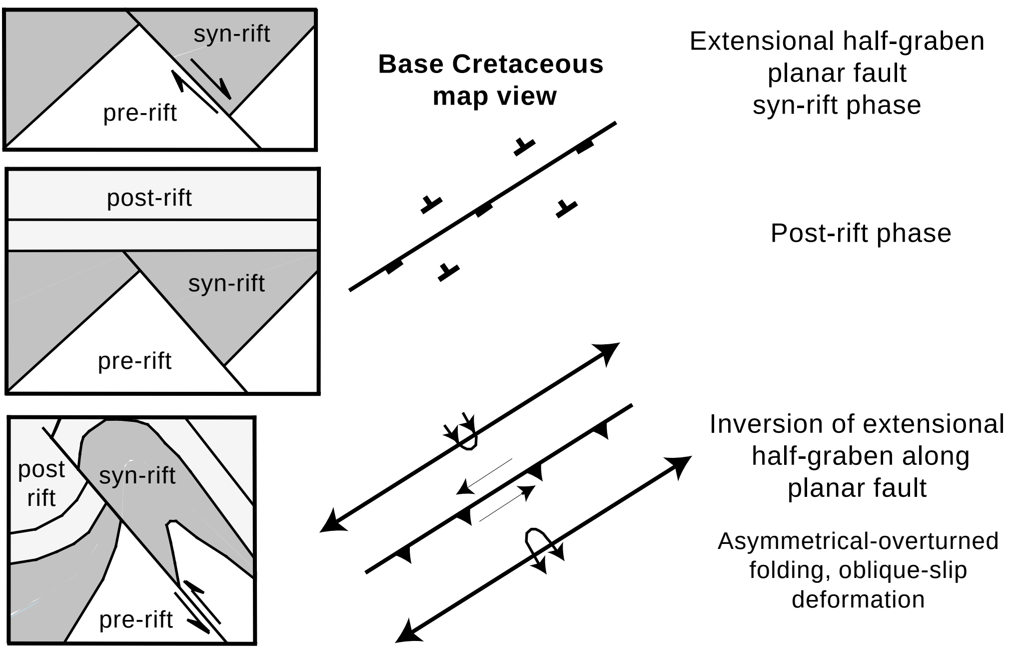
The combination of both high-angle and listric normal faults in the Missour basin and Atlas mountains result in a complex tectonic history following inversion. Horizontal stress applied to the syn-rift fault systems of

Figure 2.20 Faults associated with the Atlas rift system occur as both listric and planar normal faults. Fault bend folds are frequently formed by the reactivation of listric faults, when compression is applied normal to the fault plane. Fault propagation folds can form when compression is normal to a fault plane of reactivated planar faults. Listric faults reactivate by dip-slip movement when subjected to normal compression, while planar faults commonly reactivate by oblique-slip movement.

Inversion of Listric Normal Faults



Inversion of Planar Normal Faults



the Atlas has resulted in deformation by oblique-slip reverse movement as well as fault-bend-fold, thin-skinned style of deformation resulting from inversion. Applying structural inversion models to observed structures in the Atlas mountains may present new exploration opportunities that have not previously been attempted as an exploraiton strategy in Morocco.

References

Ait Brahim, L., and P. Chotin, Oriental Moroccan Neogene volcanism and strike-slip faulting: *Journal of African Earth Sciences*, v.11, p. 273-280, 1990.

Barr, D., Lithospheric stretching, detached normal faulting and footwall uplift, in M.P. Coward, J.F. Dewey, P.L. Hancock, eds., Continental Extensional Tectonics*: Geological Society Special Publication,* London, *No. 28*, p. 75-94, 1987.

Beauchamp, Jaques, Triassic sedimentation and rifting in the High Atlas (Morocco), in W. Manspeizer, ed., *Triassic-Jurassic Rifting: Elsevier*, 477-497, 1988.

Bensaid, M., J. Kutina, A. Mahmood, and M. Saadi, Structural evolution of Morocco and new ideas on basement controls of mineralization: Global Tectonics and Metallogeny, v.3, 59-69, 1985.

Brechbuhler, Y.A., Bernasconi R. and Schaer J., Jurassic sediments of the central High Atlas of Morocco: Deposition, burial and erosion history, in V. Jacobshagen ed., The Atlas System of Morocco-Studies on its geodynamic evolution-*Lecture notes in Earth Science: Berlin, Springer-Verlag*, v.15, 201-215, 1988.

Brede R., Hauptmann, M., and Herbig, H., Plate Tectonics and intracratonic mountain ranges in Morocco-The Mesozoic-Cenozoic development of the Central High Atlas and the Middle Atlas: *Geologische Rundschau*, v.81/1, 127-141, 1992.

Coward, M.P., R. Gillcrist and B. Trudgill, Extensional Structures and their tectonic inversion in the Western Alps, in A. M. Roberts, G. Yielding and B. Freeman, eds., The geometry of normal faults: *Geological Society Publication*, London, v.56, 93-112, 1991.

Dewey, J. F., Helman, M. I. , Turco, E., Hutton, D.H.W., and Knott, S.D., Kinematics of the western Mediterranean, in M.P. Coward, D. Dietrich, R.G. and Park, eds., Alpine Tectonics*: Geological Society Publication*, London, v. 45, 265-283, 1989.

Doblas, M., and R. Oyarzun, Neogene extensional collapse in the western Mediterranean (Betic-Rif Alpine orogenic belt): implications for the genesis of the Gibraltar arc and magmatic activity: *Geology*, v.17, 430-433, 1989.

du Dresnay, R., 1971, Extension et development des phenemenes recifaux jurassiques dans le domaine atlasique marocain, particulierement au lias moyen: *Bulletin de la Société Géologique de France,* (7), v.13, 46-56, 1971.

du Dresnay, R., Recent data on the geology of the Middle-Atlas (Morocco), in V. Jacobshagen Ed., The Atlas System of Morocco-Studies on its geodynamic Evolution-Lecture notes Earth Science: Berlin, *Springer-Verlag*, v.15, 293-320, 1988.

El Alji, M., and M. Ouazzaba, Structural Styles and Hydrocarbon Potential of the Missour Basin, in D. Clark-Lowes, D. Macgregor, J. Fisher, and D. Moody eds., Hydrocarbon Geology of North Africa-First Symposium, *Geological Society of London Petroleum Group*, 21, 1995.

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 V. Jacobshagen ed., The Atlas System of Morocco-Studies on its geodynamic Evolution-Lecture notes Earth Science: Berlin, *Springer-Verlag*, v.15, 273-291, 1988.

Giese, P., and Jacobshagen, V., Inversion Tectonics of Intracontinental Ranges: High and Middle Atlas, Morocco: *Geologische Rundschau*, v.81/1, 249-259, 1992.

Hallam, A., Mesozoic marine organic-rich shales. in Brooks, J. and Fleet, A.J. eds., Marine Petroleum Source Rocks: *Geological Society Special Publication,* London, v. 26, 251-261, 1987.

Herbig, H. G., Synsedimentary tectonics in the northern Middle Atlas (Morocco) during the Late Cretaceous and Tertiary. in V. Jacobshagen, ed., The Atlas System of Morocco-Studies on its geodynamic Evolution-Lecture notes in Earth Science: Berlin-Heidelberg-New York, *Springer-Verlag*, v.15, 321-337, 1988.

Jacobshagen, V., Major fracture zones of Morocco: The South Atlas and the Transalboran fault systems, *Geologische Rundschau*, v.81, 185-197, 1992.

Jacobshagen, V., Brede, R., Hauptmann, M., Heinitz, W., and Zylka, R., Structure and post-Paleozoic evolution of the central High Atlas, in V. Jacobshagen, ed., The Atlas System of Morocco-Studies on its geodynamic Evolution-Lecture notes in Earth Science,: Berlin-Heidelberg-New York , *Springer-Verlag*, v.15, 245-271, 1988.

Katz, Barry J., A survey of rift basin source rocks, in J.J. Lambiase, ed., Hydrocarbon Habitat in Rift Basins*: Geological Society Special Publication*, London, v.80, 213-242, 1995.

Laville, E., and Pique, A. , Jurassic penetrative deformation and Cenozoic uplift in the Central High Atlas (Morocco): a tectonic model. Structural and Orogenic inversions: *Geologische Rundschau*, v. 81/1, 157-170, 1992.

Leblanc, D, Tectonic adaptation of the External Zones around the curved core of an orogen: The Gibraltar Arc: *Journal of Structural Geology*, v.12, 1013-1018, 1990.

Leblanc, D., and Ph. Olivier, Role of strike-slip faults in the Betic-Rifian Orogeny: *Tectonophysics*, v.101, 345-355, 1984.

Le Pichon, Xavier, and Jean-Claude Sibuet, Subsidence and stretching, in S. Joel *et al.* eds*., American Association of Petroleum Geologists Memoir*, v. 34, 731-741, 1982.

Loomis, T.P., Tertiary mantle diapirism, orogeny, and plate tectonics east of the Strait of Gibraltar: *American Journal of Science*, v. 275, 1-30, 1975.

Mattauer, M., P. Tapponnier, and F. Proust, Sur les mécanismes de formation des chaénes intracontinentales. L'exemple des chaénes atlasiques du Maroc: *Bulletin de la Société Géolologique de France*, v.7, 521-526, 1977.

McKenzie, D., Some remarks on the development of sedimentary basins: *Earth Planetary Science Letters*, v. 40, 25-32, 1978.

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*, v.148, 229-233, 1988 .

Michard, A., Eléments de géologie Marocaine, *Notes Mémoires Service Géolologique*, Rabat, v.252, 408, 1976.

Miles, Jenifer A., Illustrated Glossary of Petroleum Geochemistry, *Oxford University Press*, 126, 1989.

Miranda, J.M., J. Freire Luis, I. Abreu, L.A. Mendes Victor, A. Galdeano, and J.C. Rossignol, Tectonic framework of the Azores triple junction: *Geophysical Research Letters*, v.18, 1421-1424, 1991 .

Mitra, Shankar, Geometry and Kinematic Evolution of Inversion Structures: *American Association of Petroleum Geologists Bulletin*, v. 77, no. 7, 1159-1191, 1993.

Morel, J., Z. Mostafa, and A. Poisson, Relations entre la subsidence des bassins moulouyens et la creation des reliefs atlasiques (Maroc): un exemple d'inversion tectonique depuis le Neogene: *Bulletin de la Societe Geologique de France*, v.164 (1), 79-91, 1993.

Morley, C.K., Origin of a major cross-element zone: Moroccan Rif: *Geology*, v.15, 761-764 , 1987.

Morley, C.K., Developements in the structural geology of rifts over the last decade and their impact on hydrocarbon exploration, In: Hydrocarbon Habitat in Rift Basins, J.J. Lambiase: *Geological Society Special Publication*, London, v.80, 1-32, 1995.

Pique, A., Dahmani, M., Jeannette, D., and Bahi, L., Permanence of structural lines in Morocco from Precambrian to Present: *Journal of African Earth Sciences*, v. 6, no. 3, 247-256, 1987.

Rosendahl, B.R., D.J. Reynolds, P.M. Lorber, C.F. Burgess, J. McGill, D. Scott, J.J. Lambiase and S.J. Derksen, 1986, Structural expressions of rifting: lessons from Lake Tanganyika, Africa*: Geological Society of London Special Publication*, v.25, 25-43.

Schaer, Jean-Paul, and Rodgers, J., Evolution and Structure of the High Atlas of Morocco. In: Jean-Paul Schaer, and J. Rodgers, eds., The Anatomy of Mountain Ranges: *Princeton University Press*, 107-127, 1987.

Sengor, A.M.C., N. Gorurand and F. Saroglu, Strike-slip faulting and related basin formation in zones of tectonic escape; Turkey as a case study, in K.T. Biddle and N. Christie-Blick eds., Strike-slip deformation, basin formation and sedimentation: *Special Publication-Society of Economic Paleontologists and Mineralogists*, v.37, 227-264, 1984.

Schlanger, S.O., M.A. Arthur, H.C. Jenkyns and P.A. Scholle, The Cenomanian-Turonian anoxic event, 1.Stratigraphy and distribution of organic carbon-rich beds and the marine d 13C excursion, in J. Brooks, and A.J. Fleet, eds., Marine Petroleum Source Rocks: *Geological Society Special Publication*, London, v. 26, 371-399, 1987.

Stewart, I.J., R. P. Rattey, I.R. Vann, 1992, Structural style and the habitat of hydrocarbons in the North Sea, in R.M. Larsen, H. Brekke, B.T. Larsen and E. Talleraas, eds., Structural and Tectonic Modeling and its Application to Petroleum Geology, NPF Special Publication: *Elsevier*, 197-220, 1992.

Studer, M., and du Dresnay, R., Deformations synsedimentaires en compression pendant le Lias superior et le Dogger, au Tizi nÌIrhil (Haut-Atlas central de Midelt, Maroc). *Bulletin de la Societe Géologique de France Bulletin*: v.7, no. 22, 391-397,1980.

Westaway, R., Present-day kinematics of the plate boundary zone between Africa and Europe, from the Azores to the Aegean: *Earth Planetary Science Letters*: v. 96, 393-406, 1990.

Williams, G. D., C. M. Powell and M. A. Cooper, Geometry and kinematics of inversion tectonics, in M.A. Cooper and G. D. Williams eds., Inversion Tectonics, *Geological Society Special Publication,* London, v.44*,* 3-15*,* 1989*.*

Ziegler, P.A., Geological Atlas of Western and Central Europe, *Shell International Petroleum*, The Hague: 2 vol., 130 p, 40 pl., 1982.

CHAPTER THREE

Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological-geophysical transect [2]

Abstract

The High Atlas Mountains of North Africa were formed by the reactivation of a major intracontinental rift system that 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 20-30 Ma (Miocene-Oligocene), 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 kilometers wide, comparable to the width of the present-day Red Sea. Syn-rift and post-rift sedimentary rocks were uplifted by the reactivation of syn-rift normal faults, with further shortening along newly formed thin-skinned thrust faults. Structures formed by the reactivation of syn-rift faults resulted in structures

---

[2]Submitted as: Weldon Beauchamp, Richard W. Allmendinger, Muawia Barazangi, Ahmed Demnati, Mohamed El Alji and Mohammed Dahmani, Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological-geophysical transect, *Tectonics*, 1997.

with different geometries than those created by newly formed fault-bend and fault-propagation faults. 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 Mountains.

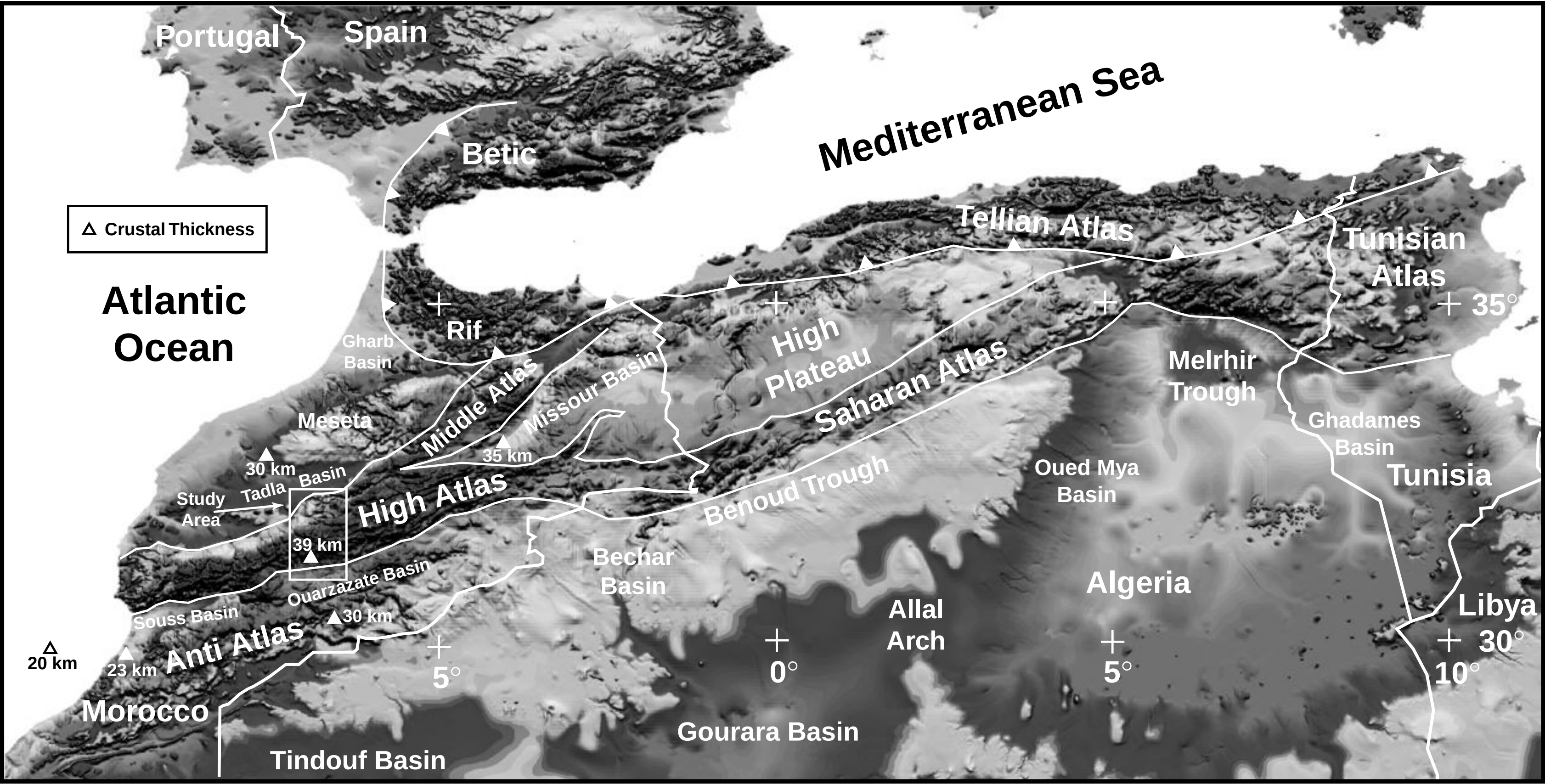
Introduction

Pre-existing intracontinental rift systems are important tectonic features as they focus strain during plate convergence along nearby plate boundaries. The style and magnitude of rifting (pure shear or simple shear) and the orientation of the rift system relative to plate convergence contribute to the magnitude of shortening and style of inversion. The overall geometries of rift systems may control the vergence or bi-vergence of thrusting during inversion. Stresses transmitted during plate convergence may result in the decoupling of the crust at different levels based upon the variation of crustal thickness and yield strength associated with rifts, resulting in multiple levels of detachment.

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). This rift system extended from what is now the Atlantic margin of Morocco, to the Mediterranean coast of Tunisia (~2000 km) (Figure 3.1). A geophysical-geological cross section (Transect A-A', Plate 3) was constructed across the central High Atlas Mountains of Morocco to develop a better understanding of how intracontinental mountain belts are formed (Figure 3.2).

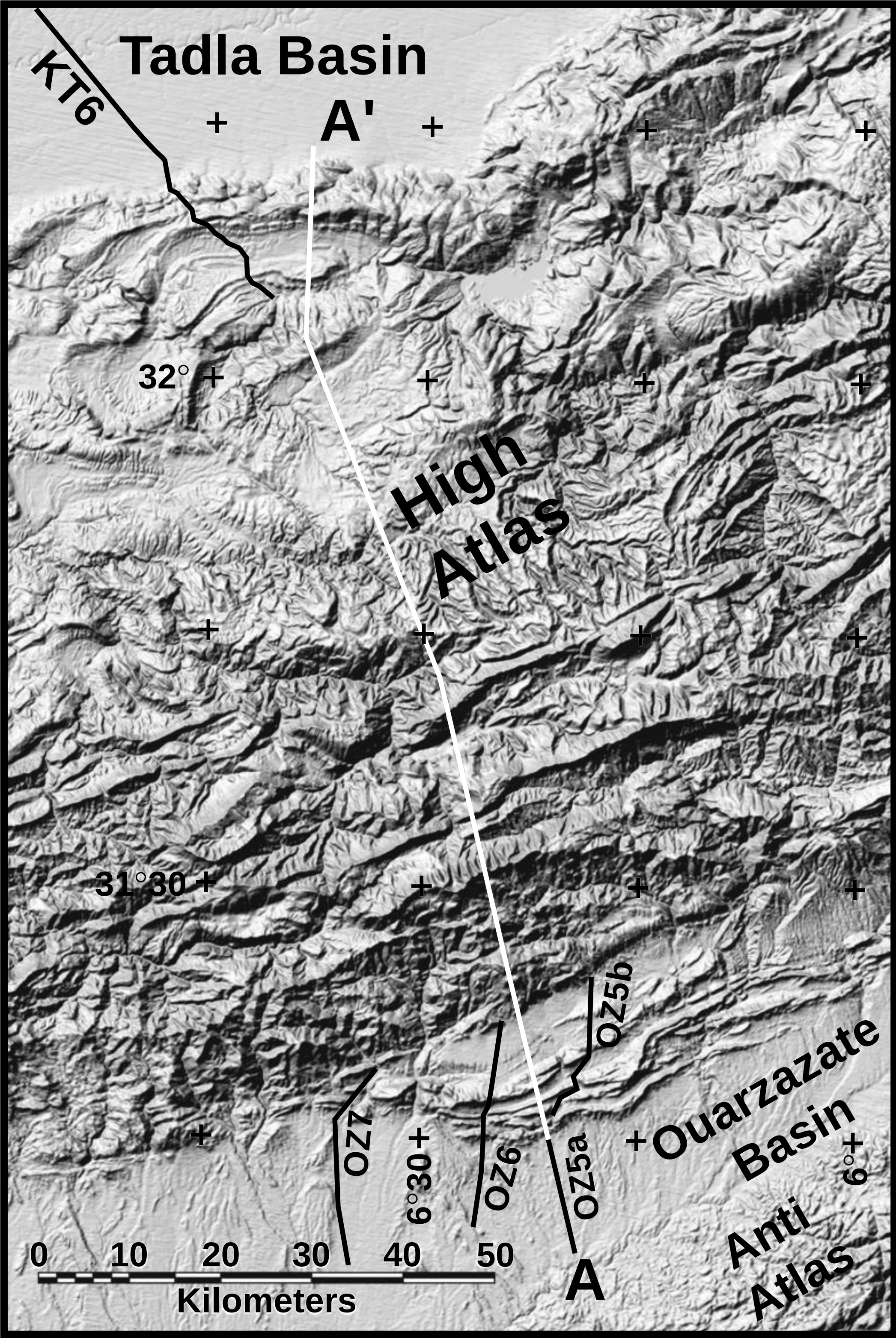
The construction of a balanced and restorable cross section across the High Atlas Mountains enables the study of the kinematic history of the mountain belt, as well as how strain resulting from convergence between the

Figure 3.1. Tectonic map of North Africa showing the location of the region of Transect A-A', the High Atlas, Anti-Atlas, Tell Atlas and Betic Rif orogens overlain on digital topography. Estimates of crustal thickness in Morocco are also shown.



Portugal
Spain
Mediterranean Sea
Betic
Crustal Thickness
Atlantic Ocean
Tellian Atlas
Tunisian Atlas
35°
Rif
Gharb Basin
High Plateau
Middle Atlas
Missour Basin
Saharan Atlas
Melrhir Trough
Ghadames Basin
Meseta
35 km
30 km
Tadla Basin
Study Area
High Atlas
Benoud Trough
Oued Mya Basin
Tunisia
39 km
Bechar Basin
Ouarzazate Basin
Algeria
Souss Basin
30 km
Allal Arch
Libya
Anti Atlas
20 km
23 km
5°
0°
5°
30°
10°
Morocco
Gourara Basin
Tindouf Basin

Figure 3.2. Map showing the location of seismic reflection profiles (KT6-Tadla Basin, OZ-5a & OZ-5b, OZ-6 & OZ-7) used in the study and the location of transect A-A'.



Tadla Basin
KT6
A'
32°
High
Atlas
31°30
OZ5b
Ouarzazate
Basin
OZ7
6°30
OZ6
OZ5a
6°
Anti
Atlas
A
0
10
20
30
40
50
Kilometers

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') are estimated to have been 36 km (Plate 3).The amount of shortening across what is interpreted as having been the original High Atlas rift basin (113 km in width, restored section) was estimated between reference lines (A) and (B) (Plate 3). Values for shortening indicate the Atlas rift basin was shortened by a minimum of 36 km (32%).Previous estimates of the amount of shortening across the High Atlas had been placed between 10-20% (*Brede*, 1992). 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. Several pin lines and loose lines were placed along transect A-A' to determine the magnitude of shortening across the High Atlas (Plate 3). Thrusting and inversion resulted in the largest magnitude of shortening along the margins of the High Atlas over a distance of 10 km (Plate 3, Transect A-A') with little shortening in the interior of the High Atlas Mountains along the cross section.

Many foreland fold and thrust belts have only one sense of thrusting and vergence. Intracontinental mountain belts are commonly bi-vergent and have a more complex history for the timing and sequence of thrusting. The bi-vergence of thrusts in intracontinental mountain belts may be related to the reactivation of rift fault systems that are often bi-vergent (e.g., *Bally,* 1984; *Fraissinet et al.*, 1988; *Hayward and Graham* 1989; *Jossen and Filali-Moutei*, 1992).A bi-vergent intracontinental mountain belt implies that at some point in the orogen there should be a null point where the sense of vergence will be reversed. Bi-vergent 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 bi-vergent mountain belt that has been affected by moderate to strong inversion (*Gauthier*, 1960; *Rolley*, 1978; *Lowell*, 1995).

Regional Setting

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 continental crust of North Africa contains significant sedimentary rift basins as a result of thinning of the crust during the Mesozoic . The convergence of the North African and European plates during the Cenozoic through Present may be partially accommodated by shear along Proterozoic zones of weakness in the crust (*Mattauer et al.,* 1972; *Milanovsky*, 1981; *Ziegle*r, 1982; *Laville and Pique*, 1991; *Giese and Jacobshagen,* 1992).

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 the crust beneath the Meseta (Figure 3.1) 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 (2-D) along two profiles across the High Atlas indicating crustal thickness between 34-38 km. *Makris et al.* (1985) noted 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 crustal root 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 3) .

Pre-Rift Phase

The Anti-Atlas south of the High Atlas Mountains expose Paleozoic through Precambrian 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 to 250 Ma).

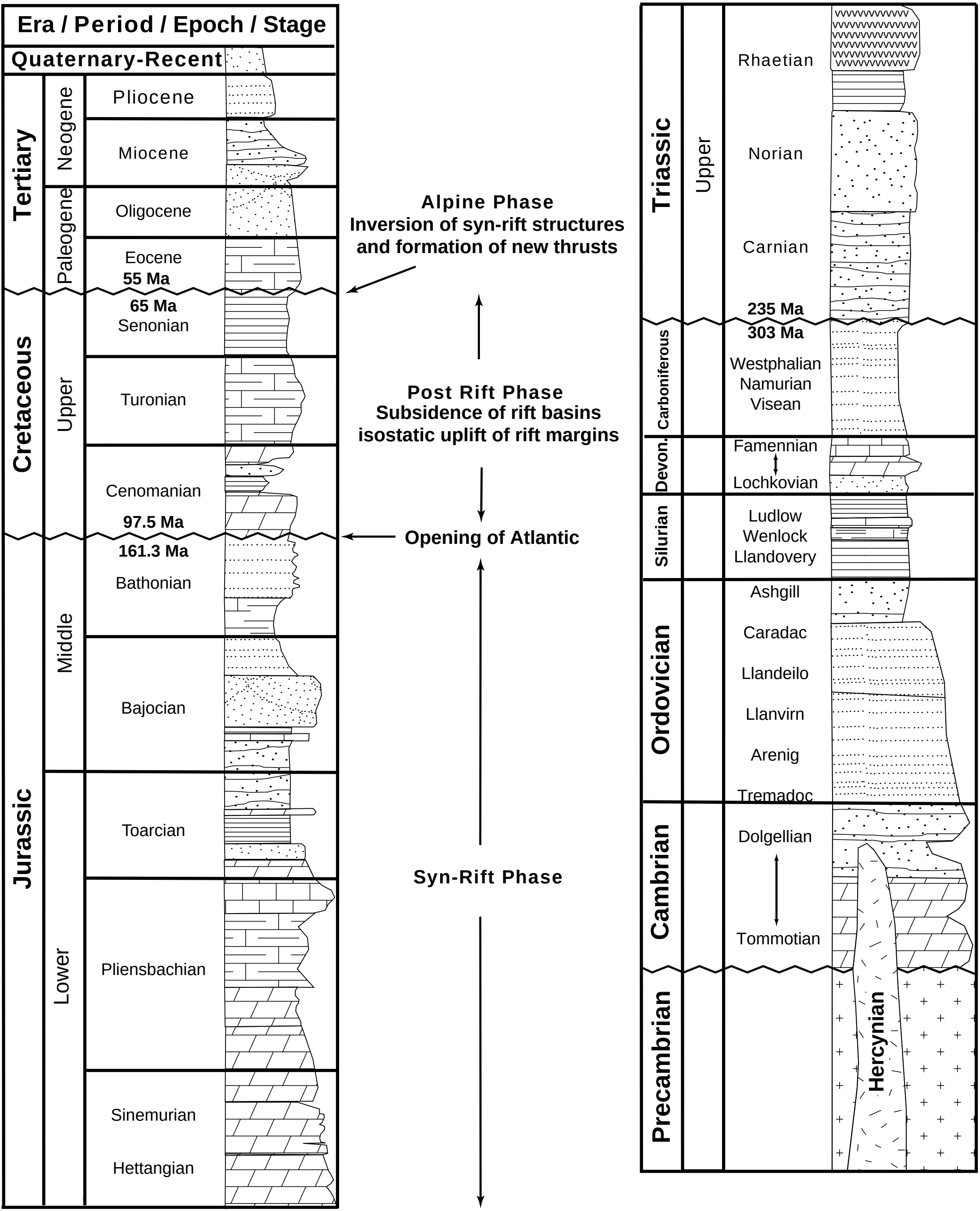
Precambrian granites and volcanics crop outin the Anti-Atlas at the southern end of the transect A-A' of this study (Figure 3.2). Similar Precambrian rocks have not been recognized to the north of the High Atlas Mountains. Igneous rocks exposed north of the High Atlas Mountains are granites of Hercynian 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 3.3). These Silurian shales range from 200 to 1000 meters in thickness, and provide an important sedimentary sequence for the nucleation of thrusts and duplexes within the Paleozoic. 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 3.2).

Syn-Rift Phase

The Atlas rift system began in the Triassic and was active through the

Figure 3.3. Generalized Stratigraphic section of the central High Atlas Mountains compiled from geological maps (*Saadi et al.*, 1977; *Saadi et al.*, 1979, 1985; *Fetah et al.*, 1990), measured sections (*Michard*, 1976; *Rolley*, 1978; *Jabour*, 1988; *Ensslin*, 1992), and well data (Tadla basin DRZ-1 and KMS-1).



Era / Period / Epoch / Stage
Quaternary-Recent
Tertiary
Neogene
Pliocene
Miocene
Paleogene
Oligocene
Eocene
55 Ma
Cretaceous
Upper
65 Ma
Senonian
Turonian
Cenomanian
97.5 Ma
Jurassic
Middle
161.3 Ma
Bathonian
Bajocian
Lower
Toarcian
Pliensbachian
Sinemurian
Hettangian
Alpine Phase
Inversion of syn-rift structures
and formation of new thrusts
Post Rift Phase
Subsidence of rift basins
isostatic uplift of rift margins
Opening of Atlantic
Syn-Rift Phase
Triassic
Upper
Rhaetian
Norian
Carnian
235 Ma
303 Ma
Carboniferous
Westphalian
Namurian
Visean
Devon.
Famennian
Lochkovian
Silurian
Ludlow
Wenlock
Llandovery
Ordovician
Ashgill
Caradac
Llandeilo
Llanvirn
Arenig
Tremadoc
Cambrian
Dolgellian
Tommotian
Precambrian
Hercynian

Jurassic (*Michard,* 1976; *Manspeizer et al.,* 1978; *Schaer and Rodgers*, 1987; *Laville*, 1988). The syn-rift sequence of the Atlas began with the deposition of continental Triassic clastics composed of interbedded sandstones, shales, anhydrites, and volcanics (*Manspeizer et al.*, 1978; *Beauchamp*, 1988; *Medina*, 1988). The basal Triassic is composed of continental red beds which are overlain by thoeliitic basalts (~400 m, western Atlas) (*Manspeizer et al.,* 1978). Thicknesses of Triassic strata range between 4000-5000 meters in the western High Atlas, with thicknesses of less than 1000 meters in the central High Atlas, eastern High Atlas, Middle Atlas Mountains, and Missour basin (*Manspeizer et al.*, 1978, *Beauchamp et al.,* 1996) (Figure 3.1). Triassic age rocks are absent in the Ouarzazate basin and Anti-Atlas south of the High Atlas Mountains, along the southern extent of the cross section A-A' (Figure 3.2). Well data in the Tadla basin (DRZ-1 and KMS-1) indicate that Triassic rocks north of the High Atlas have thicknesses between 350-450 meters. 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 meters in thickness.

Triassic evaporites provide a weak stratigraphic interval that controlled the development of Cenozoic contractional structures. In the field, Triassic evaporite 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).

Syn-rift Jurassic sedimentary rocks were deposited only within the Atlas rift system within the present High Atlas Mountains, with the exception of the Missour basin and High plateau (*Laville*, 1988; *Beauchamp et al.,* 1996)

(Figure 3.2). 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 sedimentary rocks display complex facies distribution, varying both normal to the Atlas rift basin axis as well as along strike. Jurassic sedimentary rocks of the eastern High Atlas (Figure 3.1) 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 that are more clastic and often contain thick sandstone and shale intervals (*Rolley*, 1978). The Late Jurassic sedimentary rocks are composed of thick massive limestones and dolomites (Liassic). Upper Jurassic (Oxfordian-Kimmeridgean-Tithonian) through Lower Cretaceous (Berriasian-Albian) sedimentary rocks in the High Atlas Mountains and margins are not well preserved, and may never have been deposited (Figure 3.3).

The syn-rift 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 E.,* 1988; *Beauchamp*, 1988; *Laville and Pique*, 1991). Extension combined with strike-slip movement has been used to explain different syntectonic sedimentary relationships such as facies changes, angular disconformities and abrupt changes in thickness. 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).

Post-Rift Phase

Post-rift 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 non-deposition.

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 meters (*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 European 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*, 1993). Thrusting and inversion during the Oligocene formed the current topographic relief for the High Atlas Mountains, yet Tertiary foreland basins to the north and south of the High Atlas are lacking. Tertiary 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 meters. The lack of developmentof foreland basins may indicate a high rigidity for the upper crust along the margins of the Atlas.

Structural/Geophysical Transect of the High Atlas

Geological/Geophysical Data

The location of transect A-A' (Plate 3) was chosen to parallel the direction of thrust transport (Figure 3.4) and as near as possible to the existing seismic reflection profiles along the northern and southern margins of the High Atlas (KT-6, north; OZ-5, south; Plate 4). 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 structures were mapped 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 GPS 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 3.2). 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.

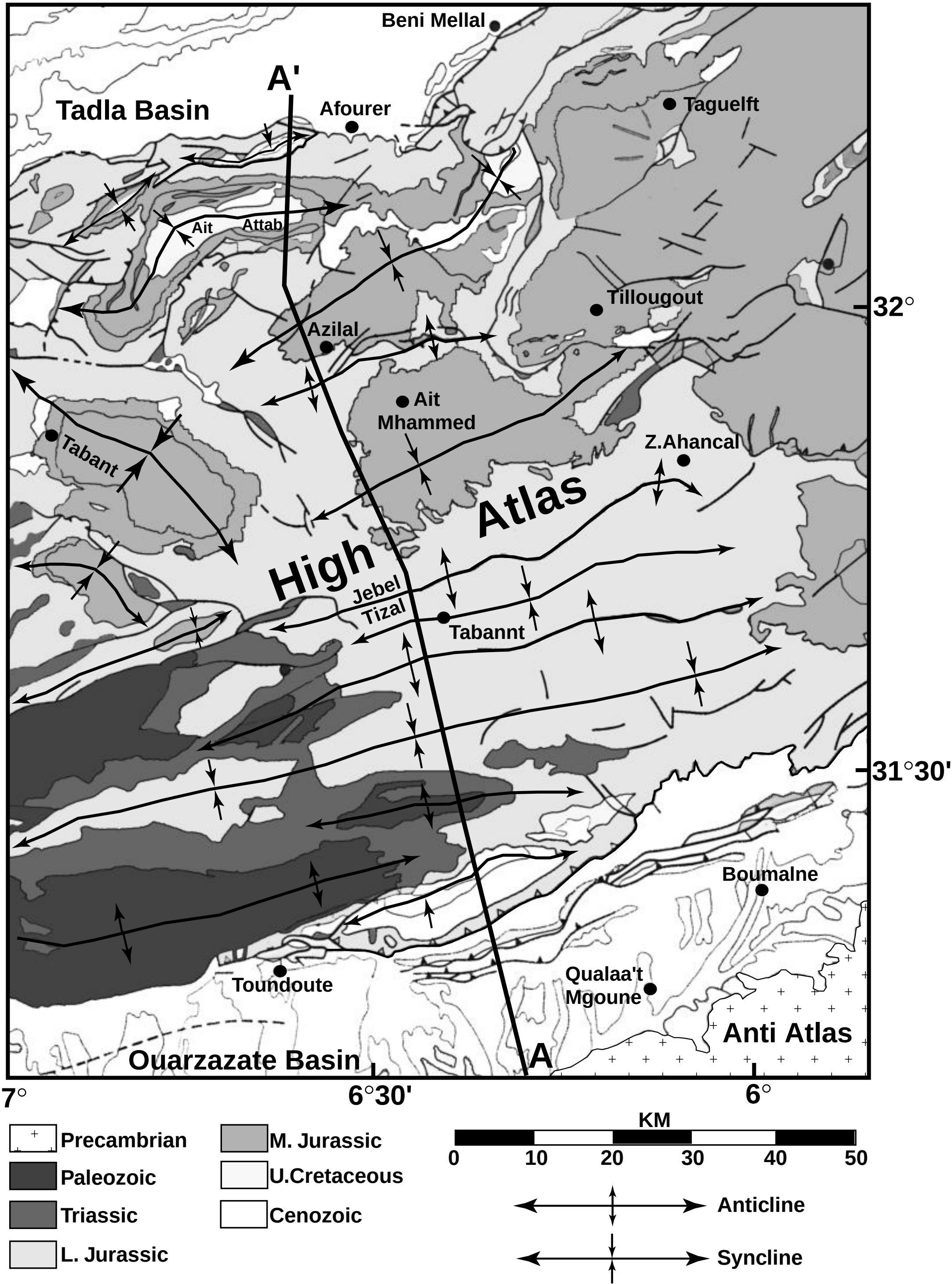
Syn-rift normal faults commonly can be identified by the presence of Jurassic basalts that were intruded along the fault plane, as well as sedimentary thickening into faults. In the field, many small Mesozoic normal faults thus identified have not been reactivated. A common characteristic of structures in the High Atlas along transect A-A' is the asymmetric style of fault-related folding. Many of the topographic ridges are directly related to structures of this nature. It is believed that these structures were generated by fault-propagation folds that originate from evaporites in the Triassic (Figure 3.5). 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 3) displayed together with digital topographic data illustrate the asymmetry of the High Atlas Mountains (Figure 3.6). 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 Bouguer gravity data show a decrease in the asymmetrical anomaly northward along the traverse A-A’ as does the topographic profile.

Given the amount of tectonic loading along the southern margin, and the lack of a significant foreland basin, the High Atlas Mountains appear to be uncompensated. Magnetic data displayed along traverse A-A’ shows a broad regional anomaly beneath the High Atlas Mountains of approximately 100 nT. Several smaller magnetic anomalies along the traverse are on the order of 10-15 km in wavelength, and show a 20-30 nT deflection in the regional magnetic gradient (Plate 3). 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 (Office National de Recherches et d’Exploitations Petrolieres) 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 post-stack processing was carried out on

Figure 3.4. 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).



Beni Mellal
A'
Tadla Basin
Afourer
Taguelft
Ait
Attab
32°
Tilllougout
Azilal
Ait
Mhammed
Tabant
Z.Ahancal
Atlas
High
Jebel
Tizal
Tabannt
31°30'
Boumalne
Toundoute
Qualaa't
Mgoune
Anti Atlas
Ouarzazate Basin
A
7°
6°30'
6°
Precambrian
Paleozoic
Triassic
L. Jurassic
M. Jurassic
U.Cretaceous
Cenozoic
KM
0
10
20
30
40
50
Anticline
Syncline

the lines south of the High Atlas Mountains (OZ-5, Ouarzazate basin) (Figure 3.2). These seismic profiles were chosen because of their orientation relative 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 OZ-5 were acquired by CGG (Compagnie Generale de Geophysique) for ONAREP using a Vibroseis source (50 meter shot spacing, 96 channels, 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 construction of the cross section A-A' (Figures 3.2 and 3.7). 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 3.7).

Description of Structures

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 3.7). 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 3). 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, Triassic and into the Ordovician without encountering Lower Cretaceous or Jurassic rocks (Plate 4). Reflections correlative to the

Figure 3.5 A photograph and corresponding geological sketch showing kink axes related to 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.

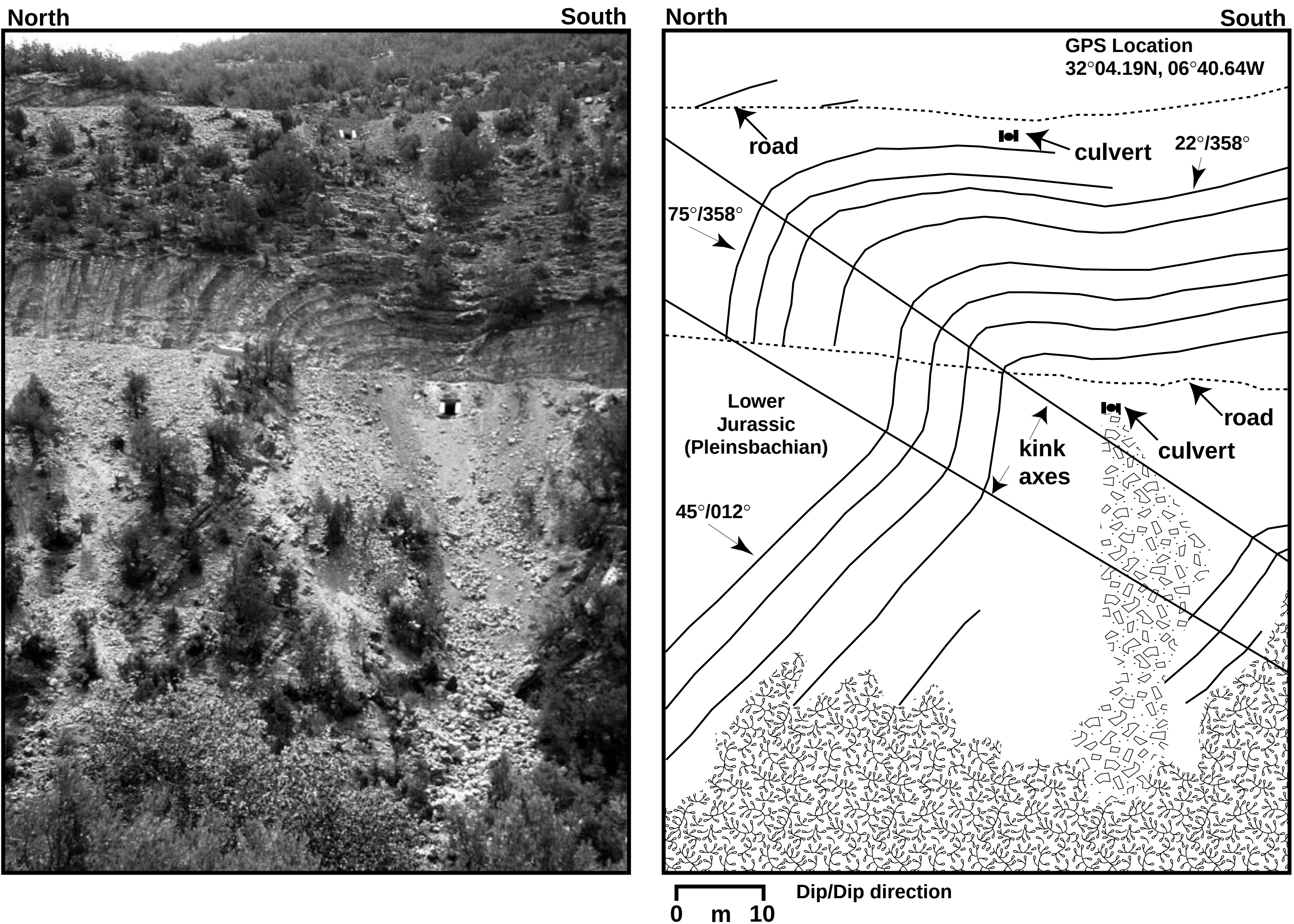
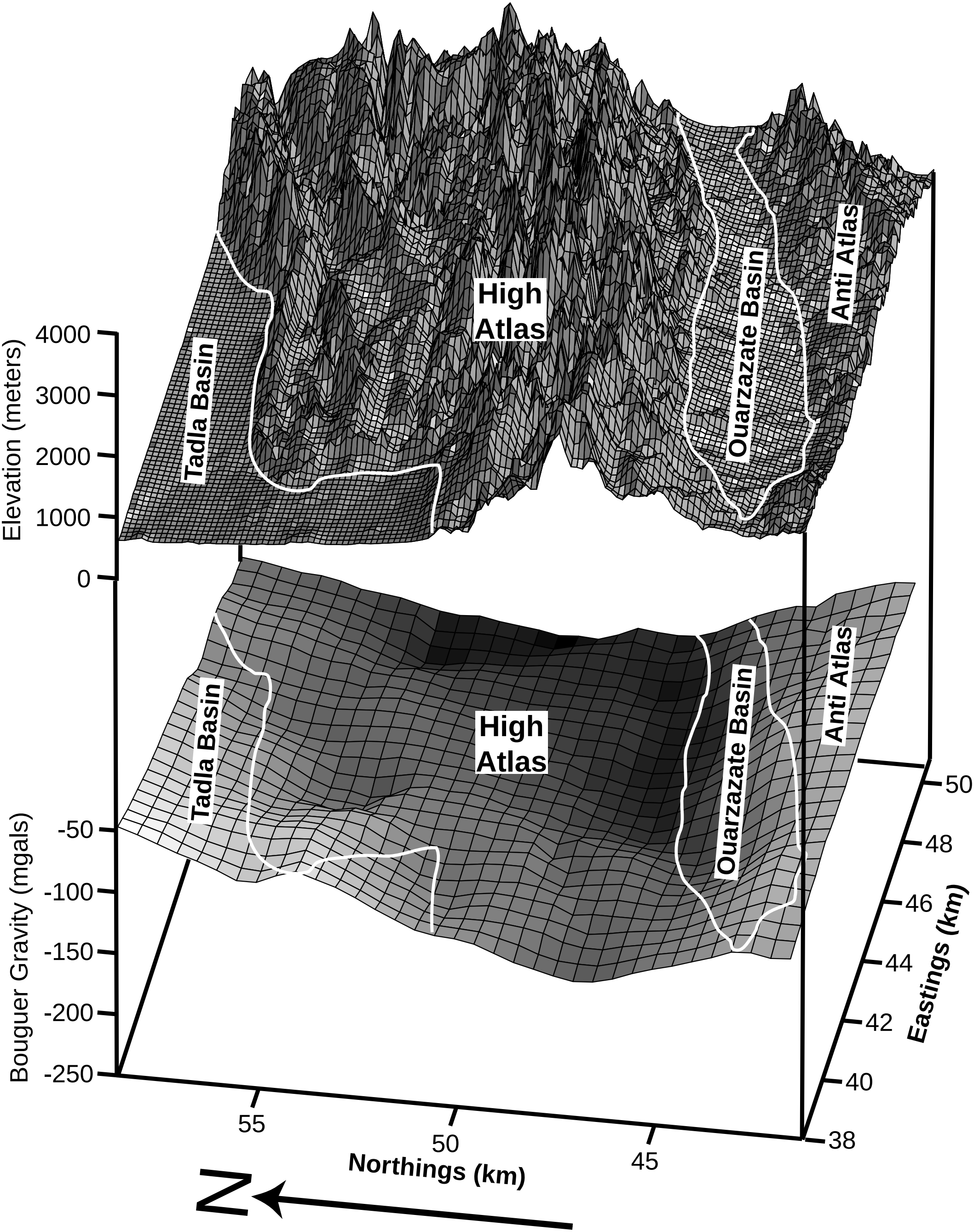

North
South
North
South
GPS Location
32°04.19N, 06°40.64W
road
culvert
22°/358°
75°/358°
Lower
Jurassic
(Pleinsbachian)
kink
axes
culvert
road
45°/012°
0 m 10
Dip/Dip direction

Figure 3.6 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.



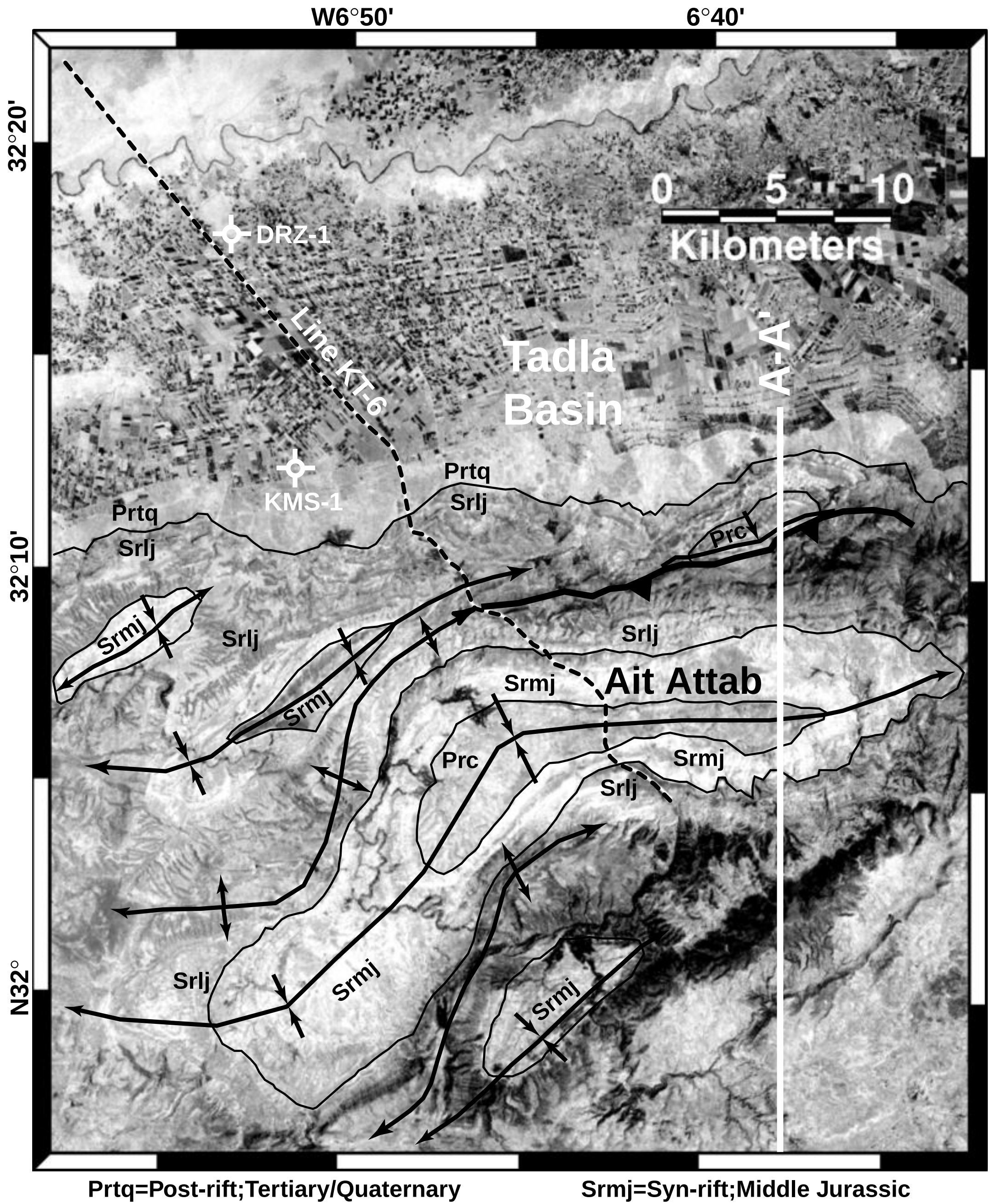
High Atlas
Tadla Basin
Ouarzazate Basin
Anti Atlas
Elevation (meters)
4000
3000
2000
1000
0
High Atlas
Tadla Basin
Ouarzazate Basin
Anti Atlas
Bouguer Gravity (mgals)
-50
-100
-150
-200
-250
55
50
45
Northings (km)
Eastings (km)
50
48
46
44
42
40
38
N

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 illustrates by a spectacular field relation located on the northern limb of the Ait Attab syncline (Figures 3.7 and 3.8). A low angle thrust fault dips 15° towards 210° (SSW). Massive Early Jurassic carbonates and Triassic clastic rocks in the hanging wall flat of this fault are thrust over Late Cretaceous red shales and sandstones Cenomanian in age (Figure 3.8). The zone is composed of breccia several meters in thickness. This breccia is composed of angular fragments of the overlying massive dolomites and limestones, and range up to 0.5 meters in size. The breccia is poorly cemented with the exception of 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 previously mentioned thrust

The Ait Attab syncline (Figure 3.7) 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 3.9). Thickening of the Middle and Lower Jurassic rocks across the syncline from southeast to northwest indicates that the syn-rift rocks were deposited in the hanging wall of an active normal fault that dipped southwards into the paleo-Atlas rift basin. The folded rocks lie above

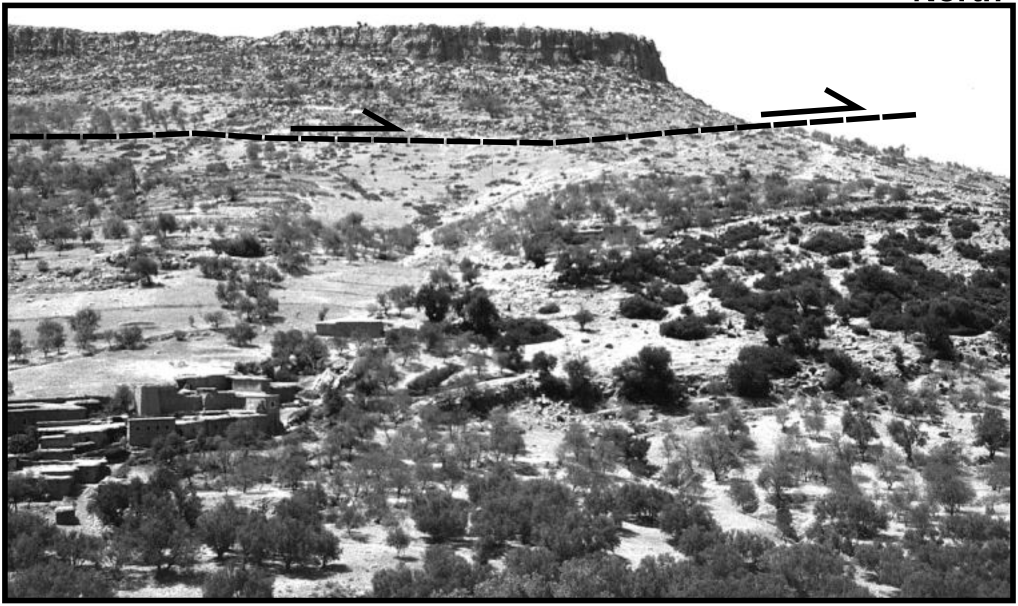
Figure 3.7 Simplified geological map of the Ait Attab syncline along the northern margin of the High Atlas Mountains. Shown are the locations of seismic line KT-6, wells DRZ-1 and KMS-1 and transect A-A'. Fold axes, fault traces and stratigraphic contacts are overlain on a LANDSAT-TM image (near infrared band 5).

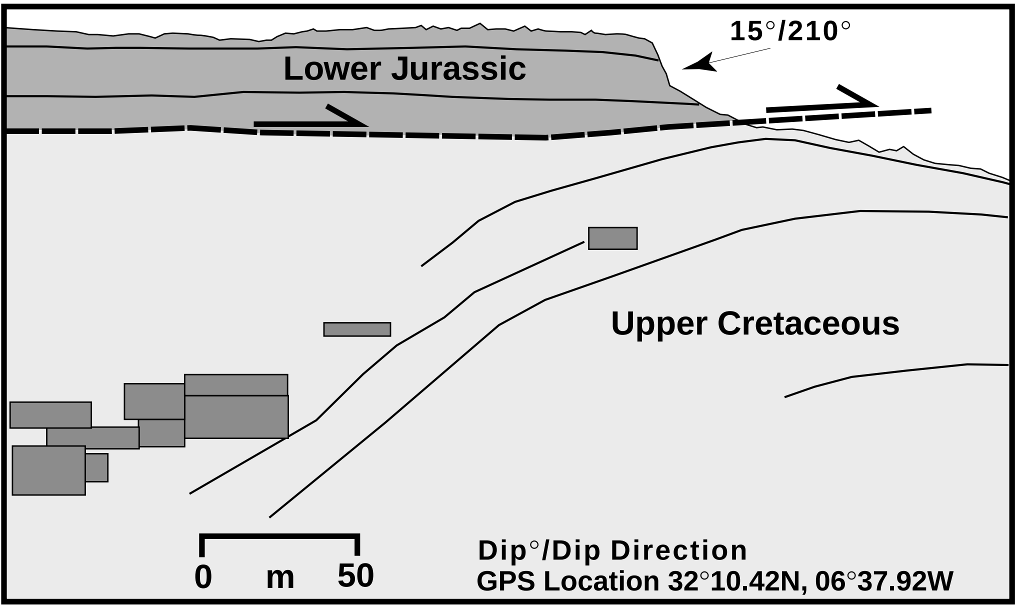


Prtq=Post-rift;Tertiary/Quaternary
Prc=Post-rift;Upper Cretaceous

Srmj=Syn-rift;Middle Jurassic
Srlj=Syn-rift;Lower Jurassic

Figure 3.8 A photograph showing a field location along the northern margin of the Ait Attab syncline (fault is shown on Figure 3.7). 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.


South
North

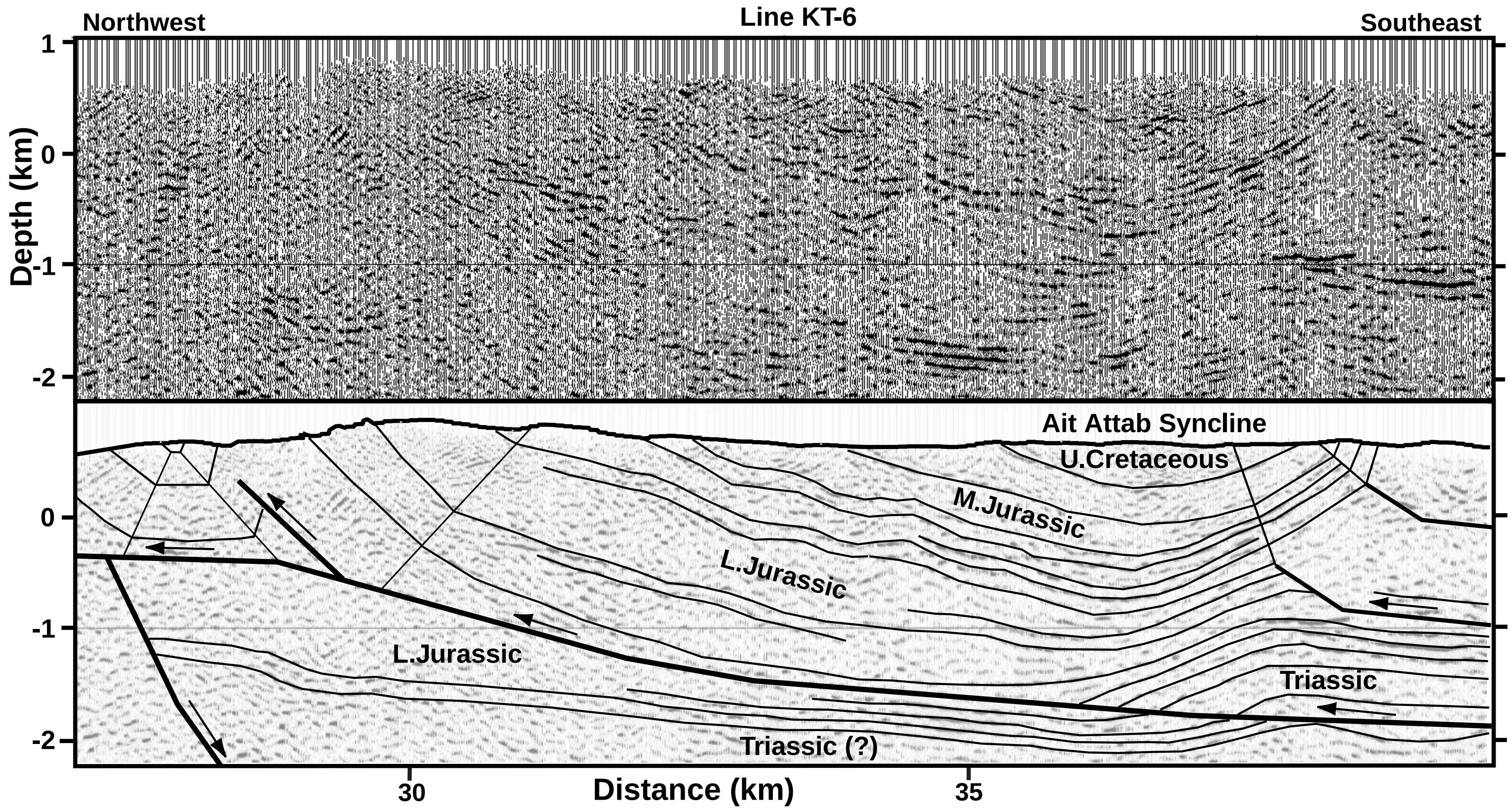

15°/210°
Lower Jurassic
Upper Cretaceous
0
m
50
Dip°/Dip Direction
GPS Location 32°10.42N, 06°37.92W

high amplitude reflectors that dip more gently than the overlying exposed rocks in the Ait Attab syncline (Figure 3.9). The lower reflections are interpreted as being the footwall flat of the major fault, described above, along the northern margin of the Ait Attab syncline (Figure 3.7). This broad, open syncline has a wavelength of ~12 km and is interpreted to have formed between two fault-bend anticlines (Figure 3.9, Plate 3). Locally, the geometry of the limbs has been complicated by small fault-propagation folds (Figure 3.5 and Plate 4). In three dimensions, the syncline displays a classic type-2 fold interference pattern (*Ramsey and Huber,* 1987). The older of the two fold axial surfaces strikes NE to E and the younger fold axial surface strikes NW and is near vertical. It is believed these two phases of folding were generated by one phase of deformation, with the thrusting of the rift sedimentary rocks northwestward over a lateral ramp.

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 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. 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 ramp from an interpreted décollement in the Triassic (Plate 3). The central region of the High Atlas known as Ait Mohammed is bounded by two of these large scale structures. The gently dipping syncline of Ait Mohammed is approximately 20 km wide. Beneath Ait Mohammed, a fault duplex system

Figure 3.9 A detailed portion of seismic line KT-6 across the Ait Attab syncline (see Figures 3.2 and 3.7 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 3.5). Bedding above the thrust fault is not parallel to the fault along the southeast end of the line (previously deformed), that may be syn-rift rocks inverted along a reactivated normal fault.



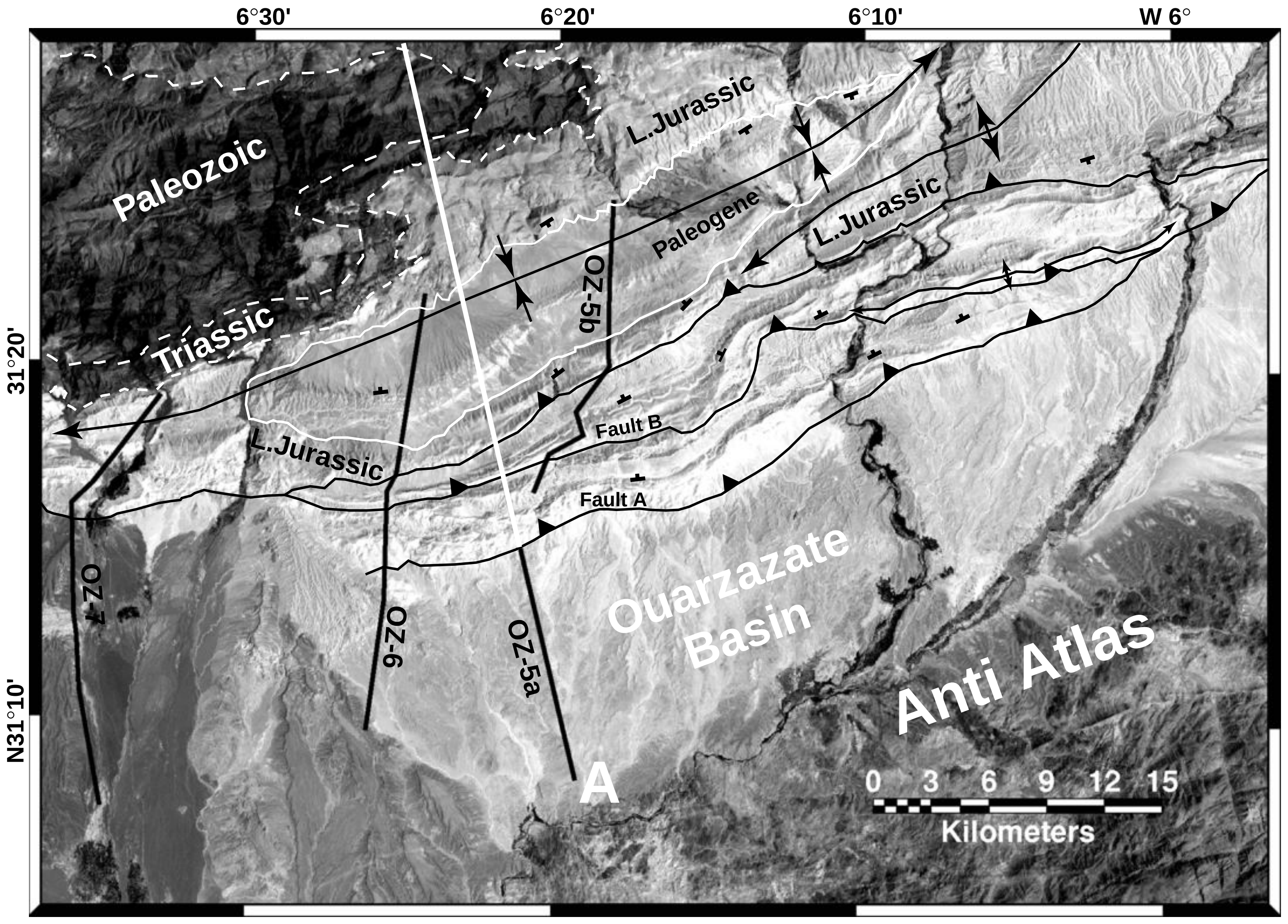
Northwest
Line KT-6
Southeast
Depth (km)
1
0
-1
-2
Ait Attab Syncline
U.Cretaceous
M.Jurassic
L.Jurassic
L.Jurassic
Triassic
Triassic (?)
0
-1
-2
30
35
Distance (km)

was interpreted in the Paleozoic to achieve the amount of uplift present above the décollement (~12 kilometers 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 3 probably does not represent the true geometry of the deformed beds in the Paleozoic, 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.

Southern High Atlas Mountains

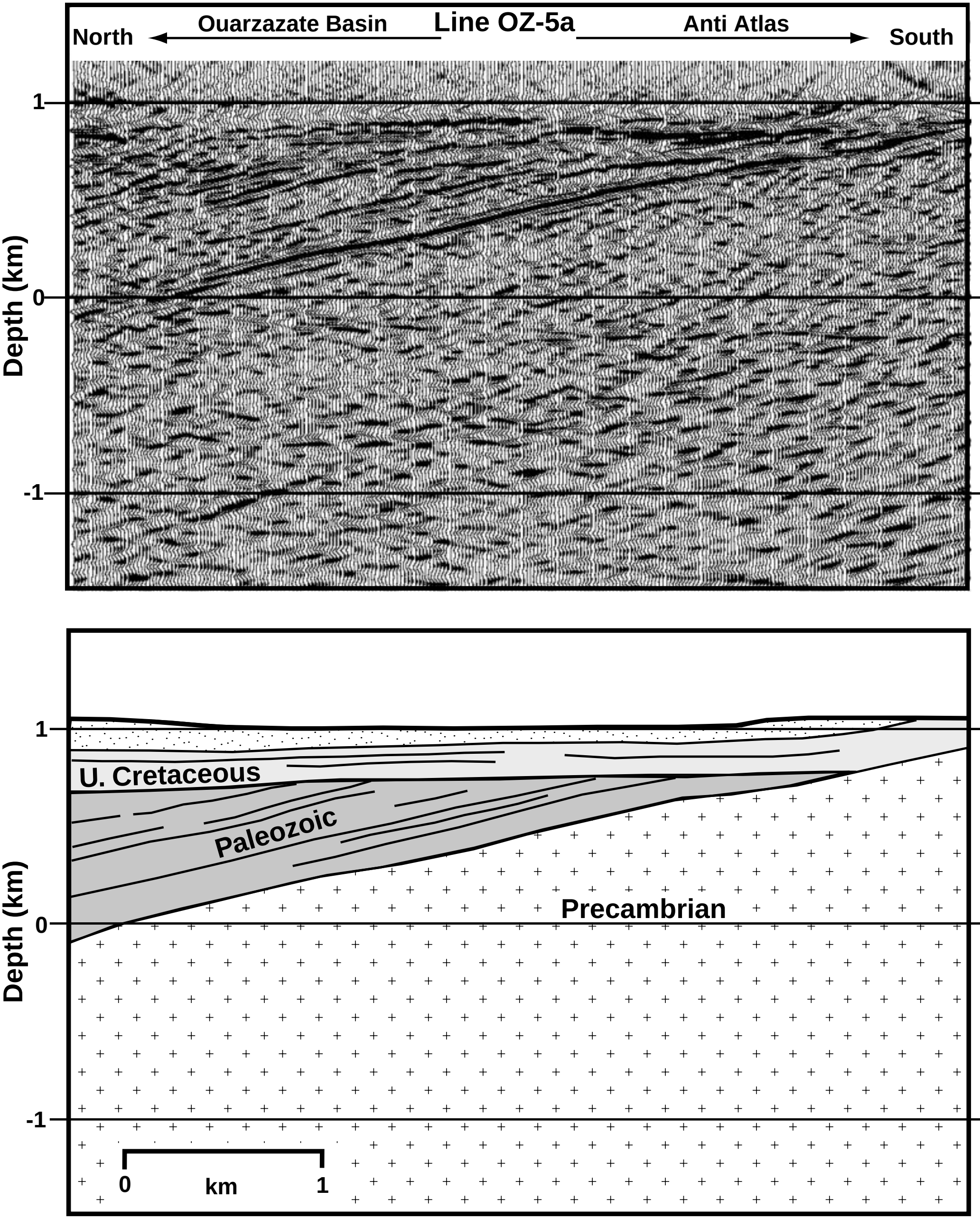
Three seismic lines from the Ouarzazate basin were supplied by ONAREP for this study. Line OZ-5 also extends completely across the Ouarzazate basin from the exposed Precambrian of the Anti-Atlas into the foothills of the southern High Atlas Mountains (Figure 3.10). The southern end of line OZ-5 shows a high amplitude reflector that dips 15° northward (Figure 3.11). 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 syn-rift 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

Figure 3.10 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 are 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 meters) that been uplifted by the reactivation of syn-rift faults. The Paleozoic rocks in the core of this anticline have been intensely deformed by what is thought to be a series of fault duplexes (Plate 3).



6°30'
6°20'
6°10'
W 6°
31°20'
N31°10'
Paleozoic
Triassic
L.Jurassic
Paleogene
L.Jurassic
L.Jurassic
OZ-5b
Fault B
Fault A
OZ-7
OZ-6
OZ-5a
Ouarzazate Basin
Anti Atlas
A
0 3 6 9 12 15
Kilometers

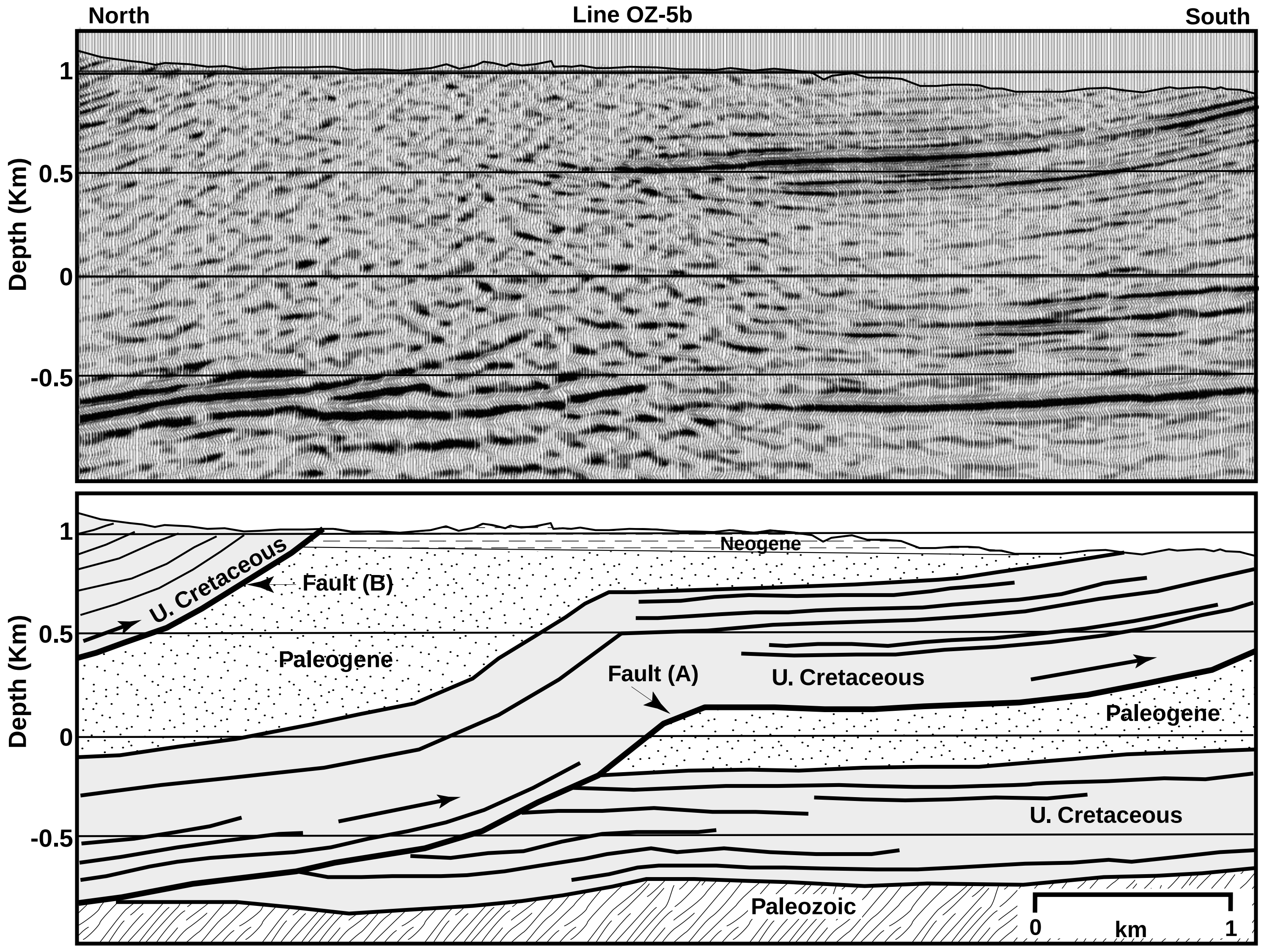
Figure 3.11 Seismic line OZ-5a located along the southern margin of the Ouarzazate basin and the Anti-Atlas (see Figure 3.10 for location). The high amplitude reflection that dips 15° from south to north across the section is interpreted as Precambrian basement. The reflectors above the basement are interpreted as Paleozoic rocks that are truncated by the Upper Cretaceous unconformity.


Line OZ-5a
North
Ouarzazate Basin
Anti Atlas
South
Depth (km)
1
0
-1
U. Cretaceous
Paleozoic
Precambrian
Depth (km)
1
0
-1
0
km
1

identified on LANDSAT-TM images (Figure 3.10). Fault-bend folds deformed the post-rift rocks of the Ouarzazate basin during uplift and inversion of the Atlas rift basin. Based on interpretation of Line OZ-5b (Figure 3.12), we suggest that the base of the Upper Cretaceous is the principal decollement 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 post-rift rocks to the surface in the Ouarzazate Basin. Because Neogene rocks unconformably overlie the hanging wall of fault (A) (Figure 3.12) but are cut by fault (B), we infer an out-of-sequence progression of faulting. Further evidence for out-of-sequence thrusting is present by the fact that faults closer to the High Atlas Mountains are not folded by faults to the south in the Ouarzazate basin (Plate 4).

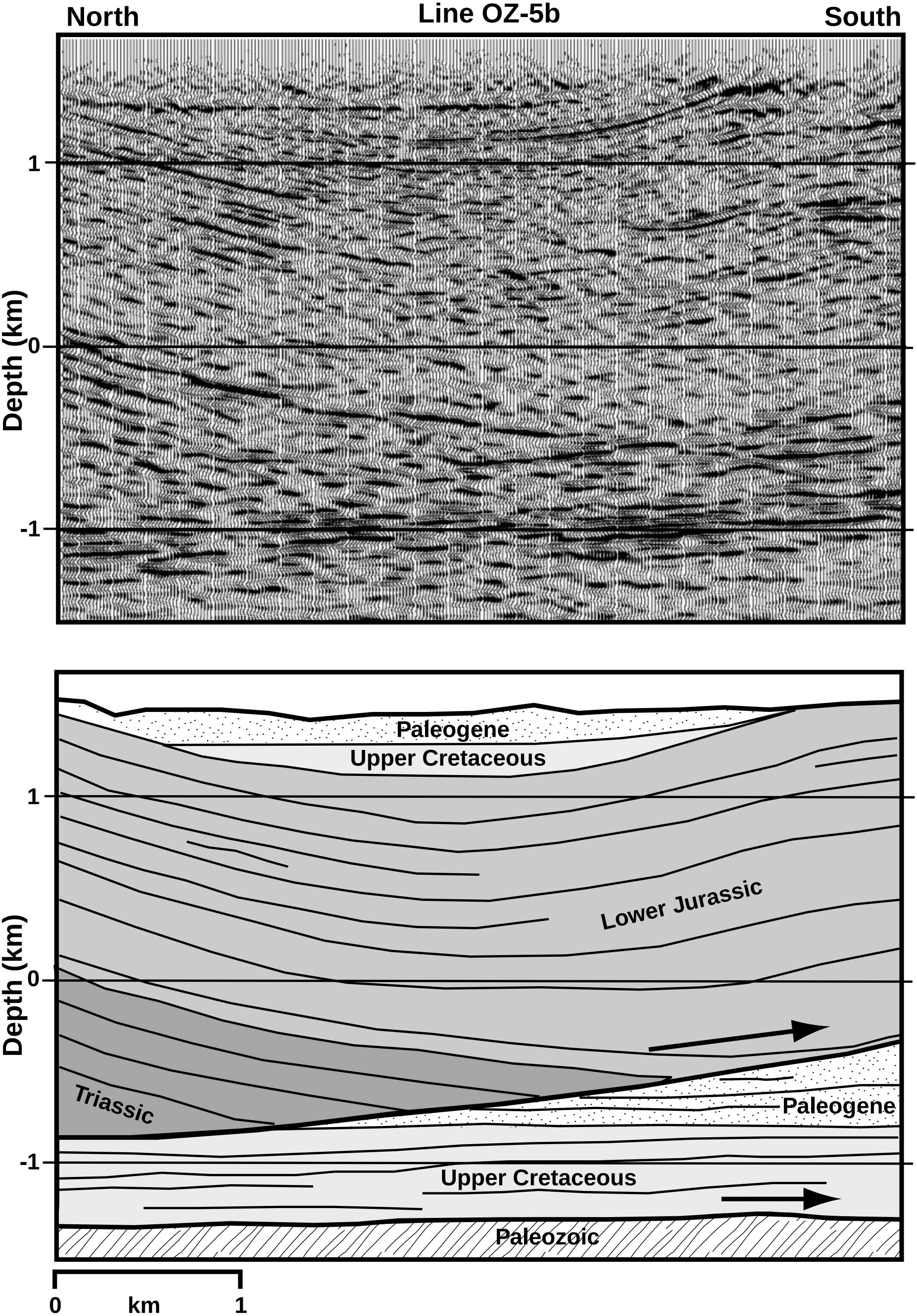
Northward along line OZ-5b the décollement at the base of the Upper Cretaceous is cut by a younger thrust that places syn-rift sedimentary rocks over post-rift Upper Cretaceous and Paleogene rocks (Figure 3.13 and Plate 4). The rocks in the footwall of this thrust were originally deposited along the platform margin of the Atlas rift (no syn-rift age rocks). The syn-rift Triassic and Lower Jurassic rocks were thrust in the hanging wall of a thrust fault from within the Atlas rift basin, and then 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

Figure 3.12 Seismic line OZ-5b located along the southern margin of the High Atlas Mountains (see Figure 3.10 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).



North
Line OZ-5b
South
Depth (Km)
1
0.5
0
-0.5
Neogene
U. Cretaceous
Fault (B)
Paleogene
Fault (A)
U. Cretaceous
Paleogene
U. Cretaceous
Paleozoic
0
km
1

Figure 3.13 Seismic line OZ-5b located to the north of Figure 3.12 shows what is interpreted as syn-rift sedimentary rocks transported in the hanging wall of a new thrust (see Figure 3.10 for location). These syn-rift strata are thrust out-of-sequence over an earlier thrust formed between the Upper Cretaceous and Paleozoic.


North
Line OZ-5b
South
1
0
-1
Depth (km)
Paleogene
Upper Cretaceous
Lower Jurassic
Triassic
Paleogene
Upper Cretaceous
Paleozoic
0
km
1

hanging wall of the thrust, possibly related to syn-rift deposition during extensional faulting (e.g., growth fault). The rocks in the hanging wall of this reactivated fault were previously deformed during the syn-rift extensional phase (hanging wall rollover) of deformation, and were subsequently transported up the same normal fault until a new thrust fault was formed transporting syn-rift rocks over the platform margin of the rift (Plates 3 and 4).

Construction of a Geological-Geophysical Transect

This study set forth to generate a transect across the High Atlas Mountains along a transect(A-A') that would optimize the data available, and would for the first time generate a cross section across the High Atlas Mountains that is balanced and restorable. Such a transect would 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 syn-rift normal faults, while new thrusts are 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, and that line lengths remain the same before and after shortening (*Woodward et al.*, 1989). Refolded folds are common throughout the High Atlas and violate the 2-D assumptions inherent in cross section balancing. However, because our transect is nearly parallel to the second fold axial surface, the effect of the refolding is minimal. A depth to detachment calculation was made (approximately 12 km depth) based upon a composite thickness from well data, and measured thicknesses in the High Atlas, Anti-Atlas. This depth to detachment calculation is in agreement with a proposed detachment based on previous geophysical studies (10-15 km) (*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 3.11). The post-rift Late Cretaceous age rocks were assumed to not thicken 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 3). The construction of the transect also assumesthere was no pre-existing 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).

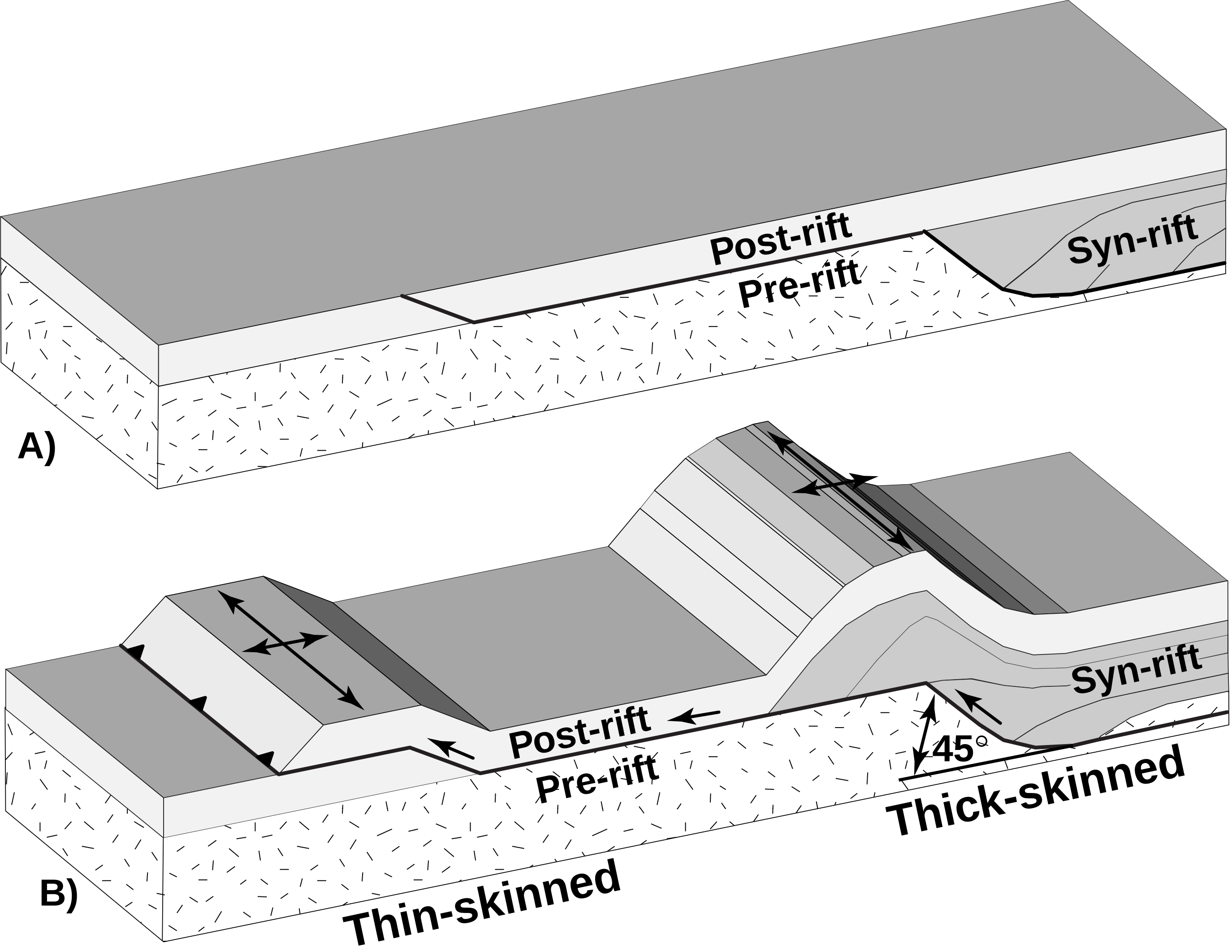
Estimates of shortening attained from this study based upon transect A-A' (Plate 3) are a minimum estimate of the amount of shortening across the High Atlas. A local pin line was placed outside of the orogen in the Tadla basin to the north, and a regional pin line in the Ouarzazate basin south of the High Atlas Mountains. Based upon a depth of décollement calculation of 12 km a template was created for the footwall of the original Atlas rift system (Plate 3). The depth of basement in the footwall of the template (Plate 1, deformed and restored sections) was derived from seismic reflection data (Figure 3.11). Whereas the original rift basin was comprised of many normal faults with various displacements, vergence and magnitudes, only the main rift bounding faults are considered for reactivation along transect A-A'. Syn-rift sediments deformed in the hanging wall of the rift basin were transported up the southern rift bounding fault (Figure 3.14 and Plate 3). Thrusts with a fault-bend fold geometry are north verging (Figures 3.8 and 3.9) along the northern margin of the High Atlas, ramping upwards from evaporites in the Triassic (Plate 3).

Discussion

Historical Evolution

The Atlas mountain belt is the largest 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 regarding 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 several hundred kilometers through the crust to an intraplate region where strain was accommodated along crustal weaknesses formed by rifting in the Mesozoic and probably earlier Paleozoic orogenic events The evolution of the Atlas

Figure 3.14 The tectonic evolution of the High Atlas Mountains involved newly formed fault-bend folds and fault-propagation folds combined with reactivated syn-rift faults and new thrust faults (A-B). 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 upwards until a new thrust or footwall short cut fault is formed. Syn-rift rocks transported in the hanging wall of these thrusts will not be parallel to the fault plane due to previous deformation. The newly formed thrusts and footwall short cut faults are mechanically more efficient.



Post-rift
Pre-rift
Syn-rift
A)
Syn-rift
Post-rift
Pre-rift
45°
Thick-skinned
B)
Thin-skinned

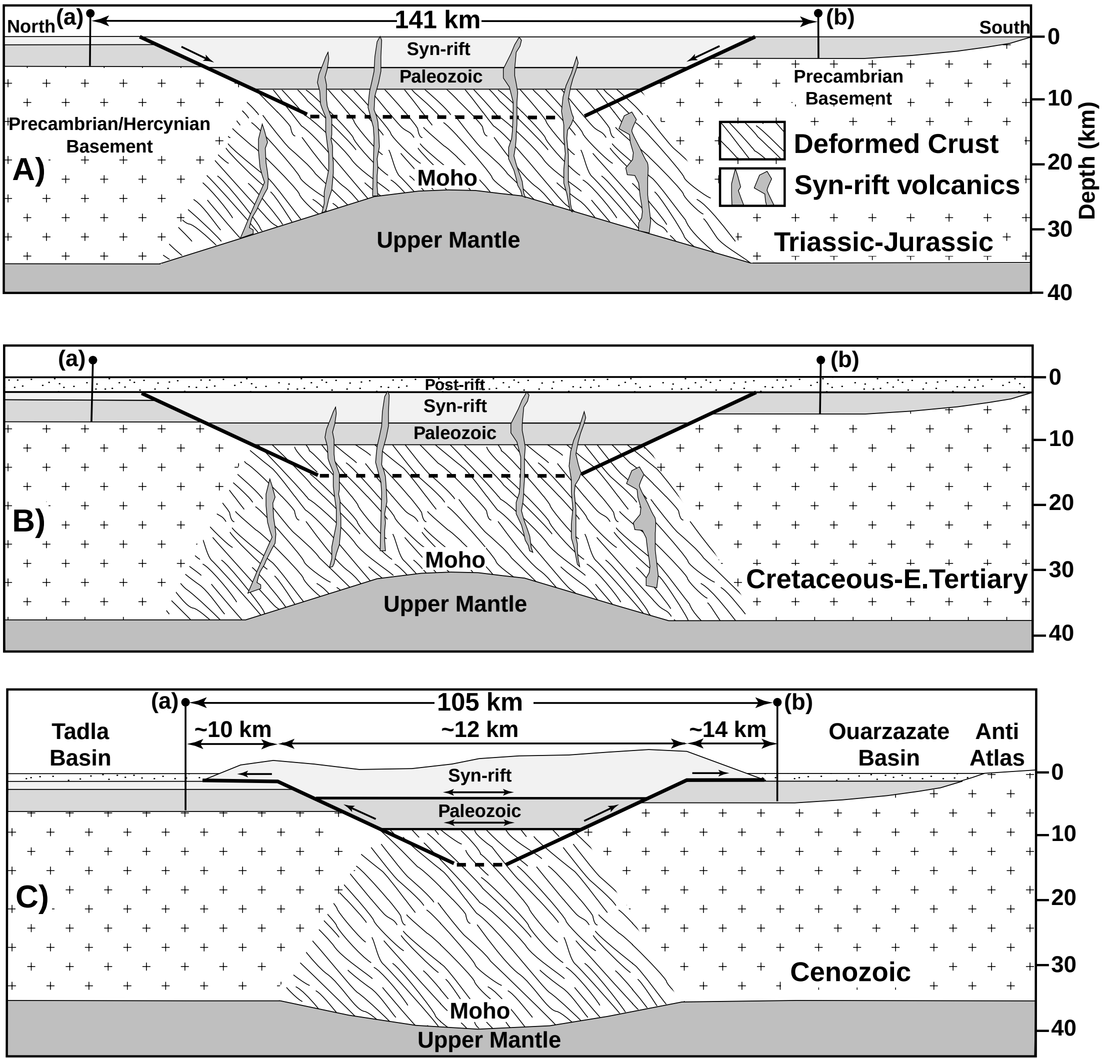
Mountains and the resulting structural styles were predetermined by conditions ensuing from earlier tectonic events. Triassic age basaltic intrusives found in the Central High Atlas Mountains yield ages of 210-182 Ma, and are of a MORB-like composition indicating a subcontinental to suboceanic composition, with crustal contamination (*Fiechtner et al.*, 1992). The presence of volcanic rocks intercalated with the rift strata of the Atlas Mountains indicates crustal thinning was accomplished by a pure shear or a similar rifting mechanism (Figure 3.15a).

The post-rift phase of the Atlas rift basin coincided with the successful opening of the Atlantic Ocean (Figure 3.15b). During the Late Jurassic and Early Cretaceous the northwestern margin of the African plate was uplifted due to thermal expansion and rifting in the Atlantic. Subsidence along the Atlantic margin was amplified during the Late Jurassic Early Cretaceous with continued rifting and spreading. Though sedimentary rocks of Late Jurassic and Early Cretaceous are preserved offshore along the Moroccan passive margin, few sedimentary rocks of these ages (161.5-97.5 Ma) are preserved in the Atlas Mountains and margins onshore Morocco (Figure 3.3).

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 (Figure 3.14). 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,

Figure 3.15 Schematic cross sections showing the tectonic history of the Atlas syn-rift phase (A), the post-rift phase (B), and the final uplift and inversion of the Atlas rift system (C) to form the present day Atlas Mountains. Two regional pin lines are shown which are equivalent to those in Plate 3 (a) and (b). The distance between these pin lines after restoration is approximately 141 km (Figure 3.15A), which yields the original width of the Atlas rift system. Subsidence during the Late Jurassic to Tertiary is also shown (Figure 3.15B). Convergence between the African and Iberian plates in the Tertiary (Miocene-Oligocene) resulted in the inversion of the Atlas rift basin by bi-vergent 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 mid-lower crustal levels (Figure 3.15C).


North
(a)
141 km
(b)
South
Syn-rift
Paleozoic
Precambrian/Hercynian Basement
Precambrian Basement
Deformed Crust
Syn-rift volcanics
A)
Moho
Upper Mantle
Triassic-Jurassic
Depth (km)
0
10
20
30
40
(a)
(b)
Post-rift
Syn-rift
Paleozoic
B)
Moho
Upper Mantle
Cretaceous-E.Tertiary
0
10
20
30
40
(a)
105 km
(b)
~10 km
~12 km
~14 km
Tadla Basin
Ouarzazate Basin
Anti Atlas
Syn-rift
Paleozoic
C)
Cenozoic
Moho
Upper Mantle
0
10
20
30
40

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 (141 km to 105 km) based on a restoration of the deformed section(Plate 3 and Figure 3.15). The greatest shortening (10-14 km) is confined to a narrow region(~10-15 km) along the margins of the High Atlas mountains (Figure 3.15c). The central High Atlas Mountains display far less shortening in the shallow crust (syn/post rift 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 mid-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 mid-lower crust. The transfer of shortening and the accommodation of strain to the mid-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, 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 additional 25-40 % of deformation in extensional terrains, and that these same estimates may apply to shortening in compressional terraines (*Marrett and Allmendinger*, 1992). Hence, 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.

It has been shown in this study that, while thin-skinned deformation along newly formed faults occurred in the post-rift sequence of the Moroccan Atlas, further deformation by the reactivation of syn-rift normal faults (thick-skinned) has resulted in a "hybrid" thick and thin-skinned style of deformation (Figure 3.14). Faulting along the southern margin of the High Atlas Mountains and northern Ouarzazate basin (Figures 3.10 and 3.12) involves thin-skinned faults similar to those reported in the Algerian Saharan Atlas Mountains (*Outtani et al.*, 1995). Thick-skinned deformation inverts syn-rift strata along pre-existing faults, often with continued transport of strata along newly formed thin-skinned faults towards the exterior of the basin (Figures 3.14a and 3.15c). The lower angle thrusts allow for larger displacements and greater shortening than the reactivated syn-rift faults. The overall process of inversion has the effect of concentrating shortening where the syn-rift strata are extruded over the margins of the rift and onto the adjacent platform. If reactivation of a listric syn-rift 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 short cut 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 syn-rift normal fault (*Beauchamp et al.*, 1996).

Conclusions

The objective of this study was to develop a better understanding of how intracontinental mountain belts evolve. 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 ONAREP (Office National de Recherches et d'Exploitations Petrolieres) 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 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 non-deposition. The major uplift phase in the Atlas Mountains occurred between the Miocene-Oligocene (20-30 Ma). Syn-rift and post-rift sedimentary rocks were uplifted by the reactivation of syn-rift normal faults, with further shortening along newly formed thin-skinned thrust faults. These inverted syn-rift rocks have different geometries and exhibit significant changes in thickness over a short distance, that is different 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 margins of 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 in the interior of the mountain belt. Thrusting in the High Atlas Mountains is bi-vergent, with thrusts dipping to the south along the northern margin (Tadla basin), and northward dipping faults to the south (Ouarzazate basin). This bi-vergence may have been influenced by the original rift configuration. The minimum shortening in the central part of the topographically High Atlas Mountains suggests significant mid-lower crust shortening. The results of this research clearly demonstrates the remarkable difference between the evolution and present structural architecture of this intracontinental mountain belt, in comparison to those associated with plate boundary processes.

References

Bally, A.W., Tectogenese et sismique reflexion*, Bulletin Societe Geologique France*, 7, 279-285, 1984 .

Beauchamp, J., Triassic sedimentation and rifting in the High Atlas (Morocco), in W. Manspeizer, ed., Triassic-Jurassic Rifting: *Elsevier*, 477-497, 1988.

Beauchamp, W., M. Barazangi, A. Demnati and M. El Alji, Intracontinental Rifting and Inversion: the Missour Basin and Atlas Mountains of Morocco, *American Association of Petroleum Geologists Bulletin*, 80/9, 1459-1482,1996.

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*, Geologische Rundschau*,81/1, 127-141, 1992.

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

Broughton, P., and A. Trepanier, Hydrocarbon Generation in the Essaouira Basin of Western Morocco, *American Association of Petroleum Geologists Bulletin*, 77/6, 991-1015, 1993.

Coward, M.P., R. Gillcrist and B. Trudgill, Extensional Structures and their tectonic inversion in the Western Alps, in A. M. Roberts, G. Yielding and B. Freeman, eds., The geometry of normal faults: *Geological Society Publication*, 56, 93-112,1991.

Crevallo, P.D., J. E. Warme, M Septfontaine, R. Burke, Evolution of Jurassic carbonate platforms in an active transtensional rift, *American Association of Petroleum Geologists Bulletin*, 71/5, 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 M.P. Coward, D. Dietrich, and R.G. Park, eds., Alpine Tectonics: *Geological Society Publication*, 45, 265-283, 1989.

Ensslin, R., Cretaceous synsedimentary tectonics in the Atlas system of Central Morocco, *Geologische Rundschau*, 81/1, 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 G. Choubert ed., *Tectonique de l'Afrique, Unesco*, Paris, 1971.

Fetah, M., M. Bensaid, M. Dahmani, Carte Géologique Du Maroc-Zawyat Ahancal, 1:100,000, Royaume Du Maroc, *Ministere De L'Energie Et Des Mines*, 355, 1990.

Fiechtner, L., H. Friedrichsen and K. Hammerschmidt, Geochemistry and geochronology of Early Mesozoic thoeleiites from Central Morocco, *Geologische Rundschau*, 81/1, 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 V. Jacobshagen ed., The Atlas System of Morocco-Studies on its geodynamic Evolution-Lecture notes Earth Science: Berlin, *Springer-Verlag*, 15, 273-291,1988.

Gauthier, H., Contribution l'étude géologique des formations post-liasiques des bassins du Dadés et du Haut Todra, *Notes Mémoires du Service Géologique*, Rabat, 119, 212 pp., 1960.

Ghissassi, A., M. Saadi, T. Skalli, E. A. Hilali, A. Boudda, Carte Géologique Du Maroc-Afourar, 1:100,000, *Royaume Du Maroc, Ministere De L'Energie Et Des Mines*, 247, 1977.

Gibbs, A. D., Balanced cross-section construction from seismic sections in areas of extensional tectonics, *Journal of Structural Geology*, 5, 153-160, 1983.

Giese, P., and V. Jacobshagen, Inversion Tectonics of Intracontinental Ranges: High and Middle Atlas, Morocco: *Geologische Rundschau*, 81/1, 249-259, 1992.

Hayward, A. B. and R. H. Graham, Some geometrical characteristics of inversion, in M. A. Cooper and G. D. Williams eds., Inversion Tectonics, *Geological Society Special Publications*, 44, 17-39, 1989.

Jabour, H. and K. Nakayama, Basin modeling of Tadla basin, Morocco, for hydrocarbon potential, *American Association of Petroleum Geologists Bulletin*, 72, 1059-1073, 1988.

Jacobshagen, V., R. Brede, M. Hauptmann, W. Heinitz, and R. Zylka, Structure and post-Paleozoic evolution of the central High Atlas, in V.

Jacobshagen, ed., The Atlas System of Morocco-Studies on its geodynamic Evolution-Lecture notes in Earth Science, Berlin-Heidelberg-New York , *Springer-Verlag*, 15, 245-271, 1988.

Jenny, J., Carte Géologique du Maroc au 1/100,000 Feuille Azillal (High Atlas central), *Notes Mémoires du Service Géologique*, Rabat, 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, *Geologische Rundschau*, 81, 143-156, 1992.

Laville, E., and A. Pique, Jurassic penetrative deformation and Cenozoic uplift in the Central High Atlas (Morocco): a tectonic model. Structural and Orogenic inversions: *Geologische Rundschau*, 81/1, 157-170, 1992.

Laville, E., and A. Pique, La distension crustale atlantique et atlasique au Maroc au début du Mésozoique: le rejeu des structures hercyniennes, *Bulletin de la Societe Geologique de France*, 162/6, 1161-1171, 1991.

Laville, E., A multiple releasing and restraining stepover model for the Jurassic strike-slip basin of the Central High Atlas (Morocco). In W. Manspeizer, ed., Triassic-Jurassic Rifting, Continental Breakup and the Origin of the Atlantic Ocean, Amsterdam: *Elsevier*, 1988, 499-523.

Lowell, J.D., Mechanics of basin inversion from worldwide examples, in J. G. and P. G. Buchanan,eds., Basin Inversion, *Geological Society Special Publication*, 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, *Annales Geophysicae*, 3, 369-380, 1985.

Manspeizer, W., J. Puffer and H. Cousminer, Separation of Morocco and eastern North America: a Triassic-Liassic stratigraphic record, *Geological Society of America Bulletin*, 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 K. R. McClay, ed., Thrust Tectonics, *Chapman and Hall*, London, 93-104, 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 Mémoires du Service Géologique*, Rabat, 252,pp. 408, 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, 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, Ch. Hoepffner, Southern margin of the Variscan belt: the north-western Gondwana mobile zone (eastern Morocco and Northern Algeria), *Geologische Rundschau*, 82, 432-439, 1993.

Ramsay, John G., and M. I. Huber, The techniques of modern structural geology; Volume 2, Folds and fractures, p.494, *Academic Press*, London, 1987.

Rolley, J.-P., Carte Geologique du Maroc au 1/100,000 Feuille Afourer (High Atlas central), *Notes Mémoires du Service Géologique*, Rabat, 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*: Geological Society of London Special Publication*, 25, 25-43, 1986.

Saadi, M., M. A. Hilali, A. Demnati, Carte Magnetique du Maroc, Interpretation du Leve Aeromagnetique, Bechar and Marrakech, 1:1,000,000, *Royaume du Maroc, Ministere de l'Energie et des Mines*, 320-C, 1979.

Saadi, M., E. A. Hilali, M. Bensaid, Carte Géologique du Maroc-Jebel Saghro-Dadés, 1:200,000*, Royaume du Maroc, Ministere de l'Energie et des Mines*, 161, 1977.

Saadi, M., M. Bensaid, M. Dahmani, Carte Géologique du Maroc-Azilal, 1:100,000*, Royaume du Maroc, Ministere de l'Energie et des Mines*, 339, 1985.

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

Schmidt, K.-H., The tectonic history of the Pre-Saharan depression (Morocco)-a geomorphological interpretation, *Geologische Rundschau*, 81/1, 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, *Geologische Rundschau*, 81/1, 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.

Schaer, Jean-Paul, and J. Rodgers, Evolution and Structure of the High Atlas of Morocco. In: Jean-Paul Schaer, and J. Rodgers, eds., The Anatomy of Mountain Ranges: *Princeton University Press*, 107-127, 1987.

Suppe, J., Geometry and kinematics of fault-bend folding*, American Journal of Science*, 283, 684-721, 1983.

Suppe, J., and D. A. Medwedeff, Geometry and kinematics of fault-propagation folding, *Eclogae Geologicae Helvetiae*, 83/3, 409-454, 1990.

Tadili, B., and M. Ramdani, Computer file of Moroccan earthquakes, *Bulletin of the Seismological Society of America*, 73, 653-654, 1983

Wang, G. M., M. P. Coward, W. Yuan, S. Liu and W. Wang, Fold growth during basin inversion-example from the East China Sea Basin, in J. G. and P. G. Buchanan, eds., Basin Inversion, *Geological Society Special Publication*, 88, 492-522, 1995.

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, *Geologische Runschau*, 81/1, 237-248, 1992

Woodward, N. B., S. Boyer and J. Suppe, Balanced Geological Cross-Sections: An Essential Technique in Geological Research and Exploration, in M. L. Crawford and E. Padovani, eds., Short Course in

Geology, *American Geophysical Union*, Washington, D.C., 6, 132 pp., 1989.

Ziegler, P.A., Geological Atlas of Western and Central Europe, *Shell International Petroleum, The Hague*: 2 vol., 130 pp., 40 pl., 1982.

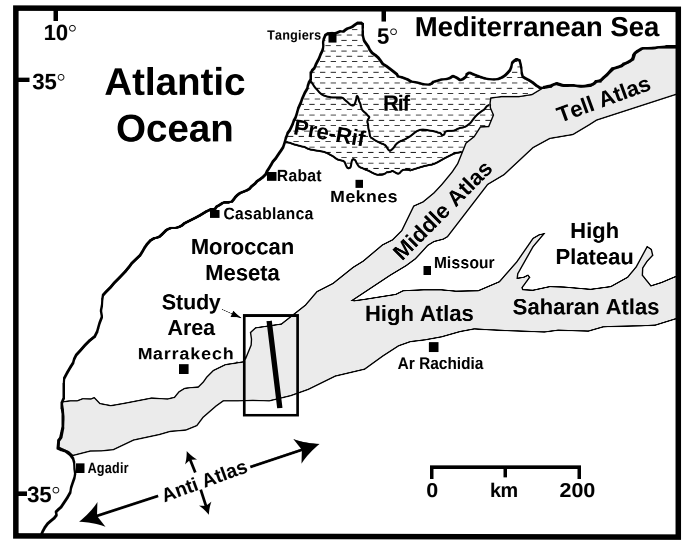
Plate 3

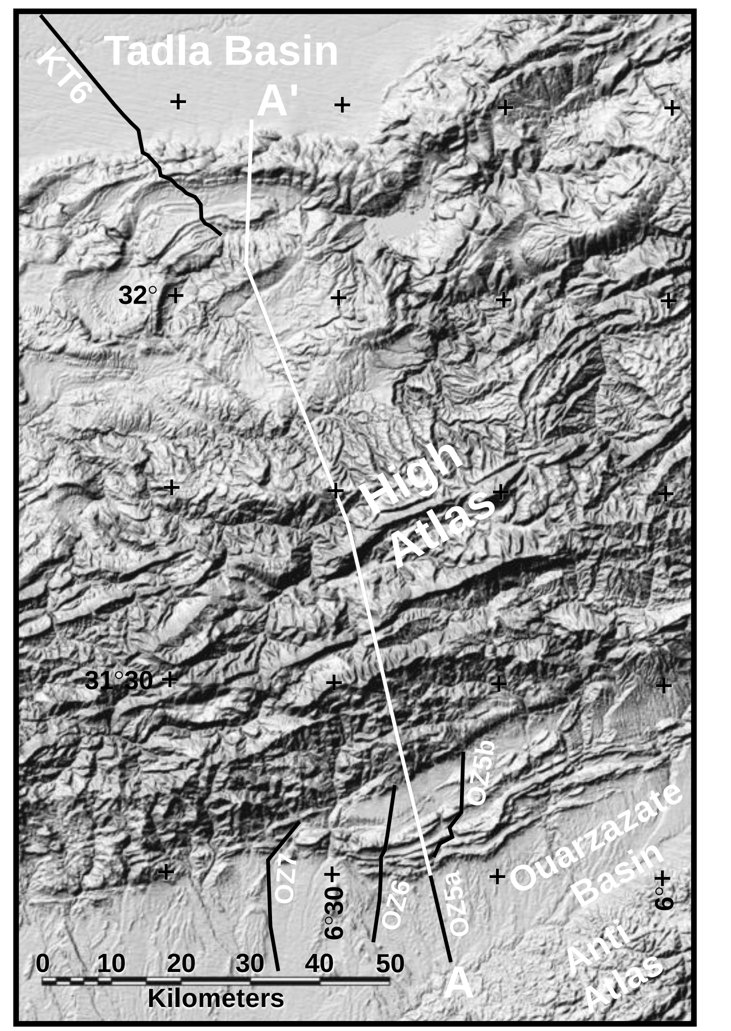
A Balanced Geological-Geophysical Transect of the High Atlas Mountains, Morocco

Weldon Beauchamp, Richard Allmendinger, Muawia Barazangi,
Institute for the Study of the Continents,
Cornell University, Ithaca, New York

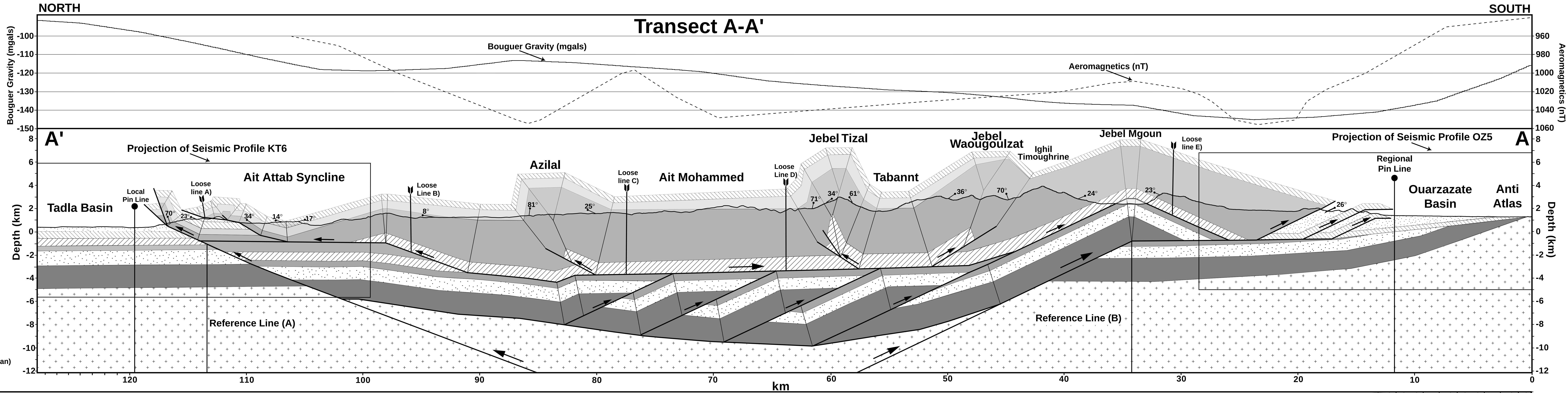
Ahmed Demnati, Mohamed El Alji,
Office National de Recherches et d'Exploitations Petrolieres
Rabat, Morocco

Mohammed Dahmani
Ministere de l'Energie et des Mines
Rabat, Morocco





*No vertical Exaggeration



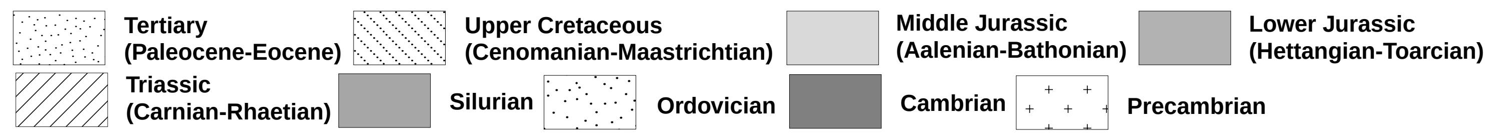


Plate 4

Northwest
Line KT-6
Southeast
DRZ-1
Tadla Basin
KMS-1
Ait Attab Syncline
Neogene
Upper Cretaceous
Triassic
Lower Jurassic
Triassic
Lower Jurassic
Paleozoic
Triassic
Precambrian
Paleozoic
Depth (km)
Distance (km)
Northwest
Line KT-6
Southeast
DRZ-1
Tadla Basin
KMS-1
Depth (km)
Distance (km)
North
Line OZ-5a & b
South
High Atlas
Paleogene
Upper Cretaceous
Lower Jurassic
Triassic
Ouarzazate Basin
Anti Atlas
Paleogene
Neogene
Upper Cretaceous
Precambrian
Paleozoic
Depth (km)
Distance (km)
North
Line OZ-5b
Line OZ-5a
South
High Atlas
Ouarzazate Basin
Anti Atlas
Depth (km)
Distance (km)

CHAPTER FOUR

Superposed folding resulting from inversion of a syn-rift accommodation zone, Atlas Mountains, Morocco

Abstract

A conspicuous offset of the northern margin of the High Atlas Mountains in Morocco is comprised of several large superposed folds, the largest of which is known as the Ait Attab syncline. The original northeast trending syncline (F1) was folded by a second set of fold axes (F2) that trend northwest. The superposed folding was generated by one phase of Cenozoic compression, with thrusting of syn-rift rocks northwestward over a prior accommodation zone formed during Mesozoic rifting. This accommodation zone is expressed in the exposure of syn-rift rocks, the exposure of Paleozoic strata in the footwall, and a coincident offset of topography. Inversion was accomplished by the transport of syn-rift strata along reactivated normal faults and newly formed thrusts. The unique spatial pattern of refolding is believed to be characteristic of inversion that is influenced by pre-existing rift structures.

Introduction

Existing models of inversion are predominantly 2D accordion style models that assume extension and subsequent compression are coaxial. Rift systems characterized by long straight faults bounding the rift system result from orthogonal extension (*McClay and White*, 1995). The inversion of rift systems where compression and extension are orthogonal produce structures that trend parallel to the preexisting rift structures.

However, many rift systems show patterns of *en echelon* normal faulting and segmented faults (*McClay and White*, 1995), that are the product of oblique extension during rifting. It can be expected that compression relative to structures produced by oblique extension would produce structural geometries that are a result of 3D strain. The study area in the High Atlas Mountains indicates extension was oblique during rifting, displaying accommodation zones, en echelon normal faults and pull-apart basins. During inversion in the Tertiary, compression resulted in a unusual pattern of superposed folding, influenced by preexisting extensional structural geometries.

Intracontinental mountain belts such as the Atlas Mountains often form by the inversion of pre-existing intracontinental rift systems (*Bally*, 1984, *Beauchamp et al.*, 1996). The High Atlas Mountains evolved from a major Mesozoic rift system (~2000 km) that was uplifted and inverted in the Cenozoic (*Beauchamp et al.,* 1996, 1997a, b). The convergence of the African and Iberian plates in the Tertiary resulted in the inversion of Mesozoic strata along pre-existing syn-rift faults, and by the transport of these rocks along newly formed low angle thrusts.

The Jebilet Accommodation Zone

The northern margin of the High Atlas Mountains in Morocco contains a conspicuous offset (~90°) of the topography and exposure of the syn-rift Mesozoic rocks (Figure 4.1). Syn-rift strata thicken dramatically to the south and east into the High Atlas Mountains. The present day Tadla, Haouz and Ouarzazate basins were the shelf/platform margin areas of the paleo-Atlas rift (Figure 4.1). Mesozoic strata are absent along the eastern margins of the

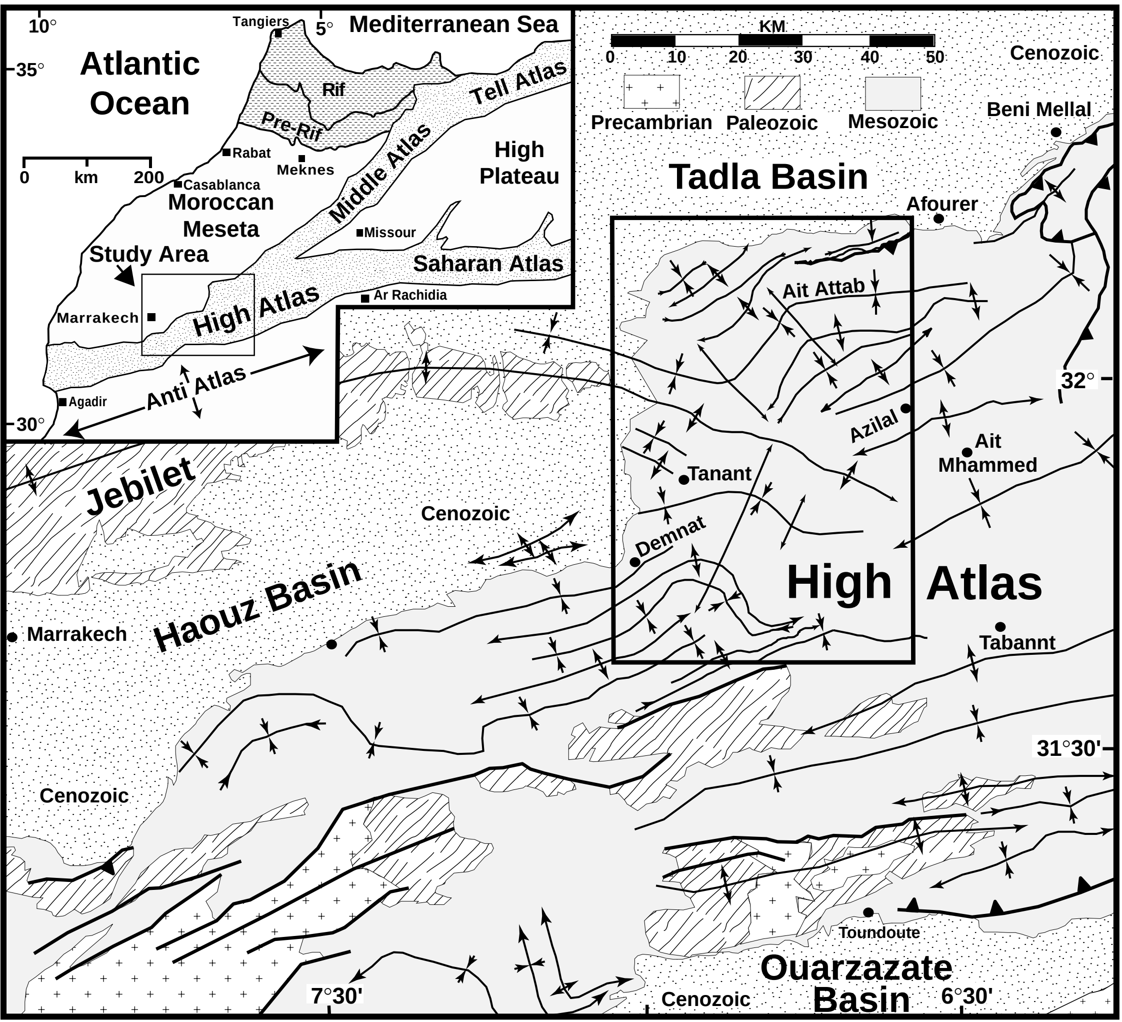
Jebilet, where Paleozoic age rocks crop out (Figure 4.1). Immediately to the south and east of the Haouz Basin and the exposed Paleozoic rocks of the Jebilet, are 2-3 kilometers of preserved syn-rift strata (*Beauchamp et al.*. , 1997). North of the Ait Attab syncline (Figure 4.1) well data show that syn-rift Jurassic strata are absent and Triassic strata are condensed to less than 500 meters in thickness (*Jabour and Nakayama,* 1988). The Jebilet and offset in topography are interpreted to be related to a syn-rift accommodation zone. The accommodation zone affected the deposition of syn-rift strata during rifting, and later, influenced the thrusting of syn-rift strata northward from the rift basin to the shelf margins. The geometry of this accommodation zone may have controlled the thrusting and subsequent superposed folding.

The Jebilet extends eastward across the accommodation zone where the anticlinorium plunges to the southeast (Figure 4.1). This anticline separates the Guettioua syncline and the Ait Attab syncline (Figure 4.2) and is not affected by a second phase of folding. The Paleozoic strata of the Jebilet is believed to be in the hanging wall of a thrust system which verges northward from the Atlas Mountains. This thrust may have formed as a footwall short-cut fault that transported Paleozoic rocks from the margin of the rift basin.

Cenozoic Folding Patterns

Prominent large scale refolded folds bound the proposed accommodation zone (Figure 4.1). These folds were formed during uplift and inversion in the Oligocene-Miocene related to convergence between the African and Iberian plates. Fault related folding is believed to have been

Figure 4.1 Location and simplified tectonic/structural map of the study area in the High Atlas Mountains, Morocco. Fold axes in the High Atlas are approximately parallel to the orogen and normal to the direction of thrusting, except where preexisting accommodation zones have influenced the regional stress field, resulting in polyphase deformation. Also shown (box) is the location of a more detailed geological map that displays superposed folding (Figure 4.2). Compiled from geological maps (*Saadi et al.*, 1977; *Rolley*, 1978; *Saadi et al.,* 1985; *Jenny*, 1988).



10°
Tangiers
5°
Mediterranean Sea
35°
Atlantic Ocean
Rif
Tell Atlas
Pre-Rif
Rabat
Meknes
0 km 200
Casablanca
Moroccan Meseta
Middle Atlas
High Plateau
Missour
Study Area
Saharan Atlas
Ar Rachidia
Marrakech
High Atlas
Anti Atlas
Agadir
30°
KM
0 10 20 30 40 50
Cenozoic
Precambrian
Paleozoic
Mesozoic
Beni Mellal
Tadla Basin
Afourer
Ait Attab
32°
Azilal
Ait Mhammed
Tanant
Jebilet
Cenozoic
Demnat
Haouz Basin
High Atlas
Marrakech
Tabannt
31°30'
Cenozoic
Toundoute
Ouarzazate Basin
7°30'
Cenozoic
6°30'

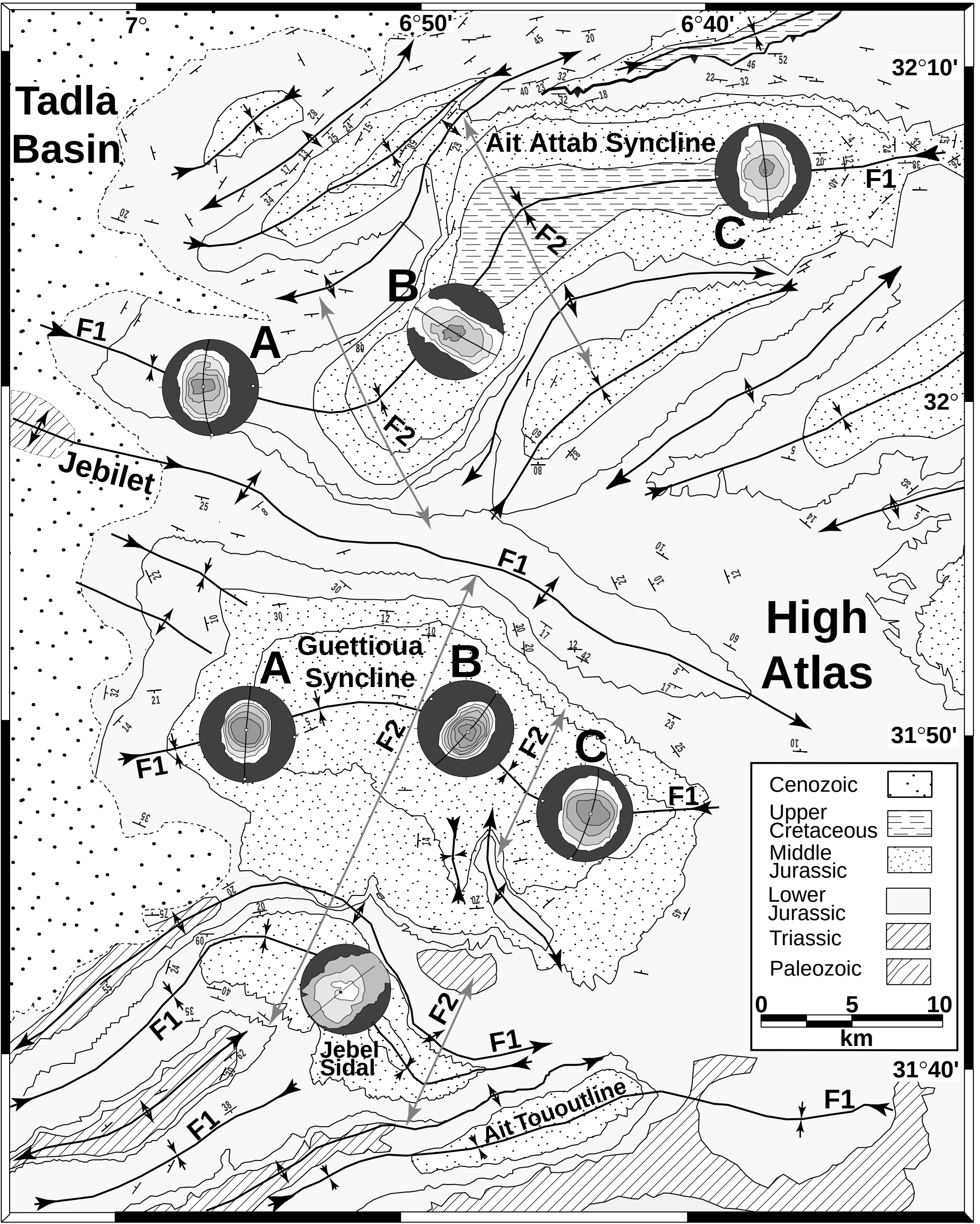
controlled by a preexisting geometry formed during the syn-rift phase of the Atlas system.

Superimposed folding is a common structural occurrence in orogenic belts. However, the timing and sequence of folding in the Atlas Mountains display a unique pattern of folding. Folds trend approximately parallel to the margins of the Atlas mountain belt (Figures 4.1 and 4.2) on either side of the proposed accommodation zone, indicating the direction of transport (thrusting) was normal to the orogen. However, folding is also parallel to the proposed accommodation zone and coaxial to the fold trends on either side of the transfer zone.

Ait Attab Syncline

Folding within the accommodation zone (Figure 4.2) display two phases of folding (F1 and F2). The Ait Attab syncline is north of the accommodation zone, and demonstrates an early phase of folding (F1) oriented NE, characterized by upright cylindrical folds that are refolded by second phase (F2) oriented NW. Folding of the F1 axial plane about a vertical F2 axis yields a Type-2 interference pattern (*Ramsay*, 1987). Strike and dip data were divided into separate regions (A,B & C) for the Ait Attab syncline that represent homogeneous domains of folding (Figure 4.2). The cylindrical best fit through poles to bedding for each domain (F1) was plotted on a stereogram to define the interlimb angle (Figure 4.3). Fold domain “A” shows a fold trend of 089° and a plunge of 13° (Figure 4.3), compared to domain “B”, with a fold trend of 031° and a plunge of 03°. Since these folds are upright cylindrical folds, the trace of the F2 axial plane (330°) is the bisector of these two fold trends (Figure 4.2).

Figure 4.2 Detailed geological map of the accommodation zone along the northern margin of the High Atlas (see Figure 4.1 for location). The first phase of folding (F1) of the Ait Attab and Jebel Sidal structures is normal to the first phases of folding in the Guettioua structure. Likewise, the second phase of folding (F2) of the Ait Attab and Guettioua synclinal structures are also normal to one another. The sequence and orientation of folding necessitates that folding was influenced by a preexisting accommodation zone during tectonic inversion. The map was compiled from field work, Landsat-TM images and geological maps (*Huvelin*, 1972;*Saadi et al.*, 1985; *Jenny*, 1988). Stereoplots of poles to bedding show the dip domains of folding.



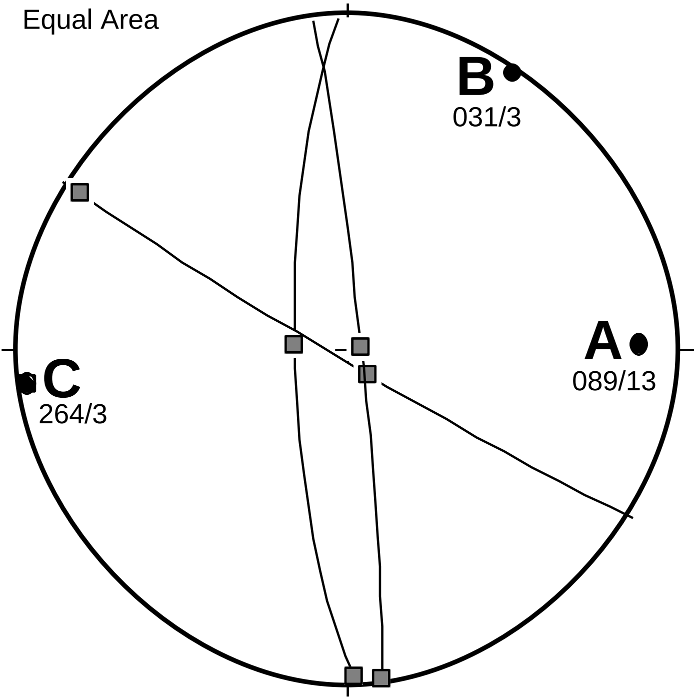
7°
6°50'
6°40'
32°10'
32°
31°50'
31°40'
Tadla Basin
Ait Attab Syncline
Jebilet
High Atlas
Guettioua Syncline
Jebel Sidal
Ait Tououtline
A
B
C
F1
F2
Cenozoic
Upper Cretaceous
Middle Jurassic
Lower Jurassic
Triassic
Paleozoic
0
5
10
km

Domain “B” of the Ait Attab syncline has a fold trend of 031° and a plunge of 3° (Figures 4.2 and 4.3). The trend (264°) and plunge (03°) of domain “C’” vary from domain “B” as a result of a similar axial plane of refolding (328°).

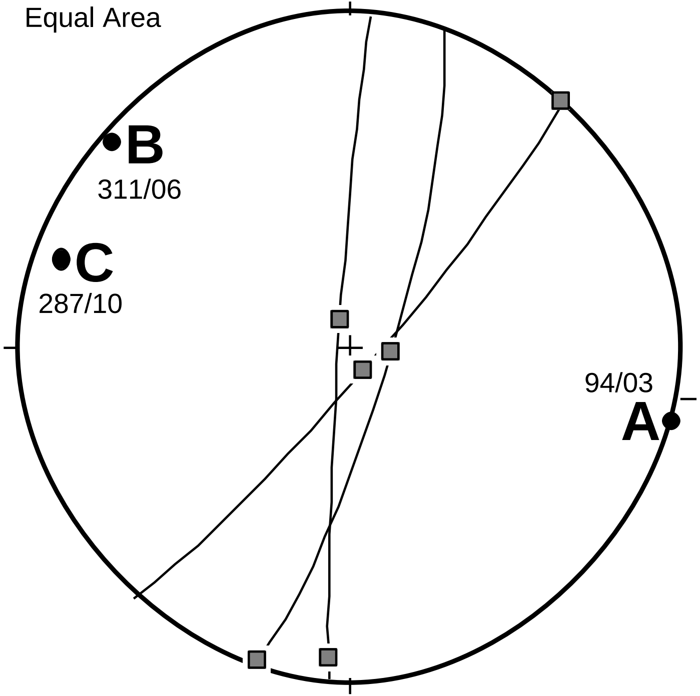
Jebel Guettioua-Sidal Synclines

Farther to the south, the Guettioua syncline displays an early phase of folding (F1) oriented NW followed by a later phase of folding (F2) oriented NE. The Jebel Sidal syncline was folded first (F1) along a NE trend and folded later by a second phase (F2) NE. The sequence and timing of folding in the area of study (Figure 4.2) indicate that the first phase of folding (F1) seen in the Ait Attab and Jebel Sidal structures was oriented to the NE. These fold axes (F1) are normal to the first phase of folding (F1) of the Guettioua structure (NW) (Figures 4.2 and 4.3). The orientation of the second phases of folding illustrates a similar discrepancy. The second phase of folding (F2) of the Ait Attab syncline (Figure 4.2), is normal to the second phase (F2) of folding of the Guettioua and Jebel Sidal structures. Data for the Guettioua syncline was also analyzed by separate domains (A, B and C) (Figure 4.3). Fold domain “A” of the Guettioua syncline shows a fold trend of 094°, plunging 03° to the east. Domain “B” has a fold trend of 311° and a plunge 06° to the NW. The Guettioua syncline is also an upright cylindrical fold, and the bisector of fold trends “A” and “B” is the trace of the F2 axial fold plane (023°) (Figures 4.2 and 4.3). Domain “C” is separated from “B” by a similar F2 axial fold plane. The trace of the F2 axial fold plane (023°) is strongly oblique to the F2 axial trace of the Ait Attab syncline (330°). The dip domain “B” for both the Ait Attab syncline and the Guettioua syncline are nearly coaxial, as were the initial F1 fold axes of the two synclines (Figure 4.3).

Figure 4.3 Stereograms displaying the cylindrical best fit through poles to bedding, and the trend of the fold axes of the separate fold domains of the refolded Ait Attab and Guettioua synclines.



Ait Attab Syncline



Guettioua Syncline

Summary: Paradox of different orders of superposition across the accommodation zone

The sequence of folding seen in the region of study implies that two separate phases of deformation could not have generated the unique pattern of folding, without the presence of pre-existing structural elements. Offsets in the basin margins result in 3D strain across the accommodation zone. During compression and subsequent inversion across these accommodation zones, a 3D strain is produced coupled to the preexisting accommodation zone. The structures formed by compression across these accommodation zones are affected by 3D strain, and result in unique refold patterns.

The Paleozoic exposure of the Jebilet anticlinorium trends eastward and then plunges to the southeast (Figure 4.2). The expression of this anticline can be seen by the gentle dips in the Jurassic on both limbs of the fold. The trend of the Jebilet anticline is not affected by the second phase of folding (F2) that refolds the Ait Attab and Guettioua synclines. This unusual fold relationship was likely generated as a result of the fold forming parallel to the ramp of an accommodation zone, rather than forming over a preexisting normal fault.

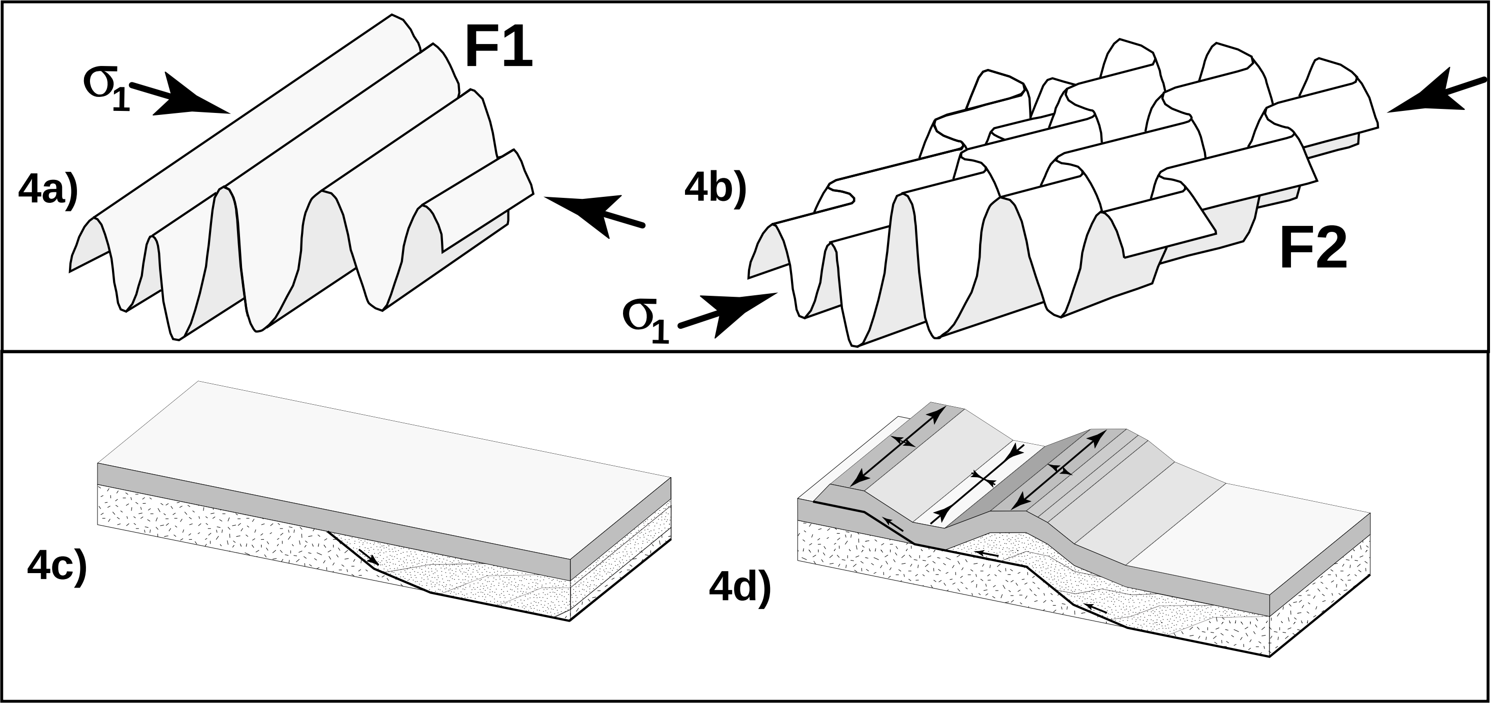
Interpretation of the structural evolution of the refolding geometry

Inversion of the down-to-the-basin normal faults resulted in the transport of the rocks in the hanging wall of the normal faults northward out of the basin. Inversion may have been achieved by the transport of strata from the hanging wall of the normal fault northwards along low angle thrust faults (Figure 4.4, a and b). One such thrust is exposed on the northern limb of the Ait Attab Syncline (Figure 4.2) (*Beauchamp et al.*, 1997a, b). This thrust transports syn-rift Lower Jurassic strata in the hanging wall of the thrust over Upper Cretaceous strata in the footwall. The uplift and inversion of the syn-rift strata may also be achieved by footwall short-cut faults that transport part of the footwall in the hanging wall away from the rift basin onto the margins (e.g. Jebilet).

We propose that in the High Atlas Mountains a single continuous phase of deformation across a preexisting structural feature, such as a extensional accommodation zone, resulted in the unique pattern of superposed folding. The sequence of superposed folding in the High Atlas Mountains of Morocco show that this style of folding are an important characteristic of inversion.

Accommodation zones are common features of rift systems. The geometries of these extensional structures result from a change in strike of the rift basin margin produced by a ramp/relay fault system or transfer fault. The step-like accommodation of two parallel normal faults by a higher angle fault forming a ramp, is a common feature in extensional rift basins (*Rosendahl*, 1986). It was found that *en echelon* normal faults form in rift systems when extension is oblique to the margin of the rift system (*McClay and White*, 1995). Transfer faults are also a common characteristic of extensional terrains, and allow for the transfer of extensional slip between faults. These faults are analogous with lateral ramps in thrust tectonic terrains. Extensional transfer faults, as with lateral ramps, transport rotational strike and dip components during rifting (*Gibbs*, 1984). These transfer faults occur as oblique or lateral accommodation zones and may involve a change in fault polarity. A transfer fault in cross-section may have a high angle flower geometry during rifting (*Gibbs*, 1987), in which case the fault may be normal to the basin margin, or

Figure 4.4 Schematic models showing the formation of superimposed folding as a result of two different stress orientations (4a and 4b). The shortening and inversion of the accommodation zone may have occurred by the transport of strata from the hanging wall of the syn-rift half graben northward along low angle thrust faults (4c and 4d).



σ1
F1
4a)
4b)
F2
σ1
4c)
4d)

more oblique, and involve a lower angle extensional ramp. Faults offset by these transfer faults can be parallel to the rift basin and have a planar, listric or extensional fault ramp geometry. The geometry of transfer faults often influences the development of younger folds upon inversion (*Alonso*, 1989). The transfer faults or ramps may be oblique or normal to the basin margin, and the orientation of the extensional/compressional ramp will result in differential movement in the hanging wall (in the sense of the transport direction) on the oblique ramp during inversion (*Casas-Sainz*, 1993).

Superimposed folding is often associated with a change in the orientation of a regional stress field through time. Superimposed folds may occur by successive deformational events separated by long time intervals, multiple deformational phases in one orogenic cycle, continuous deformation in one orogenic cycle, and simultaneous folding from several directions in one orogenic phase (*Ramsay*, 1987).

The Cobar Basin of Australia is an inverted Paleozoic basin that exhibits superposed folding (*Smith and Marshall*, 1992). Superposed folding in the Cobar basin is proposed to have occurred due to margin-normal shortening, followed by progressive oblique deformational shear. The Davenport Province of Central Australia (*Stewart*, 1987), is another region where superposed folding of Proterozoic age rocks resulted from two episodes of deformation that utilized pre-existing syn-sedimentary normal and transfer faults reactivated in reverse and strike-slip senses, respectively. In this region, major sedimentary faults such as those associated with accommodation zones in a rift system, controlled the structural domains of each fold trend.

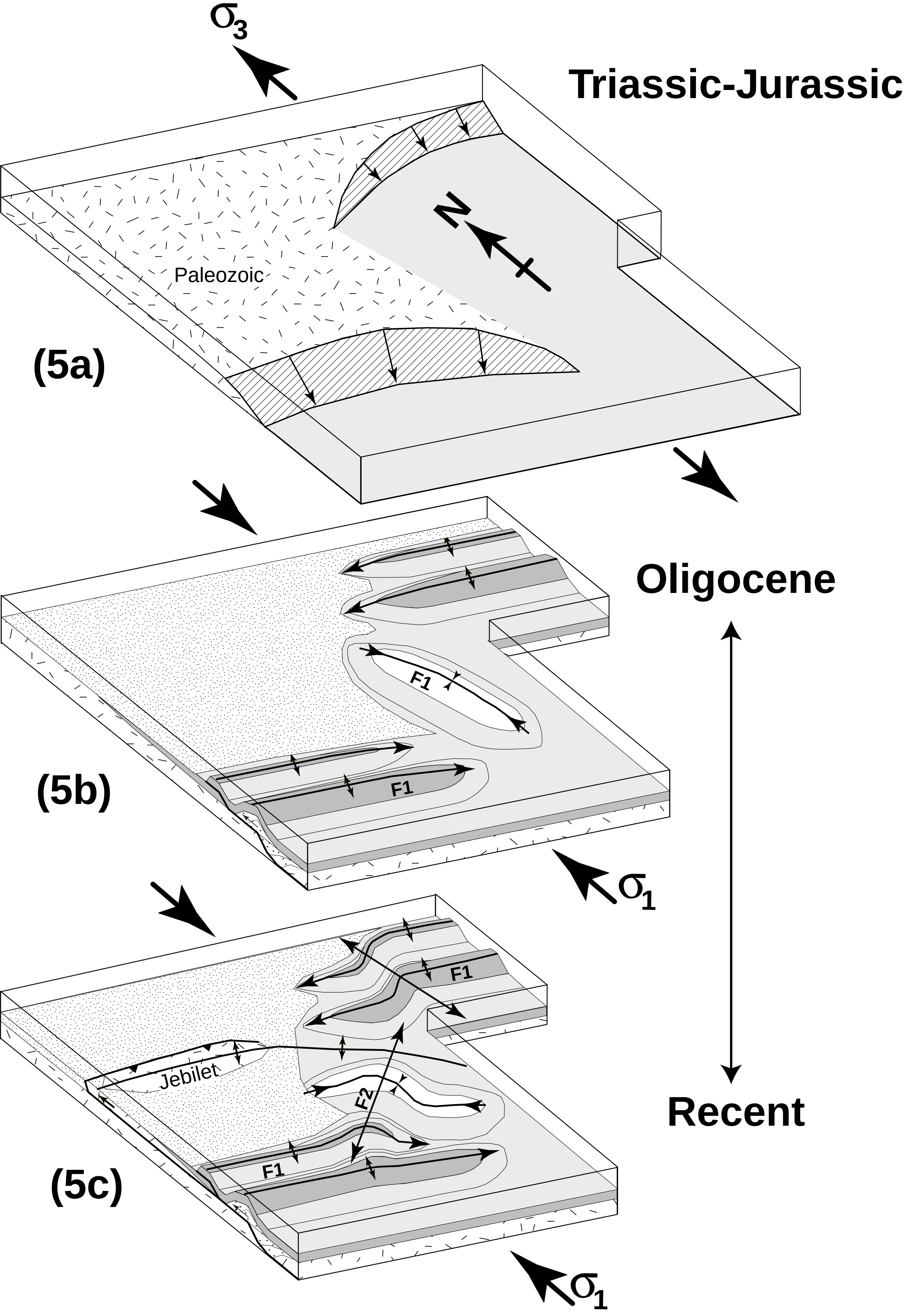
Steeply dipping faults offer greater resistance to reverse dip-slip movement during compression, and may form buttresses where displacement is concentrated (*Velasque et al.,* 1989). Folding can result from pre-existing structures (e.g., folds, faults or diapirs) which provide buttresses that concentrate strain. The geometry or configuration of pre-existing structures will control or alter the stress field affecting folding.

Discussion and Conclusions

The superposed folding along the northern margin of the High Atlas Mountains in Morocco was attributed by *Laville* (1978) to have occurred by successive rotation of compressional phases from WSW-ENE to NNE-SSW. Superposed folding in the eastern High Atlas Mountains was recognized by *De Sitter* (1960), and was attributed to separate tectonic phases and stress orientations.

However, we relate the unusual sequence and relationship of folding to pre-existing structural control (Figure 4.5). Separate tectonic phases and stress orientations would have yielded a different pattern of folding than is found in this region of the High Atlas Mountains. Superposed folding resulting from two different stress orientations often yields two phases of folds (F1 and F2), each phase having parallel fold axes (Figure .4.4a and 4.4b). In the Atlas mountains, as in the Davenport Province of Australia, oversteepened fold limbs (common in the Atlas) were probably caused by the second phase of folding. Both examples from Australia exhibit Type 1 and 2 interference patterns *(Ramsay*, 1987). The folding in the study area of the Atlas Mountains exhibit a Type 2 interference pattern (*Ramsay*, 1987), where the axial surfaces of the first folds are folded along with the limbs of the first folds. The superposed folding in the Atlas Mountains may have resulted from a syn-rift accommodation zone made

Figure 4.5 The inversion of a syn-rift accommodation zone (4.5a) by margin-normal shortening created first folds (F1) that are oriented parallel to pre-existing ramps and down-to-the-basin normal faults (4.5b). Continued compression across the accommodation zone results in the interference of the first F1 folds and the formation of F2 folding (4.5c).



σ3
Triassic-Jurassic
N
Paleozoic
(5a)
Oligocene
F1
F1
(5b)
σ1
Recent
F1
Jebilet
F2
F1
(5c)
σ1

up of a ramp/fault relay, or a transfer fault normal to the basin margin. This ramp transfers slip laterally between two major down-to-the-basin (southward dipping) normal faults (Figure 4.5a). Folding in the area of this study yield folds that are normal to one another in both phases (F1 and F2) (Figure 4.5b and 4.5c). Continued shortening across the accommodation zone may have created interference of the first F1 folds. The refolding of the F1 folds resulted in the F2 phase fold axes.

The presence of pre-existing structural geometries such as accommodation zones, fault ramps, fault relays, *en echelon* folding and other features formed by rift processes will have an effect on subsequent compressional stress fields generated by plate convergence and other tectonic processes. These structural geometries formed during rifting will affect the 3D strain field which may often result in superposed folding that is disharmonic. Superposed folding which is disharmonic may be a unique characteristic to inverted rift systems that result in intracontinental mountain belts.

References

Alonso, J.L., Fold reactivation involving angular unconformable sequences: theoretical analysis and natural examples from the Cantabrian Zone (Northwest Spain), *Tectonophysics*, 170,57-77, 1989.

Bally, A.W., Tectogenese et sismique reflexion, *Bulletin de la Societe Géologique de France*, 7,279-285, 1984.

Beauchamp, W., M. Barazangi, A. Demnati and M. El Alji, Inversion of synrift normal faults in the High Atlas Mountains, Morocco, *The Leading Edge*, 1171-1175, 1997a.

Beauchamp, W., Richard W. Allmendinger, Muawia Barazangi, A. Demnati and M. El Alji, Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological-geophysical transect, *Tectonics*, American Geophysical Union, (*in review*), 1997b.

Beauchamp, W., M. Barazangi, A. Demnati and M. El Alji, Intracontinental Rifting and Inversion: the Missour Basin and Atlas Mountains of Morocco, *American Association of Petroleum Geologists Bulletin*, 80, 1459-1482, 1996.

Casas-Sainz, A.M., Oblique tectonic inversion and basement thrusting in the Cameros Massif (Northern Spain), *Geodinamica Acta*, Paris, 6,3, 202-216, 1993.

De-Sitter, L. U., Plissement Croise Dans Le Haut-Atlas*, Geologie en Mijnbouw*, 5, 277-282, 1960.

Gibbs, A.D., Structural evolution of extensional basin margins, *Journal of the Geological Society*, London, 141, 609-620, 1984.

Gibbs, A.D., Development of extension and mixed-mode sedimentary basins, in M.P. Coward, J.F. Dewey and P.L. Hancock, eds., Continental Extensional Tectonics: *Geological Society Publication*, London, 28, 19-33, 1987.

Huvelin, P., Carte Géologique et Gitologique des Jebilet, 1:200,000, *Royaume du Maroc, Ministere de l'Energie et des Mines,*. 232C, 1972.

Jabour, H. and K. Nakayama, Basin modeling of Tadla basin, Morocco, for hydrocarbon potential*, American Association of Petroleum Geologists Bulletin*, 72, p. 1059-1073, 1988.

Jenny, J., Carte Géologique du Maroc au 1/100,000 Feuille Azillal (High Atlas central), *Notes Mémoires du Service Géologique*, Rabat, 339,104 pp., 1988.

Laville, Edgard, Incidence des jeux sucessifs d'un accident synsédimentaire sur les structures plicatives du versant nord du Haut Atlas central (Maroc), *Bulletin de la Société Géologique de France*, no.3, 329-337, 1981.

McClay, K.R. and M. J. White, Analogue modelling of orthogonal and oblique rifting, *Marine and Petroleum Geology,* 12, 137-151, 1995.

Ramsay, John G., and M. I. Huber, The techniques of modern structural geology; Volume 2, Folds and fractures, *Academic Press*, London, p. 494, 1987.

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*: Geological Society of London Special Publication*, 25, 25-43, 1986.

Rolley, J.-P., Carte Geologique du Maroc au 1/100,000 Feuille Afourer (High Atlas central), *Notes Mémoires du Service Géologique*, Rabat, 247, 103 pp., 1978.

Saadi, M., E. A. Hilali, M. Bensaid, Carte Géologique du Maroc-Jebel Saghro-Dadés, 1:200,000*, Royaume du Maroc, Ministere de l'Energie et des Mines*, 161, 1977.

Saadi, M., M. Bensaid, M. Dahmani, Carte Géologique du Maroc-Azilal, 1:100,000, *Royaume du Maroc, Ministere de l'Energie et des Mines*, 339, 1985.

Smith, J.V., and B. Marshall, 1992, Patterns of folding and fold interference in oblique contraction of layered rocks of the inverted Cobar Basin, Australia, *Tectonophysics*, 215, 319-334, 1992.

Stewart, A.J., Fault reactivation and superposed folding in a Proterozoic sandstone-volcanic sequence, Davenport Province, central Australia, *Journal of Structural Geology*, 9, 4, 441-455, 1987.

Velasque, P.C., L. Ducasse, J. Muller, R. Scholten, The influence of inherited extensional structures on the evolution of an intracratonic chain: the example of the Western Pyrenees, *Tectonophysics*, 162, 243-264, 1989.